SPECTRAL WAVE-DRIVEN SEDIMENT TRANSPORT ACROSS A FRINGING REEF

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Abstract

A laboratory experiment was conducted to investigate the dynamics of cross-shore sediment transport across a fringing coral reef. The aim was to quantify how the highly bimodal spectrum of high-frequency (sea-swell) and low-frequency (infragravity and seiching) waves that is typically observed on coral reef flats, influences the various sediment transport mechanisms. The experiments were conducted in a 55 m wave flume, using a 1:15 scale fringing reef model that had a 1:5 forereef slope, a 14 m long reef flat, and a 1:12 sloping beach. The initial 7 m of reef flat had a fixed bed, whereas the back 7 m of the reef and the beach had a moveable sandy bed. Four seven-hour irregular wave cases were conducted both with and without bottom roughness elements (schematically representing bottom friction by coral roughness), as well as for both low and high still water levels. We observed that the wave energy on the reef flat was partitioned between two primary frequency bands (high and low), and the proportion of energy within each band varied substantially across the reef flat, with the low-frequency waves becoming increasingly important near the shore. The offshore transport of suspended sediment by the Eulerian mean flow was the dominant transport mechanism near the reef crest, but a wide region of onshore transport prevailed on the reef flat where low-frequency waves were very important to the overall transport. Ripples developed over the movable bed and their properties were consistent with the local high-frequency wave orbital excursion lengths despite substantial low-frequency wave motions also present on the reef flat. This study demonstrated that while a proportion of the sediment was transported by bedload and mean flow, the greatest contributions to cross-shore transport was due to the skewness and asymmetry of the high and low-frequency waves.

Keywords

Fringing reef, Sediment transport, Laboratory model, Infragravity waves, Bottom roughness, Wave skewness

1 Introduction

There is a growing body of literature on the hydrodynamic processes generated by the interaction of waves with coral reef structures, including the evolution of incident swell wave fields, the dynamics of low-frequency (infragravity) waves, and the generation of mean wave-driven flows (see reviews by Monismith (2007) and Lowe and Falter (2015)). Reef systems display very different bathymetric characteristics from sandy beaches; they have a steep forereef slope, a rough shallow reef crest (often located far from the shore) and are connected to the shoreline via a shallow rough reef flat and sandy lagoon. At the reef crest, high-frequency waves (i.e. sea-swell waves with periods 5 – 25 s) are dissipated in a narrow surf zone via wave breaking and bottom friction (e.g. Lowe et al., 2005a). In this region, low-frequency waves (i.e., infragravity waves with periods 25 – 250 s) are generated by the breaking of incident high-frequency waves (Péquignet et al., 2014; Pomeroy et al., 2012a). In some cases low-frequency wave motions with periods even larger than the infragravity band (i.e., periods exceeding 250 s) can also be generated at the natural (seiching) frequency of coral reef flats, which may also be resonantly forced by incident wave groups (Péquignet et al., 2009; Pomeroy et al., 2012b). This disparity between high and low-frequency waves often results in a bimodal spectrum of wave conditions on coral reef flats and lagoons, where wave energy is partitioned between distinct high-frequency (sea-swell) and low-frequency (infragravity) wave bands (Pomeroy et al., 2012a; Van Dongeren et al., 2013). Recent hydrodynamic studies have shown how these different wave motions interact with the rough surfaces of reefs, and cause rates of bottom friction dissipation to be highly frequency dependent (e.g. Lowe et al., 2007; Pomeroy et al., 2012a; Van Dongeren et al., 2013). The extent to which bimodal spectra of hydrodynamic conditions on reefs affect sediment transport processes has yet to be investigated and is the focus of this paper.

It is customary to decompose the total sediment transport into two primary modes (bedload and suspended load), which enables a more detailed description of the physical processes involved, and can more readily be used to distinguish between the effects of currents and waves. For bedload transport, initial work focused on steady (unidirectional) flow in rivers and coastal systems where mean currents dominate (e.g. Einstein, 1950; Engelund and Hansen, 1967; Meyer-Peter and Müller, 1948). These descriptions relate the transport of sediment to the exceedance of a flow velocity or shear stress threshold. Current-driven suspended load is deemed to occur when the flow velocity (or bed shear stress) generates sufficient turbulent mixing to suspend particles in the water column (e.g. Bagnold, 1966). Traditionally, to incorporate these suspended load processes into predictive formulae, a vertically varying shape function describing the sediment diffusivity (e.g., constant, exponential, parabolic, etc.) is assumed, along with a reference concentration usually located near the bed (e.g. Nielsen, 1986; Soulsby, 1997; Van Rijn, 1993) .

The extension of sediment transport formulae to wave-driven (oscillatory flow) conditions was initially considered within a quasi-steady (wave-averaged) framework analogous to current-driven transport formulations, albeit extended to account for the enhancement of the bed shear stress induced by the waves. This approach has also primarily concentrated on high-frequency waves, despite low-frequency waves (as well as wave groups) also being important in the nearshore zone (e.g. Baldock et al., 2011). The importance of the shape of a wave form on sediment transport (i.e., due to the skewness or asymmetry of individual waves) has also been considered using a half-cycle volumetric approach, where the separate contribution of shoreward and seaward wave phases to the net transport is considered (e.g. Madsen and Grant, 1976). In general, these suspended sediment transport formulations are all sensitive to how the sediment is distributed vertically in the water column, with a number of different shape functions and empirical diffusion parameters

proposed that depend on the wave conditions (e.g. Nielsen, 1992; Van Rijn, 1993). It is also important to note that for rough beds, such as those with ripples, sediment suspension can be further enhanced by vortices generated at the bed (e.g. O'Hara Murray et al., 2011; Thorne et al., 2003; Thorne et al., 2002). Finally, in contrast to the more widely-used steady or wave-averaged approaches, instantaneous (or intra-wave) sediment transport models have also been proposed that attempt to directly model the transport over each phase of an individual wave cycle (e.g. Bailard, 1981; Dibajnia and Watanabe, 1998; Nielsen, 1988; Roelvink and Stive, 1989).

Irrespective of how the sediment transport is described, in wave applications these formulations tend to either assume that the transport can be described with properties of a single idealized monochromatic (regular) wave, or alternatively for the case of spectral (irregular) wave conditions, that the spectrum is narrow enough in frequency space that energy is concentrated near a well-defined (i.e., unimodal) peak and hence can be described by a single representative wave condition (height and period). For reef environments, there are distinct differences in how high and low-frequency waves transform across reefs, and hence in their relative importance over different zones of the same reef. As a result, sediment transport on reefs is still poorly understood and the applicability of existing sediment transport formulations to the distinct hydrodynamic conditions on reefs is not known. Consequently, an important first step is to understand the relative importance of suspended load and bedload to cross-shore transport, and more specifically how the mean flow and the distinct spectral wave conditions on reefs influence sediment transport.

Few detailed laboratory studies have been utilised to investigate physical processes on fringing reef systems, and all of those have exclusively investigated hydrodynamic processes and not sediment transport (e.g. Demirbilek et al., 2007; Gourlay, 1994; Gourlay, 1996a; Gourlay, 1996b; Yao et al., 2009). The objectives of this paper are thus to use a scaled

physical model of a fringing coral reef to: (1) quantify the spectral evolution of high and low-frequency wave fields across a reef; (2) understand the mechanisms that drive suspension of sediment within the water column; (3) identify how high and low-frequency waves affect the magnitude and direction of cross-shore sediment transport processes; and (4) determine the relative importance of suspended load versus bedload to the overall sediment transport. As the characteristics of a reef flat can vary from reef to reef, in this study we focus on the impact of both the reef flat water depth and bottom roughness on these processes. In section 2, we commence with an overview of the experimental design, instrument setup and the methods adopted to analyze the results. The results are presented in Section 3. Finally in Section 4, we assess the relative importance of the various sediment transport mechanisms, as well as the role of suspended versus bedload to changes in the overall cross-shore sediment fluxes. We conclude with a discussion of the implications of this study for the relative importance of different cross-shore transport mechanisms in fringing coral reef systems.

