Wave-driven Geomorphology of Pacific Carbonate Coastlines: From Landscape to Wavelength Scale

by

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Abstract

The shallow marine ecosystems of coral atolls and the human communities they support are among the most vulnerable to anthropogenic climate change. Sea-level rise threatens to inundate low-lying reef islands, tropical cyclone intensification threatens islands with flooding and erosion, and ocean warming and acidification threaten the health of coral reefs. Unfortunately, the sediment dynamics that shape the morphology of coral reefs and atoll reef islands are poorly understood, hindering predictions of coral atoll responses to climate change forcing. Here, I apply an eclectic set of methods, including numerical modeling, physical lab experiments, and sedimentological analysis, to produce insights into the ways tropical cyclones and waves move sediment on fringing reefs. First, I use a numerical model of hydrodynamics to predict the influence of sea-level rise and wave climate change on sediment transport across a coral atoll fringing reef. I demonstrate that by the end of the century, sea-level rise will reduce sediment transport rates from the fore reef to the beach, but increase transport rates from the reef flat to the beach. Wave climate change will have relatively negligible influence on cross-reef sediment transport. Additionally, I use the weathering of foraminifera tests to produce a sediment proxy of transport duration and direction across atoll reef flats, but demonstrate that the proxy does not clearly identify storm deposits. Second, I execute a series of experiments in an oscillating flow tunnel to constrain the rate at which sediment erodes reef surfaces under waves. I find that the erosion rate increases as a power law of wave orbital velocity, and that amount of sediment has a second-order influence. Finally, I establish grain size in a sediment core retrieved from a blue hole in the Marshall Islands as a proxy for tropical cyclone genesis and, using the results from an ensemble of climate models, demonstrate that enhanced tropical cyclogenesis during the Little Ice Age may have been driven by an anomalously negative Pacific Meridional Mode. This thesis demonstrates the importance of sediment dynamics on the morphology of fringing reefs and atoll reef islands and the sensitivity of those dynamics to centennial climate variability.

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Chapter 1

Introduction

All currently living Scleractinian (stony) coral reefs began growing less than 14,000 years ago (Montaggioni, 2005). Due to their symbiotic relationship with photosynthetic microorganisms, Scleractinian coral are generally restricted to ocean depths shallower than c. 50 m (Kleypas et al., 1999). Thus, during every glaciation over at least the past few million years, corals have retreated with sea level, leaving behind reefs that were constructed from their skeletons during the previous interglacial. And during every de-glaciation, corals have returned with rising sea level, re-populating the sides of their old reefs from the bottom up and thereby expanding the areal extent of individual reefs with each glaciation. Through the cycle of abandonment and re-population with glaciation-related sea-level fluctuations, such expanding reefs can form coral atolls (Toomey et al., 2016a; Dickinson, 2004), roughly donut-shaped, intertidal carbonate platforms that enclose shallow lagoons (Fig. 1-1). Thus, modern coral reef environments gain their form through the rapid (geologically speaking) adjustment of coral communities to changing environmental conditions. The dynamic nature of coral reefs extends to the reef islands they maintain.

For many of the currently 8 million Pacific islanders (GCRMN, 2018), coral reefs provide the very land beneath their feet. Atoll reef islands are constructed almost entirely from bits of dead coral, coralline algae, foraminifera, and other organisms that grow on atoll reefs. Additionally, the reefs that fringe these islands break and dissipate most ocean waves before they reach the shoreline, thereby protecting the islands

Figure 1-1: An aerial photograph of the eastern atoll rim on Jaluit Atoll, Republic of the Marshall Islands and an inset satellite image of the entire atoll, with the aerial photo extent outlined in red. The fringing reef can be broken into reef flat, reef crest, and fore reef (offshore of the reef crest) regions. Sediment is produced across the fringing reef, tends to be sparse on the reef flat and reef crest, and accumulates on the reef island and in the atoll lagoon.

from constant reworking and flooding. Despite this protection, and because they consist of unconsolidated sediment, the shorelines and geometry of reef islands can vary substantially on decadal timescales (Ford and Kench, 2015; Webb and Kench, 2010), and occasionally storms have eroded whole islands from the atoll rim in a matter of hours (Ford and Kench, 2014). Thus, atoll reef islands and their inhabitants are considered particularly vulnerable to sea-level rise (SLR) and the intensification of tropical cyclones expected from global climate change (Knutson et al., 2019). Determining the likely response of reef and reef island morphology to climate change should thus be a priority.

Unfortunately, the physical processes and biophysical interactions governing the formation, evolution, and maintenance of atoll reefs and reef islands are poorly understood. For example, only recently have field studies identified the source of large wave bores on reef flats due to resonant amplification, and the origin of some low frequency waves remains unknown. Despite recent flume studies, the processes governing sediment transport across fore reefs and reef flats are poorly understood, and only a few limited field studies have empirically determined even the primary direction of transport across atoll reefs. Some processes thought to govern reef morphology, such as wave-driven abrasion, have never been quantified. Finally, independent reconstructions of climatic and environmental conditions during reef island formation and evolution are sparse, constraining our ability to confidently reconstruct reef island geomorphological processes. In this thesis I attempt to increase our understanding of a selection of these processes.

1.1 Wave dynamics and sediment transport on fringing reefs

Despite the protection from large waves afforded by their fringing reefs, reef islands still flood occasionally due to wave-driven hydrodynamics. The waves breaking on the reef crest induce a radiation stress gradient between the fore reef and reef flat that

raises the water surface elevation on the reef flat relative to the mean ocean surface (Vetter et al., 2010), a phenomenon known as wave setup (Fig. 1-2). Additionally, variations in the height of high-frequency waves (wave period = 1-30 s) that break on the reef crest generate low-frequency waves (wave period > 30 s) through a process called breakpoint forcing (Pequignet et al., 2014; Symonds et al., 1982). The low-frequency waves can form standing waves on the reef flat or grow across the reef flat through resonant amplification (Fig. 1-2) (Gawehn et al., 2016; Becker et al., 2016; Pequignet et al., 2014) and occasionally form tsunami-like wave bores that flood reef islands (Cheriton et al., 2016; Roeber and Bricker, 2015; Hoeke et al., 2013). These wave dynamics have received heightened scrutiny over the past decade, but still there are gaps in our knowledge. In particular, although their role in shaping island morphology through overwash is well documented (e.g. Tuck et al., 2019b), very little attention has been given to the delivery of sediment from the fringing reef to the beach.

On atoll reef islands, the beaches are made of sediment that is produced on the fringing reef flat and fore reef and then transported across the reef to the beach. Sediment already on the beach is reworked and can be removed from the beach by alongshore transport and overwash, wherein water overtops the beach crest and floods the island interior, carrying beach sediment with it. Thus, reef island shorelines erode due to an increase in alongshore transport gradients, enhanced overwash, reduced sediment production on the surrounding reef flat, or reduced onshore transport of that sediment (Fig. 1-3). The sensitivity of reef island shorelines to these processes contributes to the dynamic nature of reef islands. Some reef island shorelines exhibit seasonal cycles of morphological change (Dawson et al., 2014). Many reef islands have exhibited significant increases in island area over the 20th century, though many more have exhibited decadal shoreline change with no net change in area (Kench et al., 2018b; Webb and Kench, 2010). The increase in island area synchronous with SLR has led to some optimism regarding the fate of reef islands in the future (McLean and Kench, 2015). However, reef island responses to SLR and other aspects of climate change will not necessarily be linear, and experimentation is required to predict future

Figure 1-2: Wave evolution across a fringing reef as modeled with XBeach. A) Model bathymetry. B) Significant wave height of high frequency (0.3-0.03 Hz) waves. C) Significant wave height of low frequency (0.03-0.001 Hz) waves. D) Elevation of the water surface relative to mean sea level (MSL).

Figure 1-3: Schematic illustration of the sediment balance maintaining the ocean-facing beaches of atoll reef islands. Sediment is produced on the outer reef flat and fore reef, transported cross-shore to the island beach, and from there is transported in the alongshore direction or onto the island surface through overwash.

reef island responses.

Although an important component of the maintenance of reef island shorelines, the response of cross-shore sediment transport to climate change remains unexamined. Recently, physical models in a wave flume and large wave tank demonstrated that reef island beach crests respond to SLR by retreating lagoon-ward and increasing in elevation primarily through overwash (Tuck et al., 2019b,a). Recent studies used numerical models to examine the response of alongshore sediment transport gradients to SLR and wave climate change and found that existing gradients could be expected to intensify over the next century, resulting in extension of reef island spits and erosion of windward shorelines (Shope et al., 2017; Shope and Storlazzi, 2019). Relative to these components of the reef island sediment budget, however, sediment production and transport cross-reef to reef island shorelines has received less attention. Flume and field studies suggest that cross-shore sediment transport on reefs is driven by high frequency waves and that the high frequency waves are modulated by low frequency waves (Pomeroy et al., 2015, 2018), but have not explored how these dynamics change with SLR or wave climate. In Chapter 2 we use a numerical modeling case study to answer the question: How will cross-shore transport respond to changes in wave

climate and SLR forecast for the next century?

In part, research into sediment transport processes on atoll reefs is hindered by the sparse distribution of sediment across these reef environments. Sediment transport algorithms often used for beaches and shorefaces along e.g. passive margins are inapplicable, as they assume sediment availability for transport is constant and often uniform (Soulsby and Damgaard, 2005; McCall et al., 2014; Bowen, 1980). However, on coral reefs, loose sediment is often contained in scattered deposits and can be virtually absent on atoll reef flats the majority of the time. Additionally, deposits of sediment on atoll reefs can vary over very short time periods. For example, on the fringing Ningaloo reef in Australia, the central reef flat often has a thin layer of sediment that is transported in bedload, but when large storm waves propagate across the reef, they clear the outer and middle reef flat of sediment, leaving it bare (Rosenberger et al., 2019). Storms can also deposit large amounts of coarse sediment on the reef flat that is then reworked across the reef flat and to the beach (Shannon et al., 2013; Blumenstock et al., 1961b). Attempts to identify transport pathways across the reef have included tracking painted cobbles on the fore reef (Kan, 1994), installing passive integrated transponder (PIT) tags in large coral cobbles and relocating them periodically (Ford, 2014; Kench et al., 2017), and using the degree of weathering of sediment with known geographic origins as a proxy for transport time or distance (Dawson et al., 2014; Fellowes et al., 2017). To determine the dominant sediment transport pathways across the northwestern fringing reef flat and blue hole on Jaluit Atoll, Republic of the Marshall Islands, I analyze the degree of weathering of foraminifera tests produced at reef margins in Chapter 3. Additionally, I explore the possibility of using foraminifera weathering as a proxy for storm-deposited sediment layers in a blue hole sediment core.

1.2 Wave-driven abrasion of submarine bedrock

The transport and fate of sediment on atoll reefs has implications not just for sediment sinks like atoll reef islands and lagoons, but also for the sediment-producing atoll reefs

Figure 1-4: Aerial photograph of grooves on the eastern rim of Jaluit Atoll, Republic of the Marshall Islands. Grooves are channel-like structures found on the fore reefs of most living coral reefs globally, and are hypothesized to form through wave-driven abrasion.

themselves. Sediment that collects on fore reefs can smother corals (Erftemeijer et al., 2012) or contribute to wave-driven abrasion, in which sediment is rubbed against and erodes coral and the reef surface by waves, like sandpaper. Wave-driven abrasion on reefs produces spatial variability in net reef accretion rates (Engels et al., 2004) and likely limits mean net reef accretion rates (Grossman and Fletcher, 2004). Despite its influence on the morphology of coral reefs (Fig. 1-4), wave-driven abrasion rates have not been quantified for reef environments. Additionally, exactly how wave-driven abrasion rates vary with environmental conditions like wave characteristics and sediment availability remains unknown. In Chapter 4, I describe a series of physical lab experiments I conducted to quantify the influence of wave orbital velocity (the speed of near-ocean-bottom back-and-forth motions under waves) and sediment availability on abrasion rates.

1.3 Reconstructing tropical cyclone activity in the western North Pacific

Tropical cyclones are low pressure storms whose winds generate the largest waves incident on atoll reefs. Over just a few hours, the waves generated by tropical cyclones can effect geomorphological change with existential implications for reef islands. Where an atoll reef flat is initially bare, tropical cyclones can rip large coral boulders off of the atoll fore reef and deposit them on the reef flat to act as the seeds for new reef islands, with finer-grained sands and cobbles collecting in the lee and interstices of such boulders to form persistent landforms (e.g. Kench et al., 2018b). Where an atoll reef flat supports an island, tropical cyclone-generated waves can erode the island completely from the reef flat (e.g. Ford and Kench, 2014), erode large channels through the island (e.g. Blumenstock et al., 1961a), or deposit a large quantity of coarse sediment on the reef flat that eventually adds to island volume (e.g. Shannon et al., 2013). Tropical cyclones also shape fringing reef morphology, contributing to prominent, crustose coralline algae-covered reef crests and conglomerate platforms that underlie some reef islands (Woodroffe, 2008).

To determine how tropical cyclones may have influenced atoll and reef island formation, it would help to have some notion of spatial and temporal variability in tropical cyclone activity over the Late Holocene. However, the observational record of tropical cyclones is short, and prior to satellite remote sensing in the 1970's is also quite sparse. Geological proxy records of tropical cyclone activity can extend our observational records over thousands of years. Unfortunately, although many proxy records have been constructed in the North Atlantic (e.g. Donnelly et al., 2015), few have been produced for the western North Pacific, despite the fact that the latter ocean basin is the most active globally for tropical cyclones. In Chapter 5 I develop a sediment-based proxy reconstruction of tropical cyclone genesis in the western North Pacific and examine the output from a series of experiments with global climate models to identify likely climate factors responsible for temporal variability in the proxy record.

Chapter 2

Modeling the impacts of a changing climate on cross-shore sediment transport: Kwajalein Atoll, Marshall Islands

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Abstract

Atoll reef islands primarily consist of unconsolidated sediment, and their ocean-facing shorelines are maintained by sediment produced on their fringing reefs, which primarily consist of coral and consolidated carbonate. Changes in the waves that propagate across the fringing reef can alter net cross-shore sediment transport rates and thus affect the sediment budget and morphology of atoll reef islands. Here we investigate the influence of sea-level rise and projected wave climate change over the coming century on potential cross-shore sediment transport across the fringing reef on Kwajalein Island, Republic of the Marshall Islands using a phase-resolving hydrodynamic model. For the current reef flat geometry, sea level, and wave climate, on- and off-shore components of the potential bedload transport are nearly balanced. Mean annual wave energy incident on the island is projected to decrease by 12% by the end of the

century, and we find this decrease has negligible influence on potential cross-shore sediment transport rates on the fringing reef flat and beach, as waves on the reef flat are strongly depth-limited. However, 0.5-2.0 m of sea-level rise leads to greater high-frequency wave heights, skewness, and associated shear stress on the fringing reef flat, with a decrease of these parameters on the fore reef. These changes in hydrodynamics increase potential sediment transport from the outer reef flat to the beach, but decrease potential sediment inputs onshore from the fore reef. The effect of projected wave climate change and sea-level rise on reef flat and reef island morphology will depend on the availability and production of sediment on the fore reef and reef flat itself.

2.1 Introduction

Anthropogenic climate change over the coming centuries threatens the habitability of atoll reef islands and is expected to alter island sediment transport dynamics. Predictions of atoll reef island response to climate change have relied on extrapolation from recent trends in remotely sensed images and examination of reef- and island-building timelines in the geological record (e.g. McLean and Kench, 2015). However, the processes governing sediment production, transport, and fate on atolls remain poorly understood. Here we focus on Kwajalein Island in the Republic of the Marshall Islands, using wave and hydrodynamic modeling to investigate how changes in wave climate and increased sea level driven by climate change are likely to alter cross-shore sediment transport and affect the sediment budgets of ocean-facing reef island beaches.

Atoll reef islands are dynamic landforms typically composed of unconsolidated carbonate sediment that reach an elevation of no more than 2-5 m above mean sea level (aMSL) (Woodroffe, 2008). Reef islands are distributed non-uniformly on the annular, intertidal carbonate platforms of mid-ocean coral atolls and are vulnerable to oceanic natural hazards such as tropical cyclones and tsunamis, which can form or erode islands entirely (Ford and Kench, 2015). Due to their low elevation, unconsolidated composition, and natural hazard risk, atoll reef islands and their human inhabitants are considered particularly vulnerable to hazards caused by anthropogenic climate change.

The International Panel on Climate Change (IPCC) forecasts climate-change-associated global sea-level rise (SLR) of up to 1 m by 2100 (IPCC, 2013). Due to the sea-level fingerprint of melting ice sheets, SLR in the tropical Marshall Islands could exceed 2 m under the same IPCC emissions scenarios (Kopp et al., 2017). Currently, the fringing fore reefs and reef flats that separate reef islands from the open ocean dissipate most waves before they reach island shorelines (Quataert et al., 2015). However, SLR will allow larger waves to impinge on the shoreline, resulting in more frequent flooding of atoll reef islands (Cheriton et al., 2016; Storlazzi et al., 2015b; Merrifield et al., 2014); episodic flooding of the islands and saltwater intrusion through the karstified atoll platform can disrupt subsistence agriculture and the freshwater lenses on which it depends (Oberle et al., 2017; Storlazzi et al., 2018; Hoeke et al., 2013).

As global atmospheric circulation patterns change with climate, so too will the directional distribution and height of waves incident on fringing reefs and island shorelines (Shope et al., 2016), likely forcing island morphological change (Shope et al., 2017; Shope and Storlazzi, 2019). Additionally, the combination of SLR and climate change-driven increases in the frequency and intensity of tropical cyclones (e.g. Zhang et al., 2017) will likely result in more frequent and disruptive modification of atoll reef island morphology (Ford and Kench, 2016). Although these changes may make human habitation difficult, they do not necessarily pose an existential threat to the island landforms themselves.

The majority of atoll reef islands in the Pacific formed during or after a sea level highstand that lasted 5-1.5 kya, during which sea level was 1-3 m aMSL in the equatorial Pacific Ocean (Dickinson, 2009; Grossman et al., 1998). With the premise that few modern islands existed on atoll platforms during rapid post-glacial SLR and the sea-level highstand, it has been proposed that climate change-driven SLR could cause erosion of atoll reef island shorelines and eventually lead to the permanent removal of entire islands from atoll rims (Dickinson, 1999, 2009). However, research using mid-Holocene histories of atoll reef islands as proxies for island response to higher sea level demonstrate that some islands formed, grew, and remained stable under the

sea-level highstand (Kench et al., 2012, 2014). Furthermore, under recent, accelerating SLR (0.013 mm yr^{-2} , Church and White, 2006), many islands have expanded in area (McLean and Kench, 2015). However, trends in island size are mixed, with some island accretion decelerating or reversing in recent decades (Rankey, 2011; Ford, 2013; Ford and Kench, 2015). In addition, even as island planform area increases with SLR, this accretion is not uniform, with the ocean-facing shorelines often experiencing erosion (Webb and Kench, 2010). In the Marshall Islands in particular, islands have increased in planform area over the past 70 years, although since the 1970’s this trend has reversed for many islands (Ford, 2013; Ford and Kench, 2015). The absence of a consistent shoreline response to modern, higher rates of SLR on atoll reef islands suggests that either climate-change-driven SLR has so far not had measurable net impact on shoreline change or that shoreline response is moderated by local conditions (Rankey, 2011).

The specific drivers of observed shoreline changes remain unknown, in part due to the low temporal resolution of the satellite and aerial image record used to track shoreline change (Ford, 2013) and the complexity of atoll island shoreline dynamics. The shorelines of atoll reef islands are dynamic, often responding to variability in wave climate on seasonal and inter-annual timescales with little to no net change in island area or volume (Rankey, 2011; Dawson and Smithers, 2010). However, in a time-mean sense, the ocean-facing shorelines of atoll reef islands are maintained by a balance in which sediment produced on the fringing fore reef and reef flat is delivered cross-shore to the beach and sediment on the beach is removed through alongshore transport and episodic overwash (Tuck et al., 2019a; Perry et al., 2011). The rate of cross-shore sediment input to the shoreline is in turn a function of sediment production on the fringing reef and the hydrodynamic flows that deliver sediment to the shoreline. To investigate the hydrodynamic flows responsible for sediment delivery, we use the concept of potential sediment transport, i.e. the expected sediment transport rate assuming an inexhaustible volume of sediment. Shoreline erosion thus occurs due to a sediment budget deficit, which from a balanced initial state can be caused by one or a combination of the following: enhanced net alongshore sediment transport

gradients, enhanced overwash, reduced reef sediment production, or reduced potential cross-shore sediment transport.

The processes composing the sediment budgets that maintain island morphology are understudied. In terms of sediment removal from the beach, studies have nearly unanimously predicted greater overtopping and, thus, inferred overwash frequency on atoll island shorelines due to SLR over the next century (e.g. Merrifield et al., 2014; Cheriton et al., 2016; Shope et al., 2017; Storlazzi et al., 2018). Recent wave flume and wave tank studies suggest that reef island morphodynamics exhibit a negative feedback relationship with overtopping, in which island freeboard elevation keeps up with SLR through overwash-induced landward migration and vertical accretion (Tuck et al., 2019a). Additionally, Shope et al. (2017) and Shope and Storlazzi (2019) examined how changing wave climate and SLR may affect net alongshore transport on atoll reef islands and predict that changing wave direction and SLR-enhanced alongshore transport gradients will redistribute sediment along ocean-facing and lagoon shorelines, resulting in net erosion of windward, ocean-facing shorelines. Similar to other nearshore environments, the cross-shore transport component of the sediment balance is particularly poorly understood. A wave flume experiment found that cross-shore sediment transport resulted from skewness and asymmetry of both high- and flow-frequency waves, with the relative dominance of these contributors varying over the reef profile (Pomeroy et al., 2015). Additionally, recent field studies have highlighted the problems with applying conventional sediment transport flux equations to reef environments (Pomeroy et al., 2018, 2017). Recent hydrodynamic modeling using hypothetical reef flat geometries suggests that internal feedbacks can control reef flat elevations and widths for a given constant sea level (Ortiz and Ashton, 2019).

In this study we quantify the impact of SLR and wave climate change on atoll island hydrodynamics and potential cross-shore sediment transport. First, we quantify changes in wave climate over the 21st century for Kwajalein Atoll, in the Republic of the Marshall Islands, as forecasted by an ensemble of global climate models. Then, we calibrate the phase-resolving hydrodynamic wave model, XBeach, for Kwajalein’s windward fringing reef and verify the model’s accuracy. Finally, we use the XBeach

model to quantify the response of bottom shear stress and inferred potential sediment transport to both wave climate change and SLR.

2.2 Study Area

Kwajalein Atoll is a large coral atoll in the Ralik island chain of the Republic of the Marshall Islands (RMI), in the central North Pacific (Fig. 2-1). It has the largest lagoon in the world with a surface area of over 2300 km² (Sugerman, 1972). The lagoon is surrounded by a 300-1500 m wide intertidal reef platform, broken periodically by deep channels connecting the ocean to the lagoon. Beyond the reef platform is a fore reef consisting of coral-algal boundstone and unconsolidated or weakly consolidated sediment (Hunt Jr. et al., 1995).

Kwajalein Atoll is partially sheltered from wave energy by its position within the Marshall Islands. Wave generation by trade winds incident on Kwajalein Atoll is fetch-limited by the Ratak island chain 250-300 km to the east, lowering mean incident wave height relative to atolls in the eastern chain (Fig. 2-1a). Similarly, northern RMI atolls block some swell generated by extratropical cyclones from reaching Kwajalein Atoll, but northerly swell waves occasionally flood islands on the atoll (Quataert, 2015). Tropical cyclones locally generate large waves, but historically these large wave events have been rare. Twelve storms recorded in the International Best Track Archive for Climate Stewardship (IBTrACS) have come within 100 km of Kwajalein Island since 1850. At least two of those tropical cyclones, Typhoons Roy (1988) and Zelda (1991), generated waves that overtopped the ocean-facing beach ridges on Kwajalein Island and nearby Ebeye Island (Center, 1988, 1991). Colonial records also indicate that a tropical cyclone inundated Kwajalein Atoll in 1875, reportedly sweeping the island of all inhabitants (Spennemann, 2004).

Figure 2-1: Maps of the study area showing a) mean significant wave height (H_s) from the CFSRR reanalysis, b) Kwajalein atoll bathymetry, and c) the location of LIDAR data and wave sensors used to construct and evaluate the XBeach model.

2.3 Data and Methods

2.3.1 Wave climate analysis

To examine the wave climate incident on Kwajalein Atoll, we used two sources of modeled incident wave data. First, we used the Climate Forecast System Reanalysis Reforecast (CFSRR) Wavewatch III hindcast (Chawla et al., 2012) to classify the source of incident waves. The National Oceanic and Atmospheric Administration (NOAA) used CFSRR winds generated at 0.5 degree resolution to drive Wavewatch III reanalysis hindcasts of significant wave height (H_s), peak wave period (T_p) and mean wave direction (θ_{mean}) globally for the 1979-2009 period. We compiled all thirty years of wave records from the grid cell centered at (8.75°N, 168.25°E), the open ocean cell nearest to the southeastern (windward) rim of Kwajalein Atoll, offshore of Kwajalein and Ebeye Islands.