2 Methods and data analysis

2.1 Experimental design and hydrodynamic cases

The experiment was conducted in the Eastern Scheldt Flume (length: 55 m, width: 1 m, depth: 1.2 m) at Deltares (The Netherlands), which is equipped with second-order (Stokes) wave generation and active reflection compensation (Van Dongeren et al., 2001) (Figure 1a). The laboratory fringing reef model was constructed to a scale ratio of 1:15, corresponding to a Froude scaling of 1:3.9. The latter represents the balance between inertial and gravitational forces, and was used to maintain hydrodynamic similitude of essential processes, such as wave steepness, shoaling and breaking (Hughes, 1993). The reef model (Figure 1b) consisted of a horizontal approach, a 1:5 forereef slope from the bottom of the flume to a height of 0.7 m, a horizontal reef flat of 14 m length (7 m was a fixed bed and 7 m was a movable sediment bed) and a 1:12 sandy beach slope from the reef flat to the top of the flume. The

forereef slope and fixed (solid) reef flat were constructed from marine plywood, while the movable bed consisted of a very well-sorted and very fine quartz sand with a median diameter $d_{50} = 110 \mu\text{m}$ and standard deviation $\sigma = 1.2 \mu\text{m}$. This sand was chosen to be large enough to be non-cohesive (Hughes, 1993) and, when geometrically scaled, is equivalent to a grain size of 1.70 mm at prototype (i.e., field) scale – thus, comparable to the medium to coarse sand observed on many coral reefs (e.g. Harney et al., 2000; Kench and McLean, 1997; Morgan and Kench, 2014; Pomeroy et al., 2013; Smith and Cheung, 2002).

Four cases were simulated experimentally: a rough reef at low and high water, and a smooth reef at low and high water (Table 1). The low water condition consisted of a $h_r = 50$ mm still water level (SWL) over the reef flat, whereas the high water condition had a SWL of $h_r = 100$ mm. This corresponds to prototype SWLs of 0.75 m and 1.5 m, respectively. All four cases were conducted with a repeating ten minute TMA-type wave spectrum (Bouws et al., 1985) with an offshore (incident) significant wave height H_{m0} of 0.2 m (prototype 3.0 m) and a peak period T_p of 3.2 s (prototype 12.4 s). These conditions were selected to be representative of relatively large (but typical) ‘storm’ conditions that most wave exposed reefs experience (Lowe and Falter, 2015). Each case was run for 7 h and was partitioned into three sub-intervals (A: 1 hr, B: 2 hrs, C: 4 hrs).

To assess the impact of bottom roughness on the hydrodynamics and sediment transport across the reef, an idealised bottom roughness was used. Although it is not possible to capture the full complexity of natural three-dimensional roughness of coral reefs in a laboratory model, the roughness properties were nevertheless carefully chosen to match the bulk frictional wave dissipation characteristics observed on reefs (i.e., bottom friction coefficients for waves and currents that are typically of order 0.1 (e.g. Lowe et al., 2005b; Rosman and Hench, 2011)). For the rough reef cases, ~18 mm concrete cubes were glued to

the plywood (Figure 1c) with a spacing of 40 mm in a staggered array to achieve an estimated wave friction factor of $f_w \approx 0.1$, based on oscillatory wave canopy flow theory (Lowe et al., 2007). In contrast, smooth marine plywood was used for the smooth reef cases.

2.2 Hydrodynamic measurements

Synchronized hydrodynamic measurements were obtained at 18 locations along the flume (Table 2, with reference to Figure 1b). Surface elevations were measured at all 18 locations with resistance wave gauges (Deltares GHM wave height meter, accuracy ± 0.0025 m). Six of the wave gauges were collocated with electromagnetic velocity meters (Deltares P-EMS E30, accuracy within ± 0.01 m/s $\pm 1\%$ of measured value) and one was collocated with a Nortek Vectrino 2 profiler (Nortek AS). All measurements were obtained at 40 Hz, with the exception of the Vectrino that sampled at 100 Hz in 1 mm bins over 30 mm.

The hydrodynamic data were analyzed spectrally (using Welch's approach) with a 10 minute segment length (the repetition interval of the wave maker time-series) that were overlapped (50%) with a Hanning window applied (31, 63 or 127 approximate degrees of freedom depending on the run sub-interval considered). It is customary to separate the variance at prototype (field) scale into distinct frequency bands, i.e. high-frequency or sea-swell (1 – 0.04 Hz, 1 – 25 s), infragravity (0.004 – 0.04 Hz, 250 – 25 s) and very low-frequency motions (< 0.004 Hz, < 250 s) (e.g. Brander et al., 2004; Hardy and Young, 1996; Lugo-Fernández et al., 1998; Pomeroy et al., 2012a). In this study, we restrict our analysis to the relative importance of two prototype scale frequency bands, which we refer to as 'high' (1 – 0.04 Hz or 1-25 s) and 'low' (~ 0.0026 – 0.04 Hz or 25-380 s); in model scale, these correspond to ~ 3.9 – 0.16 Hz (~ 0.26 – 6.25 s) and 0.01 – 0.16 Hz (6.25 – 100 s) for the high and low-frequency bands, respectively (Figure 1d). In this study, the low-frequency band includes both the infragravity and the very low-frequency motion that contains the first mode

(quarter wave length) seiche period based on the geometry of the reef (e.g. Wilson, 1966).

We note that a seiche may form on reefs when the eigenfrequency of the reef falls within the incident wave or wave group forcing frequencies (Péquignet et al., 2009). Wave setup was also computed as the mean water level at each instrument location after an initial adjustment period (1 min) to allow the wave maker to start-up.

In the shallow nearshore zone, waves can become both skewed and asymmetric, which can influence material transport. Skewed waves are characterized by having narrower crests and wider troughs, while asymmetric waves have a forward-leading form with a steeper frontal face and a gentler rear face. We evaluated the skewness Sk and asymmetry As of the waves across the model, which was defined based on the near bed oscillatory velocity ($\tilde{u}$):

$$Sk = \frac{\langle \tilde{u}^3 \rangle}{\langle \tilde{u}^2 \rangle^{3/2}} \cong \frac{\langle \tilde{\eta}^3 \rangle}{\langle \tilde{\eta}^2 \rangle^{3/2}} \quad (1a)$$

$$As = \frac{\langle H(\tilde{u})^3 \rangle}{\langle \tilde{u}^2 \rangle^{3/2}} \cong \frac{\langle H(\tilde{\eta})^3 \rangle}{\langle \tilde{\eta}^2 \rangle^{3/2}} \quad (1b)$$

where $\langle \rangle$ is the time average over the time series, the ‘ $\sim$ ’ denotes the oscillatory (unsteady) component of the velocity u and water elevation η , and H is the Hilbert transform of the signal. In both Eqs. (1a) and (1b), we relate the water elevation to wave velocities assuming the wave motions are shallow (a reasonable assumption on the reef flat, which we also test below). This enables the evolution of the waveform to also be investigated with both the higher cross-shore resolution provided by the wave gauges, as well as locations where the wave velocities were directly measured.

The wave skewness Sk , represented by the third velocity moment, is recognised as an important driver of cross-shore sediment dynamics in nearshore systems (e.g. Ribberink and Al-Salem, 1994). It is often used as an indicator of the role of nonlinear wave processes on the transport, which can be conceptually viewed as describing a relationship between the bed stress (proportional to u^2) that is responsible for stirring sediment into the water column, and an advective velocity (proportional to u) that transports it (e.g. Bailard, 1981; Guza and Thornton, 1985; Roelvink and Stive, 1989; Russell and Huntley, 1999). Thus, forward-leaning (asymmetric) waves introduce a phase lag effect that can transport sediment in the direction of wave propagation. The rapid transition from the maximum negative velocity to the maximum positive velocity enhances the bed shear stress (and hence sediment entrainment) due to the limited development of the bottom boundary layer. In contrast, the relatively longer transition from the positive (onshore) velocity to negative (offshore) velocity enables the suspended sediment to settle (e.g. Dibajnia and Watanabe, 1998; Ruessink et al., 2011; Silva et al., 2011).

In order to assess the relative importance of the high and low-frequency bands (as well as their interactions), including both the magnitude and direction, we decompose the velocity skewness term on the reef flat further (c.f. the energetics approach by Bagnold, 1963; Bailard, 1981; Bowen, 1980). The velocity (u) measured at 20 mm above the bed was bandpass filtered in frequency space to obtain mean ($\bar{u}$), high ($\tilde{u}_{hi}$) and low ($\tilde{u}_{lo}$) frequency velocity signals that were then substituted into the numerator of Eq (1a) and expanded to produced 10 terms that are described in Table 3 (e.g. Bailard and Inman, 1981; Doering and Bowen, 1987; Russell and Huntley, 1999). The total (original) velocity signal was then substituted into the denominator of Eq. (1a) to normalize the terms.

2.3 Suspended sediment measurements

2.3.1 Sediment concentrations

Suspended sediment concentration (SSC) time series were measured at 5 locations across the reef flat with near-infrared fiber optic light attenuation sensors (Deltares FOSLIM probes). The sensor measures point concentrations and consists of two glass fibers mounted on a rigid rod separated by the sample volume with the difference in near-infrared intensity (due to absorption and reflection) between the two fibers related to the SSC (e.g. Tzang et al., 2009; Van Der Ham et al., 2001). A filter prevents ambient light from influencing the measurements. Prior to the experiments, the FOSLIMs were calibrated with the same sand used in the study. Known concentrations of sediment were sequentially added to a suspension chamber with a magnetic stir rod and measured with each instrument to form instrument-specific calibration curves (relating the concentration to an instrument output voltage) with a linear response ($R^2 > 0.99$). The FOSLIM signals from each case were despiked to remove data that exceeded the voltage range (e.g. due to bubbles or debris in the sample volume), were further subjected to a kernel-based despiking approach (Goring and Nikora, 2002), and then low-pass filtered (4 Hz cutoff) with a Butterworth filter to remove high-frequency noise. The background concentration measured 15 minutes after the completion of an experiment was also removed from the signal, as this represented the suspension of a minimal amount very fine sediment fractions (e.g., dust in the flume that can slightly affect optical clarity).

Time-averaged SSC profiles were obtained on the reef flat where two of the FOSLIMs ($x = 8.80$ m and 12.29 m) were sequentially moved vertically at 10 minute intervals after an initial spin up time (20 min). The length of each time series was consistent with the repetition interval of input wave time series applied to the wave maker (10 min) and thus enabled the concentration time series to be synchronised at each vertical sample location. With only single point velocity measurements available at most cross-shore locations, we

used linear wave theory to extrapolate the wave velocity profiles through the water column to compute suspended sediment flux profiles. This is a reasonable assumption as wave velocities for the shallow water waves on the reefs had minimal vertical dependence, and moreover, the largest source of vertical variability in the sediment fluxes was the much more substantial vertical variation in sediment concentrations. During each experiment, the bedform properties did not change substantially over the sampling period but did migrate horizontally during the experiment (see Section 3.3).

The optical SSC time series were supplemented with vertical profiles of time-averaged suspended sediment concentrations that were obtained by pump sampling at one location ($x = 9.83$ m) over the sandy bed. A collocated array of five 3 mm diameter intakes were vertically positioned with logarithmic spacing (Table 2) and oriented perpendicular to the flume side walls (c.f. Bosman et al., 1987). The 1 L synchronous pump samples collected over ~2 minutes were vacuum filtered onto pre-weighed membrane filters (Whatman ME27, $0.8 \mu\text{m}$), dried (100°C for 24 hrs) and weighed. Based on the intake diameter and volume flow rate, the intake flow velocity ranged from 0.85 m/s to 1.56 m/s, and hence was consistently greater than three times the measured root-mean-squared (RMS) velocity; therefore, errors due to inefficiencies in particle capture are expected to be very small (Bosman et al., 1987).

Traditionally, advection-diffusion models have been used to describe the time-averaged as well as wave-averaged vertical distribution of suspended sediment within a water column (see Thorne et al. (2002) for a recent review) and this forms the basis of many sediment transport formulations. In this approach, it is assumed that the vertically downward gravity-driven sediment flux is balanced by an upward flux induced by vertical mixing:

$$w_s \langle C(z) \rangle + \epsilon_s \frac{\partial \langle C(z) \rangle}{\partial z} = 0$$

where C is the instantaneous concentration at elevation z above the bed, w_s is the settling velocity of the sediment, ϵ_s is the vertical sediment diffusivity and the effects of both horizontal advection and horizontal diffusion are assumed to be comparatively small. To quantify the differences in the SSC profiles in each case, we estimated the depth dependence of the sediment diffusivity in Eq. (2) (since all other variables are known) for the pump sample concentration profile data based on the settling velocity ($w_s = 0.0081$ m/s for this sediment). The pump sampler data had higher vertical spatial resolution than the FOSLIM data and enabled the vertical structure of the sediment diffusivity on the reef flat to be assessed in finer detail.

2.3.2 Sediment fluxes

Profiles of the time-averaged horizontal suspended sediment flux $\langle uC \rangle$ were calculated using data from the two FOSLIM profile locations. The time series of u and C were initially decomposed into a mean (steady) and oscillatory (unsteady) component:

$$\langle uC \rangle = \langle (\bar{u} + \tilde{u})(\bar{C} + \tilde{C}) \rangle = \bar{u}\bar{C} + \langle \tilde{u}\tilde{C} \rangle \quad (3a)$$

where the $\langle \rangle$ denotes the time-averaging operator, the overbar indicates mean quantities and the ' $\sim$ ' denotes the oscillatory component. The first term $\langle \bar{u}\bar{C} \rangle$ on the right-hand side of Eq. (3a) is the suspended sediment flux driven by the mean (wave-averaged) Eulerian flow. The second term $\langle \tilde{u}\tilde{C} \rangle$ is the oscillatory flux and is non-zero when fluctuations in cross-shore velocity and sediment concentration are correlated. This oscillatory component was further decomposed into high- and low-frequency contributions, corresponding to the first two terms

on the right-hand side of Eq. (3b), respectively. The cross-product terms (the last two terms in Eq. 3b) represent interactions between high and low-frequency oscillations.

$$\langle \tilde{u}\tilde{C} \rangle = \langle (u_{hi} + u_{lo})(C_{hi} + C_{lo}) \rangle = \langle u_{hi}C_{hi} \rangle + \langle u_{lo}C_{lo} \rangle + \langle u_{hi}C_{lo} \rangle + \langle u_{lo}C_{hi} \rangle \quad (3b)$$

The frequency dependence of the oscillatory suspended flux was also investigated by a cross-spectral analysis of the sediment concentration and velocity time series (e.g. Hanes and Huntley, 1986). Cross-spectral estimates (S_{uC}) were obtained from detrended, Hanning windowed (50% overlap) data with a 5 minute segment length and 55-70 degrees of freedom. The magnitude and direction of the oscillatory fluxes at different frequencies were determined from the co-spectrum (the real part of the cross-spectrum), while the phase spectrum and coherence-squared diagram provided information on the phase lags and the (linear) correlation between u and C , respectively.

2.4 Bed measurements

An automatic bed profiler (van Gent, 2013) measured changes in the bed elevation along the flume at regular intervals throughout each experiment. The profiler was mounted on a motorized trolley that traversed the flume along rails mounted on top of the flume. It simultaneously measured three lateral transects ($y = 0.25$ m, 0.5 m and 0.75 m) at ~ 1 mm vertical resolution and 5 cm horizontal resolution without the need to drain the flume. At each measurement location, the profiler lowered a vertical rod until the bed was detected. The elevation of the bed at each point was then determined relative to the reference profile that was conducted prior to the commencement of each experiment. The sediment erosion and deposition rates within the model were then estimated from the difference between successive bed surveys over an elapsed time period.

The evolution of the bedforms (ripples) on the reef flat were measured with a Canon EOS 400D camera (3888 x 2592 pixels) that obtained images at 0.5 Hz between $x = 12.8$ m and 13.6 m on the reef flat. The images were projected onto a single plane with known targets in the images that were surveyed to ~ 1 mm accuracy and also had an equivalent pixel resolution of ~ 1 mm. A Canny edge detection algorithm (Canny, 1986) was used to detect the location of edges in the image based on local maxima of the image intensity gradient at the sediment-water interface. A peak and trough detection algorithm was used to determine the height of each ripple (η_r), which was defined as the absolute vertical distance from a trough to the next peak. The ripple length (λ_r) was defined as the distance between two successive peaks. The ripple crests were followed through the ensemble of the images (Figure 2) to determine the approximately constant ripple propagation velocity.