Second, to examine changes in wave climate we used Wavewatch III forecasts driven by four general circulation models (GCM) from the Coupled Model Intercomparison Project 5 (CMIP5) (Storlazzi et al., 2015a). The four GCMs used were the Beijing Climate Center Community Systems Model 1.1 (BCC), Institute for Numerical Mathematics Coupled Model 4 (INMCM), Model for Interdisciplinary Research on Climate 5 (MIROC), and the Geophysical Fluid Dynamics Laboratory Earth Systems Model 2M (GFDL). For each GCM, waves were simulated for three time periods: historical (1976-2005), mid-21st century (2026-2045), and late 21st century (2085-2100). To capture the most extreme potential changes in wave climate, we examined only GCM results for the Representative Concentration Pathway 8.5 (RCP8.5) climate scenario representing unchecked carbon emissions and resultant radiative forcing over the 21st century. The GCM forecasts generated wave characteristics at 1.00×1.25 degree spatial resolution; we used data extracted from the grid cell centered at (8.75°N, 167.75°E). For each of the time periods, we combined the wave data from all four GCMs to create a model ensemble.

To quantitatively describe wave climate, we separated waves into classes using k-means unsupervised classification (Xu and Wunsch II, 2005) of linear wave energy

density, T_p , and θ_{mean} . The number of classes produced, k , was chosen as that which explained 90% of the variance in the three wave characteristics. Prior to classification, the values of the three wave characteristics were transformed to z-scores so their values would all be of the same order of magnitude. This classification procedure was performed separately for each time period (historical, mid-21st century, and late-21st century) and each dataset (CFSRR reanalysis, individual GCMs, and the model ensemble).

We identified the source for each wave class for the CFSRR hindcast by viewing basin-wide animations of wave characteristics. Pacific-wide results for the GCM forecasts were not available to similarly identify wave classes. Therefore, each GCM wave class was inferred to have the same source as the CFSRR wave class with the most similar wave characteristics, determined with nearest neighbor classification.

2.3.2 Wave model setup

To examine the influence of incident wave climate and its future variability on cross-shore sediment transport, we used XBeach to model wave propagation and associated hydrodynamics across Kwajalein’s fringing reef. XBeach is a nonlinear shallow water equation solver forced by time-varying spectral wave and flow boundary conditions (Roelvink et al., 2009) and has been tested and applied to fringing reef environments similar to Kwajalein Island (Lashley et al., 2018; Quataert et al., 2015). We ran XBeach in 1D non-hydrostatic mode, such that the model resolved both high- and low-frequency wave motions across a cross-shore transect of the reef. When compared with low-friction laboratory flume models of fringing reefs, XBeach accurately simulates wave height, water level, and runup with default model parameters (Lashley et al., 2018). However, when used to simulate rough fringing reefs, XBeach requires tuning of a coefficient of friction (c_f) that is used to calculate bottom shear stress (τ_b , Nm^{-2}) as:

$$\tau_b = \rho c_f u |u| \quad (2.1)$$

Figure 2-2: Model bathymetry, tidal datum, the location of wave sensors on Kwajalein Island's fringing reef, and identification of reef zones.

where ρ is water density (kg m^{-3}) and u is cross-shore water velocity (m s^{-1}) (Roelvink et al., 2009).

To tune c_f and verify that our model produced realistic results, we used data from an array of wave sensors deployed across the fringing reef at Kwajalein Island. One Nortek acoustic wave and current (AWAC) profiler and four pressure sensors (RBR Virtuosos) were deployed on the eastern fringing reef of Kwajalein Island from 9 November 2013 to 19 April 2014 (Fig. 2-1). The AWAC and pressure sensors were set to sample for 2048 s every hour at 1 Hz and 2 Hz, respectively. Sensors were geolocated using post-processed Global Navigational Satellite System (GNSS) data, and sensor elevations on the reef flat were surveyed to the Kwajalein Island tide gauge and calculated relative to local mean sea level during deployment (Fig. 2-2). For model tuning and verification, 17 d of sensor data (8-25 March 2014) were selected to cover the transition from neap tide to spring tide. Waves during this interval were characteristic of trade wind waves and were representative of the entire deployment with a mean ($\pm$ standard deviation) $H_s = 1.5 (\pm 0.5)$ m, $T_p = 7.7 (\pm 1.75)$ s, and $\theta_{mean} = 97^\circ (\pm 9^\circ)$.

We tuned model c_f using 48 h (13-15 March 2014) of AWAC data transformed to Joint North Sea Wave Project (JONSWAP) spectra defined by hourly H_s , T_p , and θ_{mean} . We then compared hourly H_s and water surface elevation between the model

and sensors on the fore reef and reef flat using root mean squared error (RMSE) and the Pearson product-moment correlation coefficient (r) as objective functions. Hourly H_s were calculated from 2048 s of sensor or model data as:

$$H_s = 4 \sqrt{\int_{f_1}^{f_2} S(f) df} \quad (2.2)$$

where f is frequency (Hz), f_1 and f_2 are frequency bounds defining high (0.3-0.04 Hz) or low (0.04-0.001 Hz) frequency waves, and S is the 1D wave-energy spectra ($\text{m}^2 \text{s}$) calculated using the Welch's modified periodogram method with a Hanning window of 1024 s. The tuning data included the largest recorded wave height of the deployment at $H_s = 2.85$ m, and tides of intermediate magnitude. The remaining 15 d of selected wave data were used to verify model accuracy over a broader range of wave and tide conditions.

High-resolution bathymetric data were unavailable along the transect of wave sensors fronting Kwajalein Island, so model bathymetry was derived from an airborne LIDAR transect across nearby Ebeye Island (Fig. 2-1c), where reef flat width and fore reef slope were equal to that along the wave sensor transect. The LIDAR data were linearly interpolated to a uniform grid with 0.5 m spatial resolution (Fig. 2-2). Topographical variability on the fore-reef was removed as in (Quataert et al., 2015) to make results more generalizable. Depths were uniformly tied to local mean sea level during deployment by lowering the cross-section until it intersected the elevation of the outer reef flat sensor. Sensor locations were matched to grid cells based on their distance from the reef crest. The inner reef flat sensor aligned well with model bathymetry, but the mid-reef flat sensor was deployed in an anomalous depression and was located 22 cm below model bathymetry (Fig. 2-1). Despite this discrepancy, we used data from all sensors for model evaluation.

2.3.3 Model scenarios and statistics

To determine the potential impact of wave climate change on cross-reef hydrodynamics and sediment transport, we ran the XBeach model under six scenarios. First,

as a baseline we forced the model with every wave class from the historical GCM ensemble (“Historical”). Second, we forced the model with every wave class from the late 21st century GCM ensemble, with sea level held at historical levels (“Wave Climate Change”). Finally, we forced the model with the historical GCM ensemble wave classes for each of four SLR magnitudes: +0.5, +1.0, +1.5, and +2.0 m relative to local MSL.

The SLR scenarios considered in this study cover the range of projected SLR by 2100 for Kwajalein Island. The IPCC Fifth Assessment Report (AR5) projected likely global SLR by 2100 of +0.74 (+0.53) m for representative concentration pathway (RCP) 8.5 (4.5) (IPCC, 2013). However, due to feedbacks between water storage, gravity, lithospheric loading, and planetary rotation, SLR can vary widely across the globe for a given increase in ocean volume (Hsu and Velicogna, 2017; Mitrovica et al., 2009). This variability, or sea-level fingerprint, results in greater forecast SLR rates in the tropics relative to the rest of the globe. After accounting for the sea-level fingerprint and incorporating expert opinion into a statistical model of SLR, Kopp et al. (2014) forecast mean (90% confidence interval) SLR at Kwajalein Island by 2100 CE of +0.89 (+0.42-1.58) m for RCP8.5 and +0.67 (+0.31-1.21) m for RCP4.5. However, these forecasts were recently updated with physical modeling of ice sheet dynamics, producing mean SLR at Kwajalein Island of +1.72 (+1-2.93) m for RCP8.5 and +1.07 (+0.54-1.89) m for RCP4.5 (Kopp et al., 2017). Thus, Kwajalein Island is expected to experience at least +0.5 m and up to nearly +2 m of SLR by 2100.

Using 4096 s-long, 1-Hz time series of wave velocity and water depth after 5 h of model spin-up for each wave class, $H_{s,hf}$ and $H_{s,lf}$ were calculated from the output water depth time series using conventional spectral analysis as in Section 4. We then separated wave velocity into low- and high-frequency components using ensemble empirical mode decomposition (EEMD, Wu and Huang, 2009). EEMD uses an iterative peak detection and curve fitting analysis to separate a discrete time series into a number of intrinsic mode functions (IMF), containing periodic components of the signal, with a non-periodic residual. The individual IMFs and residual do not share information, such that they can be summed linearly to reproduce the original

time series (Huang and Wu, 2008). The IMFs tend to contain information from within narrow frequency bands, but do not explicitly account for signal frequency. Thus, low-frequency IMFs can be combined to reproduce the highly nonlinear and irregular low-frequency wave shapes found on reef flats more accurately than conventional band-pass filters (Gawehn et al., 2016). We used EEMD to decompose each velocity time series into 10 IMFs and then applied the Hilbert-Huang Transform (Huang and Wu, 2008) to determine the time-dependent frequency content of each IMF. An IMF was classified as low-frequency if its mean frequency fell below 0.05 Hz (> 20 s wave period). We then constructed the high-frequency component (u_{hi}) by summing all high frequency IMFs with the EEMD residual and the low-frequency component (u_{lo}) by summing the remaining IMFs.

Distributions of shear stresses were built through bootstrap sampling of model results from all of the wave classes. The entirety of each water velocity time series was included multiple times according to the proportion of the year for which its wave class accounted. For example, all model output from a wave class incident on Kwajalein Island 0.1% of the year would be included once, while all model output from a wave class accounting for 30% of the year would be included 300 times. The resulting bootstrapped distributions had a total sample size of $n = 4.01 \times 10^7$. Shear stress was then calculated using Eq. 2.1, and a probability density function (PDF) estimated from the full samples.

To determine the influence of wave non-linearity on shear stress distributions, we diagnosed wave skewness (S_u) using the third-order moment of water velocity and asymmetry (A_u) using the third-order moment of the Hilbert transformation of water velocity (Ruessink et al., 2012):

$$S_u = \frac{\overline{u^3}}{\sigma_u^3} \quad (2.3)$$

$$A_u = \frac{\overline{H(u)^3}}{\sigma_u^3} \quad (2.4)$$

where σ_u is the standard deviation of velocity, $H()$ is the Hilbert transform, and the overbar represents time averaging, both calculated over an entire model output time

series.

2.3.4 Calculating potential sediment transport

To quantify potential net sediment transport rates and their response to changing wave climate and SLR, we use empirical expressions for bedload flux and suspended transport velocity. We estimated potential bedload transport rates using the general sediment bedload flux equation of Soulsby and Damgaard (2005):

$$q_j = A_2 \frac{\sqrt{\tau_b} \tau_{e,j}}{g \rho^{3/2} (s - 1)} \quad (2.5)$$

$$\tau_{e,j} = \frac{\tau_b}{|\tau_b|} \max(0, |\tau_b| - \tau_{c,j}) \quad (2.6)$$

where q is volumetric bedload flux per unit width ($\text{m}^2 \text{s}^{-1}$), j is a subscript denoting different grain sizes, τ_e is excess shear stress (N m^{-2}), g is gravitational acceleration (m s^{-2}), s is the sediment specific gravity, $A_2 = 12$ is an empirical constant, and τ_c is a critical shear stress (N m^{-2}) threshold above which sediment is mobilized. Soulsby and Damgaard (2005) found that when Eq. 2.5 was calculated using instantaneous shear stress and then phase-averaged, it produced accurate estimates of net bedload flux for both currents and nonlinear waves.

We estimate the critical shear stress threshold for a given sediment grain size and density using the Shield's parameter:

$$\theta_c = \frac{\tau_{c,j}}{\rho(s - 1)gD_j} \quad (2.7)$$

where D is characteristic grain diameter (m). For turbulent, steady flow over a horizontal bed, θ_c has been empirically determined to be $O(0.05)$ and varies weakly with the Reynolds number of the flow (Fredsoe and Deigaard, 1992). Komar and Miller (1975) demonstrated that mobilization thresholds predicted using the Shield's parameter generally hold for oscillatory flow as well, and thus we use $\theta_c = 0.06$ to calculate the critical shear stress for sediment specific gravity of 1.85 and for six median grain sizes: 0.062 mm, 0.5 mm, 2 mm, 16 mm, and 64 mm, which respectively

represent the silt/sand, medium/coarse sand, sand/pebble, medium/coarse pebble, and pebble/cobble grain size transitions. We neglect correction of Eq. 2.5 and Eq. 2.7 for bed slope effects, as slopes are small and we hold bed shape (and slope) constant for all scenarios.

Unlike bedload transport, suspended sediment transport is sensitive to phase differences between shear stress and near-bottom orbital velocity (Fredsoe and Deigaard, 1992). These two processes, in addition to wave skewness, often result in net onshore suspended sediment transport even when mean water velocity is directed offshore due to return flow (Pomeroy et al., 2015). To quantify changes in potential suspended sediment transport with wave climate change and SLR, we estimated the mean direction and magnitude of suspended transport using a representative transport velocity (van Thiel de Vries, 2009; Pender and Karunaratna, 2013) computed from wave statistics:

$$u_{rep} = u_m + u_{lo} + \alpha u_{rms}(S_u - A_u) \quad (2.8)$$

where u_m is the mean flow velocity (m s^{-1}); u_{lo} is the low frequency wave component of velocity; u_{rms} is the root-mean-square of the high frequency wave component of velocity, u_{hi} ; S_u and A_u are the wave skewness and asymmetry defined above; and $\alpha = 0.1$ is an empirical coefficient whose value was estimated through calibration runs with XBeach (van Thiel de Vries, 2009). To account for possible covariance between low frequency and high frequency wave motions, u_{rms} , S_u , and A_u were all calculated within a 120-s moving window on the u_{hi} time series produced by EEMD.

Equations 2.5 and 2.8 were derived and calibrated for low gradient beds of sediment where the quantity of bed sediment is assumed inexhaustible. However, sediment is relatively sparse on the reef flat and fore reef slope at our site, and thus the sediment bedload transport flux found with Eq. 2.5 represents an estimate of the maximum potential bedload transport rate, as our model assumes a constant volume of sediment available for transport. We use Eq. 2.8 to quantify direction and only relative changes in suspended transport rates.

2.4 Results

2.4.1 Wave sources and forecasts

Using the CFSRR wave hindcasts to characterize historical wave climate at Kwajalein Atoll, we identified four primary sources of waves: local trade wind-driven waves (TW waves; mean $H_s = 1.8$ m, $T_p = 8.4$ s, $\theta_{mean} = 76^\circ$), swell produced by northern hemisphere extratropical cyclones (NH waves; mean $H_s = 2.0$ m, $T_p = 11.4$ s, $\theta_{mean} = 21^\circ$), swell produced by southern hemisphere extratropical cyclones (SH waves; mean $H_s = 1.3$ m, $T_p = 12.6$ s, $\theta_{mean} = 171^\circ$), and wind waves produced by nearby tropical cyclones (TC waves; mean $H_s = 2.7$ m, $T_p = 8.5$ s, $\theta_{mean} = 160^\circ$). TW waves dominate incident wave energy, followed closely by NH waves (Fig. 2-3). Due to the rarity of tropical cyclones passing near Kwajalein over the 1979-2009 CE time period, TC waves make up only 2% of mean wave energy, despite their large height. SH waves tend to be small and account for only 3% of the mean wave energy.

The historical GCM ensemble produced waves similar to the CFSRR waves. Although the ensemble tended to underestimate wave height and wave energy, the classification analysis identified wave sources in the ensemble with the same relative proportions as in the CFSRR data (Fig. 2-3). However, there was considerable inter-model variability in the GCM wave data. Waves generated with BCC and GFDL GCMs had greater energy than the reanalysis, whereas waves generated with INMCM inputs had much smaller wave heights, decreasing the mean wave energy of the ensemble. Additionally, TW waves accounted for proportionally far greater wave energy in BCC and GFDL historical datasets than NH waves, and vice versa for the INMCM historical dataset.

Despite inter-model variability, a trend of decreasing mean wave energy over the next century was consistent across all models. The GCMs unanimously forecast a 4-16% decrease in mean wave energy by 2100, primarily due to a 12-58% reduction in NH wave energy. The ensemble mean forecasts a 12% reduction in mean wave energy over the next century, which is equivalent to one standard deviation of annual mean wave energy in the historical GCM ensemble. In the ensemble mean, TC and SH wave

Figure 2-3: Mean wave energy density categorized by primary wave source, dataset, and time period. Percentages indicate the proportion of mean wave energy density accounted for by each wave source within a time period.

energy were forecast to increase slightly, and TW wave energy to decrease slightly, although there was less inter-model agreement on these changes. The wave directional distributions were forecast to narrow for all wave sources except TC waves, and to shift slightly counter-clockwise for all wave sources except TW waves. As the forecast changes to wave climate are similar for Majuro and Bikini Atolls (Storlazzi et al., 2015b), the analysis presented here is likely applicable across the Marshall Islands.

The substantial projected decrease [$O(10\%)$] in wave energy incident on the windward rim of Kwajalein Atoll is consistent across models. This decrease may affect cross-shore sediment transport and sediment budgets maintaining the atoll’s reef islands.

2.4.2 Wave model evaluation

After tuning our model with 2 days of hourly wave data, the best fit friction coefficient values we identified were in accordance with c_f values from other published investigations of fringing reefs. Our model reproduced wave heights and water levels over the fringing reef most accurately with $c_f = 0.15$ on the fore reef and $c_f = 0.015$ on the reef flat and beach. For a similar reef cross-section at Roi-Namur, 82 km northwest of our site, Quataert et al. (2015) optimized XBeach model performance with $c_f = 0.1$ on the fore reef and $c_f = 0.01$ on the reef flat. The reef flats fronting Kwajalein Island and Roi-Namur are very shallow at low tide, which prevents coral growth and reduces surface roughness on the reef flat relative to the fore-reef, requiring separate friction coefficient values for those two regions of the reef (Quataert et al., 2015). Kwajalein’s fore reef has greater coral cover, and thus roughness, than Roi-Namur, which we assume is consistent with the larger friction coefficient in our optimized model. Our reef flat c_f value is less than half that of van Dongeren et al. (2013) for a 2D hydrostatic model of Ningaloo Reef ($c_f = 0.04$), likely due to the relative smoothness of the Kwajalein Island reef flat.

Wave refraction and blocking in the vicinity of the offshore AWAC possibly resulted in discrepancies between modeled and measured wave heights across the fringing reef. Modeled wave heights on the fore-reef consistently matched wave heights

at the offshore AWAC used to force the model (Fig. 2-4a), but underestimated wave heights measured by the fore-reef pressure sensor. The offshore AWAC is located on a cross-shore transect roughly 45° clockwise from the transect of wave pressure sensors on the reef flat (Fig. 2-1c). θ_{mean} at the offshore AWAC was 111° during the first 72 h of model verification and then shifted northerly to 94° for the next two weeks. As incident θ_{mean} shifted, refraction along the curving coast between the two sensors would have decreased wave heights at the AWAC relative to the fore reef sensor. Additionally, the eastern sensor transect may have recorded oblique northerly waves that were subsequently blocked by the reef before they could reach the AWAC. These discrepancies in forcing contributed to our model underestimating wave heights and water level across the fringing reef.

Comparison of model results to the mid-reef flat sensor reveals considerable bias in wave heights, but better agreement in water surface elevation (Figs. 2-4 & 2-5). We speculate that the deeper water between the inner-reef flat and mid-reef flat sensors depressed frictional dissipation of wave energy relative to the model bathymetry, resulting in higher measured wave heights but lower measured water level.

Despite some discrepancies resulting from sub-optimal model bathymetry and forcing, our model accurately simulates the important wave transformation processes occurring across the fringing reef. Incident high-frequency waves break on the shallow fore reef and reef crest, resulting in much smaller, tidally-modulated wave heights on the reef flat relative to the fore reef (Fig. 2-4). $H_{s,hf}$ decreases across the reef flat, where friction contributes to wave dissipation. Break-point forcing on the fore reef (Symonds et al., 1982; Becker et al., 2016) generates low-frequency waves on the outer reef flat that are larger than those over the fore reef (Fig. 2-5). For higher water levels and also for larger fore-reef waves, $H_{s,lf}$ increases across the reef flat (Gawehn et al., 2016; Cheriton et al., 2016). Our model tends to underestimate reef flat water levels at low tide or during large incident wave heights (Fig. 2-5), a discrepancy that is consistent with previous studies applying 1D XBeach non-hydrostatic models to both idealized reefs in laboratory flumes (Lashley et al., 2018) and the fringing reef at Roi-Namur (Quataert et al., 2015). Lashley et al. (2018) demonstrated that this

Figure 2-4: Measured and modeled high frequency significant wave heights ($H_{s,hf}$) on the (a) fore reef, (b) outer reef flat, (c) mid reef flat, and (d) inner reef flat. Sensor elevations are displayed in parentheses. $H_{s,hf}$ recorded at the offshore AWAC is plotted in (a). The objective functions, RMSE, r^2 , and mean bias (model minus observations), report the correspondence between measured and modeled data, excluding the optimization period.

Figure 2-5: Measured and modeled (a-d) low frequency significant wave height ($H_{s,lf}$) and (e-h) hourly mean water surface elevation. The objective functions, RMSE, r^2 , and mean bias (model minus observations), report the correspondence between measured and modeled data, excluding the optimization period.

water level bias likely occurs because the model does not account for wave roller contributions to kinetic energy and that the bias is common among models making shallow water or Boussinesq assumptions. Our verification exercise suggests that our model is suitable to examine changes in wave-driven hydrodynamics with varying wave climate and sea level.

2.4.3 Forecasts of wave propagation and shear stress on the fringing reef

After tuning friction coefficients in our XBeach model, we forced it with each of six scenarios representing Historical, Wave Climate Change, and four SLR conditions. Under Historical and Wave Climate Change conditions, most high-frequency wave energy dissipates onshore across the reef through breaking and frictional dissipation (Fig. 2-6a), while $H_{s,lf}$ increases onshore due to breakpoint forcing and the formation of standing waves, progressive-growing waves, and resonant amplification on the reef flat (Fig. 2-6b) (Gawehn et al., 2016). Mean water velocity is offshore over the fringing reef, with the strongest flows under breaking waves at the reef crest (Fig. 2-6c). We infer that this mean offshore flow is balanced by onshore-directed wave-induced mass flux as evidenced by the strong negative correlation between mean velocity and mean incident wave dissipation across the reef profile (Pearson’s $r = -0.92$, $p < 0.001$). Mean shear stress on the reef flat is near zero or slightly negative at modern sea level (Fig. 2-6d), which could suggest the shoreline is near steady-state, with no net erosion/accretion, as occurs in numerical modeling of motu formation on reef platforms (Ortiz and Ashton, 2019).

For the Wave Climate Change scenario, the only future changes considered are the projected 12% decrease in incident wave energy and changes to wave period and direction. The change in incident wave energy produces a corresponding decrease in fore-reef $H_{s,lf}$ (-12%), mean water velocity (-11%), and mean shear stress (-18%). However, there was a relatively smaller effect on reef flat $H_{s,lf}$ (-9%), flow velocities (-2%), and shear stress (-6%) (Fig. 2-6), likely because waves on the reef flat are so

Figure 2-6: Changes in mean wave and flow characteristics for the six modeling scenarios: left columns show Historical and Wave Climate Change scenarios and right columns show the 4 SLR scenarios with the Historical scenario included for reference. Mean (a),(e) $H_{s,hf}$ (0.3-0.04 Hz), (b),(f) $H_{s,lf}$ (0.04-0.001 Hz), (c),(g) mean cross-shore water velocity, and (d),(h) bottom shear stress under.

strongly controlled by water level.

In contrast, the 4 SLR scenarios result in large changes to hydrodynamics across the reef. With rising sea levels, less incident wave energy is dissipated by breaking on the fore reef, and the reef flat $H_{s,hf}$ increases (Fig. 2-6e). $H_{s,lf}$ on the beach and reef flat also increase with SLR, though the effect is greatest for the first 1.5 m of SLR, with little increase in $H_{s,lf}$ with 2.0 m SLR (Fig. 2-6f). With 2.0 m of SLR, the beach ridge (maximum island elevation) was flooded for 24% of the model runtime across all wave classes, but because the model underestimates mean water level, this is likely an underestimate of the amount of flooding that would occur with this level of SLR. Thus, substantial flooding of the island under ambient wave conditions is likely to initiate between 1.5 and 2.0 m of SLR, assuming constant island morphology. However, physical modeling of island response to SLR suggests island oceanfront ridges aggrade with SLR, potentially mitigating flooding hazard over time (Tuck et al., 2019a). At the reef crest, reduced wave breaking leads to reduced mean offshore flow, while on the reef flat the increased wave heights increase the offshore flow and shear stress; with $SLR \geq 0.5$ m, the reef flat shear stress becomes onshore-directed (Fig. 2-6). In total, these model results suggest that over the next century SLR will have a much larger impact on the hydrodynamics of Kwajalein Island's fringing reef than the projected reduction in wave heights.