3 Results

3.1 Hydrodynamics

Offshore of the reef, the significant wave heights of the high-frequency waves $H_{m0,hi}$ were identical for all cases (Figure 3b). On the forereef, a confined region of wave shoaling occurred before the waves broke in a narrow surf zone just seaward of the reef crest near $x=0$ m, leading to a rapid reduction in $H_{m0,hi}$. The high-frequency waves continued to gradually dissipate across the reef, but $H_{m0,hi}$ eventually became roughly constant for $x > 5$ m. The difference in still water level had the greatest effect on $H_{m0,hi}$ on the reef, by increasing the depth-limited maximum height. The presence of bottom roughness only slightly attenuated the high-frequency waves across the reef.

Offshore of the reef, the significant wave heights of the low-frequency waves $H_{m0,lo}$ were also identical between the cases, and shoaled substantially on the forereef (Figure 3c). Near the reef crest ($x=0$ m), there was a rapid decrease in $H_{m0,lo}$; however, $H_{m0,lo}$ then

gradually increased further shoreward across the reef. The presence of bottom roughness had more influence on the low-frequency waves on the reef than the differences in water level, thus opposite to the response of the high-frequency waves. A detailed investigation of the processes driving this variability will be reported in a separate paper focused on the hydrodynamics, including for a broader range of conditions (Buckley, et al., in prep). However, we clarify here that this response is most likely due to the low-frequency waves propagating both shoreward and seaward as partial standing waves, due to their much stronger reflection at the shoreline (not shown), which implies that they will propagate on the reef and hence dissipate energy over a greater distance. In addition, the increased importance of these waves towards the back of the reef flat also implies that any change in bed friction will proportionally influence the low-frequency waves more than the high-frequency waves. Overall, the contrasting cross-shore trends in the high and low-frequency waves resulted in the low-frequency waves eventually becoming comparable or larger than the high-frequency waves towards the back region of the reef flat ($x > 7.5$ m), which is very similar to field observations on fringing reefs (e.g. Pomeroy et al., 2012a).

Wave setup on the reef flat decreased when the still water level was increased from 50 mm to 100 mm (Figure 3e). As a consequence, the total water depth on the reef (i.e., still water + wave setup) was comparable between these cases (~0.11 m vs ~0.14 m). Bottom roughness had a minimal effect on the observed setup.

The wave transformation across the reef led to distinct changes in wave spectra across the reef (Figure 4). Seaward of the reef, the surface elevation spectrum is unimodal, with a very dominant peak located in the high-frequency band (Figure 4b). Further across the reef, the spectrum becomes bimodal (Figure 4c). Further still across the reef and near the shoreline, the low-frequency waves eventually become dominant (Figure 4d).

The waves offshore were weakly nonlinear, with some skewness Sk (Figure 5a) but little asymmetry As (Figure 5b), consistent with the characteristics of finite amplitude deep-to-intermediate water waves. The magnitude of Sk and As estimated from the velocity and surface elevation followed similar trends; however, there were some small differences in magnitude. On the forereef slope, both As and Sk increased rapidly as the waves began to break. As the waves propagated shoreward out of the surf zone they initially remained both highly skewed and asymmetric from $x \sim 0$ -5 m. Further shoreward ($x > 5$ m), the waves remained highly skewed, but their asymmetry decayed across the reef. With As describing how saw-toothed the wave forms are, this decay in asymmetry is due to the waves transitioning from a bore-like form in the vicinity of the surf zone, to an increasingly symmetric (but still nonlinear) form on the reef flat as the waves reformed.

Decomposition of the velocity skewness into the high and low-frequency wave contributions provides an indication of how the nonlinear characteristics of the waves should influence sediment transport processes on the reef (Figure 6). Near the reef crest ($x \sim 0$ -5 m), the $3\langle u_{hi}^2 \rangle \bar{u}$ term, a proxy for high-frequency wave stirring and transport by the Eulerian flow, was largest and directed seaward. As $\langle u_{hi}^2 \rangle$ is a positive quantity, this term is seaward as a result of the mean flow being directed offshore, which originates from the balance of wave-induced mass flux that leads to a return flow that is commonly observed on alongshore uniform beaches (e.g. Svendsen, 1984). The dominant shoreward term in this region was the high-frequency wave skewness $\langle u_{hi}^2 u_{hi} \rangle$ that was comparable, but slightly weaker than the $3\langle u_{hi}^2 \rangle \bar{u}$ term.

Towards the back of the reef and adjacent to the shore (i.e., $x > 10$ m), the influence of both the $3\langle u_{hi}^2 \rangle \bar{u}$ and $\langle u_{hi}^2 u_{hi} \rangle$ terms decreased substantially. As a result, most of the terms were of comparable importance. Most importantly, from these results it can be implied that

the low-frequency waves should play an important (or even dominant) role on the transport. The shoreward-directed low-frequency wave skewness term $\langle u_{lo}^2 u_{lo} \rangle$ grew across the reef, and became large in this back reef region; for the shallow reef cases, this was the dominant shoreward-directed term. However, the shoreward directed $\langle u_{hi}^2 u_{hi} \rangle$ and $3\langle u_{hi}^2 u_{lo} \rangle$ terms were also significant. The seaward transport in this back region was partitioned almost equally between the $3\langle u_{hi}^2 \rangle \bar{u}$ and $3\langle u_{lo}^2 \rangle \bar{u}$ components, representing the interaction of the high and low-frequency wave stirring, respectively, with the seaward-directed mean flow. The presence of roughness tended to influence only the magnitude of the terms, but not the relative importance of each (Figure 6).

3.2 Suspended load

3.2.1 Suspended sediment concentrations

The mean SSC profiles varied in response to the presence of roughness as well as the water level over the reef flat. Higher SSCs were observed for the deep water cases (Figure 7a-c, g-i) relative to the equivalent shallow cases (Figure 7d-f, j-l), at all locations across the reef flat. For each water level condition (e.g. R10 vs S10), SSCs were lower when the reef was rough relative to when it was smooth. Across the reef flat (Figure 7, left to right), the magnitude of the SSCs increased, particularly near the bed.

We used the higher resolution sediment concentration profiles derived from the pump sampler, with the known sediment fall velocity w_s , and Eq. (2) to estimate a wave-averaged sediment diffusivity ϵ_s profile for each case. For all cases, ϵ_s increased away from the bed but then reached a roughly constant value higher in the water column (Figure 8). For the shallow cases, ϵ_s was similar for both smooth and rough cases near the bed, but further away from the bed ϵ_s was slightly greater for the smooth case. For the deep cases, ϵ_s was substantially larger relative to the shallow cases. Lastly, we note that the resolution of our

data does not extend fully into the near bed sediment mixing layer, which is approximated to be ≈ 18 mm when defined as $\delta_s = 3\eta_r$ (e.g. Van Rijn, 1993) for rippled beds. It is therefore not possible to determine the complete form of the sediment diffusivity very near the bed; however, ε_s has usually been observed to be roughly constant in this narrow region (e.g. Van Rijn, 1993).

3.2.2 Suspended sediment fluxes

The magnitude and vertical structure of the decomposed sediment flux terms computed with Eq. (3a,b) differed between the four cases. The high-frequency wave term $\langle u_{hi} C_{hi} \rangle$ contributed a weak, fairly depth-uniform net shoreward flux of sediment, and this was generally similar in magnitude for all cases, except for S10 where the value was slightly higher near the bed (Figure 9). For the deep water cases (R10 vs S10), the presence of roughness reduced the high-frequency contribution $\langle u_{hi} C_{hi} \rangle$ near the bed, but there was very little difference for the shallow cases (R05 vs S05). The low-frequency term $\langle u_{lo} C_{lo} \rangle$ contributed most to the overall flux and varied with depth: the flux was nearly zero or weakly seaward high in the water column, but directed strongly shoreward near the bed. The presence of roughness reduced this low-frequency term for the shallow cases, but there was only a slight reduction for the deep cases. The cross-product terms ($\langle u_{hi} C_{lo} \rangle$ and $\langle u_{lo} C_{hi} \rangle$) were both negligible, which indicates that high and low-frequency variability in velocities and concentrations did not interact. The mean seaward-directed sediment flux induced by the mean current and concentration $\langle \bar{u} \bar{C} \rangle$, which was measured at a single height 20 mm above the bed (as the velocity was not measured vertically), was small for the rough reef cases but larger for the smooth cases. Nevertheless, the contribution of this $\langle \bar{u} \bar{C} \rangle$ term to the overall

sediment flux at this elevation was still small; the flux was more influenced by the oscillatory wave motions than by the mean-flow.

The further decomposition of the sediment fluxes into frequency space through spectral analysis elucidates both the frequency and phase dependence of the interactions between the concentrations and velocities. On the reef flat, substantially more velocity variance S_{uu} was observed within the low-frequency band (0.01 – 0.16 Hz) compared to the high-frequency band (~3.9 – 0.16 Hz) and was most energetic at $f = 0.03$ Hz (prototype: $f = 0.007$ Hz, ~142 s, Figure 10). The magnitude of S_{uu} at this frequency was particularly affected by the presence of roughness for deep-water conditions (Figure 10a vs. g) but not for shallow water conditions (Figure 10d vs. j). Consistent with the velocity variance, S_{uC} (the spectral decomposition of the sediment flux) was small within the high-frequency band (~3.9 – 0.16 Hz) but exhibited substantial variance within the low-frequency band (Figure 10, second row). Near the bed ($z = 20$ mm) the suspended sediment fluxes were shoreward ($\text{Re}(S_{uc}) > 0$), while higher in the water column the transport was directed offshore, consistent with the profiles in Figure 9. Near the bed, the concentration and velocity signals had near zero lag (Figure 10, bottom row), which indicates that variations in concentration responded nearly simultaneously to the low-frequency waves. Higher in the water column the concentration signal progressively lagged the near bed velocity away from the bed (Figure 10, bottom row). The velocity profile measured above the wave boundary layer (estimated to be <2 cm thick) exhibited very little structural variation, especially for the low-frequency waves (not shown). Thus, if we assume the flow to be irrotational at the three concentration measurement heights, there is no velocity phase difference over this region of the water column. As a consequence, the vertical lag between the concentration and velocity would be primarily due to a lag in the diffusion of sediment up through the water column. This results in a reversal in the direction of sediment transport high up in the water column

(Figure 9), as the concentrations and velocities become inversely correlated. We note that a weak seiche mode ($f \approx 0.015$ Hz) was observed in both the hydrodynamics and the cross-spectral analysis except for R05. While the statistical certainty of this frequency is less than for the infragravity peak, it indicates that the seiche mode contributes slightly to onshore transport, particularly near the bed.

3.3 Bedforms and bed profile development

On the reef flat, the bed ripples became fully developed across the movable bed within approximately 15 minutes, or equivalent to roughly 450 high-frequency wave periods. The development of ripples on the bed in this experiment is consistent with a wave-formed rippled bed-state (and consequently the bedload regime), which is defined by a Mobility Number $\psi < 240$ (e.g. Dingle and Inman, 1976):

$$\psi = \frac{U^2}{gsD_{50}} \quad (4)$$

where U is the amplitude of the near-bed horizontal orbital velocity, D_{50} is the characteristic grain size and s is the relative density of the sediment. While the equilibrium height and length of the ripples were similar for all experiments (Table 4), the shoreward migration velocity of the ripples u_r was faster by 10-20% for the smooth cases relative to the rough cases. A similar migration velocity difference was also observed between the shallow cases and the deep cases.

Just shoreward of the fixed bed ($x=7.3 - 10$ m), sediment was eroded at a greater rate for the smooth cases (Figure 11b,d) relative to the rough cases (Figure 11a,c). The beach face at $x \sim 15-17$ m experienced the greatest erosion and over a larger distance for the deep cases. Some sediment was deposited slightly off the beach ($x \sim 14.5$ m) forming a small bar, while a well-defined swash bar ($x \sim 16.5 - 17$ m) formed up the beach. The swash bar in particular was larger in magnitude for the deep cases relative to the shallow cases, and also

slightly larger for the smooth cases. Some sediment from the movable bed was also visually observed to be transported seaward over the rigid (plywood) reef flat (i.e., $x < 7$ m) and formed a thin deposition layer; while these rates were not quantified in this study, this indicates that some seaward-directed sediment transport did occur in this region. In Section 4.3 we evaluate the contribution of the various sediment transport processes measured in this study in order to determine the relative contribution of each process to the magnitude and direction of sediment transport across the reef flat.

4 Discussion

The results from this study have provided the first quantitative insight into how a wide range of different wave-driven hydrodynamic processes drive cross-shore sediment transport on fringing coral reef flats. We found that the waves that propagate onto the reef out of the surf zone were both highly skewed and asymmetric. While the skewness remained relatively constant across the reef flat, the asymmetry of the waves gradually decreased towards shore. The wave energy on the reef flat was distributed between two primary frequency bands (high and low) and the proportion of energy within each band varied substantially across the reef flat. The decomposition of the wave velocity skewness indicates that on the reef crest and the first half of the reef flat, the primary transport mechanism was due to the interaction of the mean Eulerian return flow with the high-frequency wave stirring, which was seaward directed. Further across the reef flat, both the high and low-frequency waves were an important shoreward transport mechanism, with the proportion of transport by low-frequency waves growing across the reef and in some cases becoming dominant. These trends in the velocity skewness decomposition were consistent with the trends observed in the decomposed sediment fluxes $\langle uC \rangle$. Overall, the onshore suspended sediment transport was further supplemented by onshore bedform migration.

Most wave-driven sediment transport formulae have been conceptually developed by relating the transport to the properties of an idealised monochromatic wave, which are then extended to irregular wave conditions by assigning a representative wave height and single (peak) frequency. Given the distinct and highly-spatially variable hydrodynamic conditions that occur across reefs, it is of particular interest to determine if established sediment transport modelling approaches can still assist in the prediction of sediment transport under the strongly bimodal spectral wave conditions we observed in this study. We emphasise here, that while many formulations have been developed to estimate both suspended and bedload transport, our aim is not to conduct a detailed review of these approaches, nor to attempt to determine which formulations perform better. Instead we focus more broadly on assessing how sensitive transport predictions can be to the fundamental assumptions built into these approaches, demonstrating this with commonly used formulations as examples. In particular, we assess how the dichotomy between high and low-frequency wave motions on the reef influence sediment transport rates by affecting: (i) the mechanics of sediment suspension via vertical advection-diffusion balances, (ii) the net magnitude and direction of suspended sediment transport, and (iii) bedform properties and migration rates. Finally, we evaluate the overall sediment budget on the reef in order to quantify the relative importance of these processes, and compare these rates to the observed bed profile changes.

4.1 Suspended sediment transport

4.1.1 Suspension

Suspended sediment transport depends on the ability to accurately describe how sediment is distributed within the water column, with most approaches quantifying this with a sediment diffusivity (ϵ_s) or related parameter. In our study, the observed vertical structure of ϵ_s had two distinct regions: a linear region increasing away from the bed and a constant region higher up in the water column. Thus, despite the spectral complexity of the wave field on the

reef, this structure is still consistent with the multiple layered diffusion profile proposed by Van Rijn (1993) for the simpler case of unimodal non-breaking waves in sandy beach environments.

The Van Rijn (1993) sediment diffusivity is defined by three layers: (i) a near bed layer ($z \geq \delta_s$), where ϵ_s is a function of a peak orbital velocity and a length scale (Eq. 5a); and (ii) an upper layer ($z \geq 0.5h$), where ϵ_s is based upon the assumption that vertical diffusion of sediment is proportional to the mid-depth velocity that (from linear wave theory) is proportional to H_s/T_p (Eq. 5b); and (iii) and a middle layer where ϵ_s varies linearly between the near-bed and upper regions, i.e.

$$\epsilon_{sw,bed} = \alpha_b \hat{U}_{\partial w} \delta_s \quad (5a)$$

$$\epsilon_{sw,max} = \alpha_m \frac{H_s h}{T_p} \quad (5b)$$

Here $\epsilon_{sw,bed}$ is the near bed sediment diffusivity, $\epsilon_{sw,max}$ is the maximum sediment diffusivity, α_s, α_m are empirical coefficients, $\hat{U}_{\partial w}$ is the peak orbital velocity at the edge of the wave boundary layer δ_w , δ_s sediment mixing layer of thickness (which is assumed by Van Rijn (1993) to be $\delta_s = 3\delta_w = 3\eta_r$ for rippled beds), H_s is the significant wave height, T_p is the peak period and h is the water depth. Although this approach has been directly applied to unimodal wave conditions having a well-defined spectral peak, the choice for these variables become more complex and even arbitrary for a bimodal spectrum where (for example) a second large spectral peak period may be located within a low-frequency band; or in our case, across a reef platform where the peak period may shift gradually from high to low-frequency wave bands.