If sediment mobilization and bedload transport are primarily functions of shear stress, the sign change in mean shear stress across the reef flat with SLR above 0.5 m could have major implications for sediment transport rates across the fringing reef. To better understand how the modeled changes in wave characteristics affect mean shear stress, we examined wave skewness for the full spectrum of wave frequencies, high-frequency waves, and low-frequency waves (Fig. 2-7). Under Historical conditions, the high-frequency wave skewness ($S_{u,hf}$) and low-frequency wave skewness ($S_{u,lf}$) are large and positive (peaked waveforms) on the fore reef, contributing to positive (onshore) mean shear stress on the shallow fore reef (Fig. 2-7b,c). Additionally, $S_{u,hf}$ increases onshore across the fore reef as waves shoal, such that mean shear stress increases in magnitude in the onshore direction until the waves begin breaking (Figs.

Figure 2-7: Mean wave skewness for (a),(d) full spectrum, (b),(e) high-frequency, and (c),(f) low-frequency modeled wave motions for the six model scenarios. Left hand column shows Historical and Wave Climate Change scenarios, right hand column shown the 4 SLR scenarios plus Historical for reference.

2-6d & 2-6b). On the reef flat, where high-frequency wave motions are generated primarily by broken wave bores, both high- and low-frequency wave motions are less positively skewed (Fig. 2-7b,c) and $H_{s,hf}$ is much smaller, resulting in near-zero mean shear stress there (Fig. 2-6d).

For the Wave Climate Change scenario, the impact on wave skewness by the decreased incident wave energy is minimal, with slightly reduced $S_{u,hf}$ and slightly increased $S_{u,lf}$ across the fringing reef transect (Fig. 2-7a-c). However, increasing SLR causes $S_{u,hf}$ to decrease on the fore reef and increase on the reef flat (Fig. 2-7e), while $S_{u,lf}$ decreases substantially on the fore reef and slightly on the inner reef flat, but increases slightly on the outer reef flat (Fig. 2-7f). On the reef flat, full spectrum wave skewness increases 100% for 1.0 m of SLR and then increases less

sharply, by 10% for an additional 1 m of SLR. The increase in reef flat full-spectrum wave skewness with SLR is driven primarily by increases in $S_{u,hf}$ (Fig. 2-7e,f).

In all scenarios, $S_{u,hf}$ increases sharply on the beach and $S_{u,lf}$ sharply decreases slightly further onshore (Fig. 2-7). XBeach does not explicitly simulate all swash zone processes important for sediment transport, such as turbulent swash-swash interactions or complex vertical flow structure at the beach toe. Additionally, it is unlikely that our model is adequately resolving the entire swash zone and the shallow water depths characteristic of swash. However, the fact that full-spectrum wave skewness becomes negative and appears to agree more with $S_{u,lf}$ than $S_{u,hf}$ on the beach is characteristic of swash zones with high- and low-frequency wave energy (Elfrink and Baldock, 2002).

Across the fringing reef, wave skewness contributes to skewed probability distributions of shear stress magnitudes. Offshore-directed mean flow results in shear stress that is more frequently negative than positive. However, under positively skewed waves, onshore velocity under the shorter-duration crest is greater in magnitude than offshore velocity under the longer-duration trough. Thus, under skewed waves in the Historical scenario, shear stress on the fore reef and reef flat has greater magnitudes in the onshore direction than offshore (Fig. 2-8).

SLR alters shear stress distributions non-uniformly across the fringing reef. SLR inundates the beach to a higher elevation and increases absolute shear stress on the lower portion of the beach that was historically only inundated at high tide. On the reef flat, velocities become more extreme in both onshore and offshore directions with SLR, but with increasing wave skewness, the resulting shear stress extremes are ultimately greater in the onshore direction. On the fore reef, shear stress magnitudes decrease, resulting in a narrower distribution (Fig. 2-8j). Thus, our model results suggest that with 1 m of SLR, mean absolute shear stress increases on the beach and across the reef flat by over 100% but decreases on the fore reef by 13%.

Figure 2-8: Modeled shear stress distributions and their projected change with 1 m SLR. (a)-(e) full spectrum bottom shear stress for Historical and 1 m SLR scenarios, and (f)-(j) the change in distribution between the two scenarios. The x-axes of the graphs are flipped to indicate that positive shear stresses are directed onshore.

2.4.4 Potential bedload sediment transport

To quantify the net effect wave climate change and SLR will have on net bedload transport rates by the end of the century, we calculated potential bedload flux from the shear stress distributions for the Historical, Wave Climate Change, and all SLR scenarios. The process of estimating bedload transport rates from bottom shear stress is illustrated in Fig. 2-9. Sediment transport only occurs when bottom shear stress magnitudes exceed the critical shear stress (Fig. 2-9c), which results in a contraction of the distribution of τ_e (Fig. 2-9b). Because bedload transport rate varies nonlinearly with shear stress, even if mean τ_b and τ_e are negative, net bedload transport can be positive (Fig. 2-9a, 2-10). Thus, the extremes in a shear stress distribution influence bedload transport disproportionately more than moderate values. Because wave skewness tends to amplify the positive extremes of the shear stress distribution relative to the negative, wave skewness generally results in net onshore bedload flux.

In the Historical scenario, the average shear stress mobilizing silt-to-medium sand-sized sediment is offshore-directed on the beach and outer-to-mid reef flat, whereas coarser sediment, with larger sediment mobilization thresholds, are mobilized by onshore-directed mean shear stress (Fig. 2-11a-e). On the inner reef flat and fore reef, the shear stress probability distributions are skewed such that τ_e is onshore-directed for all sediment size classes that are mobilized. However, under the Historical scenario, net bedload flux is directed onshore for all sediment size classes across the fringing reef (Fig. 2-11f-j). Although our calculations suggest net onshore bedload transport, the magnitude of this transport is small, particularly in comparison to calculated potential fluxes for higher sea levels (Fig. 2-11). For coarse-grained sediment (> 16 mm), computed fluxes are low because τ_c is rarely exceeded; however, for finer grained sediment the reef flat remains active, and both onshore- and offshore-directed sediment transport components appear to be balanced (Fig. 2-10). This suggests that the current reef flat configuration, in combination with modern sea level and wave climate, is in a near steady state configuration (Ortiz and Ashton, 2019).

With increasing sea level, the magnitude of bedload transport increases on the reef

Figure 2-9: Schematic illustration of the calculation of (a) potential bedload transport for pebble-sized material ($\tau_c \geq 1 \text{ N m}^{-2}$), given (b) excess shear stress, τ_e , calculated from (c) a hypothetical distribution of near-bottom shear stress, τ_b . Zero values are excluded from the probability distribution functions (PDF) in (a) and (b) for ease of display, but they are included in the calculation of the mean. Mean values and direction are displayed for qualitative illustration.

Figure 2-10: Modeled probability distributions of potential bedload transport for Historical and +1 m SLR scenarios. Each row of histograms represents bedload transport for a different sediment grain size. For sediment with $D_{50} > 16$ mm, mobilization only occurs with outlier shear stress values. These histograms indicate that near-zero Historical net bedload transport for most grain sizes is due to a balance between offshore and onshore transport, not failure to mobilize sediment.

Figure 2-11: Excess shear stress and resulting net bedload flux for Historical, Wave Climate Change, and 4 SLR scenarios. (a)-(e) Excess shear stress and (f)-(j) potential bedload sediment transport rate by grain size and SLR scenario. Negative values indicate net offshore-directed shear stress or potential transport.

flat and beach due to larger $H_{s,hf}$ and $S_{u,hf}$ (Figs. 2-7, 2-8, 2-11). On the beach, SLR of up to 0.5 m produces a small net offshore transport for fine sediment but net onshore transport for coarser material. Further SLR results in net onshore transport for all sediment size classes, at a higher magnitude than the Historical scenario. Onshore-directed bedload transport across the reef flat is forecasted to increase 50-100% with just 0.5 m of SLR and 200-900% with 1.0 m of SLR. On the fore reef, bedload flux remains directed onshore with up to 2 m of SLR, but decreases in magnitude by over 20% for all sediment size classes with 50 cm of SLR and by over 50% with 2 m of SLR (Fig. 2-11).

2.4.5 Suspended sediment transport trends

Although most large waves break on the reef crest, and only highly-diminished waves and wave bores propagate across the reef flat, coarse sediment is still be mobilized and transported in suspended load (Pomeroy et al., 2015). Once sediment is suspended, flow acceleration contributes to sediment momentum, and wave skewness and asymmetry contribute to net directional transport, which we quantify with u_{rep} . Under historical conditions, modeled u_{rep} is directed onshore across the reef flat and the shallow fore reef and is directed offshore over the reef crest, where return flow is strongest, and the outer reef flat (Fig. 2-12c). The modeled changes in u_{rep} with wave climate change and SLR are similar to the changes in bedload transport. On the outer reef flat u_{rep} increases in magnitude and is directed more onshore with increasing SLR (Fig. 2-12d). Across most of the fore reef, representative transport becomes less onshore-directed with increasing SLR, even changing sign to be mean offshore-directed over a portion of the shallow fore reef for > 1 m SLR. On the inner reef flat, however, predicted u_{rep} decreases with increasing SLR, and becomes offshore-directed for > 1 m SLR. Although $S_{u,hf}$ is forecasted to increase over the reef flat, high-frequency wave asymmetry decreases in magnitude substantially (by 72% for 1 m SLR, 93% for 2 m SLR) over the reef flat (Fig. 2-12b), and mean return flow increases in magnitude, resulting in decreased onshore forcing of suspended sediment transport.

2.5 Discussion

First, the wave climate changes forecasted for Kwajalein Atoll were some of the largest of all locations examined in the Pacific (Shope et al., 2016). Assuming that SLR exceeds $+0.5$ m across the tropical Pacific Ocean basin by 2100, our modeling suggests that SLR by 2100 will have a larger influence on cross-shore transport than projected wave climate change on fringing reefs across the Pacific. Second, SLR will cause prolonged inundation of the lower beach, enhancing shear stress magnitudes across the subaqueous portion of the beach. From this we infer an increase in the volume of

Figure 2-12: Changes in wave asymmetry (a-b) and representative transport velocity, u_{rep} , (c-d) under 6 modeling scenarios: Historical and Wave Climate Change scenarios (left panels) and 4 SLR scenarios with Historical included for reference (right panels). Plots of full spectrum wave asymmetry and low frequency wave asymmetry can be found in Fig. 2-13.

Figure 2-13: Mean wave asymmetry for (a),(d) full spectrum, (b),(e) high-frequency, and (c),(f) low-frequency modeled wave motions for the six model scenarios. Left hand column shows Historical and Wave Climate Change scenarios, right hand column shown the 4 SLR scenarios plus Historical for reference.

beach sediment that can be mobilized and greater geomorphological activity of island margins, similar to that projected by Shope and Storlazzi (2019). Third, sediment on the fore reef will be less likely to be transported onto the reef flat than historically, but sediment on the reef flat will be much more likely to be transported onshore to the beach. The net direction and magnitude of morphological changes induced by these alterations to shear stress and potential sediment transport depend on the amount (and thus production that is related to coral reef health) and distribution of sediment on the fringing reef available for transport. Where sediment is produced may also depend on SLR, as SLR moves the point at which waves break shoreward, potentially influencing the spatial distribution of coral growth and erosion. Fourth, frequent flooding of the island by ambient (non-storm) waves will occur between +1.5 and +2.0 m of SLR. This flooding will likely move sediment from the beach to the island surface, potentially increasing the elevation of the beach ridge with sea level, but also possibly resulting in lagoon-ward migration of the island (Tuck et al., 2019a).

Moving onshore across the reef flat, the potential bedload transport decreases substantially from the outer to mid-reef flat and then is constant or decreases slightly from the mid- to inner reef flat (Fig. 2-14). This gradient is relatively constant with SLR. If the reef flat consisted entirely of mobile sediment transport or if sufficient mobile sediment were present, these flux gradients would suggest a mechanism for the reef flat to accrete vertically. However, most reef flats (including Kwajalein Atoll), are sparsely covered with sediment, confounding predictions of future morphological change. At the beach, for +0.5 m of SLR, our model forecasted an offshore shift in both potential bedload and suspended sediment transport rates for all sand-sized sediment. Thus we project that with SLR, sand will initially erode from the beach and either accumulate on the inner reef flat or be removed by alongshore transport. This erosion could move the beach toe in the offshore direction. However, with additional SLR, the bedload transport on the beach reverses to onshore, while u_{rep} becomes more offshore directed, potentially continuing beach erosion, but only of finer grained sediment. Therefore, SLR-driven morphological change on atoll reef islands should be assumed to be neither linear nor monotonic with water level, even for lower-end SLR

Figure 2-14: Gradients in potential bedload transport across the reef flat and beach for sediment with $D_{50} =$ (a) 0.5 mm, (b) 2.0 mm, (c) 16.0 mm, and (d) 64.0 mm.

projections for the next century. Additionally, morphological changes would have feedbacks on wave propagation and hydrodynamics not captured by our model, and thus the evolution of the beach and resultant morphological changes are not clear.

The windward fringing reef on Kwajalein Island has a relatively steep fore reef and narrow and smooth reef flat. However, we expect the effects of SLR on Kwajalein to be similar to other fringing reefs, as long as they are at similar elevations relative to mean sea level. The main trends in potential sediment transport with SLR are driven by increasing $H_{s,hf}$ and $S_{u,hf}$ on the reef flat, and decreased bottom shear stress on the fore reef. Increasing sea level should cause similar changes on fringing reefs irrespective of reef flat width, roughness, and fore reef slope. However, reef flat width, roughness, and fore reef slope do influence the dynamics of low-frequency waves on the reef flat. Very wide or rough reef flats can cause smaller $H_{s,lf}$ (e.g. Pomeroy et al., 2018; van Dongeren et al., 2013, possibly because the low-frequency waves are dissipated through friction before interacting with the shoreline to generate

standing or resonant waves. Steeper fore reefs tend to generate larger breakpoint-driven low-frequency waves (Quataert et al., 2015). Although the direct contributions of low-frequency wave motions to shear stress and sediment transport are likely small, they modulate $H_{s,hf}$ on the reef flat and beach and flooding of the shoreline. SLR may have a larger effect on low-frequency wave characteristics on fringing reefs with different morphology than on Kwajalein Atoll. For example, on rough reef flats where low-frequency waves tend to be frictionally dissipated before reaching the shoreline under Historical conditions, SLR may allow low-frequency waves to forming standing or resonant waves on the reef flat, causing a large increase in $H_{s,lf}$ there. We speculate that the rapid increase of low-frequency wave action on the reef flat would enhance the trends our modeling predicts for Kwajalein Atoll, as resonant or standing low-frequency waves on the reef flat would increase time-mean water surface elevation relative to dampened low-frequency waves.

We emphasize that we use the two measures of sediment transport, q_b and u_{rep} , only as proxies for transport rates. In particular, q_b scales with shear stress in excess of a mobilization threshold. However, Kench and McLean (1996) demonstrated that, for non-spherical bioclastic carbonate sediment, the mobilization threshold can be lower than that predicted by the critical Shield’s parameter. Additionally, in an environment where sediment is sparse, i.e. the Kwajalein Island reef flat, isolated sediment grains or small collections of grains not interlocking with a bed of sediment would experience lower stabilizing frictional force, resulting in a lower mobilization threshold (Aubert et al., 2016). Finally, above some shear stress threshold, all sediment size classes are mobilized and a more sophisticated transport parameterization would be required to account for inter-particle interactions and suspended sediment-flow feedbacks (Roelvink et al., 2009). Additionally, u_{rep} was tuned for low-gradient, sandy beaches, and it assumes a particular balance between wave asymmetry and skewness that could be weighted differently for different topography or sediment characteristics, though we assume that a proper calibration of u_{rep} for our environment will not change with SLR.

The opposing trends in potential sediment transport between the fore reef and

reef flat might explain some of the variability in atoll reef island shoreline response to modern SLR. For example, Webb and Kench (2010) found that the majority of ocean-facing shorelines eroded over the past 40-80 years on Funafuti Atoll, Tuvalu, and Pingelap and Mokil Atolls, Federated States of Micronesia, potentially due in part to SLR. However, on Tarawa Atoll, Kiribati, ocean-facing shorelines were observed to have universally accreted (Webb and Kench, 2010). A more recent review found substantial variability in rates of atoll reef island area change across the Western Pacific, though they didn't separate ocean-facing and lagoonal shoreline change (McLean and Kench, 2015). Assuming our modeled changes in potential sediment transport are applicable to other Pacific atolls, variability in shoreline response to SLR could also be caused by variability in the relative importance of the fore reef and reef flat for sediment production. For example, if a beach is supplied solely by sediment produced on the reef flat (i.e. foraminifera tests and crustose coralline algae fragments), our modeling results would suggest that, all else held equal, the shoreline could prograde with SLR due to enhanced shoreward potential sediment transport on the reef flat (Fig. 2-11g-i). However, an island shoreline supplied solely with sediment generated on the fore reef (e.g. coral cobbles and *Halimeda* plates) would likely erode due to decreased onshore potential sediment transport from the fore reef. Most islands are supplied with sediment from both regions of the reef, but their relative dependence on one source region over another could affect the shoreline response to SLR.

2.6 Conclusions

In this study, we used forecasts of wave climate change from a global climate model ensemble and numerical modeling of hydrodynamics and wave propagation over a fringing reef to predict changes in potential cross-shore transport likely to occur due to SLR and wave climate change. The climate model ensemble forecasted a 12% reduction in wave energy incident on the fringing reef at Kwajalein Island by the end of the century, equivalent to a reduction of one standard deviation of annual mean wave energy in the instrumental record. However, this marked change in wave energy

had negligible impact on the hydrodynamics and potential sediment transport across the fringing reef relative to SLR.

Our numerical model analysis forecasted that for SLR of +0.5 to +2.0 m, wave height and skewness will increase across the reef flat, resulting in more onshore-directed shear stress and potential bedload transport. On the fore reef, waves will be less positively skewed, such that the magnitude of net onshore potential bedload transport decreases substantially. On the beach, our model projected more negatively skewed wave motions for +0.5 m of SLR, resulting in offshore directed potential bedload transport. From these trends, we infer that SLR should initially cause the beach toe to shift offshore and that with greater SLR net erosion (progradation) of the beach may occur if the beach's sediment budget is dominated by sediment generated on the fore reef (reef flat). Predicting the morphological evolution of a particular island due to SLR and wave climate change requires incorporation of local sediment production and coupling between sediment transport processes and morphological evolution.

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Chapter 3

Foraminifera taphonomy as proxy for large wave events: A case study in the Marshall Islands

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Abstract

Observational records of tropical cyclones can be extended over the past few thousand years using sediment proxies of transport and overwash by large waves. However, in high-energy carbonate environments, conventional methods of identifying large wave events in sediment proxy records can fail due to homogeneous, coarse grain size and homogeneous chemical makeup. We investigated the relative weathering of foraminifera tests as an alternative proxy for large wave events in a blue hole sediment core retrieved from Jaluit Atoll, Republic of the Marshall Islands. We found clear evidence of gradients in weathering under ambient hydrodynamic conditions across the fringing and back reef flats bracketing the blue hole. However, despite the existence of these gradients, event layers in the sediment core were not clearly distinguishable from non-event layers using weathering as a proxy.

3.1 Introduction

Observational records of tropical cyclones are temporally short and geographically sparse before the satellite era, stymying efforts to quantify long-term variability in tropical cyclone characteristics required for estimating current and future coastal hazard risk (Hong et al., 2017). Sediment proxies of transport and overwash associated with large wave events allow us to extend the historical record across millennia. However, in high-energy carbonate environments, it can be difficult to identify large wave event layers in sediment records using conventional methods such as grain size, bulk density, and elemental analysis. In these environments, grains are often coarse (median grain diameter, D_{50} , exceeds $125\ \mu\text{m}$), resulting in uniform bulk density down core and event beds whose grain size distribution is difficult to distinguish from the already coarse background. Additionally, in carbonate environments where there is little contrast between potential sediment sources (e.g. atolls or uplifted reef complexes), it is difficult to distinguish the chemical composition of event deposition from ambient deposition.

Foraminifera taphonomy is one alternative to more conventional methods to identify sediment provenance and transport history. Large benthic foraminifera are calcifying organisms often found near the reef crest of tropical reef environments. Large benthic foraminifera, so-named for their relatively complex internal structure, can contribute significantly to the total volume of unconsolidated sediment islands on reef platforms and coral atolls (Dawson et al., 2014). Degradation of foraminifera tests typically corresponds to their transport history. Tests that have undergone transport over longer distances or time periods experience greater weathering, especially in shallow reef environments where tests can be mechanically abraded or experience chemical weathering when exposed to air and rainwater (Kosciuch et al., 2017; Fellowes et al., 2017; Dawson et al., 2014). Large wave events induce stronger than normal currents across the reef, resulting in greater mobilization of benthic foraminifera and more rapid transport before deposition in deep water. Ostensibly, more rapid transport and deposition can result in lower overall weathering of foraminifera tests and less

rounding of test fractures when abrasion does occur (Kosciuch et al., 2017).

Here we test the ability of taphonomic analysis of foraminifera tests to identify large wave events in a blue hole sediment record from Jaluit Atoll, Republic of the Marshall Islands.

3.2 Methods

Jaluit Atoll consists of a 250-1000 m wide, intertidal carbonate platform encircling a deeper lagoon, with vegetated islands interspersed on top of the platform. In November, 2015, we extracted a sediment core from a 30-m deep, mildly stratified blue hole located at the northwestern tip of the atoll, where the carbonate platform is particularly wide (>1 km, Fig. 3-1). The sediment core, LTD2, was extracted with a Rossfelder P-3 vibracoring system as a 1-m surface drive that preserved the sediment-water interface and a 3.25-m drive that over-penetrated the sediment surface (Fig. 3-3). Around the blue hole we manually collected surface samples along three transects extending from the blue hole to the atoll lagoon, the ocean-facing reef crest, and along a sand spit to a nearby island (Fig. 3-1b).

To identify large wave events, we performed grain size analysis on LTD2. In the lab, the core was sampled every centimeter, and samples were dried and sieved at 1 mm grain size to produce a dry-weight-normalized coarse fraction. Anomalously high peaks in coarse fraction were interpreted as large wave events. Four large wave event peaks and four nearby non-peak layers were sampled from the core for taphonomic analysis of foraminifera tests. Organic material and solitary mollusk shells were sampled, where available, for radiocarbon dating.

For taphonomic analysis, the selected core and surface sediment samples were dried, sieved, and split to isolate the $>500\text{ }\mu\text{m}$ size fraction. Then, under a binocular dissection microscope, 300 foraminifera per sample were identified by genus and rated by degree of abrasion and corrosion as described in Fellowes et al. (2017). Under this rubric, individual tests are categorized as Pristine, Minor alteration, Major alteration, and Severe alteration depending on the degree of surface alteration (through

Figure 3-1: Study area with (a) true color Landsat 8 image of Jaluit Atoll and (b) the location of the surface samples, blue hole, and sediment core LTD2.

fracturing and pitting), and loss of spines where applicable (Fig. 3-2). Additionally, for tests with minor to major alteration, fractures were categorized as rounded or angular following Kosciuch et al. (2017).

To evaluate the ability of the taphonomic classification to distinguish large wave event deposits from background sedimentation, we performed two simple tests. First, we determined if any of the taphonomic classifiers were consistently more or less common in the event layers relative to the background sediment (e.g. whether every event layer had a smaller proportion of foraminifera classified as Severe alteration than every background interval). Second, in case there were relevant covariance between classifiers that could be missed with that simple analysis, we performed a principle component analysis of all taphonomic classifiers. We then examined the ability of the first two principal components to distinguish events and non-events.

Figure 3-2: Scale used to evaluate taphonomy of *Calcarina* and *Amphistegina* tests. Classification scheme adapted from Fellowes et al. (2017).

3.3 Results and Discussion

The sediment core recorded two historical tropical cyclones and multiple other layers we interpret as pre-historical large wave events. A post-bomb date from the first, particularly prominent coarse event bed in the surface drive dates to -7.8 cal. yr BP (Fig. 3-3), which is synchronous with the 1958 Typhoon Ophelia that made landfall on Jaluit Atoll, causing significant change and damage to many of its islands (Blumenstock, 1958; Blumenstock et al., 1961b). Another prominent peak in both drives lies between the 1958 peak and a carbonate radiocarbon date with a median calibrated age of 84 cal yr BP. This peak likely corresponds to an unnamed 1905 typhoon that also severely affected islands on Jaluit Atoll (Spennemann and Marschner, 1994). In addition to these identifiable historical storm events are several more prominent peaks that we infer also correspond to large wave events. The sedimentation rate in the core was 7.5 mm/yr, meaning the coarse fraction proxy has nearly annual resolution and the event layers are easily distinguished from background sedimentation.