We evaluated the sensitivity of the measured sediment diffusivity ϵ_s profiles with the profiles estimated with Eq. (5) and the wave parameters using the observations at $x \approx 10$ m (Figure 12). Initially, we consider the conventional approach where ϵ_s is predicted using $H_{m0,total}$ based on the total energy in the full spectrum and the peak period of the full spectrum $T_{p,total}$ (i.e., following Van Rijn (1993)). Then, due to the bimodal form of the wave spectrum, we also estimate the profile that would be predicted for ϵ_s assuming first that the high-frequency waves dominate (i.e., $H_{m0,hi}/T_{p,hi}$) and then assuming that the low-frequency waves dominate (i.e., $H_{m0,lo}/T_{p,lo}$). Notably, the estimated ϵ_s profiles for all three approaches substantially under-predicted the observed profiles (Figure 12a-d). Specifically, the ϵ_s profile estimated by the high-frequency wave parameters ($H_{m0,hi}/T_{p,hi}$) predicted mixing near the bed that increased above the bed for the deep cases but not for the shallow cases. Using the low-frequency wave parameters ($H_{m0,lo}/T_{p,lo}$), this predicts comparable mixing near the bed, but with minimal mixing higher in the water column. The ϵ_s profile using the total spectrum (i.e., $H_{m0,total}/T_{p,total}$) predicted slightly more mixing than the other cases, but still grossly underpredicted ϵ_s . The substantial breakdown in the predicted ϵ_s is due to the presence of both high and low-frequency waves over the reef, but with somewhat more low-frequency wave energy that causes the spectral peak ($T_{p,total}$) to occur within the low-frequency band. This reduces the predicted mixing higher up in the water column (cf. Eq 5b). Therefore, the difficulty in applying conventional approaches to these bimodal spectral conditions are very clear; on the back of the reef, the suspension processes represented by ϵ_s would be predicted to be characterised by low-frequency waves but would ignore the efficient suspension of sediment by the high-frequency waves that also are present.

With the largest of the two spectral peaks falling within the low-frequency band, we assess how the ε_s profiles would change by replacing T_p with T_{m02} in the denominator of Eq. (5), where T_{m02} represents a weighted mean wave period based on the second moment of the wave spectrum. While this approach is non-standard, it has the effect of assigning the total wave energy to the mean period, thus shifting the representative period towards the high-frequency band. Physically this makes sense, as the total wave energy would contribute to mobilisation of the sediment, and the relevant time scale should fall somewhere between the high and low-frequency wave peaks. The ε_s profiles estimated with T_{m02} were considerably different from those predicted with T_p (Figure 12e-h), and much closer to the values we observed. This highlights how important both the low and high-frequency waves are to the sediment suspension on reefs, and that predictions can be quite sensitive to what is assumed to be the representative period of motion. Further work is clearly required to understand the precise mechanisms and interactions that drive sediment suspension under these complex spectral wave conditions; however, the results do indicate that the mean period T_{m02} rather than a peak period is a more appropriate choice when applying existing diffusivity formulations.

4.1.2 Advection

The processes that drive sediment suspension are only one component of predicting rates of suspended load; mechanisms that drive the net advection of the suspended sediment (averaged over a wave cycle) are equally important, as they determine the rate and direction of the transport. In this study we observed some similarities, but also a number of key differences, in how suspended sediment is transported in reef environments relative to what is typically observed on beaches. In a region extending from the forereef slope and through the surf zone, the dominant offshore transport mechanism was the seaward-directed Eulerian

mean flow, which was partially offset by shoreward transport by the high-frequency wave skewness; this is similar to the pattern observed on beaches seaward of the surf zone (e.g. Ruessink et al., 1998; Russell and Huntley, 1999). However in contrast to beaches, where the surf zone is often wider and located close to the shoreline, the presence of the wide reef flat caused these two transport mechanisms to decrease substantially in importance across the reef. Instead, within the back reef region, transport by the low-frequency waves became increasingly important (and for shallow water depth case became dominant), which is a distinct difference to beach environments where low-frequency waves typically become dominant primarily within the swash zone (e.g. Van Dongeren et al., 2003). These trends in the hydrodynamics are consistent with our direct observations of the suspended sediment fluxes, which together emphasize the importance of low-frequency waves to cross shore sediment transport on reefs.

4.2 Bedload and ripple properties

Sediment ripples are usually observed in the lagoon and back region of coral reefs in the field (e.g. Storlazzi et al., 2004) and the fraction of the bedload induced by their migration could represent a substantial proportion of the total sediment transported in reef environments. This is due to the relatively low wave energy in the back regions of wide coral reef flats (which leads to relatively high Rouse numbers), which can favor sediment transport along the bed but may not be sufficient to suspend it into the water column. The proportion of the bedload transport rate due to bedform migration can be estimated from the ripple height and propagation velocity (Eq 6):

$$Q_r(x) = u_r (\eta_r(x) - \eta_{r,0}) \quad (6)$$

where $Q_r(x)$ is the local volumetric bedload transport rate due to ripple migration per unit width, based on the ripple migration velocity u_r and the ripple surface profile $\eta_r(x)$ relative

to the ripple trough elevation $\eta_{r,0}$. To obtain a mean bedload transport rate, Eq (6) can be integrated over the ripple wave length giving:

$$\bar{Q}_r = \frac{u_r}{\lambda_r} \int_0^{\lambda} (\eta_r(x) - \eta_{r,0}) dx = \alpha \eta_r u_r \quad (7)$$

where $\alpha = V/\eta_r \lambda_r$ is a shape function that relates the ripple geometric dimensions to the ripple volume per unit width V . Eq. (7) can then be modified for porosity (n_p) and re-expressed in terms of the dry weight of the sediment based on the sediment density (ρ_s):

$$Q_b = \alpha \rho_s (1 - n_p) \eta_r u_r \quad (8)$$

Implicit in this analysis is the assumption that within the ripple regime (Section 3.3), bedload is confined to a thin layer of sediment that is transported with the migrating bed forms, which is a conventional approach for estimated bedload transport rates (e.g. Aagaard et al., 2013; Masselink et al., 2007; Traykovski et al., 1999; van der Werf et al., 2007). Given that the accurate prediction of properties of the bedforms are essential for predicting the bedload transport in this regime, we compare the ripple properties we observed to the equations of Malarkey and Davies (2003) (the non-iterative form of the equations by Wiberg and Harris (1994)), which were derived from a large number of data sets, although again focusing on monochromatic or unimodal spectral wave conditions. We thus evaluate whether these established equations are able to predict the ripple dimensions in our experiment as a function of A/d_{50} , where A is an orbital excursion of a representative wave motion (discussed below). This parameter has been selected as it has been shown to collapse a wide range of data onto a single curve (Soulsby and Whitehouse, 2005).

If the analysis is restricted to a representative wave motion in the high-frequency band, the equations of Malarkey and Davies (2003) predict the ripple dimensions reasonably well even in this reef environment where there is also a large amount of low-frequency wave

energy superimposed (Table 4). This suggests that the addition of substantial low-frequency wave motions has limited influence on the bed forms, i.e. the properties appear to be the same as what would occur for a pure high-frequency wave field of the same magnitude.

4.3 Suspended vs. bedload transport on coral reef flats

A goal of this experiment was to determine how the bimodal spectral wave conditions that are generated across coral reef flats influence cross-shore sediment transport processes. To evaluate the relative importance of the different transport mechanisms, we constructed a sediment budget (Figure 13) based on measurements of the low (Q_{lo}) and high (Q_{hi}) frequency contribution to the suspended sediment fluxes), the suspended sediment flux due to the mean Eulerian flow (Q_m), and the bedload transport (Q_b) contribution. The net transport of sediment derived from the sum of these components ($\sum Q$) was compared to an independent estimate of the net transport of sediment (a flux) obtained via cross-shore integration of bed profile changes both shoreward and seaward of $x = 8.8$ m (the point closest to the midpoint of the movable bed on the reef, and where co-located velocity, surface elevation and sediment concentration measurements were obtained):

$$Q_p(x) = \int_{x=23.8}^{x=8.80} \frac{\partial z_b}{\partial t} (1 - n_p) dx - \int_{x=8.80}^{x=7.00} \frac{\partial z_b}{\partial t} (1 - n_p) dx \quad (9)$$

where Q_p is the cross shore sediment transport associated with profile change across the movable bed ($x = 7.0 - 23.8$ m), $\partial z_b / \partial t$ is the cross-flume averaged bed level change between adjacent profile measurement locations, and n_p is bed sediment porosity (which for very well sorted sand is ≈ 0.4).

The analysis indicates that suspended load driven by both high and low-frequency wave motions (Q_{hi} and Q_{lo} , respectively) made the greatest contribution to the cross shore transport (Figure 13). Both Q_{hi} and Q_{lo} were of comparable magnitude, although in most

cases Q_{lo} was slightly larger. Suspended transport by the Eulerian flow Q_m was the dominant seaward transport mechanism, and while it was relatively small, it was greater for the smooth reef cases. The estimated onshore bedload transport rate Q_b tended to be smaller than both Q_{hi} and Q_{lo} . At first this may seem contradictory, given that for the Rouse numbers, bedload would be expected to play an important role. The dominance of the suspended load in this experiment could in part be due to not fully capturing all of the true bedload that occurred. Very accurate measurements of bedload are notoriously challenging or impossible (e.g. Aagaard et al., 2013), as it is due to sediment grains rolling or saltating along the bed that may not be completely associated with ripple migration. To gain additional confidence that the ripples are representative of those observed in the field, we compare the ripple dimension and migration rates measured in this experiment with those measured by Becker et al. (2007) at Waimea Bay (Hawaii) on a carbonate sediment dominated pocket beach surrounded by reefs. Although the field site is quite different morphologically to the present laboratory study, these field measurements were obtained for very similar prototype water depths ($h = 1 - 2$ m) and wave heights ($H_s = 0.2 - 1$ m). The laboratory ripple dimensions were comparable to those measured in the field ($\eta_r = 9$ cm vs. $10 - 20$ cm, $\lambda_r = 0.6$ m vs. $0.4 - 1.2$ m in prototype scale) and if we assume that the ripple migration at that site was primarily driven by oscillatory wave motion, ripple migration rates were also comparable at prototype scale ($U_r \sim 2.8$ m day⁻¹ for the lab vs. $-3.3 - +4.5$ m day⁻¹ measured in the field). This provides additional confidence that the laboratory-derived ripple dimensions and bedload contribution are both of comparable magnitude to real field-scale observations.

The comparison of the net transport ($\sum Q$) with estimates from the integration of the bed level changes (Q_p), shows that both are positive quantities, i.e. consistent with a shoreward accumulation of sediment for $x > 8.8$ m, although $\sum Q$ tended to be consistently smaller than Q_p . There was reasonable agreement between $\sum Q$ and Q_p for the deep cases

(both smooth and rough) indicating that it is possible to roughly close the sediment budget; however, ΣQ was much lower for the shallow cases. The source of this discrepancy is unclear, but indicates that some additional shoreward transport was missing from the budget. The most likely source is due to the absence of very near-bed sediment transport measurements (i.e., below the lowest sampling heights of the FOSLIM and pump sampler, including between the crest and trough of the ripples). This is the region of sediment transport that is universally the most difficult to define and experimentally quantify, as it represents a transition region between what is clearly suspended versus bedload (Nielsen, 1992). Despite being a very small region ($< \sim 1$ cm), sediment concentrations would be high and this would likely influence the total rate of transport. This is also consistent with the sediment budget being more nearly closed for the deep cases, as this thin nearbed region would have less impact on the overall budget and there is also greater vertical mixing of sediment into the water column (i.e., higher ϵ_s ; Figure 12). Nevertheless, despite this discrepancy, the results clearly show the relative importance of the shoreward transport mechanisms induced by both the low and high-frequency waves on both the suspended and bedload transport; the bed profile observations suggest that these mechanisms would only be of greater importance. Most importantly, the results suggest that the bedload is of the same order of magnitude, but still smaller than the suspended load, which is different from some recent high energy beach studies that suggest that suspended load can be more than an order of magnitude greater (e.g. Aagaard, 2014; Masselink et al., 2007).

We finally note that cross-shore sediment transport would likely be enhanced if the results of this study are extended to two-dimensional reef-lagoon systems with open lagoons or channels perforating the reef (i.e., barrier reefs or atolls). While these systems are driven by the same hydrodynamic processes, these more complex systems are unconstrained in the alongshore direction and have channels where wave-driven flows across the reef can return to

the ocean. For these reef morphologies, a shoreward Eulerian cross-reef flow is thus often present (e.g. Lowe et al., 2009), which was weakly seaward directed in that closed one-dimensional reef used in the present study that represents a fringing reef morphology.

Therefore, while the transport processes induced by the nonlinear waves on the reef should be the same within these other types of reefs, the suspended transport due to the Eulerian mean flow would be different. In this case, the shoreward mean flow would enhance transport towards the shore and, if there is sufficient energy within the lagoon, would drive sediment back out the channel. Nevertheless, we would expect that the wave-induced transport mechanisms induced by the high and low-frequency waves would operate similarly, as they are primarily influenced by the morphology of the reef flat (i.e., independent of whether a channel is present or not). As such, there should only be an enhancement of the shoreward transport for cases where the reef morphology varies alongshore.

5 Conclusions

While there is already a large and growing literature on how coral reef structures modify a wide range of nearshore hydrodynamic processes, as well as a limited number of observations of sediment concentrations and rough estimates of transport rates on reefs, detailed studies of the mechanisms that drive sediment transport on reefs have been severely lacking. In this study, we utilized a physical model of a fringing reef to examine these sediment transport processes in detail for the first time, which has revealed the following key results:

1. As waves break on the steep reef slope and continue to propagate across the wide reef flat, the wave spectrum changes considerably, from initially being dominated by high-frequency (sea-swell) waves on the forereef and in the vicinity of the surf zone, to gradually being dominated by low-frequency (infragravity) waves towards the back of the reef. This trend is consistent with a number of recent field observations conducted on fringing reefs.

2. The skewness and asymmetry of both the high and low-frequency waves on the reef flat make the major contribution to shoreward suspended sediment transport. On the seaward portion of the reef, the high-frequency waves play a more important role on this transport, whereas on the back portion of the reef the low-frequency waves eventually become dominant. Some of this shoreward transport on the reef is offset by the seaward Eulerian mean flow. But overall, the net suspended sediment transport was directed towards the shore.
3. Due to the bimodal characteristics of the wave spectrum that also evolves in space, existing wave-averaged suspended sediment transport formulations will likely breakdown in reef applications. This is due to most approaches having been derived assuming a single representative wave motion, which cannot be readily defined under these complex spectral conditions. We found that predictions of the suspended sediment concentration profiles could vary widely depending on how this representative wave was chosen (i.e., whether it focused on the high-frequency waves, the low-frequency waves, or the total wave energy). In our study we found the best agreement when the total wave energy was used to determine a representative wave height, and the mean wave period (specifically T_{m02}) was chosen as the representative period. Nevertheless, the development of intra-wave sediment transport formulations that can account for the strong interactions between the high and low-frequency waves would no doubt help improve suspended sediment transport predictions on reefs considerably.
4. Bedload transport on the reef was shoreward-directed and associated with the shoreward migration of bed ripples. The geometry of these ripples was controlled by the high-frequency waves; despite the presence of the substantial low-frequency wave motions that occur on the reef, these appear to have little influence on the properties of the ripples. While transport by bedload appeared to make a smaller contribution than the suspended

load, the bedload still made a substantial contribution and enhanced the net shoreward transport of sediment across the reef.

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Table 1. Simulation cases with parameters expressed in model scale.

ID	H_{m0} [m]	T_p [s]	h_r [m]	Reef State
R10	0.20	3.20	0.10	Rough
R05	0.20	3.20	0.05	Rough
S10	0.20	3.20	0.10	Smooth
S05	0.20	3.20	0.05	Smooth

*Bulk parameters used in the generation of the TMA spectrum

Table 2. (left) Location and type of measurement, with the distance x measured relative to the reef crest. (right) The vertical position z of the pump sample intakes relative to the initial bed.