The surface sample transects displayed clear taphonomic gradients from both the reef crest and back reef to the blue hole, demonstrated by the decreasing proportion of severely altered foraminifera with distance from the blue hole (Fig. 3-4). This gradient is consistent across the two primary genera, *Calcarina* and *Amphistegina*, which together account for 84% of all foraminifera counted in the surface samples. The gradient is less steep in slope for *Amphistegina spp.* (Fig 3-4b,c). Interestingly, the gradients for Transect 1 (from offshore towards the blue hole) and Transect 4 (from the lagoon towards the blue hole) have very similar magnitudes despite their different environments.

From these results we infer that the foraminifera tests are produced on the fringing reef flat and in the back-reef lagoon before being transported to the blue hole. Additionally, *Amphistegina spp.* are altered more slowly than *Calcarina spp.*, indicating greater resistance to alteration, which is unsurprising given their morphological differences.

Despite the clear taphonomic gradients along the surface transects, in the sedi-

Figure 3-3: Depth-grain size plots of both drives of LTD2, with calibrated radiocarbon dates and the location of foraminifera samples indicated.

Figure 3-4: Taphonomic gradients of surface samples along the three surface transects in Fig. 3-1 for (a) all foraminifera genera, (b) *Calcarina* spp., and (c) *Amphistegina* spp.

ment core, large wave events could not be distinguished from non-events using any taphonomic index, for all genera as well as for *Calcarina spp.* or *Amphistegina spp.* considered separately (Fig. 3-5). Additionally, principle component analysis did not reveal any better classifiers of large wave events in the sediment core record.

To explain the failure of our taphonomic analysis to distinguish event beds despite the surface taphonomic gradients, we hypothesize two explanations. First, rapid transport during storms results in alteration similar to that of slow transport under ambient conditions. The invariance of test degradation relative to deposition rate in Fellowes et al. (2017) suggests this is a possibility, albeit over centennial timescales. Second and more likely, the sediment deposited in the middle of the blue hole has the same source whether deposited under storm or ambient conditions. For example, the majority of all core sediment could have come from deposits along the shallower rim of the blue hole or uniformly degraded island sediments (e.g. Transect 3), or could represent a spatial average over the entire reef flat. The lack of distinction between storm and background deposition in the sediment core demonstrates at the very least that storms do not introduce appreciable quantities of sediment from otherwise inaccessible deposits (e.g. from the fore reef slope). More generally, foraminifera weathering may be a poor proxy for large wave events, especially where there are persistent reservoirs of altered foraminifera tests nearby.

3.4 Conclusions

Foraminifera weathering provides a viable proxy for transport distance on atoll reef flats, as we have demonstrated on Jaluit Atoll's leeward rim, which has strong linear gradients in foraminifera degradation from source to sink, both from the ocean and lagoon shore. However, we found that foraminifera weathering was a poor proxy for hydrodynamic conditions during deposition, as foraminifera deposited during large wave events were not clearly distinguishable from those deposited during ambient conditions. Either taphonomic indexes may be similarly influenced by rapid, energetic transport and slower transport, or sediment from persistent deposits on the reef flat

Figure 3-5: Taphonomic indexes for (a) all genera, (b) *Calcarina* spp., and (c) *Amphistegina* spp. Note that “Angular fractures” is measured separately from alteration.

at our study site may contaminate what would otherwise be a strong signal in event layers.

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Chapter 4

Experimental quantification of bedrock abrasion under oscillatory flow

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Abstract

Wave-driven abrasion of submarine bedrock is hypothesized to drive the geomorphic evolution of reefs and rocky coasts globally, but the dependence of abrasion rate on wave forcing and sediment availability remains poorly quantified. We performed a series of laboratory experiments in which an artificial substrate was abraded by coarse-grained sediment subjected to oscillatory flows with orbital velocities of 0.4-1 m s⁻¹ and varying sediment loads. From these experiments, the bedrock incision rate scaled by bedrock tensile strength (I , m yr⁻¹ MPa²) varies with mean root-mean-square velocity ($\langle u_{rms} \rangle$, m s⁻¹) according to a power law, $I = 1.0 \langle u_{rms} \rangle^{4.2}$. Additionally, the relationship between sediment load and bedrock incision rate demonstrates a measurable tools-and-cover effect similar to abrasion in fluvial environments, such that incision is fastest at intermediate sediment loads. However, because oscillatory flows accumulate sediment into bedforms, bedrock remains exposed in the bedform troughs, reducing the efficiency of the cover effect for high sediment loads relative to unidirectional flow. Our results quantify the relationship between submarine bedrock abrasion and wave forcing, providing an empirical model to predict bedrock incision rates in nearshore environments.

4.1 Introduction

Wave-driven abrasion of exposed bedrock drives coastal geomorphological change across many environments globally. On rocky shores, wave-driven abrasion contributes to the downwearing of shore platforms (Blanco-Chao et al., 2007; Trenhaile, 2000) and cliff retreat (Robinson, 1977; Sunamura, 1976). On coral reefs, wave-driven abrasion of the reef surface by loose sand and coral cobbles can reduce net reef accretion rates and produce spatially variable accretion (e.g. Engels et al., 2004; Grossman and Fletcher, 2004). Additionally, field observations (e.g. Blanchon and Jones, 1995; Cloud Jr., 1954) and geostatistical analysis (Duce et al., 2016) suggest wave-driven abrasion is the primary driver of the formation of fore reef grooves, which are longshore-periodic channel-like structures that are potentially important for reef ecosystem health (Storlazzi et al., 2011). Explaining the formation and accurately predicting the evolution of hardbottom coastal environments requires knowledge of the relationships between abrasion rate and wave and sediment characteristics.

Despite the prevalence of wave-driven abrasion along coasts worldwide, few studies have measured the dependence of the erosional process on wave characteristics and sediment availability. Wave tank experiments have shown that abrasion of coastal cliff faces, which can produce cliff retreat rates of up to 1 m yr^{-1} (Sunamura, 2018), is modulated by the presence and width of a beach at the base of the cliff (Sunamura, 1976). Numerical modeling suggests that beach sediment can also induce morphological feedbacks responsible for the evolution of headlands and embayments on rocky coastlines (Limber and Murray, 2014). On rocky shore platforms at the base of cliffs, vertical erosion by abrasion of up to 0.02 m yr^{-1} has been measured (Robinson, 1977; Blanco-Chao et al., 2007). Robinson (1977) observed that abrasion of the bedrock platform under beaches decreases with thickness of the overlying sediment, but had insufficient data to quantify the effect.

Abrasion by unidirectional flow in terrestrial bedrock channels is analogously important to the evolution of rivers and their surrounding valleys and hillslopes. Laboratory experiments have constrained rate laws governing river bed abrasion as a func-

tion of sediment flux and bedrock tensile strength. Experiments using rotary abrasion mills (Sklar and Dietrich, 2001; Scheingross et al., 2014) demonstrated the existence of the tools-and-cover effect hypothesized by Gilbert (1877): sediment impacts on the bed enhance abrasion, but as sediment load increases, inter-granular collisions dissipate an increasing proportion of sediment kinetic energy and sediment armors the bed, reducing the abrasion rate (Sklar and Dietrich, 2004). From these experimental results, mechanistic models have been developed to predict bedrock incision as a function of channel geometry, bedrock strength, and sediment characteristics (Sklar and Dietrich, 2004; Turowski, 2007; Lamb et al., 2008). Scaled appropriately, these incision rate laws can be used in landscape evolution models to explore how fluvial topography might depend on sediment, flow, and bedrock characteristics (e.g. Pelletier, 2007) and to test the rate laws by comparing model predictions with observed topography.

The unsteadiness of flow and sediment motion under waves may cause rate laws for wave-driven abrasion to deviate from those derived for steady flow in fluvial environments. Theoretical and numerical models suggest that abrasion rate is a function of the flux rate and kinetic energy of sediment, which are determined by shear stress and sediment characteristics in steady flows (Sklar and Dietrich, 2004; Lamb et al., 2008; Aubert et al., 2016). However, under oscillatory wave motions, normal stresses due to fluid acceleration also contribute to the transport and kinetic energy of sediment (Fredsoe and Deigaard, 1992; Drake and Calantoni, 2001). Also in contrast to fluvial environments, shear stress under waves varies over seconds-long timescales, the boundary layer is not fully developed, and sediment in bedload often comes to a rest every oscillation period (Fredsoe and Deigaard, 1992; Soulsby and Damgaard, 2005)). Finally, bedforms commonly form under oscillatory flow (e.g. Pedocchi and Garcia, 2009), and it is unknown what effect these may have on abrasion.

Here we present results from a series of experiments in which we use a laboratory oscillatory flow tunnel to quantify the abrasion rate of bedrock as a function of oscillation velocity and sediment load. We find that abrasion increases exponentially with oscillation velocity and exhibits a tools-and-cover effect different from that observed

in unidirectional abrasion mill experiments.

4.2 Oscillatory Abrasion Experiments

We hypothesized that rates of wave-driven abrasion are primarily a function of oscillation velocity at the bed, which can be characterized by near-bed root-mean-square velocity (u_{rms}), derived from linear Airy wave theory as:

$$u_{rms} = \frac{H\omega}{\sqrt{8} \sinh kh} \quad (4.1)$$

where H is wave height (m), ω is wave frequency (radians s^{-1}), k is wavenumber (radians m^{-1}) and h is water depth (m). To produce oscillatory flow with a range of u_{rms} at the scale of prototype waves, we constructed a prototype oscillating u-tube (POUT) that consisted of a 2.45 m-long horizontal duct with rectangular cross-section and two vertical reservoir pipes (Figs. 4-1A, 4-2). Using two thrusters and a control system, sinusoidal water-level fluctuations were generated and maintained in the vertical reservoirs. Water level in one reservoir was recorded at 10 Hz with a pressure sensor and was used to calculate u_{rms} in the horizontal duct through conservation of mass. When the POUT was filled, the water oscillated with a natural period of $T_s \approx 4.75$ s and u_{rms} of up to 1 $m s^{-1}$. Given a particular u_{rms} and T_s , the distance a water parcel travels during half of an oscillation is the orbital diameter,

$$d_o = \frac{\sqrt{2}}{\pi} u_{rms} T_s \quad (4.2)$$

To generate measureable erosion during relatively short experiments, we used closed-cell polyurethane foam with a range of low tensile strengths ($\sigma_T = 0.34 - 0.71$ MPa) to simulate reef or bedrock substrate (Scheingross et al., 2014). After an experiment, the foam was removed from the duct, dried in an oven at 100°C for at least 8 hours, and then reweighed. Post-experiment weight (m_{post}) was subtracted from pre-experiment weight (m_{pre}) and normalized by the experiment duration (t_{exp}) and the foam’s exposed surface area (A_{sub}) and density (ρ_{sub}) to calculate incision

Figure 4-1: Experimental design and the influence of bedforms. (A) Schematic drawing of the oscillatory flow tunnel; and (B) video stills from two experiments illustrating a fully mobile sediment band (left column) and a partially static isolated ripple (right column). The blue boxes in B indicate two layers of sediment grains that remain stationary during the entire wave period and armor the bed.

Figure 4-2: Photos of the prototype oscillating u-tube (POUT), including (A) assembled oscillatory flow tunnel, (B) computer controller, (C) close-up of a thruster installed in one arm of the u-tube, and (D) close-up of the acrylic duct with bedforms visible. The erodible foam substrate was bolted flush with the duct floor, which consisted of removable acrylic panels to allow for variable foam length. The tensile strength of each batch of foam was measured using a dual column tabletop testing system (Instron 3360, Norwood, MA, USA).

rate:

$$I = \frac{m_{pre} - m_{post}}{A_{sub} \Delta \rho_{sub} \Delta t_{exp}} \quad (4.3)$$

Because bedrock subjected to the low velocity impacts of rolling or saltating sediment fails in tension (Sklar and Dietrich, 2001), and following previous experiments investigating the volumetric erosion rate of rock, concrete, and foam, we multiplied the incision rate by the square of the substrate tensile strength (σ_T^2) (Sklar and Dietrich, 2001; Scheingross et al., 2014). (In situations where sediment instead slides on top of exposed bedrock, shear strength may be a more appropriate quantity for scaling abrasion rates.)

Sub-rounded siliciclastic gravel was used as the abrasion medium. The gravel had a median (95% confidence interval) sieve diameter of $D_{50} = 6.1$ (4.2-7.6) mm and density $\rho_s = 2570$ (2300-2840) kg m⁻³. Dry gravel was scattered uniformly across the duct floor and foam before initiation of each experiment. We refer to the average volume of gravel per unit area of the duct floor as sediment load, S (cm³ cm⁻²).

Upon initiation of oscillatory motion in each experiment, most of the sediment in the duct rapidly self-organized into one or more bedforms, leaving most of the duct floor covered only by scattered sediment grains. The morphology of the bedforms varied with sediment load. At low S (0.0075-0.015 cm³ cm⁻²), the bedforms consisted of mobile bands one or two grain diameters thick and 10-30 grain diameters wide that migrated back and forth over distances much greater than their width (Videos DR1 and DR2). In experiments with high S (≥ 0.12 cm³ cm⁻²), the bedforms consisted of mounds several grain diameters high with appearance similar to rolling-grain ripples (Perillo et al., 2014). These latter bedforms resembled isolated ripples of sand-sized sediment observed in the field. For example, on the Ningaloo, Australia fringing reef, bare substrate is exposed when coarse sand forms ripples after large wave events remove most sediment from the reef flat (Rosenberger et al., 2019). On mixed sand-gravel beaches, sand forms isolated ripples on otherwise immobile beds when flows are insufficiently strong to mobilize gravel (Hay et al., 2014).

Motion of the isolated ripple bedforms varied with sediment load and u_{rms} , which

Figure 4-3: Variability in number of bedforms with oscillation velocity and sediment load. Data from test runs not listed in Table 4.1 are included.

we categorize as low ($<0.75 \text{ m s}^{-1}$) or high ($\geq 0.75 \text{ m s}^{-1}$). In low- u_{rms} /moderate- S and high- u_{rms} /high- S experiments, the mounds had a stationary core, and sediment saltated from the stoss face of a mound, over the mound crest, to the bare substrate on the lee side every half-oscillation (Fig. 4-1B, Videos DR3 and DR4). In high- u_{rms} /moderate- S experiments, the mounds migrated back and forth in their entirety, primarily through saltation (Fig. 4-1B, Video DR5).

The number and spacing of bedforms in our experiments tended to increase with increasing S and decrease with increasing u_{rms} (Fig. 4-3). When 2-3 bedforms formed, their spacing scaled with d_o (range: $0.6d_o$ - $1d_o$) and was consistent with field observations of orbital ripple spacing in coarse sediment beds (Pedocchi and Garcia, 2009) (Fig. 4-4). We hypothesized that by focusing sediment, these bedforms were likely to produce spatial variability in abrasion rate along the path of fluid flow. Thus, to capture the full range of this variability, the bedrock simulant was cut to a length of $0.8d_o$, one full bedform spacing, for the target u_{rms} of each experiment. Usually only one bedform would be located on the bedrock simulant, while sediment sequestered in other bedforms could not contribute to incision. To account for this sequestration, we normalized S in each experiment by the number of bedforms, which we will refer

Figure 4-4: The relationship between bedform spacing (λ) and orbital diameter (d_o). Field data were compiled from the same literature sources as in Pedocchi and Garcia (2009). Only data with a particle Reynolds number, $Re_p > 13$ are displayed (see Pedocchi and Garcia (2009) for definitions).

to as normalized sediment load (S^* , $\text{cm}^3\text{cm}^{-2}$).

We performed two series of experiments using the POUT (Table 4.1). In the first set of experiments, we held sediment load constant at either $0.03 \text{ cm}^3\text{cm}^{-2}$ or $0.12 \text{ cm}^3\text{cm}^{-2}$ and varied u_{rms} in the range $0.4\text{--}1.0 \text{ m s}^{-1}$. In the second set of experiments, we held u_{rms} constant at 0.5 , 0.6 , or 0.75 m s^{-1} and varied sediment load over $0.0075\text{--}0.48 \text{ cm}^3\text{cm}^{-2}$.

4.3 Controls on Bedrock Incision

Across all experiments, mean incision rate of the bedrock simulant increased with u_{rms} (Fig. 4-5). The relationship between u_{rms} and incision rate was described well by a power law of the form

$$I = C_t K \langle u_{rms} \rangle^n, \quad (4.4)$$

where I is incision rate scaled by σ_T^2 ($\text{m yr}^{-1} \text{ MPa}^2$); $C_t = 3.15 \times 10^7 \text{ s yr}^{-1}$ is a conversion factor for time; K is a constant incorporating the influence of sediment

Figure 4-5: Relationship between oscillation velocity and bedrock incision rate. Vertical error bars represent ± 1 standard deviation in mean incision rate as estimated from a Monte Carlo uncertainty analysis in which measurement uncertainty and variability in material properties (foam weight, foam density, and foam tensile strength) were propagated through the incision rate calculation 10,000 times for each experiment. Horizontal bars indicate the inter-quartile range of oscillation velocity during each experiment. The black line is a nonlinear least-squares fit of a power law to all of the data. Gray shading indicates the 95% confidence interval of the fit.

characteristics and the nature of the abrasion process ($s^{(n-1)} \text{ m}^{-(n-1)} \text{ MPa}^2$); angle brackets $\langle \rangle$ indicate time-averaging over an entire experiment; and n is a dimensionless exponent. A nonlinear least-squares fit provided estimates (95% confidence interval) of $n = 4.2$ (3.6, 4.8) and $C_t K = 1.0$ (0.7, 1.3), with $R^2 = 0.86$. Thus, incision rate scales with u_{rms} raised to a power of approximately 4.

For a given u_{rms} , the relationship between the incision rate and sediment load was characterized by a tools-and-cover effect (Fig. 4-6). The incision rate increased from low S^* to a maximum at an intermediate S^* (tools effect). From this maximum value, the incision rate decreased with increasing S^* (cover effect). The shape of our inferred tools-and-cover effect is similar to that found in rotary mill experiments (Sklar and Dietrich, 2001), but has a slower decay of incision rate, indicating a weaker cover effect. The tools and cover effects can be modeled with a power function of sediment

mass (T17, Turowski and Hodge, 2017). Applied to the abrasion mill experiments, T17 collapses to a special case in which additional sediment falls equally on exposed and already covered bedrock (Turowski, 2007). In our experiments, however, the best fit of T17 indicates that sediment is preferentially sequestered in bedforms, reducing the cover effect (Fig. 4-6).

The tools-and-cover effect in our experiments was influenced by bedform dynamics, such that abrasion rates varied with position relative to the bedforms. Under the “tools” regime, incision rates reached a maximum at the center of the full range of bedform motion and steeply decreased away from that point (Fig. 4-7A). However, when sediment load was high enough that isolated ripple bedforms armored a portion of the bed, incision rates were lowest under the center of the bedform range of motion and highest at the bedform edges, where sediment saltated vigorously (Fig. 4-7A). The spatial variability in abrasion rates was also a function of u_{rms} . For constant sediment load, a large portion of the bed under the bedform was armored at low u_{rms} , whereas the same region experienced maximum incision rates at higher u_{rms} (Fig. 4-7B).

4.4 Discussion

4.4.1 Sediment controls on wave-driven abrasion and reef morphology

Our experiments quantify how orbital velocity and sediment load affect wave-driven abrasion of submarine bedrock. We demonstrate that abrasion rate increases exponentially with orbital velocity and that sediment load provides a secondary control through a tools-and-cover effect. Sediment in our experiments self-organizes into bedforms that are either fully mobile, in which all sediment composing the bedform moves every oscillatory period, or partially static, in which stationary sediment forms the core of the bedform. For constant u_{rms} the location of the maximum abrasion rate moves with increasing sediment load from the mean center of fully mobile bedforms

Figure 4-6: Influence of sediment availability on bedrock incision rate. Vertical error bars represent ± 1 standard deviation of the mean incision rate as estimated in the Monte Carlo uncertainty analysis. Symbol definitions as in Fig. 4-5. Lines are non-linear least squares fits of $I = k_1 S^* (1 - (1 - k_2) k_3 S^*)^{1/(1-k_2)}$ (T17; Turowski and Hodge, 2017) to the data in the figure. k_1 is a free parameter accounting for sediment characteristics; k_2 accounts for the probability (P) that additional sediment falls on exposed or covered bedrock; and k_3 accounts for the maximum instantaneous bedload flux at a given velocity. Single values of k_1 and k_2 were optimized for all data in the figure, while k_3 was allowed to vary for each of the three orbital velocities. For the dashed line, k_2 was set constant at 0.999, equivalent to the abrasion mill case (Turowski, 2007). For the solid line, $k_2 = 1.46$ was optimized from all data.

to the margins of partially static bedforms. We hypothesize that the transition of bedforms from fully mobile to partially static marks the transition in dominance from the tools effect to the cover effect.

The power-law dependence of abrasion rate on oscillation velocity implies a strong feedback between water depth and abrasion rate. Orbital velocity under waves decreases at an accelerating rate with depth (Fig. 4-8). Thus, as a surface lowers through wave-driven abrasion, the abrasion rate will decrease rapidly. The strength of this feedback implies that surfaces formed primarily through abrasion are likely restricted to shallow depths. For example, from 4 m to 8 m depth, the hypothetical reef abrasion rate outlined above drops 85%. This negative feedback relationship between water depth and abrasion rate likely influences the temporal evolution and equilibrium state of coastal landforms like shore platforms and fore reef grooves.

The incision rates observed in our experiments demonstrate the potential impor-

Figure 4-7: Variability in abrasion rate with position relative to bedforms. Incision rate measured along the length of the foam substrate demonstrating the influence of bedforms in three experiments (A) with $\langle u_{\text{rms}} \rangle = 0.75 \text{ m s}^{-1}$, including those in Fig. 4-1B, and (B) with $S^* = 0.12 \text{ cm}^3 \text{ cm}^{-2}$. Incision rates are normalized by the maximum incision rate in each experiment profile. Lines (shading) indicate mean (± 1 standard deviation) of four replicate outside micrometer measurements made at 1 cm intervals down the length of foam substrate.

Figure 4-8: Visualization of the variability in near-bed orbital velocity, u_{rms} , as a function of offshore wave height and water depth for four different wave periods. The parameter space of u_{rms} covered in our experiments is outlined in red. Offshore wave heights were shoaled to a given depth using linear Airy wave theory to calculate u_{rms} . Black contours indicate intervals of 0.2 m s^{-1} .

tance of abrasion for the evolution of nearshore environments. For example, the tensile strength of reef-building corals is typically 1.0-3.2 MPa, while the reef substrate on which the corals grow has even lower tensile strength, typically 0.2-0.8 MPa (Madin, 2005; Madin et al., 2013). Typical trade wind-driven waves in the central North Pacific ($H_s = 1.5$ m, $T_p = 8$ s) account for over 50% of annual mean wave energy (Shope et al., 2016) and generate $u_{rms} = 0.76$ m s⁻¹ in water 4 m deep. If sediment availability were within the range of our experiments, these waves would produce incision rates of 0.3-10 m yr⁻¹ for reef substrate and 0.02-0.4 m yr⁻¹ for healthy coral. In contrast, maximum measured accretion rates of reefs and linear extension rates of individual coral colonies are 0.020 m yr⁻¹ and 0.2 m yr⁻¹, respectively (Montaggioni, 2005). However, care should be taken when applying our results to field settings as abrasive tools erode themselves with impacts against bedrock and other tools, especially if the tools have similar tensile strength as the bedrock.

Assuming that abrasive tools are rapidly replaced on productive reefs, the incision rates implied by our experiments would overwhelm reef accretion and coral growth. That net reef accretion occurs at all implies that sediment capable of generating appreciable abrasion is scarce on reef surfaces, particularly as sediment cover would also hinder coral growth. However, shore-normal grooves common to fore reefs often contain a thin veneer of coarse sediment and show signs of abrasion (Duce et al., 2016; Cloud Jr., 1954). Thus, not only are grooves potentially formed through abrasion, they may also promote reef growth by sequestering abrasive tools.

4.4.2 Comparison to Fluvial Bedrock Abrasion

The dependence of abrasion rate on orbital velocity in our experiments is broadly similar to that found for abrasion under unidirectional flow. Numerical modeling of bedrock abrasion under unidirectional flow suggests that the abrasion rate exhibits a power-law dependence on excess shear stress, with the exponent ranging from 1 to 3 depending on sediment load (Aubert et al., 2016). Sediment in that numerical model had a negligible threshold of motion, so assuming that excess shear stress varies roughly with shear stress and that shear stress scales with velocity squared

(Fredsoe and Deigaard, 1992), the abrasion rate in the numerical model varied with flow velocity raised to a power of 2 to 6, bracketing our estimate. Additionally, the numerical exponent increased with increasing sediment load, which is consistent with our results (Fig. 4-5). However, due to the development of bedforms under oscillatory flow, the cover effect in our experiments was weaker than in the abrasion mill experiments for unidirectional flow. Theoretically the cover effect in bedrock rivers, can differ from the abrasion mill experiments, but here we demonstrate a causal link between bedform formation and a reduced cover effect. The decrease in abrasion rate with distance from the edge of the bedforms implies that there is also spatial variability in the tools effect, which has been assumed absent in fluvial abrasion.

4.5 Conclusions

Wave-driven abrasion has long been hypothesized to be an important driver of coastal change, but its dependence on wave characteristics and sediment availability was poorly quantified. Our experiments demonstrate that abrasion rates increase as a power function of wave orbital velocity and exhibit a tools-and-cover effect with sediment load that is distinct from fluvial abrasion. We provide an empirical relationship between oscillation velocity and incision rate that predicts coastal bedrock abrasion rates from typical wave characteristics. Additional study is required to determine if asymmetrical velocity and acceleration common to shallow water waves cause bedform-cover dynamics and abrasion rates to differ from our experimental conditions.

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Table 4.1: Summary of experiments

Run ID	$\langle u_{rms} \rangle$ (m s ⁻¹)	$\langle T \rangle$ (s)	Incision rate (m yr ⁻¹)	Experiment duration (hr)	Substrate σ_T (MPa)	Substrate density (kg m ⁻³)	Sediment load (cm ³ cm ⁻²)	Number of bedforms
S11	0.95	4.7	1.69	12.6	0.71	146.6	0.12	1
S12	0.77	4.88	0.61	12.55	0.71	146.6	0.12	1
S13	0.64	4.9	0.14	23.53	0.71	146.6	0.12	2
S17	0.88	4.5	1.9	3	0.71	146.6	0.12	1
S18	0.6	4.75	0.78	6	0.34	79	0.12	2
S20	0.52	4.71	0.26	11.23	0.34	79	0.12	2
S21	0.6	4.73	1.21	8	0.34	79	0.06	1
S24	0.61	4.61	0.52	11.44	0.34	79	0.24	2
S25	0.61	5.14	0.26	10	0.34	79	0.48	2
S26	0.64	4.57	1.21	10	0.34	79	0.03	1
S32	0.51	4.54	0.47	4.37	0.46	102.2	0.06	2
S33	0.51	4.69	0.19	8.05	0.46	102.2	0.48	3
S34	0.52	4.63	0.47	9.73	0.46	102.2	0.03	1
S35	0.5	4.74	0.19	10.02	0.46	102.2	0.12	2
S36	0.5	4.77	0.43	8.85	0.46	102.2	0.015	1
S37	0.51	4.71	0.38	14.81	0.46	102.2	0.0075	1
S40	0.5	4.73	0.19	10	0.46	102.2	0.24	2
S41	0.8	4.71	1.94	5.38	0.46	102.2	0.12	1
S42	0.75	4.73	0.95	4	0.46	102.2	0.0075	1
S43	0.76	4.71	1.51	3	0.46	102.2	0.03	1
S44	0.74	4.75	1.32	3.75	0.46	102.2	0.06	1
S45	0.72	4.75	1.32	4.08	0.46	102.2	0.24	2
S46	0.75	4.71	0.95	4.25	0.46	102.2	0.48	2
S47	0.81	4.72	1.42	5.71	0.46	102.2	0.015	1
S48	0.81	4.71	2.27	4.11	0.46	102.2	0.12	1
S49	0.83	4.72	1.89	4	0.46	102.2	0.12	1
S50	0.91	4.75	5.26	2.24	0.39	77.5	0.03	1
S51	0.77	4.75	0.67	6.73	0.71	146.6	0.09	1
S52	0.79	4.76	2.37	4	0.39	77.5	0.09	1
S53	0.8	4.78	2.63	3.39	0.39	77.5	0.03	1
S54	0.77	4.75	0.77	8.38	0.71	146.6	0.06	1
S55	0.44	4.76	0.12	21.97	0.71	146.6	0.03	2
S56	0.52	4.75	0.33	6.2	0.39	77.5	0.06	1
S57	0.57	4.62	0.72	10	0.39	77.5	0.06	2
S58	0.54	4.74	0.46	10	0.39	77.5	0.06	2
S59	0.53	4.72	0.39	6	0.39	77.5	0.06	2
S63	0.5	4.78	0.14	24.19	0.71	146.6	0.09	2

4.A Appendix: POUT Physics

4.A.1 Basic physics

The Prototype Oscillating U-Tube (POUT) is similar to u-tubes used for pedagogical purposes or as manometers, and is governed by the same basic physics. The idealized u-tube has two vertical pipes ("reservoirs" herein) joined by a straight or curving length of pipe ("duct" herein), all with the same cross-sectional area. It is partially filled with water to some level in the reservoirs. As long as pressure on the water surface in both reservoirs is equal, the water level in both reservoirs will remain equal. If the pressure on the water surface in one reservoir is increased relative to the other, the water level in the high pressure reservoir will drop to compensate according to

$$\Delta p = 2\rho g\Delta z, \quad (4.5)$$

where Δz is the drop in water surface elevation (m), ρ is water density (kgm^{-3}), g is gravitational acceleration (ms^{-2}), and Δp is the water surface pressure difference between the two reservoirs (Pa).

If the differential pressure between the two reservoirs were removed while $\Delta z \neq 0$, the water levels in the reservoir would begin to oscillate. Assuming absence of friction, the water levels would oscillate forever about the mean water level with oscillation amplitude Δz and a frequency that can be derived from the conservation of energy, as below.

First, potential energy in the system is a function of the difference in elevation of the water level to the equilibrium (mean) water level, and kinetic energy is defined by the velocity of water flowing through the u-tube. Assuming incompressibility and uniform cross-section, the velocity of the water is uniform throughout the u-tube:

$$E = PE + KE = \rho g A(Z)^2 + \frac{1}{2}\rho ALw^2 \quad (4.6)$$

where Z is instantaneous water surface elevation relative to the equilibrium or mean

water surface, A is cross-sectional area, L is the length of water in the u-tube (so total water mass in the tube is ρAL), and w is vertical water velocity in the reservoir (Fig. 4-9). Assuming absence of friction, energy is conserved and:

$$\frac{dE}{dt} = 0 = 2\rho g A Z \frac{dZ}{dt} + \rho A L w \frac{dw}{dt} \quad (4.7)$$

And simplifying, remembering that $w = dZ/dt$,

$$\frac{dE}{dt} = 0 = \frac{2g}{L} Z + \frac{d^2 Z}{dt^2} \quad (4.8)$$

which describes harmonic motion with a frequency of

$$\omega = \sqrt{\frac{2g}{L}}, \quad (4.9)$$

where ω is angular frequency in $\text{rad}\cdot\text{s}^{-1}$.

To understand instantaneous behavior in the system, we have to look at the Navier-Stokes equations. Let's assume incompressibility and irrotational flow:

$$\begin{aligned} \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} &= \frac{1}{\rho_w} \left(\frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \tau_{xy}}{\partial y} + \frac{\partial \tau_{xz}}{\partial z} \right) - \frac{1}{\rho_w} \frac{\partial p}{\partial x} + \frac{F_x}{\rho} \\ \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} &= \frac{1}{\rho_w} \left(\frac{\partial \tau_{xy}}{\partial x} + \frac{\partial \sigma_{yy}}{\partial y} + \frac{\partial \tau_{yz}}{\partial z} \right) - \frac{1}{\rho_w} \frac{\partial p}{\partial y} + \frac{F_y}{\rho} \\ \frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} &= \frac{1}{\rho_w} \left(\frac{\partial \tau_{xz}}{\partial x} + \frac{\partial \tau_{yz}}{\partial y} + \frac{\partial \sigma_{zz}}{\partial z} \right) - \frac{1}{\rho_w} \frac{\partial p}{\partial z} + \frac{F_z}{\rho} - g \end{aligned} \quad (4.10)$$

$$\nabla \vec{u} = 0$$

where τ and σ are shear and normal stresses (Pa) and F is the sum of any exogenous forcing per unit length ($\text{Pa}\cdot\text{m}^{-1}$). If we assume that the exogenous forcing cancels with viscous effects (as the thrusters do in the POUT) and that flow is radially symmetric about the central streamline, then uniform flow in the reservoirs is governed by

$$\frac{\partial w}{\partial t} = -\frac{1}{\rho_w} \frac{\partial p}{\partial z} - g, \quad (4.11)$$

and in the duct by

$$\frac{\partial u}{\partial t} = -\frac{1}{\rho_w} \frac{\partial p}{\partial x}. \quad (4.12)$$

The conservation of volume holds everywhere,

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (4.13)$$

Flow through any bends or changes in cross-section requires inclusion of the advection terms from Eqs. 4.10, but this can be ignored if one focuses on flow in the reservoir and duct, far from these transitions.

Equation 4.11 states that for flow to occur (i.e. acceleration $\neq 0$), there has to be a pressure gradient in excess of the hydrostatic pressure gradient. For our idealized u-tube, we know acceleration of the fluid in the u-tube from Eq. 4.8 and thus

$$\frac{\partial p}{\partial z} = -\rho g + \rho \frac{2gZ}{L} \quad (4.14)$$

Integrating from the water surface to a given depth in a reservoir, instantaneous pressure is a function of the water surface elevation deviation in that reservoir:

$$p_{res}(s) = \rho g s \left(1 - \frac{2Z_{res}}{L} \right) + p_{air}, \quad (4.15)$$

where s is distance (m) along the center streamline from the water surface in one reservoir, p_{air} is air pressure (Pa), and the subscript $_{res}$ indicates quantities evaluated in the same reservoir.

4.A.2 Design of the POUT

The POUT is a u-tube with two vertical, PVC reservoirs, connected horizontally at their base by an acrylic duct (Fig. 4-10). The square cross-section of the duct allowed for the installation of a lid over the length of the duct so that substrate and sediment could be easily changed between experiments. However, during experiments the U-Tube was filled with water to roughly halfway up the reservoirs, causing hydrostatic

Figure 4-9: Sketch of an ideal u-tube and the physical quantities describing its oscillation.

pressure in the duct to exceed 19 kPa . Twenty-six clasps were installed to hold the lid down against a nylon foam gasket (generic weather-proofing for doors/window) to counteract this pressure. During experimental runs, 440 kg of mason blocks were stacked on top of the duct lid to assist in preventing leakage. Despite this precaution, there was usually some very slow leakage from the far ends of the lid. To prevent cracking of the square acrylic joints, a partially removable exoskeleton of angle-iron was constructed for the duct.

Water level oscillations in the POUT were driven by a reversible underwater thruster (Blue Robotics T200) located in each of the lower reservoirs. These thrusters generate the initial pressure gradient necessary for oscillation and subsequently compensate for frictional energy losses to maintain the amplitude of the oscillation. The thrusters were set to modulate their force output as a cosine function at the system frequency. A pressure sensor (Measurement Specialties MS5837-30BA) with 20 Pa resolution and located at the base of one of the upper reservoirs was used to track water level and derive RMS velocity in the duct. The POUT was designed with reservoirs that had a larger cross-sectional area than the duct to increase the maximum achievable duct RMS velocity without overtopping the reservoirs or sending water

level below the pressure sensor. Unfortunately, the apparent precision of the pressure sensor was $\sim \pm 100 Pa$, making it difficult to precisely target very low RMS velocity.

Closed-cell high density urethane foam (Coastal Enterprises PBLT-4 and PBLT-8) was used as the experimental abrasion surface. This foam was supposedly manufactured with precise densities and strength specifications, but I conducted independent measurements of density and tensile strength and found that a couple of the batches I used departed from the manufacturer specifications. Therefore, I independently verified the tensile strength and density of each sheet (8x4 ft) of foam I used. The substrate was bolted to the acrylic duct floor by pairs of nylon machine screws and washers spaced every 11 cm down the length of the foam. Removable acrylic floor panels (11-cm in width) were used to elevate the surface on either side of the foam and keep it flush with the duct bottom.

Figure 4-10: Scaled diagram of the POUT. The scale bars represent inner diameters, widths, and heights.

4.A.3 Adjustments to ideal u-tube physics for the POUT

Unfortunately, for the purposes of calculating governing physical quantities, the POUT differs from an idealized U-Tube in one very important way: cross-section varies throughout the u-tube. The upper reservoirs have greater cross-sectional area than the lower reservoirs containing the thrusters, and the lower reservoirs have greater cross-sectional area than the horizontal duct. Assuming incompressibility, the conservation of volumetric flux from the reservoirs to the duct requires that the velocity and excursion length of a particular water parcel in the duct or lower arms be greater than a similar parcel in the upper arms by the ratios of their cross-sectional areas. For example:

$$u_{duct} = \frac{A_{res}}{A_{duct}} w_{res}, \quad (4.16)$$

where u and w are instantaneous, cross-sectionally-averaged water velocities and the subscripts $_{res}$ and $_{duct}$ refer to the reservoir and duct, respectively. Of course, with variable water velocity comes a different kinetic energy, requiring rebalancing of Eq. 4.6:

$$E = PE + KE = \rho g A_{res} (Z)^2 + \frac{1}{2} \rho \sum_i A_i L_i U_i^2, \quad (4.17)$$

where $U_i = w_{res} A_{res} / A_i$ is along-streamline velocity, and the subscript i refers to every portion of the POUT with a different cross-sectional area. This changes the system frequency:

$$\omega = \sqrt{\frac{2g}{\sum_i L_i \frac{A_{res}}{A_i}}} = \sqrt{\frac{2g}{L_{apparent}}}, \quad (4.18)$$

where $L_{apparent}$ is the apparent system length. With changing kinetic energy and system frequency comes a different pressure gradient in the u-tube, and

$$p_{res}(s) = \rho g s \left(1 - \frac{2Z_{res}}{L_{apparent}} \right) + p_{air}, \quad (4.19)$$

4.A.4 Equations for running the POUT

The easiest way to calculate characteristic velocity, u_{rms} (ms^{-1}) in the duct is to determine the minimum and maximum water level excursions in the reservoirs. Because the u-tube is a harmonic oscillator, the water level in the reservoir varies according to

$$Z(t) = Z_0 \cos \omega t \quad (4.20)$$

where Z_0 is the average of the minimum and maximum water level excursion from equilibrium each oscillation. From this equation we can derive velocity in the reservoir

$$\frac{dZ}{dt} = w_{res} = Z_0 \omega \sin \omega t \quad (4.21)$$

and RMS velocity in the duct as well using Eq. 4.16 and performing the root mean square over one period:

$$u_{rms} = \frac{1}{\sqrt{2}} \frac{A_{res}}{A_{duct}} Z_0 \omega \sin \omega t \quad (4.22)$$

However, determining Z_0 requires measurements from the pressure sensor. As Eq. 4.19 indicates, pressure measured at the sensor requires correction for the pressure gradient through the entire POUT. If we define p_u as the measured pressure when the water level is at its maximum, p_d as the measured pressure when the water level is at its minimum, and h as the distance between the pressure sensor and mean water level (Fig. 4-9), then we can calculate Z_0 through:

$$\begin{aligned} p_u &= \rho g (h + Z_0) \left(1 - \frac{2Z_0}{L_{apparent}} \right) + p_{air} \\ p_d &= \rho g (h - Z_0) \left(1 + \frac{2Z_0}{L_{apparent}} \right) + p_{air} \\ \frac{p_u - p_d}{2\rho g} &= Z_0 \left(1 - \frac{2h}{L_{apparent}} \right) \end{aligned} \quad (4.23)$$

Hydrostatic pressure can be assumed when measuring h just after filling the reservoirs, remembering to subtract out air pressure (or the pressure measured immediately before filling). Equation 4.23 can be applied using a moving average filter or between

adjacent peaks and troughs in the pressure sensor data. In either case, using Eq. 4.23 is advantageous because it does not require estimating the non-hydrostatic pressure gradient in the u-tube and automatically removes air pressure from the signal, assuming p_{air} does not change over the averaging time interval.

Applying Eq. 4.23 requires knowing the apparent system length, $L_{apparent}$, as defined in Eq. 4.18. In the absence of friction, $L_{apparent}$ could be estimated by evaluating the denominator in Eq. 4.18 piecewise for each portion of the POUT with a different cross-section. This yields a value of $L_{apparent} - 2h = 8.20$ m. However, friction, especially induced by expansion, contraction, and turning of flow, can increase this value. To derive $L_{apparent}$ with friction, I used two methods:

1. At the beginning of each experiment I measured the system frequency of the POUT by raising the water level, turning off the rotors, and using spectral analysis on the resulting oscillations. I then inverted Eq. 4.18 to calculate $L_{apparent}$ for each system frequency and calculated the mean. This yielded $L_{apparent} - 2h = 9.0$ m.
2. I used a camera to track water level in one reservoir arm during a calibration run (using the triangle similarity technique with a calibrated camera), and compared the distance between the upstroke and downstroke measured by the camera to pressure at the pressure sensor and inverting Eq. 4.23. This yielded $L_{apparent} - 2h = 8.78$ m.

Because friction may also have affected my estimates of the system frequency at the beginning of each run (resulting in longer observed system frequency than actual), I used $L_{apparent}$ estimated in method 2 above. Note, applying $L_{apparent} - 2h = 8.20$ m versus $L_{apparent} - 2h = 9$ m in Eq. 4.23 results in mean differences in measured Z_0 , and thus velocity, of $< 2\%$. Thus, calculated velocity is relatively invariant to choice of $L_{apparent}$.

Chapter 5

Tropical cyclogenesis in the Pacific deep tropics during the Little Ice Age

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Abstract

The instrumental record demonstrates that tropical cyclone activity is sensitive to inter-annual and decadal variability in oceanic and atmospheric conditions. However, our understanding of the influence of climate on tropical cyclones is restricted by the short historical record and sparse prehistorical reconstructions, particularly in the western North Pacific. Here we provide a proxy reconstruction of tropical cyclogenesis over the past three millennia in the western North Pacific. The concurrence of peak cyclogenesis in our reconstruction and hydroclimate proxy variability during the Little Ice Age suggests a centennial-scale link between tropical Pacific hydroclimate and tropical cyclone climatology. Using an ensemble of global climate models and a synthesis of west Pacific tropical cyclone landfall proxies, we demonstrate that migration of the Pacific Walker circulation and variability in two tropical Pacific climate modes during the Little Ice Age contributed to enhanced cyclogenesis and increased landfall frequency in the tropical western North Pacific.

5.1 Expanding the observational record

Understanding the influence of large-scale climate variability on tropical cyclone (TC) activity is critical for determining the hazard risk of coastal communities as climate changes. From 2010 to 2030, global population growth is estimated to increase the number of people exposed annually to TCs from 133.7 million to 149.3 million, 90% of whom live along the Asian Pacific coastline (Peduzzi et al., 2012). Simultaneously, anthropogenic climate change is expected to alter TC climatology over the next century. Climate provides the boundary conditions that determine the frequency, geographic distribution, and intensity of TCs (Emanuel et al., 2004; Gray, 1968). However, the influence of global climate is mediated by regional oceanic/atmospheric circulation that varies on seasonal to centennial timescales (Wu et al., 2012; Brandon et al., 2013; van Hengstum et al., 2016; Gao et al., 2018; Huang and Oey, 2018). To understand future TC risk and how it might change in response to anthropogenic radiative forcing, we must understand how TCs have varied in response to past forcing and the regional mechanisms governing that response.

Our understanding of climate controls on TCs is limited by the observational record. Satellite records provide comprehensive spatial coverage of TC formation and propagation, but only extend to the early 1970's (Camargo et al., 2007a). Historical records have longer temporal coverage (e.g. International Best Track Archive for Climate Stewardship, IBTrACS, 1848-present), but are less spatially comprehensive than satellite records. Paleoclimate proxies lack the spatial coverage of satellite and historical records, but produce much longer records at comparable temporal resolution. Recently, records of TC landfall have been constructed with high temporal resolution from speleothem proxies of cave flooding (Denniston et al., 2015) and proxies of sediment transport over coastal barriers, over fringing reefs, or into coastal karst basins (Woodruff et al., 2015; Toomey et al., 2016b; Donnelly et al., 2015; van Hengstum et al., 2016, 2014; Lane et al., 2011; Gischler et al., 2008; Donnelly et al., 2001). These reconstructions extend the observational record by several millennia and reveal large centennial variability in TC activity. However, although the western North Pacific

(WNP) is the most active basin for TCs globally, few pre-modern reconstructions with annual-decadal resolution have been developed there (Liu et al., 2001; Yu et al., 2009; Woodruff et al., 2015) relative to the western North Atlantic (Donnelly et al., 2015; van Hengstum et al., 2016, 2014; Lane et al., 2011; Boldt et al., 2010; Gischler et al., 2008; Donnelly et al., 2001).

The dearth of long-term TC reconstructions in the western Pacific hinders identification of the processes driving low-frequency variability in TC statistics, including their response to external forcing. The farther TCs travel, the more they are influenced by post-genesis environmental conditions (Emanuel et al., 2004). Thus, it is difficult to separate the influence of different aggregate TC characteristics (i.e. genesis frequency, genesis location, storm track, and landfall intensity) on reconstruction variability when a reconstruction is located far from genesis locations. In the western North Atlantic, recent research has overcome this difficulty by contrasting records at many sites across the basin. For example, comparisons revealed that regional ocean circulation likely drove changes to the dominant storm track, which in turn resulted in a cross-basin shift in TC landfall frequency from the Gulf of Mexico to the northeastern US about 550 cal. yr BP (Donnelly et al., 2015; van Hengstum et al., 2016). In the WNP, however, the few TC reconstructions that exist are located far from the primary regions of cyclogenesis. Thus, attempts to contrast records on the western margin of the basin have failed to identify the TC characteristics and drivers responsible for a clear southward shift in landfall frequency about 500 cal. yr BP (Woodruff et al., 2015; Yan et al., 2016), for example.

Here we introduce a sediment proxy reconstruction of TC landfall frequency that captures genesis frequency variability in the deep tropical central Pacific—a key hotspot of tropical cyclogenesis that feeds the broader WNP basin. Tropical cyclogenesis potential in this region is expected to increase with warming sea surface temperatures (SST) and weakening of vertical wind shear forecast over the 21st century (Zhang et al., 2017). The short travel distance between genesis location and proxy sites mitigates the influence of storm track and post-genesis intensification on our reconstruction. By focusing on TC genesis and using a multi-model ensemble

of general circulation models (GCMs), we can identify the large-scale climate patterns influencing cyclogenesis variability near our site and identify the climate modes driving those patterns.

5.2 Methods

5.2.1 Study site and field methods

We developed a sediment proxy reconstruction of TC activity at Jaluit Atoll in the southern Marshall Islands. Jaluit Atoll is a mid-ocean atoll with a 250-1000 m wide intertidal reef flat encircling a shallow lagoon. The mildly stratified blue hole from which we extracted sediment cores LTD2 and LTD3 is located on the northwestern tip of Jaluit Atoll, where the atoll rim is widest (Fig. 5-1). The blue hole is flanked by three small, vegetated islands between which smooth channels connect it to the ocean-facing reef flat, which is covered in crustose coralline algae. The blue hole is backed by a deeper reef flat, 400 m-wide with actively growing coral.

The sediment cores were retrieved using a Rossfelder P-3 vibracoring system. For core LTD2, a separate surface drive of the upper 1-1.5 m of sediment was retrieved to preserve the sediment surface, because the primary drive over-penetrated. Coring sites were selected based on bathymetry and absence of hard substrate such as large coral heads. Bathymetry and seismic surveys were recorded using an Edgetech 3100 Chirp sub-bottom sonar system with a 4–24 kHz fish towed behind a small outboard-motor craft. Cores were sectioned in the field before transportation to the lab.

5.2.2 Laboratory analysis

Cores were split and described in the lab (Sec. 5.3.1), then sampled at 1-cm intervals. Samples were wet-sieved at 63 μm and dry-sieved at 250 and 2000 μm to obtain dry weight-normalized grain size fractions at each sieve size.

We established age controls for each core with a series of accelerator mass spectrometer (AMS) radiocarbon dates (Table 5.1) and assumed core-top dates. Where

Figure 5-1: Maps of the reconstruction site: a) Map of the tropical western Pacific, with the storm tracks of every tropical cyclone in the IBTrACs dataset to pass within 100 km of Jaluit Atoll; b) map of the site showing location of cores as green asterisks; c) bathymetric profile of the basin from which the sediment cores were extracted, with a salinity profile.

available, we sampled terrestrial organic macrofossils for dating, but they were rare, especially below the first few meters. Most of our core dates were measured from detrital inorganic carbonate, and thus have reduced precision and greater uncertainty due to the marine reservoir effect. All radiocarbon dating was performed at the National Ocean Sciences Accelerator Mass Spectrometer (NOSAMS, Woods Hole, MA, USA) facility. Sediment at the top of each core was assumed deposited immediately prior to core extraction in November 2015. The top core section of the LTD2 primary drive did not capture the sediment-water interface, so the location of the core-top was inferred from the separate surface drive using a near-surface tie point. Using tie points this way assumes deposition rate in the vicinity of the core-top was equal between the primary and surface drive. For LTD2, we used the base of a large coarse event bed as a tie point (Fig. 5-2). The base of a large storm event bed common to all LTD2 and LTD3 drives was used as an additional tie point for LTD3 age control (Sec. 5.3.1).