Location	x [m]	η	u	C	$\langle C \rangle$		
1	-15.06	X	X	-	-		
2	-5.30	X	-	-	-		
3	-3.10	X	-	-	-		
4	-1.80	X	X	-	-		
5	-1.10	X	-	-	-		
6	-0.50	X	-	-	-		
7	-0.23	X	-	-	-		
8	-0.04	X	-	-	-		
9	0.35	X	X	-	-		
10	1.00	X	-	-	-		
11	1.60	X	-	-	-		
12	2.90	X	X	-	-	z [mm]	v_p [m/s]
13	4.25	X	-	X	-	94	1.13
14	5.80	X	X	-	-	74	0.85
15	8.80	X	-	X	-	56	1.56
16	9.83	-	-	-	X	40	1.56
17	10.54	X	X	X	-	29	1.56
18	12.29	X	-	X	-	20	1.56
19	13.99	X	X [#]	X	-		

η is the surface elevation

u is the cross reef velocity

indicates the location of the Vectrino 2

C is the instantaneous concentration (of suspended sediment)

$\langle C \rangle$ is the time averaged concentration obtained by pump sampling

z is the pump sample intake elevation relative to the initial bed position

v_p is the velocity of the pump sampling intake

Table 3: Third order moment decomposition of the velocity u . We note that the dominant terms are later found to be those highlighted by the *. Refer to Section 3.1 for details.

Term	Composition	Description
M1	$\bar{u}^3$	mean velocity cubed
M2*	$\langle u_{hi}^2 u_{hi} \rangle$	skewness of high-frequency waves
M3*	$3\langle u_{hi}^2 u_{lo} \rangle$	correlation of high-frequency wave variance and low-frequency wave velocity
M4*	$3\langle u_{lo}^2 u_{hi} \rangle$	correlation of low-frequency variance and high-frequency wave velocity
M5*	$\langle u_{lo}^2 u_{lo} \rangle$	skewness of low-frequency waves
M6*	$3\langle u_{hi}^2 \rangle \bar{u}$	stirring by high-frequency waves and transport by mean flow
M7*	$3\langle u_{lo}^2 \rangle \bar{u}$	stirring by low-frequency waves and transport by mean flow
M8	$6\langle u_{hi} u_{lo} \rangle \bar{u}$	six way correlation
M9	$3\langle u_{hi} \rangle \bar{u}^2$	time-average of high-frequency wave oscillatory component
M10	$3\langle u_{lo} \rangle \bar{u}^2$	time-average of low-frequency wave oscillatory component

Table 4: Ripple characteristics on the reef flat calculated by image analysis and predicted with the equations of Malarkey and Davies (2003).

Case [#]		R10	R05	S10	S05
Measured					
η_r	[mm]	6 (3)	5 (1)	6 (1)	6 (1)
λ_r	[mm]	40 (4)	37 (3)	41 (4)	42 (5)
η_r/λ_r	[-]	0.15	0.14	0.15	0.14
N_r	[-]	10	13	20	20
u_r	[mm/hr]	25 (2)	28 (1)	29 (2)	32 (3)
N_p	[-]	8	10	13	10
Predicted					
$\eta_{r,hi}$	[mm]	7	7	7	5
$\lambda_{r,hi}$	[mm]	38	41	40	31

[#] interval C of the respective cases

all ripples progressed ‘shoreward’ across the reef flat

N_r is the mean number ripples analyzed per image

N_p is the mean number peaks followed in the experiment interval

The numbers in the parentheses are the standard deviation.

Figure 1. (a) A side view photograph of the experimental setup. (b) A diagram of the experimental setup, with the distances x relative to the reef crest. The dotted blue line indicates the high water case, the red dotted line the low water case. The locations of instruments are indicated by the solid vertical lines (labels for locations 7, 8 and 10 are not shown for clarity). The solid black shading defines the fixed bed reef and the solid grey shading the movable bed. (c) A photograph of the roughness elements ($\sim 18 \text{ mm}^3$ at 40 mm spacing) on the reef crest. (d) Surface elevation spectral estimate at location 10 on the reef flat with the low and high-frequency bands adopted in the study indicated.

Figure 2. Ripple crest progression by crest tracking obtained by the image analysis. The vertical axis follows the ripple crest along the reef flat. Each line of data represents the progress of one ripple crest. Only R10c is shown, as other cases are similar.

Figure 3. Hydrodynamic observations for all cases. (a) The model bathymetry with the location of the instruments indicated by the vertical lines. (b) The high-frequency significant wave height evolution. (c) The low-frequency significant wave height evolution. (d) The total significant wave height. (e) The setup.

Figure 4. Wave spectra across the reef for case R10. (a) The model bathymetry with the location of the instruments where the indicated spectral estimates were obtained: (a) location 2 offshore (b) location 12 on the reef flat and (c) location 19 near the beach. The high (low) frequency band is indicated by the blue (red) portion of the spectral estimate. Very similar trends in spectral evolution across the reef were shown for the other three cases.

Figure 5. Cross-shore distribution of (a) skewness Sk (Eq. 1a) and (b) asymmetry As (Eq. 1b) for R10. The black line is calculated from the surface elevation measurements and the markers are calculated from the nearbed velocity measurements.

Figure 6. The dominant velocity moment terms across the reef (refer to Table 3). Positive (negative) values indicate shoreward (seaward) transport. Note that analysis of the Eulerian terms at one location ($x = 12.3$ m) in R05 is questionable, as mass conservation should require the Eulerian flow to be seaward, thus consistent with the other sites. This is most likely due to the position of the instrument being too high in the water column in this shallow experiment, as it was observed to occasionally become exposed. A Monte Carlo error analysis that incorporated the reported velocity accuracy of the instrument and 1000 realisations of the perturbed velocity time series predicts that the error for each computed moment term is small ($<1\%$ of the magnitude for all data points).

Figure 7. Time-averaged concentration profiles at location 15 by the FOSLIM (left column), middle of the reef by pump sampling (middle column) and near the beach at location 18 by the FOSLIM (right column) for R10 (a-c), R05 (d-f), S10 (g-i) and S05 (j-l). The red line indicates the mean profile and the horizontal black line is the non-dimensional mean ripple height.

Figure 8. Vertical structure of the sediment diffusivity ϵ_s estimated from the mean concentration profile for all pump sample profiles obtained. The dashed (dash-dot) horizontal line is the mean ripple height for the deep (shallow) water conditions. The sediment diffusivity for R10 was calculated with fewer measurements due to a fault in the data collection at one elevation during that simulation.

Figure 9. Time-averaged sediment flux profiles decomposed into high-frequency (black solid line) and low-frequency (black dashed) contributions. The cross terms are indicated by the colored dashed lines (the blue line lies beneath the red line and thus cannot be seen) and the time averaged Eulerian flux is indicated by the blue dot. The solid horizontal line indicates the non-dimensional mean ripple height. Positive (negative) values denote shoreward (seaward) transport.

Figure 10. Cross-spectral analysis of the velocity signal measured 20 mm above the bed with the concentration signal recorded by the FOSLIM at different elevations at $x=8.80$ m. (top row) The velocity spectra, (middle row) the real component of the cross-spectrum between velocity and concentration and (bottom row) the phase from the cross-spectral analysis focused around the peak in the low-frequency band. The solid vertical lines indicate the low-frequency range considered in this study and the vertical dotted line indicates the location of the peak frequency ($f = 0.03$ Hz). Note the different horizontal axis centered around the peak frequency in the bottom row.

Figure 11. The bed profile rate of change determined from the difference in the profile measurements for (dash) interval B and (solid) interval C of each case.

Figure 12. (red) The observed profile of the sediment diffusivity ε_s from the mean concentration profile measured at $x = 8.8$ m (on the reef flat) compared to the predicted diffusion profiles by Van Rijn (1993) for (green) the low-frequency waves, (dark blue) the high-frequency waves and (light blue) the total wave spectrum. The left column shows predictions using the peak period T_p , while the right column shows predictions using the mean period T_{m02} . Each row represents a different simulation. The horizontal black line is the mean ripple elevation.

Figure 13. Sediment transport balance of each simulation. The left column shows the relative contribution of each transport mechanism to the total transport. Q_{hi} (Q_{lo}) is the suspended transport for high (low) frequency waves, Q_m is the suspended transport by mean flow and Q_b is the sediment transport by bedform migration. The right column shows the estimated transport ($\sum Q$) and the transport measured by integration of the bed profile (Q_p) seaward and shoreward of the FOSLIM located at $x = 8.80$ m. The error bar indicates the uncertainty in the sediment volume change due to uncertainties in the bed elevation measurements propagated through the numerical integration.