The LTD3 age model (Fig. 5-2) was generated using the Bayesian age modeling software BACON (Blauuw and Christen, 2011), using the IntCal13 curve for terrestrial organic carbon dates (Reimer et al., 2013), the Marine13 curve for inorganic carbon dates from detrital carbonate sediment grains (Reimer et al., 2013), and Northern Hemisphere Zone 3 (ITCZ region) post-bomb curve for one post-bomb date (Hua et al., 2013). BACON constructs an ensemble of accumulation histories through Markov Chain Monte Carlo iteration on provided dates and thus provides a distribution of likely ages for every centimeter of the core. Inorganic carbonate dates were corrected for a marine reservoir effect using $\Delta R = 41 \pm 42$ yr, determined for the past 2500 years from branching coral samples retrieved from Ebon Atoll, Republic of the Marshall Islands, 200 km southwest of the Jaluit Atoll blue hole (Weisler et al., 2018).

5.2.3 Event bed selection

We used a simple statistical analysis of sediment grain size to identify event beds in our cores (Fig. 5-9). First, outliers were identified as samples of coarse fraction

Figure 5-2: Profiles of sediment cores collected from the Jaluit Atoll blue hole. a) BACON age model for LTD3 and b) coarse fraction profiles of LTD2 and LTD3. BACON-calibrated radiocarbon dates are displayed as green triangles. Carbonate samples that were dated using gas ion source are indicated with the acronym GIS. A tie point used in the age models and to establish core top depth by comparing drives are indicated with a blue dashed line. Storm beds were identified as those samples that exceeded 1.5 standard deviations above an 11-cm moving average, where both statistics were calculated while ignoring >2 standard deviation outliers.

that exceeded 2 standard deviations above the mean coarse fraction, both statistics calculated within an 11-cm moving window, encompassing roughly 50 years according to the age model. Then the moving average was recalculated, but this time excluding outliers. Finally, event beds were identified as peaks in the moving average-subtracted coarse fraction anomaly that exceeded 1.5 standard deviations of coarse fraction, calculated for the entire core excluding outliers. Removing outliers from the analysis prevented particularly large event beds from masking the presence of smaller peaks nearby, and using the moving window removed the influence of decadal variability in the background coarse fraction signal. We found that increasing the window size, increasing the threshold coarse anomaly, and including outliers affected the absolute number of event beds identified, but did not change the qualitative patterns in event bed frequency in core LTD3 (Figs. 5-10 and 5-11). For the statistical analysis, we used the coarse fraction with sieve diameter between 250 μm and 2 mm to account for anomalous transport characteristics of very coarse bioclastic sediment (Sec. 5.3.2).

We calculated centennial event frequency using a procedure that incorporates uncertainty in the age model. For every identified storm event bed, we randomly sampled 9000 ages from the Bayesian age model ensemble and from them constructed a probability distribution function (PDF) with annual resolution. We then summed the values of all PDFs within a 100-year moving window, resulting in a time series of continuous centennial event frequency estimates incorporating relative uncertainty and age model shape for each event (Fig. 5-9). We estimated a record-wide mean event frequency of 1 event/century at Jaluit Atoll by counting all identified event beds and dividing by the length of time captured by the core. Mean frequency was used to construct a hypothetical Poisson distribution describing the number of events expected per century. Active and passive intervals were identified as those intervals in the frequency time series where the number of events exceeded the 95% confidence interval of the Poisson distribution for at least one century (Figs. 5-10 and 5-11).

5.2.4 Best track analysis

To illustrate modern TC genesis and tracks in the vicinity of our reconstruction sites under modern climate conditions, we calculated simple statistics on 6-hourly best track data from IBTrACS-WMO, v03r10 (Knapp et al., 2010) for western Pacific and eastern Pacific basins encompassing Jan. 1848 - Jun. 2017.

5.2.5 Climate model analysis

We used the results from seven general circulation models (GCMs) to explore possible drivers of anomalous TC climatology during the Little Ice Age (LIA). We used every model involved in the fifth iteration of the Coupled Model Intercomparison Project (CMIP5) that provided results for the Last Millennium experiment (Table 5.2), except for the Model for Interdisciplinary Research on Climate Earth System Model (MIROC-ESM), which was excluded because of its long-term drift (Sueyoshi et al., 2013). We used monthly mean values for all of our analyses.

We quantified TC genesis potential using a genesis potential index (GPI; Emanuel, 2010; Emanuel and Nolan, 2004):

$$\text{GPI} = |\eta|^3 \chi^{-4/3} \max(\text{PI} - 35 \text{ ms}^{-1}, 0)^2 (25 \text{ ms}^{-1} + V_{\text{shear}})^{-4}, \quad (5.1)$$

where η is absolute vorticity at 850 hPa; χ is a measure of the moist entropy deficit of the middle troposphere and represents an entropy gradient barrier to cyclone intensification (Emanuel et al., 2008); PI is potential intensity, an estimate of maximum achievable wind speeds as a function of convectively available potential energy (Bister and Emanuel, 1998); and V_{shear} is absolute wind shear between 250 hPa and 850 hPa. As can be inferred from the sign of the exponent applied to each variable, vorticity and potential intensity tend to promote TC formation and intensification, respectively, while the moist entropy deficit and wind shear tend to depress or interrupt formation and intensification, respectively.

We calculated GPI and all of its components at each model’s native spatial resolution before interpolating linearly to a common $1^\circ \times 1^\circ$ grid. Relative anomalies

in each index between the LIA (1400-1700 CE) and the MCA (1000-1300 CE) were calculated according to $(\text{LIA} - \text{MCA}) / \text{MCA} \times 100\%$ for each model. Multi-model ensemble averaging was performed using the median statistic as the last step in any analysis where it is presented (e.g. percent change between the MCA and LIA was calculated for each model separately and then the median taken for a multi-model ensemble value and displayed in Fig. 5-5). For all comparisons, time averages include only values from each hemisphere’s primary storm season, i.e. July-November (JASON) in the North Pacific and December-April (DJFMA) in the South Pacific.

To determine the magnitude of vertical wind shear (potential intensity) anomalies during the LIA relative to variability throughout the last millennium, we conducted a bootstrap analysis in which we calculated the ensemble median, meridional mean shear (potential intensity) anomaly, relative to the last millennium mean, for every 300-year period within the last millennium experiment results. From these populations we calculated the 95% confidence intervals of the anomalies at each latitude.

The Pacific Meridional Mode (PMM) indexes were calculated from SST and meridional and zonal surface wind fields for 20°S-30°N, 150-265°E. First, the influence of ENSO variability on the fields was removed by subtracting a linear least squares fit between the fields and the Cold Tongue Index, as in Chiang and Vimont (2004). Maximum Covariance Analysis was then applied to the cross-covariance matrix between SST and both wind fields (with land cells masked out). The PMM SST and wind indexes were identified with the second mode for the 1000-1850 CE period. We measured correspondence between mean annual storm season PMM and vertical wind shear using Spearman rank correlation for the years 1000-1850 CE. Statistical significance was calculated with a two-tail Student t-test after taking multiple testing into account with application of the false discovery rate procedure and setting $q = 2.5\%$ (Hu et al., 2017).

5.3 Reconstructing tropical cyclone activity

5.3.1 Core descriptions

We retrieved primary (318 cm) and surface (127 cm) drives for core LTD2 from 30 m depth at the oceanward end of the Jaluit Atoll blue hole (Fig. 5-1c). The surface of the core is characterized by a 15 cm layer of very fine to fine sand overlying a 10-cm-thick event bed of coarse sand and shell hash. Below this first event bed the core is characterized by visually uniform medium-coarse sand down to 296 cm, below which the sediment grades to fine sand. The core is visibly interspersed by large shell fragments and shell hash.

Core LTD3 was retrieved as a single, 561-cm-long drive from 28 m depth at the lagoonward end of the main blue hole basin. The surface of the core is characterized by a thin layer of silt that rapidly grades to fine to medium sand in the first few centimeters. The rest of the core is characterized by visually uniform, very fine to fine sand interspersed with shell fragments.

The sediment in our cores consists nearly entirely of detrital carbonate shell and skeletal material from foraminifera, coral, coralline algae, *Halimeda* green algae, echinoderms, and mollusks. Loss-on-ignition analysis of the organic carbon content was conducted at 10 cm intervals down the first 120 cm of LTD2 and 250 cm of LTD3. Organic carbon made up a uniform $4.5 \pm 0.1\%$ (median $\pm$ standard deviation) of the sediment by dry weight in LTD2 and $4.3 \pm 0.1\%$ in LTD3.

We submitted 25 samples for radiocarbon dating at NOSAMS. These included 6 organic carbon samples of root or bark material and 19 inorganic carbonate samples (8 foraminifera samples, 10 mollusc shell samples, and 1 *Halimeda* calcareous algae plate sample). Following previous studies demonstrating the presence of post-bomb radiocarbon in living *Halimeda* (Holmes, 1983; Xu et al., 2015), which implies the absence or reduction of a marine reservoir effect for these shallow algae, we did not apply a marine reservoir effect for our 1 *Halimeda* sample. All but one of the inorganic carbonate samples returned calibrated ages implying a nearly constant sedimentation rate of c. 1.8 mm y^{-1} . However, three of the organic carbon samples

returned anomalously young ages. We speculate that the anomalously young samples may have been dragged from a shallower depth during coring or core splitting. An alternative hypothesis, that the three young organic dates and one outlier inorganic carbon date define the sedimentation rate of the core, would imply that all of the inorganic carbonate material is reworked older material. The presence of older, reworked material is common in carbonate environments, but is generally apparent as a wide distribution of inorganic carbonate dates at many depths, with no obvious trend in mean age with depth (e.g. Chen et al., 2019). In contrast, our inorganic carbonate dates lie close to a line of nearly constant slope, which would be improbable if they were reworked material.

5.3.2 Accounting for irregularly-shaped bioclastic sediment

In this study, we use coarse deposits as a proxy for sediment transport induced by anomalously strong currents. In our implicit model, sediment is entrained over the surrounding shallow reef environment and transported across the blue hole basin, where some portion settles out of surface currents and is deposited in the blue hole. Assuming that grain size positively correlates with settling velocity, the strength of surface currents determines the maximum grain size mobilized and the distance sediment is transported across the blue hole before settling out of the surface currents. With slow, fair weather currents, mainly fine-grained sand and silt are deposited, while fast, storm-driven currents can mobilize and deposit thick layers of coarse sand. Using grain size as a proxy for anomalous mobilization and transport assumes that settling velocity strongly and monotonically increases with grain size. A very strong relationship between grain diameter and settling velocity exists for spherical and near-spherical sediment grains (Dietrich, 1982), but coarse biogenic carbonate sediment, like that produced in shallow reef environments, have irregular shapes and can have significant effective porosity, resulting in mobilization thresholds and settling velocities lower than would be predicted from their size (Kench and McLean, 1996; Braithwaite, 1973; Maiklem, 1968). In particular, above a characteristic grain diameter of about 2 mm, the settling velocity (and inferred mobilization threshold) of some

bioclastic sediment plateaus (Braithwaite, 1973; Maiklem, 1968). If that sediment made up a substantial portion of the coarse fraction in a deposit, inferences about the strength of the current that mobilized and transported it before deposition could be invalidated.

Due to the deviation in the relationship between grain size and settling velocity for some bioclastic sediment, we identified event layers in core LTD3 using the sediment fraction with sieve diameter 250-2000 μm (herein the “corrected fraction”). We found that using the corrected fraction reduced the size of some apparent coarse layer peaks, especially below 3 m depth in the core (Fig. 5-1e). In a few instances apparent coarse peaks in the $> 250 \mu\text{m}$ fraction were absent in the corrected fraction. For at least two of those peaks a solitary, intact, anomalously large gastropod shell generated most of the uncorrected coarse fraction peak, justifying the use of the corrected fraction here. However, we found that trends in event frequency were relatively invariant to the coarse fraction used for event identification (Figs. 5-10 and 5-11). In both coarse fractions there is an LIA active interval in event frequency preceded by a long, relatively quiescent interval. With the corrected fraction, the LIA active interval has a longer duration and peaks earlier (Fig. 5-10). However, both coarse fractions identify the same coarse deposit initiating the active interval.

5.3.3 TC proxy justification

The storms recorded in the Jaluit Atoll sediment cores were likely typhoon strength and higher (>64 kt sustained wind velocity). A post-bomb date taken from the most recent event bed in LTD2 indicates the event bed, present in all LTD cores, was deposited by severe typhoon Ophelia in 1958 (Blumenstock, 1958). Age models developed for LTD2 and LTD3 suggest the second most recent event bed was deposited roughly 50 cal. yr BP, making it contemporaneous with an unnamed typhoon in 1905 known to have damaged islands and island villages across Jaluit Atoll (Spennemann and Marschner, 1994). The statistical process we used to identify storm events excluded lower magnitude peaks between these two events, implying our storm frequency reconstruction represents frequency of landfall or near misses by typhoons to

the exclusion of less severe or more distant TCs. Additionally, the location of our site far from tsunamigenic earthquake activity along the Pacific Rim and the rarity of tsunamis relative to severe TCs in the WNP make it unlikely that tsunamis contaminated our storm record.

We argue that landfall frequency variability in our reconstructions is primarily a function of cyclogenesis conditions at near-equatorial latitudes to the south and southeast of our reconstruction sites. Only 28 of 1703 TCs in the IBTrACS dataset that formed in or entered the WNP since 1848 formed east of 180°E . Only 8 of those formed below 10°N , and none crossed 180°E below 10°N , as WNP TCs very rarely move south, especially at low latitudes (Camargo et al., 2007a). Thus, with modern climatology, TCs in our proxy records are unlikely to have formed far from the reconstruction sites at 169.4°E and 172.1°E . We assume the short travel distance means variability in our reconstructions is generated by variability in genesis conditions, and not post-genesis environmental factors.

Were genesis locations significantly further east sometime in the past, TCs would have traveled further and likely obtained higher intensities before making landfall at our sites. Higher average landfall intensity would result in greater recorded landfall frequency, even if genesis frequency and intensity remained unchanged. Under these conditions, an independent change in environmental conditions that increased (decreased) post-genesis intensification would amplify (dampen) the effect of shifted genesis location. However, this post-genesis intensification still requires variability in genesis conditions, consistent with our assumption that the reconstructions reflect variability in cyclogenesis to the south and southeast of our sites.

5.4 TC genesis over the past 3,000 years

We developed a grain size proxy of anomalous cross-reef sediment transport from a sediment core retrieved from a blue hole, a 30 m deep karst basin, on Jaluit atoll (6.256°N , 169.411°E) (Fig. 5-1). Coastal basins like blue holes preserve evidence of the close passage of intense TCs as anomalously coarse layers of sediment (van

Hengstum et al., 2014; Gischler et al., 2008). We sieved the core sediment and identified coarse TC deposits as peaks in the sediment coarse fraction (250-2,000 μm sieve diameter) that exceeded a statistical threshold (see Methods). A post-bomb radiocarbon date from the youngest coarse layer indicates that it was deposited in the late 1950s and probably by Typhoon Ophelia in 1958, which passed directly over Jaluit Atoll with estimated sustained winds of over 64 ms^{-1} (Category 4 on the Saffir-Simpson scale) and caused widespread destruction of island vegetation and manmade structures (Blumenstock, 1958). The coarse fraction anomalies of all identified event beds are close to or exceed that of Typhoon Ophelia, suggesting that the frequency of coarse deposits in our core corresponds to the frequency of intense TC passage near Jaluit Atoll over the past c. 3,000 years. Due to the position of the site at the far southeastern corner of the WNP basin, near the geographical limits of observed tropical cyclogenesis, and due to the dependence of the Coriolis force on latitude, we interpret temporal variability in our record as directly corresponding to temporal variability in nearby cyclogenesis (Sec. 5.3.3).

Our proxy reconstruction of centennial TC frequency displays a mean of 1 event/century and centennial variability similar to other reconstructions from the western Pacific. The most prominent feature in our record is a peak in frequency c. 1350-1700 CE with a maximum of 3.75 events/century that encompasses the early Little Ice Age (LIA, 1400-1700 CE) and is higher than during any other period in the record (Fig. 5-3e). We identified no storm layers during three periods exceeding two centuries in length: 950-710 BCE, 290-10 BCE, and 100-850 CE. Immediately preceding the LIA peak was a relatively quiet interval encompassing the Medieval Climate Anomaly (MCA, 1000-1300 CE). The LIA peak in cyclogenesis at our sites is synchronous with decreased landfall frequency in northwestern Australia (Denniston et al., 2015) and a substantial increase in the frequency of coarse deposits at Yongshu Reef in a South China Sea sediment core, indicating enhanced landfall frequency relative to the MCA (Yu et al., 2009) (Fig. 5-3c,d). A sedimentary reconstruction of TC landfall on Tahaa, French Polynesia recorded few storms over the past 1000 years (Toomey et al., 2016b), but reef-top storm deposits collected from across the central South Pacific

basin suggest higher TC activity during the MCA than the LIA (Toomey et al., 2013; Kench et al., 2018b) (Fig. 5-3a,b).

5.5 Large-scale drivers of TC variability

In the instrumental record, the El Nino Southern Oscillation (ENSO) dominates inter-annual variability in western Pacific tropical cyclogenesis. During El Nino events, the mean TC genesis location in the WNP shifts to the southeast, and TCs tend to last longer, obtain higher intensities, and recurve more than in non-event years (Wang and Chan, 2002; Kim et al., 2011). Thus, El Nino events tend to result in more frequent cyclogenesis near Jaluit Atoll, but fewer storm tracks crossing the South China Sea near Yongshu Reef (Patricola et al., 2018). In the South Pacific, El Nino events tend to shift the mean genesis location eastward, resulting in fewer cyclone landfalls in Australia, but more in French Polynesia (Denniston et al., 2015; Toomey et al., 2013). These effects tend to scale with the intensity of ENSO-associated SST anomalies and are more sensitive to warming associated with central Pacific than eastern Pacific El Nino events (Patricola et al., 2018).

Given the dominant influence of ENSO on cyclogenesis in the WNP historically, variations in ENSO over the past few millennia could have contributed to the variability in our reconstruction. Analysis of a synthesis of annually-resolved SST proxies for the Nino3.4 region (Emile-Geay et al., 2013) demonstrates that SST variability associated with ENSO was lower over much of the last millennium relative to the last century (Fig. 5-4a). However, over this background signal there were two century-long periods of elevated ENSO variability that correspond with the beginning and end of the active interval in our cyclogenesis reconstruction. Additionally, an annually-resolved reconstruction of SSTs in the Nino4 region from Taiwanese tree ring $\delta^{18}\text{O}$ reveals a peak in ENSO-associated variability c. 1350-1425 CE, during peak cyclogenesis in our reconstruction (Liu et al., 2017). Coral-based SST proxies from the central Pacific detail similar, but shorter-duration fluctuations in ENSO variability during the LIA (Emile-Geay et al., 2015) (Fig. 5-4b). These transient peaks in ENSO-

Figure 5-3: Reconstructions of tropical cyclone activity in the Pacific Ocean over the past 3000 years. a) radiocarbon dates from storm-deposited reef blocks, coral heads, and conglomerates from across the western South Pacific (Toomey et al., 2013; Kench et al., 2018b); b) coarse storm deposits in a back-barrier reef lagoon, Taha'a, French Polynesia (Toomey et al., 2016b); c) flood frequency reconstructed from mud deposits on speleothems in cave KNI-51, northwestern Australia (Denniston et al., 2015); d) coarse fraction (black line) and centennial frequency (blue line) of large wave deposits in a back-barrier reef lagoon, Yongshu Reef (Yu et al., 2009); e) anomaly of 250-2000 μm (black line) and $>250 \mu\text{m}$ (gray line) coarse fractions in a blue hole sediment core, Jaluit Atoll, Marshall Islands and centennial frequency (blue line) of identified storm deposits (red asterisks) (this study). The horizontal dashed line in e) indicates the coarse fraction anomaly threshold used to identify storms in the sediment record. Yellow and cyan boxes indicate the MCA (1000-1300 CE) and LIA (1400-1700 CE) time periods used for our climate model analysis.

associated SST variability imply more frequent or more intense El Nino events, either of which would have increased the frequency of TC cyclogenesis near Jaluit Atoll and thus could increase the frequency of TC events in our sediment record.

However, the above SST reconstructions indicate that ENSO-associated variability was higher during the last century than during the LIA, contrary to cyclogenesis in our reconstruction. Additionally, the enhanced frequency or magnitude of El Nino events suggested by peaks in ENSO-associated variability cannot explain enhanced landfall frequency near Yongshu Reef and reduced TC activity across the South Pacific during the LIA. Thus, it is likely that other factors contributed substantially to cyclogenesis variability over the last millennium.

Figure 5-4: Comparison of our storm reconstruction with Pacific paleoclimate proxies. a) multi-proxy reconstruction of Nino3.4 region SST anomalies (SSTA; black line) and ENSO-band standard deviation (SD; red line) calculated in a 31-year moving window on the SSTA data (Emile-Geay et al., 2013); b) normalized standard deviation (SD) of ENSO-band SSTs derived from coral $\delta^{18}\text{O}$ proxies in the western and central Pacific (Emile-Geay et al., 2015) (see Cobb et al., 2013 for data processing methods); c) lipid δD proxy of precipitation from Washington Lake, Washington Island, Northern Line Islands (Sachs et al., 2009); d) ice accumulation rates in meters water equivalent (m.w.e.) in a core from the Quelccaya ice cap, Peru (Thompson et al., 2013); e) speleothem $\delta^{18}\text{O}$ proxy of precipitation from cave KNI-51, north-western Australia (Denniston et al., 2016); f) CMIP5 multi-model ensemble median (red line) $\pm$ standard error (grey shading) of decadal-averaged Pacific Meridional Mode wind index anomaly (this study); g) comparison of age model histograms from samples indicated by black asterisks in c) and h); h) same as Fig. 1e. The red circles in b) are independent calculations of ENSO-associated variance in 30-year intervals of a coral core. The error bars in c) indicate $\pm$ 1 standard deviation. The black asterisks in c) and h) indicate the locations of data used in the age model comparison in g). Yellow and cyan boxes indicate the MCA (1000-1300 CE) and LIA (1400-1700 CE) time periods used for our climate model analysis. Dashed black lines indicate 1000-1850 CE means, dotted black lines indicate zero values.

Hydroclimate proxies from the Indo-Pacific Warm Pool (IPWP) and the central and eastern Pacific demonstrate rapid, synchronous, and spatially consistent change at the onset of the LIA. Around 1400 CE, $\delta^{18}\text{O}$ records from speleothems in northwestern Australia and southern China demonstrate a rapid transition to a drier climate that then lasted to c. 1750 CE (Fig. 5-4e, Denniston et al., 2016). Simultaneously, multiple precipitation proxies in the deep tropical IPWP and South China Sea indicate the rapid onset of wetter climate (Yan et al., 2011, 2015a). These shifts have been attributed to a contraction of the Intertropical Convergence Zone (ITCZ) (Yan et al., 2015a; Denniston et al., 2016) (Fig. 2e), and a westward shift in the Pacific Walker Circulation (PWC) (Yan et al., 2011). Simultaneously, an algal lipid δD record indicates the rapid onset of dry conditions in the central North Pacific (Sachs et al., 2009) (Fig. 5-4c), and enhanced accumulation of ice in Quelccaya ice cap, Peru (Thompson et al., 2013) (Fig. 5-4d) indicate wetter conditions in the eastern South Pacific. These meridional shifts in precipitation have been attributed to a southward displacement of the ITCZ during the LIA. A radiocarbon date from a sample 5 cm above the first event layer in the LIA active interval constrains the timing of the first event in the age model such that its modeled age distribution has considerable overlap with the onset of these changes in hydroclimate proxies in the Pacific (Fig. 5-4g). Thus, the start of the LIA active interval may have been concurrent with the basin-wide transitions between wet and dry conditions.

The tropical atmospheric circulation patterns driving hydroclimate variability during the LIA could have influenced cyclogenesis across the western Pacific. Contraction of the ITCZ would have been driven by a contraction in Hadley circulation, which dictates the characteristics (latitude, width, intensity) of zonal mean precipitation in the tropics. Recent research suggests tropical cyclogenesis in the west Pacific has shifted to the north over the past few decades as the Hadley cell has expanded poleward in response to anthropogenic climate change across the region (Kossin et al., 2014, 2016; Sharmila and Walsh, 2018). Assuming Hadley cell contraction would have the opposite effect, contracted Hadley circulation during the LIA could have enhanced cyclogenesis in the deep tropics near Jaluit Atoll. Alternatively, inter-annual variabil-

ity in the Pacific Meridional Mode (PMM) has influenced PWC and correlated with zonal shifts in WNP cyclogenesis over the past few decades (Hong et al., 2018; Gao et al., 2018). Thus, westward migration of the rising limb of the PWC could have shifted cyclogenesis westward during the LIA as well.

5.6 Detection in a climate model ensemble

To explore the associations between tropical atmospheric circulation and temporal variability in our reconstruction, we analyzed the results of the last millennium experiment in an ensemble of Coupled Model Intercomparison Project phase 5 (CMIP5) GCMs. By diagnosing a genesis potential index (GPI, (Emanuel and Nolan, 2004; Emanuel, 2010)) we determined that the ensemble predicted cyclogenesis anomalies during the LIA relative to the MCA that were consistent with TC proxy reconstructions across the west Pacific. The models predicted anomalously low GPI in the South Pacific in the vicinity of French Polynesia and northwestern Australia (Fig. 5-5a), consistent with South Pacific reef-top storm deposits and a speleothem flood reconstruction from northwestern Australia (Fig. 5-3a,c). The ensemble also predicted anomalously high GPI in the vicinity of the Philippines. Assuming TCs entering the South China Sea are generated primarily to the east and southeast, these results are consistent with a TC proxy reconstruction from Yongshu Reef that recorded enhanced TC activity during the LIA relative to the MCA (Fig. 5-3d). Finally, in the near equatorial central North Pacific where most TCs recorded at our reconstruction site originate, the ensemble predicted patchy positive GPI anomalies, while potential genesis was predicted to decrease directly over Jaluit Atoll. However, shear stress (potential intensity) was predicted to decrease (increase) within the vicinity of our site with a relatively unchanged mid-troposphere entropy gradient (Fig. 5-5c-e), indicating that TCs generated in the positive GPI anomalies would be able to continue intensification over Jaluit Atoll, despite the lower GPI driven by anomalously low ambient vorticity. Additionally, El Nino events occurring on top of this centennial mean state would likely increase vorticity (Camargo et al., 2007b) and decrease the

moist entropy gradient (Patricola et al., 2018), enhancing GPI over Jaluit Atoll and its vicinity further. However, the lack of inter-model agreement in the equatorial region for GPI and its individual components suggests that the anomalies there may not be a robust response to consistent forcing.

Figure 5-5: Ensemble median relative anomaly in tropical cyclone genesis indexes during the LIA (1400-1700 CE). Relative anomaly was calculated as $\Delta = (\text{LIA} - \text{MCA}) / \text{MCA} \times 100\%$. The a) Genesis Potential Index (Emanuel, 2010) is calculated from four variables: b) low level vorticity, η (s^{-1}), c) vertical wind shear (ms^{-1}), d) potential intensity (ms^{-1}), and e) the mid-troposphere saturation deficit, χ (dimensionless) (Emanuel et al., 2008). The color palettes are aligned so red always indicates increasing cyclogenesis potential. The sign of relative vorticity in the southern hemisphere in b) was reversed so positive change indicates more cyclonic vorticity. Percent change values were calculated from storm season averages for the two time periods. In the northern hemisphere, the WNP storm season (JASON) was used, while for the southern hemisphere, the South Pacific storm season (DJFMA) was used. No data is shown for 1°S - 1°N to indicate the different months used for averaging in each hemisphere. Black stippling indicates grid cells in which at least five of seven models agreed on the direction of change. The green triangles represent the locations of storm reconstructions. Clockwise from the northwest site: Yongshu Reef, South China Sea (Yu et al., 2009), Jaluit Atoll (this study), Tahaa, French Polynesia (Toomey et al., 2013, 2016b), and cave KNI-51, northwestern Australia (Denniston et al., 2015, 2016).

Spatially coherent, large scale shifts in environmental conditions generated most of the modeled GPI anomalies during the LIA. An anomalously large mid-troposphere moist entropy deficit across most of the western Pacific suppressed potential cyclogenesis in the models, while enhanced vertical wind shear across most of the basin would have prevented TC intensification (Fig. 5-5c,e). However, modeled vertical wind shear anomalies were negative in the northwest and positive in the northeast, forming a zonal dipole in the WNP that produced positive GPI anomalies in the South China Sea and east of the Philippines (Fig. 5-5c). Indeed, at either end of the WNP the LIA vertical wind shear anomaly exceeded the 95% confidence interval of anomalies from all 300-year periods in the last millennium experiment (Fig. 5-6a) and thus far exceeded background variability in the model results. Although potential intensity (PI) anomalies formed a similar zonal gradient in the WNP and contributed similarly to GPI, their magnitudes were small relative to background variability (Fig. 5-6b).

With further analysis of the GCM ensemble, we determined that the zonal dipole in anomalous vertical wind shear was generated by a westward shift in the rising arm of the PWC during the LIA (Fig. 5-6c,d). In the Last Millennium experiment mean, modeled zonal winds converge at the surface and diverge at height within 130-180°E (Fig. 5-6c). At either end of the basin, mean zonal winds at the surface and top of the troposphere are opposed, generating high vertical wind shear. During the LIA and across the basin, zonal winds became more westerly at the surface and easterly at the top of the troposphere. These anomalies weakened shear in the west where they opposed mean wind direction and strengthened shear in the east, corresponding to modeled vertical wind shear anomalies during the LIA (Fig. 5-6a). These modeled wind anomalies would shift the rising limb of PWC westward, which is consistent with the hypothesized mechanism of Yan et al. (2011) for a zonal concentration of precipitation in the IPWP.

Figure 5-6: Relationship between vertical wind shear anomalies and the Pacific Walker Circulation. a) Vertical wind shear anomaly (red line) and b) potential intensity anomaly (red line) during the LIA relative to 1000-1850 CE. c) Mean zonal wind (1000-1850 CE) (arrows, shading) and vertical pressure velocity (arrows). d) LIA anomaly of the zonal wind (arrows, shading) and vertical pressure velocity (arrows). Black lines and gray shading in a,b) encompass the bootstrapped 95% confidence interval. Stippling in d) indicates agreement on zonal wind anomaly direction by at least five of seven models. All quantities are ensemble medians and meridional averages over 0°N-20°N.

Figure 5-7: Ensemble median correlation between annual mean vertical wind shear and the Pacific Meridional Mode wind index (PMMU) in GCM ensemble results. Stippling indicates agreement on the direction of the correlation by at least five of seven models. Model-specific correlation maps can be found in Fig. 5-12.

5.7 Influence of climate modes

The zonal wind and vertical wind shear anomalies during the LIA are associated with centennial variability in PMM. In the instrumental record, PMM is a climate mode second only to ENSO in accounting for interannual variability of east tropical Pacific SST and wind anomalies (Chiang and Vimont, 2004). Its positive (negative) phase is characterized by a strengthened (weakened) meridional SST gradient across the mean latitude of the ITCZ in the eastern Pacific, a northward (southward) shift in the eastern Pacific ITCZ, and weaker (stronger) trade winds in the eastern North Pacific. Recent observational analysis and atmospheric GCM experiments have demonstrated that PMM influences mean cyclogenesis in the WNP, with positive (negative) PMM generating anomalous vertical wind shear that causes an eastward (westward) shift in cyclogenesis (Gao et al., 2018; Hong et al., 2018). In the instrumental record positive PMM events often precede and may initiate El Nino events (Chang et al., 2007), and the eastward shift in cyclogenesis previously associated with historical El Nino events may instead be a consequence of co-occurring positive PMM (Hong et al., 2018). In our ensemble the vertical wind shear anomaly during the LIA is highly correlated (Fig. 5-7) with a negative mean PMM during the LIA, in contrast to a positive mean PMM during the MCA (Fig. 5-4f), consistent with those previously published model experiments.

Proxy reconstructions of TC activity reveal that the transition from the end of the

MCA through the early LIA was characterized by anomalously high TC activity in the deep tropical WNP. Although ENSO variability was lower over the last millennium relative to the instrumental period, transient increases in ENSO variability when centennial mean climate conditions promoted cyclogenesis and intensification near reconstruction sites could have generated this anomaly. The combination of periodic fluctuations in ENSO variability with secular changes to centennial mean climate conditions could explain spatial variability in TC activity across the west Pacific, whereas ascription to just one of these factors appears insufficient.

Although the GPI anomalies in our model ensemble replicate the spatial variability of reconstructed anomalous TC activity across the western Pacific during the LIA, their magnitude is small relative to the anomalies in individual reconstructions. Some of this discrepancy is likely explained by lower decadal-centennial variability in the results of the CMIP5 last millennium experiment relative to that expected based on paleoclimate proxy records (Consortium, 2017; Ault et al., 2013; Zanchettin et al., 2013). For example, our model ensemble does produce anomalous patterns in meridional circulation during the LIA consistent with a contraction of the annual range of the ITCZ and a southward shift during the northern hemisphere storm season (Fig. 5-8). However, the magnitude of the ITCZ-related wind anomalies is only about 1% of the last millennium mean, which falls short of multi-degree latitude shifts in ITCZ width and position interpreted from proxy records (Denniston et al., 2016; Sachs et al., 2009). A recent modeling experiment demonstrated that even a one-degree southward (northward) shift in the ITCZ following a strong volcanic eruption in the tropical northern (southern) hemisphere can increase GPI in the deep tropical North (South) Pacific and decrease it poleward (Pausata and Camargo, 2019). Thus, the model ensemble may have failed to capture additional cyclogenesis variability associated with shifts in the ITCZ during the LIA.

Figure 5-8: Zonal mean vertical pressure velocity associated with meridional overturning circulation (shading, vectors) and non-divergent meridional wind velocity (vectors) a,b) averaged over 1000-1850 CE and c,d) the LIA (1400-1700 CE) anomaly relative to 1000-1850 CE. The dashed vertical lines indicate the equator. Negative (positive) vertical pressure velocity values indicate ascending (descending) motion. Black stippling in c,d) indicates pressure/latitude coordinates where at least 5 of the 7 models agreed on the direction of change.

5.8 Acknowledgments

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5.A Appendix: Additional tables/figures

Table 5.1: Sediment core radiocarbon dates

Core/ Drive	Sample Name	Accession #	Sample contents	Drive depth (cm)	¹⁴ C age (yr)	¹⁴ C error (yr)
LTD2/D2	LTD2_D2_1:1_22-23cm	OS-133794	Plant/wood	23	-5,645	0.25
LTD2/D2	LTD2_D2_1:1_105.5cm	OS-129304	Plant/wood	106	125	15
LTD2/D1	LTD2_D1_1:3_72cm [†]	OS-128853	Coral	72	550	20
LTD2/D1	LTD2_D1_2:3_64cm [†]	OS-128854	Coral	167	545	15
LTD2/D1	LTD2_D1_3:3_35-36cm	OS-127189	Plant/wood	259	350	15
LTD2/D1	LTD2_D1_3:3_84.5	OS-127190	Plant/wood	308	330	15
LTD3/D1	LTD3_D1_1:5_45*	OS-127347	Plant/wood	45	-465	16
LTD3/D1	LTD3_D1_1:5_48-49cm	OS-133795	Plant/wood	49	110	20
LTD3/D1	LTD3_D1_1:5_62cm [†]	OS-128851	Mollusc	62	515	15
LTD3/D1	LTD3_D1_2of5_25-26cm [†]	OS-151049	Foraminifera	126.5	735	15
LTD3/D1	LTD3_D1_2:5_52	OS-127191	Plant/wood	152.5	560	15
LTD3/D1	LTD3_D1_2:5_59.5cm [†]	OS-129133	Halimeda	160	660	15
LTD3/D1	LTD3_D1_2of5_65-66cm [†]	OS-151050	Foraminifera	166.5	1040	15
LTD3/D1	LTD3-D1-3of5-3.5cm ^{†*}	OS-139903	Mollusc	223	740	160
LTD3/D1	LTD3_D1_3of5_25-26cm [†]	OS-151051	Mollusc	245	1,670	25
LTD3/D1	LTD3-D1-3of5-29cm [†]	OS-139904	Mollusc	248.5	1,460	160
LTD3/D1	LTD3_D1_3of5_52-53cm [†]	OS-151052	Foraminifera	272	1,930	25
LTD3/D1	LTD3-D1-3of5-113.5cm [†]	OS-139901	Mollusc	333	1,980	160
LTD3/D1	LTD3_D1_4of5_20-21cm [†]	OS-151053	Foraminifera	359.5	2,330	25
LTD3/D1	LTD3_D1_4of5_36-37cm	OS-149473	Plant/wood	375.5	2,110	20
LTD3/D1	LTD3_D1_4_41 [‡]	OS-125137	Mollusc	379.5	2,260	80
LTD3/D1	LTD3_D1_4of5_60-61cm [†]	OS-151054	Foraminifera	399.5	2,720	25
LTD3/D1	LTD3_D1_4_75p5 [‡]	OS-125138	Mollusc	414	2,350	80
LTD3/D1	LTD3_D1_4of5_99-100cm [†]	OS-151055	Foraminifera	438.5	2,700	30
LTD3/D1	LTD3_D1_5of5_12-13cm [†]	OS-151056	Foraminifera	481.5	2,890	20
LTD3/D1	LTD3_D1_5of5_36cm*	OS-141714	Plant/wood	496	390	20
LTD3/D1	LTD3_D1_5_38 [‡]	OS-125134	Mollusc	498.5	2,890	90
LTD3/D1	LTD3_D1_5of5_46.5cm*	OS-149409	Plant/wood	507	60	100
LTD3/D1	LTD3_D1_5of5_52-53cm [†]	OS-151057	Foraminifera	513	2,930	20
LTD3/D1	LTD3_D1_5_53p5 [‡]	OS-125136	Mollusc	514	2,710	90
LTD3/D1	LTD3_D1_5_75p5 [‡]	OS-125135	Mollusc	536	3,210	85

[†]Indicates inorganic carbonate sample processed with hydrolysis and conventional AMS.

[‡]Indicates inorganic carbonate sample processed with gas ion source.

All samples were processed at NOSAMS.

*Indicates outliers excluded from the age model (Sec. 5.3.1).

Table 5.2: Summary of CMIP5 models used in last millennium analysis

Institute	Model	Country	LM* Years CE	Atmospheric resolution	
				Latitude (°)	Longitude (°)
BCC	BCC-CSM1.1	China	850-2000	2.81	2.81
NASA	GISS-E2-R	USA	850-1850	2.00	2.50
IPSL	IPSL-CM5A-LR	France	850-1850	1.89	3.75
LASG-IAP	FGOALS-g1	China	1000-2000	4.00	5.00
LASG-IAP	FGOALS-s2	China	850-1850	1.71	2.81
MPI-M	MPI-ESM-P	Germany	850-1850	1.84	1.84
NCAR	CCSM4	USA	850-1850	0.94	1.25

*LM = Last Millennium experiment

Figure 5-9: Illustration of the method used to calculate centennial event frequency from coarse fraction. c) Coarse fraction anomaly is used to identify event deposits as in Methods. b) The annually-binned probability distribution function (PDF, blue shading) of each event deposit is extracted from the age model and summed for each year (black line). a) The sum of annual PDFs is summed over a 100-year moving window to construct a time series of centennial event frequency incorporating age model uncertainty.

Figure 5-10: Sensitivity analysis of the procedure used to identify event beds in Jaluit Atoll core, LTD3 coarse fraction data, using the 250-2000 μm coarse fraction. Coarse fraction variance over the entire core was calculated a) with a moving-average window-size of 11 cm and exclusion of outliers, b) with a moving average window size of 11 cm and inclusion of outliers, c) with a moving average window size of 31 cm and exclusion of outliers, and d) with a moving average window size of 31 cm and inclusion of outliers. For each of these four cases, event beds were flagged with 1.5 standard deviation and 2 standard deviation cutoffs. The active (passive) interval thresholds for each of these cutoffs represents the 97.5 (2.5) percentile frequency for a Poisson distribution with the core's mean frequency. Active (passive) intervals were identified as intervals lasting at least a century in which those thresholds were exceeded (not exceeded). In all four cases, active and passive intervals displayed were identified using the 1.5 standard deviation cutoff.

Figure 5-11: Same as Fig. 5-10, but using the $>250 \mu\text{m}$ coarse fraction.

Figure 5-12: Spearman rank correlation between mean storm season vertical wind shear and Pacific Meridional Mode for each of the CMIP5 models. Correlation coefficients were calculated for Last Millennium experiment results for the period 1000-1850 CE. Black stippling indicates statistical significance as determined by a two-tailed Student t-test after taking into account multiple hypothesis testing using the false discovery rate procedure and setting $q = 2.5\%$ (Hu et al., 2017).

5.B Appendix: GCM Verification

5.B.1 Introduction

In this chapter, I reconstructed tropical cyclone genesis frequency in the southeastern corner of the western North Pacific (WNP). I then used an ensemble of CMIP5 models to explore changes in climate that could have produced the most salient feature in the storm record, a peak in cyclogenesis during the beginning of the Little Ice Age (LIA). However, in the main body of the chapter, I did not establish that the climate models produce realistic variability in the variables of interest. Although there was strong inter-model agreement on the direction of changes and relationships between genesis variables and broader climate parameters, the models could be consistently biased. In this appendix I present a comparison between CMIP5 model results and NCEP reanalysis data to illustrate qualitatively the ability of the former to replicate variability in the latter. This comparison was performed before any of the analysis in the main body of the chapter.

Ideally, I would build confidence in model results by comparing them to observational/reanalysis data under similar assumed climate forcing and over the same timescales as in the ensemble analysis. However, in this chapter I examined results from the Last Millennium experiment exclusively, precluding direct comparison of model results with gridded observational or reanalysis data under the same forcing regime. Additionally, the relatively short coverage of observational and reanalysis data precluded analysis of centennial variability. Thus, I examined the distributions of variable values and seasonal-to-inter-annual variability from the Historical experimental runs for each of the CMIP5 models and compared them to the National Centers for Environmental Prediction (NCEP) reanalysis. I also compared variability between Last Millennium and Historical experiment model data to establish model consistency between experiments.

In this analysis, I examine two primary sources of variability in tropical cyclone genesis in the WNP: seasonal variability in genesis frequency, and inter-annual variability in genesis location induced by the El Nino Southern Oscillation (ENSO).

Specifically, I a.) describe the literature supporting the use of GPI and ENSO variability as diagnostics in CMIP5 models, b.) compare values of GPI and its component indices between models and reanalysis, c.) compare modeled seasonality of GPI and its components with the reanalysis, d.) determine if the models replicate reanalysis patterns of GPI anomalies induced during El Nino events relative to La Nina events, e.) compare ENSO-related variability between models and reanalysis, and f.) determine if contrasts between Historical Last Millennium experiment results capture an increase in ENSO variability inferred from coral records. I find that most of the models replicate the static distribution and seasonality of reanalysis GPI values well, replicate ENSO variability and event frequency moderately well. Most of the models capture an increase in ENSO variability from the last millennium to the past century. Some models capture ENSO-induced spatial anomalies in GPI qualitatively similar to that in the reanalysis data. None of the models capture the increase in ENSO variability dated to 400 cal. yr BP in coral proxy data.

5.B.2 Suitability of GPI and CMIP5 models

Seasonality of GPI calculated with monthly NCEP reanalysis data is strongly correlated with seasonality in NOAA National Hurricane Center (NHC) best track TC event frequency, both in separate hemispheres and separate ocean basins (Camargo et al., 2007b). Additionally, GPI is sensitive to ENSO and detects the southeastward shift in TC genesis location during El Nino events (Camargo et al., 2007b). GPI has been used to track trends in TC frequency and magnitude in GCMs, which project increases in GPI and its constituents throughout the twenty-first century, interpreted as an increase in cyclone frequency and magnitude (Emanuel, 2013; Camargo, 2013; Emanuel et al., 2008).

Previous studies have shown that CMIP5 models, especially those with lower resolution, tend to underestimate global TC frequency (Camargo, 2013). Additionally, although they model ENSO moderately better than CMIP3 models, CMIP5 models tend to overestimate the western extent of SST anomalies during Cold Tongue El Nino (Capotondi et al., 2015) and poorly replicate central Pacific precipitation anomalies

(Bellenger et al., 2014). However, CMIP5 models do replicate ENSO frequency, irregularity, and magnitude asymmetry well, especially compared with CMIP3 (Bellenger et al., 2014; Capotondi et al., 2015), and thus should be able to replicate ENSO variability over the last millennium. Despite their weaknesses, the past performance of CMIP5 models and GPI indicate they may be suitable for examining relative changes in ENSO and tropical cyclone activity.

5.B.3 Methods

Models and reanalysis data

The GCM data analyzed in this study were produced from two experiments, CMIP5 Last Millennium (LM) and CMIP5 Historical. In the LM experiment, GCMs were forced with a standardized set of orbital, greenhouse gas, solar irradiance, aerosol, and ozone forcings, and standardized land use/change, ice sheet, and topographical boundary conditions (Schmidt et al., 2011, 2012). These conditions replicated 850-1850 CE climate forcing to the best of contemporaneous scientific understanding. Eight CMIP5 models were used in the LM experiment (Table 5.3): the Beijing Climate Center Climate System Model 1.1 (BCC-CSM1.1), the NASA Goddard Institute for Space Studies ModelE/Russel (GISS-E2-R), the Institut Pierre Simon Laplace Coupled Model 5 (IPSL-CM5A-LR), the LASG Institute of Atmospheric Physics Flexible Global Ocean-Atmosphere-Land-Sea ice (FGOALS-gl, FGOALS-s2), the Model for Interdisciplinary Research on Climate Earth System Model (MIROC-ESM), the Max Planck Institute for Meteorology Earth System Model Paleo (MPI-ESM-P), and the National Center for Atmospheric Research Community Climate System Model (CCSM4). This analysis was performed before that detailed in the main chapter.

In the CMIP5 Historical experiment, the same models (except those produced by FGOALS-s2) were initialized with a preindustrial control run and forced with evolving boundary conditions reflecting historical measurements and proxies. Changing vegetation was ignored and boundary condition values were not standardized between models. Model resolution and physics were unchanged from the LM experiment. The

Table 5.3: Summary of LM model and reanalysis characteristics

Institute	CMIP5 Model/ Reanalysis	Country	Years CE	Atmospheric resolution		
				Lat (°)	Lon (°)	Vertical (# levels)
BCC	BCC-CSM1.1	China	850-2000	2.81	2.81	26
NASA-GISS	GISS-E2-R	USA	850-1850	2.00	2.50	40
IPSL	IPSL-CM5A-LR	France	850-1850	1.89	3.75	39
LASG-IAP	FGOALS-gl	China	1000-2000	4.00	5.00	26
LASG-IAP	FGOALS-s2	China	850-1850	1.71	2.81	26
JAMSTEC	MIROC-ESM	Japan	850-1850	2.81	2.81	80
MPI-M	MPI-ESM-P	Germany	850-1850	1.84	1.84	47
NCAR	CCSM4	USA	850-1850	0.94	1.25	26
NCEP	Reanalysis	USA	1948-2015	1.91	1.88	17
NOAA	ERSST	USA	1854-2015	2.00	2.00	1

Historical experiment covered the years 1850-2000 CE.

For comparison to CMIP5 output, I retrieved atmospheric data from the National Centers for Environmental Prediction (NCEP) reanalysis version 2 (Kanamitsu et al., 2002), which assimilates climate data over a grid similar to those used in the GCM experiments (Table 5.3) for the years 1979-2015 CE. For sea surface temperature data, I used the National Oceanographic and Atmospheric Administration (NOAA) Extended Reconstructed Sea Surface Temperature v.3b (ERSSTv3b) dataset (Smith et al., 2008).

To mitigate the effect of variable model resolution on index values, I interpolated all GCM and reanalysis outputs onto a $2^\circ \times 2^\circ$ grid before calculating the indices. Storm indices were calculated per grid cell of the western north Pacific (WNP) region and for time-series analyses were then averaged over that region (Table 5.4). This region is smaller than that examined in the main body of the chapter, so that this gut-check analysis would be consistent with previous research (Yan et al., 2015b, 2016). SST was calculated per grid cell and then averaged over the Nino 3.4 index region. The Nino 3.4 index was compared with coral proxies of ENSO variability taken from Palmyra Atoll and the northern Line Islands (Fig. 5-13), as the Nino 3.4 region was assumed to express ENSO variability more reliably amongst all models than the smaller, off-equator Palmyra region.

Table 5.4: Geographic extent of regions of interest.

Region	Latitude	Longitude
WNP	4°N-30°N	120°E-180°E
Nino 3.4	5°S-5°N	190°E-240°E
Palmyra	2°N-8°N	195°E-205°E
SVD	13°S-13°N	210°E-270°E

Figure 5-13: Extent of the WNP, Nino 3.4, and Palmyra regions over which storm and SST indices were calculated.

Statistical comparison

I compared the GCM-generated storm index values to those constructed from the NCEP and ERSST datasets using statistical tests and seasonality plots. For the seasonality plots, the monthly means of the NCEP storm indices from 1979-2000 were compared with the monthly storm index means of the GCM Historical output for the same period and the monthly storm index means calculated over the entire time extent of the GCM LM output. One-sample non-parametric sign tests between GCM output and NCEP/ERSST medians were used to statistically test the difference in average values between NCEP and GCM indices, and non-parametric Ansari-Bradley equal dispersion tests were used to statistically test the difference in variance between NCEP and GCM indices. Bonferroni corrections were applied to correct for multiple hypothesis testing. All storm index comparisons were performed over the WNP.

Quantifying ENSO variability

To establish the ability of the CMIP5 models to replicate storm and ENSO variability in general, I calculated GPI and its constituent indices and the Nino 3.4 index for the Historical experiment run of each GCM and compared them to the same indices calculated from NCEP and ERSST reanalysis data. All ENSO variability comparisons were performed using the Nino 3.4 region.

To diagnose ENSO events, I constructed the Oceanic Nino Index (ONI) using ERSST data. This index is defined as the SST anomaly in the Nino 3.4 region smoothed by a 3-month moving average filter. NOAA diagnoses ENSO episodes as periods of time in which five consecutive ONI values exceed 1°C (El Nino) or -1°C (La Nina). However, I cannot assume that the GCM models produce SST values with variance equal to ERSST data, so these absolute thresholds may miss ENSO events in the GCM output. Therefore, I define El Nino (La Nina) events individually for each dataset as five-consecutive-month periods during which the Nino 3.4 index exceeds 1 (-1) standard deviation of the 1950-2000 CE time-series. I then used the diagnosed events to contrast values of GPI between El Nino and La Nina events in all of the models.

ENSO variability in index time-series data was isolated using a nonparametric, one-sided power spectral density (PSD) estimate (Mathworks, 2016). Before estimating the PSD, mean values were subtracted from the data. For inter-model comparison, the PSD was estimated over the complete time series of each experiment. For comparison to paleo-coral records, the PSD was estimated for 30 year segments of the index data at a five-year interval to avoid non-stationarity effects, as in (Cobb et al., 2013). The ENSO signal was measured as the standard deviation of the portion of each PSD estimate with a period between two and seven years. For inter-model comparison, the resulting ENSO variability time series for each model were normalized by the ENSO signal averaged over 1968-1998 in the Historical experimental run of that model. This normalization mitigates the impact of model SST variability bias in the ENSO variability estimates, analogous to the intra-site normalization of (Cobb

et al., 2013).

5.B.4 Results

Historical comparison

Seasonality plots of the seven GCMs with data during the 20th century (all models but FGOALS-s2) reveal that the GCMs replicated seasonality of GPI, the saturation deficit, vertical shear, and potential intensity very well, with few major differences between models (Fig. 5-14). However, the absolute values of the modeled vorticity anomalies tended to deviate from the reanalysis data. Only two models, MPI-M and FGOALS-gl, clearly captured the dual peaks in vorticity in the reanalysis data. The other models tended to produce vorticity with a single seasonal maximum and minimum, and with larger range than in the reanalysis. Despite the poor performance on vorticity, the models produced accurate seasonal GPI anomalies with a maximum GPI occurring in September in most of the models and in the reanalysis, with a smooth increase from April and decrease to January.

Figure 5-14: Seasonality plots constructed from monthly anomalies averaged over 1979-2000. Anomalies are relative to the mean values for the entire period. GPI values were scaled by the absolute value of maximum anomaly for each model. The black lines represent NCEP seasonality and are the same for each model. The red lines indicate GCM Historical seasonality and the blue dotted lines indicate GCM LM seasonality.

Figure 5-15: Box plots of the values of GPI components calculated from GCM Historical output, compared to NCEP output. The box borders represent the 25th, 50th, and 75th percentiles, respectively. The whiskers represent minimum and maximum values. All values were averaged over the WNP. FGOALS-s2 is not displayed because results were unavailable for the Historical experiment for that model.

There is surprising fidelity in the seasonality patterns between Historical and LM GCM runs (Fig. 5-14). The GCM LM runs produced nearly identical seasonality cycles as the Historical runs despite being calculated over a time scale an order of magnitude longer. This fidelity in seasonal pattern implies that the seasonality of storm genesis in the WNP is a product of internal model mechanisms rather than external forcing.

A comparison of the distributions of storm indices between models shows less fidelity between models and observations than in the seasonality plots (Fig. 5-15). Although the variability in values is relatively constant across models for vorticity

Table 5.5: Comparison of storm index median values between models

Dataset	PI (ms^{-1})		χ		η ($\times 10^{-5} \text{ s}^{-1}$)		V_{shear} (ms^{-1})		Nino 3.4 ($^{\circ}\text{C}$)	
	$\tilde{x}$	p	$\tilde{x}$	p	$\tilde{x}$	p^a	$\tilde{x}$	p	$\tilde{x}$	p
NCEP	61.8		0.34		3.98		12.5		27.3	
FGOALS-gl	69.4	3.2E-19	0.45	4.4E-07	3.97	7.0E-01	13.8	1.0E-01	27.6	3.7E-03
NCAR	65.8	3.1E-05	0.33	6.1E-01	4.04	3.7E-03	12.97	8.0E-01	27.2	7.0E-01
MPI-M	65.7	1.3E-03	0.35	4.1E-01	4.03	1.9E-03	12.2	6.6E-01	25.6	4.1E-42
BCC	63.5	5.5E-03	0.39	3.7E-03	4.01	7.0E-01	14.2	3.2E-02	26.9	6.2E-13
NASA-GISS	60.2	1.3E-01	0.42	1.0E-03	4.05	2.4E-03	14.6	1.0E-02	28.6	7.7E-36
IPSL	58.1	1.0E-01	0.43	1.6E-03	3.86	5.5E-06	15.4	3.7E-03	26.1	7.7E-36
MIROC	61.8	9.0E-01	0.48	2.3E-16	3.78	3.5E-31	14.8	6.5E-04	25.5	1.3E-44

^ap-value for a non-parametric Sign test for equal medians between GCM and NCEP data.

p-values significant at a Bonferroni-corrected $\alpha = 0.01/70 = 1.4 \times 10^{-4}$ are in bold

and indicate significantly different medians.

Table 5.6: Comparison of storm index variance values between models

Dataset	PI (ms^{-1})		χ		η ($\times 10^{-12} \text{ s}^{-1}$)		V_{shear} (ms^{-1})		Nino 3.4 ($^{\circ}\text{C}$)	
	σ^2	p	σ^2	p	σ^2	p^a	σ^2	p	σ^2	p
NCEP	63.4		0.02		3.2		35.3		0.84	
FGOALS-gl	50.5	2.7E-02	0.03	1.7E-07	2.9	2.2E-01	58.9	3.1E-09	1.2	9.1E-03
NCAR	73.5	8.3E-01	0.02	3.6E-05	6.3	1.6E-06	38.2	4.1E-01	1.39	2.9E-04
MPI-M	63.5	9.2E-01	0.03	3.4E-07	3.0	2.2E-01	47.9	1.1E-01	0.93	2.6E-01
BCC	36.0	1.8E-11	0.03	9.4E-17	18.1	5.8E-27	48.6	3.2E-05	0.50	3.6E-05
NASA-GISS	70.9	9.6E-02	0.02	8.4E-03	3.3	2.6E-01	55.2	6.2E-08	0.52	9.3E-04
IPSL	138.1	1.16E-14	0.03	3.3E-06	5.7	3.4E-07	56.9	4.0E-11	0.9	3.1E-01
MIROC	63.8	8.2E-01	0.02	9.8E-03	4.2	3.1E-03	31.0	5.1E-02	1.2	5.5E-06

^ap-value for a non-parametric Ansari-Bradley equal dispersion test between GCM and NCEP data.

p-values significant at a Bonferroni-corrected $\alpha = 0.01/70 = 1.4 \times 10^{-4}$ are in bold

and indicate significantly different variance.

(η) and PI, there are large inter-model differences in the distribution of vertical shear and especially χ values. The models produce storm indices with median values that are statistically similar to reanalysis data (Table 5.5), but tend to have statistically significant error in Nino3.4 index values. IPSL tends to produce index distributions with significantly greater variance than in reanalysis data, especially for PI (Table 5.6). BCC also produces significantly different variance for many of the indices, but without a consistent bias.

Spatial distribution of the ENSO signal

Empirically, ENSO events in the WNP are characterized by a southeastward shift in tropical cyclone genesis location (Wang and Chan, 2002). That shift is visible in the NCEP reanalysis GPI output (Fig. 5-16), where plotting the El Nino - La Nina anomaly reveals a tongue of increased GPI extending from the southeastward extent of the WNP region towards the center and northern end, with strong negative anomalies in the southwest over the northern Philippines and the southern half of the South China Sea. This pattern is poorly replicated in the GCM Historical runs, except possibly by the NCAR model. Although individual GCM historical runs do not reproduce the ENSO anomalies, the Ensemble Mean anomaly does show positive GPI anomalies in the southeast and negative anomalies over the northern Philippines and in the South China Sea.

Three of the models, FGOALS-gl, NCAR, and NASA-GISS, produce ENSO events with similar frequency to those in the NCEP reanalysis. The remaining models underestimate the event frequency. The ability to correctly replicate ENSO event frequency does not appear to correlate well with ability to replicate spatial variability.

Figure 5-16: Anomaly plots revealing the difference in GPI values between El Nino and La Nina events in Historical model runs (El Nino minus La Nina), normalized by the standard deviation of these anomalies. The f in each plot is the frequency of ENSO events in units of ENSO event years per year of dataset considered. The NCEP frequency of 0.16 corresponds to one ENSO event every 6.25 years. In the ensemble mean plot, displayed are only grid cells in which at least five of seven GCMs (excluding FGOALS-s2) agreed on the direction of change. Dots indicate grid cells in which at least six of seven GCMs agreed on direction of change.

The pattern of ENSO related anomalies present in the GCM LM runs closely corresponds to patterns present in the Historical runs (Fig. 5-17). However, FGOALS-s2, MPI-M, and NCAR LM runs produce patterns very similar to the NCEP reanalysis ENSO anomaly (Fig. 5-16). Additionally, the Ensemble Mean ENSO-associated GPI anomaly from the LM runs is nearly identical to that of the Historical runs. The mean El Nino event frequency appears to be lower in the LM runs than in the Historical runs, although FGOALS-gl and NCAR produce an El Nino frequency close to that of the reanalysis. Although the GCMs inconsistently reproduce the spatial GPI anomaly associated with ENSO in the reanalysis, the ensemble mean anomaly does reproduce the reanalysis features that could explain the empirical southeastward shift in mean genesis location. The ensemble mean and reanalysis anomalies both show enhanced genesis potential in the southeastward corner of the WNP basin and suppressed genesis potential to the west.

Figure 5-17: Anomaly plots revealing the difference in GPI values between El Niño and La Niña events in LM model runs, normalized by the standard deviation of these anomalies. The f in each plot is the frequency of ENSO events in units of ENSO event years per year of dataset considered. NCAR was not included due to corrupt file issues. In the ensemble mean plot, displayed are only grid cells in which at least five of seven GCMs (excluding NCAR) agreed on the direction of change. Dots in the ensemble plot indicate grid cells in which at least six of seven GCMs agreed on direction of change.

ENSO frequency and variability

ENSO variability is defined over the historic period as variance in eastern equatorial Pacific SSTs that falls within the 2-7 yr period band. When looking at proxies with resolution high enough to estimate variability at these frequencies, modern ENSO variability is significantly higher than throughout the rest of the Holocene, but the timing of this change is ambiguous (Cobb et al., 2003, 2013; Sadekov et al., 2013). For some of the models and in the multi-model ensemble mean, the Historical runs do appear to have on average higher ENSO variability than the Last Millennium runs (Fig. 5-18), but the difference is much smaller than that in coral proxies. Relative to the coral records every GCM overestimated ENSO variability before 1850, but the ranges in ENSO variability estimates are consistent between GCMs and the coral record. None of the models replicate the increase in ENSO variability c. 400 cal. yr BP suggested in the proxy data, though this shift has not been tested for statistical significance. The low magnitude of change in variability from the LM to the Historical runs again illustrates the lack of expected variability in the LM model runs (Ault et al., 2013; Landrum et al., 2013; Zanchettin et al., 2013).

The characteristic frequencies of Nino 3.4 SST variability in GCM Historical and LM runs appear to agree, as LM power spectral density (PSD) is nearly identical to Historical PSD in each model (Fig. 5-19). The periodograms of ENSO variability also explain differences in ENSO event frequency between GCMs and NCEP/ERSST data. GCMs with higher event frequency than ERSST data tend to have more or higher peaks in power at higher frequencies (lower periods) than the NCEP data (e.g. BCC and FGOALS-gl in Fig. 5-19). Those GCMs with lower ENSO event frequency tend to have redder periodograms, with gradually increasing power from ENSO frequencies into lower frequency portions of the spectrum (e.g. MIROC and MPI-M in Fig. 5-19). The periodograms from MIROC show a startling red bias likely caused by long-term drift present in model results (Sueyoshi et al., 2013).

Figure 5-18: ENSO variability over the past 1000 years estimated as the standard deviation of SST in the Nino 3.4 region at at 2-7 year frequencies. The variability values have been normalized by average ENSO variability for each model between 1968-1998 and thus represent percent change from that period in time. ENSO variability was calculated this way in 30-year windows at at 5-year intervals to be consistent with the methodology of (Cobb et al., 2013) and to avoid non-stationarity. The blue lines indicate LM run results. The orange lines represent Historical run results. Coral mean data display the results of the same analysis applied to monthly-resolution coral $\delta^{18}\text{O}$ records from (Cobb et al., 2003), taken from the Northern Line Islands, near Palmyra (Fig. 5-13). The blue triangles represent the means for the individual coral records (red circles).

Figure 5-19: Periodograms of SST variability in the Nino3.4 regions for GCM LM runs (blue lines), and the Historical period (orange lines). Note the difference in y-axis scales between plots. GCM LM power spectral density curves were binned for easier interpretation, resulting in some smoothing at higher frequencies.

5.B.5 Discussion

This exercise aimed to test the ability of CMIP5 models to replicate changes in WNP storm characteristics over the past 1000 years by comparing seasonal and inter-annual variability in GCM output and reanalysis data. In particular, I examined the ability of the models to replicate a.) seasonal variability in spatially averaged storm genesis indexes in the WNP, b.) interannual variability in the spatial distribution of storm genesis related to ENSO, and c.) the frequency and magnitude of ENSO events, diagnosed using an SST index.

All CMIP5 models examined replicated seasonal variability in the components of GPI well. Additionally, seasonality in LM runs was nearly identical to that in Historical runs for GPI and its components. However, the models were less able to replicate seasonality in basin-wide GPI, producing a peak 1-5 months before the GPI peak in the NCEP reanalysis data. The MIROC model performed most poorly, producing a GPI minima during the NCEP maxima. Based on these results, the models adequately capture seasonality of environmental conditions relevant for tropical cyclone genesis. When examining temporal variability in GPI, however, one should examine annual averages or averages over long time periods encompassing the GPI peak in all models (i.e. July-November) to avoid propagating model error in the timing of GPI.

The CMIP5 models produced spatial GPI anomalies for El Nino minus La Nina that were broadly consistent with anomalies in the reanalysis data. However, three of the models produced anomalies with much smaller magnitude in the Historical model runs, and one model, MPI-M produced Historical run anomalies that contradicted the reanalysis anomalies. However, all of the models produced spatial anomalies in the LM runs that were very similar to the historical reanalysis anomalies, with far greater inter-model agreement. Thus, the poor results from the Historical model runs are likely due to averaging over a small number of ENSO events, and not necessarily due to model inadequacy.

The GCM models did produce higher ENSO variability over the last 150 years relative to the previous 1000, though this difference was only revealed by comparing

results from two separate experiments. Without analyzing further the differences in forcings between the Historical and LM runs of each model, it is impossible to diagnose the increase in ENSO variability within the past two centuries. Two models, FGOALS-gl and BCC, had LM model runs extend to the 21st century, but neither of these model runs reveal a significant shift in ENSO variability. If these two models are representative of the larger set, this would indicate that the changes in ENSO variability between LM and Historical model runs are due either to forcing with significant mean difference between the two runs or to industrial era forcing trends not present in the LM run, such as the increase in anthropogenic greenhouse gases. Attribution to either of these mechanisms would contrast with the attribution to precession-driven changes in boreal summer warming of an inferred greater magnitude shift in ENSO variability from the mid-Holocene to present (Clement et al., 2000; Donders et al., 2008).

5.B.6 Conclusions

The eight CMIP5 GCMs examined in this study reproduce several variables crucial for measuring storm genesis in the WNP fairly well. All of the models produced vorticity, vertical shear and potential intensity with the same seasonality as found in the NCEP reanalysis. Additionally, they produced those three variables with very similar variance to observations, and with medians of the correct magnitude. However, the models tended to overestimate variability in the saturation deficit, contributing to incorrectly timed GPI maxima during boreal summer and fall. Additionally, over half of the models tended to overestimate the frequency of El Nino events. Despite these failings, nearly all of the models produced ENSO variability with similar ranges and with a modern increase seen in coral proxy records. Last Millennium model runs demonstrated spatial ENSO-linked GPI anomalies very similar to those in reanalysis data. I conclude that these LM runs for these models replicate seasonal and inter-annual variability in storm genesis indexes consistently, supporting their use for examining temporal variability in paleoclimate proxies over the past millennium. Additionally, the models do replicate a centennial shift in ENSO variability between

LM and Historical model runs, though the relatively low magnitude of this modeled shift prompts caution when examining modeled centennial variability.

Chapter 6

Conclusions

6.1 Thesis Summary

Atoll reef islands and their inhabitants are considered particularly vulnerable to sea-level rise (SLR) and the intensification of tropical cyclones expected from global climate change (Shope et al., 2017). Because the islands consist of unconsolidated sediment and are thought to have formed mainly after sea level fell in the Pacific, sea-level rise associated with global climate change could destabilize atoll reef islands, eventually leading to their total erosion from atoll platforms (Dickinson, 2009). More intense tropical cyclones also pose an existential risk to atoll reef islands, as tropical cyclones have eroded atoll reef islands completely from the atoll rim in a matter of hours (Ford and Kench, 2014). However, many reef islands have expanded during modern tropical cyclone climatology and sea-level rise, suggesting future resilience to increases in these hazards (Kench et al., 2018a). Unfortunately, our understanding of the processes governing the formation and evolution of reef islands and the coral reefs on which they depend is limited, hampering our ability to predict atoll reef island response to climate change.

This thesis applied stratigraphic analysis, numerical and physical modeling, and climate model analysis to the questions:

- What role does sediment play in coral reef geomorphological evolution?

- How does sediment generated on fringing coral reefs make its way to the back reef and reef island beaches?
- How has tropical cyclone climatology over the past thousand years influenced the stratigraphic record?
- How will global climate change affect the answers to the above questions?

Using data from a cross-shore array of wave sensors, I calibrated a hydrodynamic model of wave propagation across a fringing reef on Kwajalein Atoll in Chapter 2. Using the model I demonstrated that sediment makes its way from the fore reef and outer reef flat to the beach due to a gradient in the magnitude of onshore-directed shear stress. I determined that wave climate change is likely to have negligible effects on the transport of sediment from atoll reefs to atoll reef islands beaches relative to sea-level rise. Sea-level rise will increase the energy and skewness of high-frequency waves on the reef flat, potentially increasing the rate of sediment flux from the outer reef flat to the beach. However, sea-level rise will also decrease bottom shear stress on the shallow fore reef, potentially decreasing sediment flux from the fore reef to the beach. Determining the net effect of these changes on atoll reef island volume is stymied by absence of data on reef sediment production rates and absolute sediment transport rates.

Our understanding of sediment fluxes on atoll fringing reefs is hindered by the sparse distribution of sediment in these environments and the likely episodic nature of its transport. To overcome this limitation and examine transport pathways on a coral atoll reef flat, I examined sediment composed of the foraminifera tests (skeletons) in Chapter 3. Using the degree to which these tests had been weathered as a proxy for transport duration, I demonstrated that the transport pathways of foraminifera produced on the outer reef flat and in the lagoonal back reef on Jaluit Atoll converged on a mid-flat blue hole. Despite acting as a clear indicator of transport direction and relative duration, I was unable to apply this proxy to distinguish sediment deposited by storm waves from sediment deposited under ambient conditions.

Reef sediment dynamics are important not only because they drive reef island

morphological change, but also because they drive the evolution of coral reefs themselves. One important process generating topographic variability on coral reefs is wave-driven abrasion. Despite the near ubiquity of this process on coral reefs globally, its dependence on wave characteristics and sediment availability hasn't been quantified. In Chapter 4 I designed, constructed, and used an oscillatory flow tunnel to examine how abrasion rate scales with oscillation velocity and sediment load. I found that abrasion rate increases exponentially with oscillation velocity (and thus increases exponentially with wave height and decreases rapidly with increasing wave period and depth) and exhibits a tools-and-cover effect with sediment load. Additionally, bedforms that form under oscillatory flow limited armoring of the bed at higher sediment loads, indicating that the cover effect under waves is less efficient than under unidirectional flow. The abrasion rates I generated were quite high, underscoring the importance of sediment removal and sequestration on coral reefs for sediment morphology and growth.

Finally, in Chapter 5 I examined the influence of past climate variability on tropical cyclone activity in the vicinity of the Marshall Islands. I used grain size analysis of a core collected from a blue hole on Jaluit Atoll to develop a reconstruction of tropical cyclone activity in the region over the past three thousand years. The reconstruction exhibited a peak in tropical cyclone frequency during the Little Ice Age. Through analysis of the results of a climate model ensemble experiment, I traced this variability in tropical cyclone climatology to variation in mid-tropospheric wind shear likely driven by an anomalously positive Pacific Meridional Mode during the Medieval Climate Anomaly that flipped to anomalously negative values during the subsequent Little Ice Age. However, climate models still have difficulty replicating Late Holocene variability evident in proxy records, leaving open the possibility that the ensemble did not capture the climate dynamics responsible for the variability in our reconstruction.

6.2 Future directions

6.2.1 Wave-driven geomorphology of reef islands

Since the inception of my research presented in Chapter 2, substantial strides have been made in our understanding of wave dynamics on fringing reefs and sediment transport in coral reef environments generally. For example, numerous studies have investigated the influence of low-frequency waves on high-frequency wave heights and water levels on the reef flat and have found that a) the low-frequency waves are generated through breakpoint forcing and can be treated as a dynamic component of wave setup (Becker et al., 2016), b) low-frequency waves can grow in size while propagating across the reef flat due to standing modes and resonant amplification (Gawehn et al., 2016), and c) low-frequency waves modulate water level and thus high-frequency wave height on the reef flat, and are thus major contributors to island flooding (Cheriton et al., 2016). Field studies have explored sediment transport in a thick coral canopy and across large reef flats (Pomeroy et al., 2017, 2018). Additionally, in 2019 the best flume studies to date confirmed that absent a source of input sediment, atoll reef islands respond to sea-level rise by retreating lagoonward (Tuck et al., 2019a), similar to barrier island retreat on passive margin coastlines (Lorenzo-Trueba and Ashton, 2014).

However, there still exist large gaps in our understanding of sediment transport processes on atoll fringing reefs. For example, there exist no validated algorithms for sediment transport in sediment-poor environments like atoll reef flats, which is why I used conventional algorithms as proxy estimates of transport in Chapter 2. Additionally, the actual response of atoll reef islands to sea-level rise will likely depend a lot on trends in sediment production on the reef flat and fore reef. For example, ocean acidification and rising sea surface temperatures, both projected effects of anthropogenic climate change, enhance bioerosion, and thus sediment production, of coral (DeCarlo et al., 2015). However, sustained acidification and warming could devastate coral reef communities, leading to habitat collapse and an abrupt decrease in sediment production over the long term (Perry et al., 2011). Unfortunately, it is very

difficult to predict the response of coral reefs to stressors over long timescales, much less the extent of those stressors, and thus the response of reef island morphology, which depends on reefal sediment production, may be equally unpredictable.

Further research into reef island response to anthropogenic climate change could focus on establishing algorithms for sediment transport across fringing reefs that could then be integrated into a coupled model of coral health, reef sediment production, sediment transport, hydrodynamics, and reef island morphology.

6.2.2 Wave-driven abrasion

After establishing an empirical power law between wave orbital velocity and abrasion rate in Chapter 4, I briefly outlined the potential significance of the large measured abrasion rates for coral reef environments. Briefly, the magnitude of abrasion rates in my experiments implies that abrasive sediment is either very rare on coral reefs or sequestered such that it cannot abrade the majority of the reef surface. In fact, numerous previous studies have observed that coarse sediment accumulates as a thin layer in many fore reef grooves (Duce et al., 2016; Cloud Jr., 1954; Gischler, 2010). From this line of reasoning, one can infer geomorphic feedbacks that potentially generate grooves and other topographic variability on coral reefs.

First, there is a positive feedback between the generation and deepening of a topographic low in the reef surface and accumulation of sediment. As sediment accumulates in a given location, it intensifies abrasion through the tools effect. Abrasion erodes the reef surface, deepening the location. As a location deepens relative to the surrounding reef surface the topographic gradient becomes more favorable for transport into the location and more adverse for sediment removal, and the location will accumulate more sediment. This positive feedback would enhance the topographic relief of depressions in the reef surface and thus high frequency topographic variability more generally.

Second, there are two potential negative feedbacks between the abrasion rate and deepening of a topographic low. If a topographic low accumulates too much sediment through the positive feedback outlined above, the cover effect could begin to dominate,

reducing the abrasion rate and slowing or stopping further deepening. Additionally, as depth increases, wave orbital velocity at the bed and the abrasion rate decrease. The combination of positive and negative feedbacks implies that there is a potential equilibrium morphology of grooves or other depressions in the reef surface such that abrasive sediment is rapidly moved there from the rest of the reef surface, but not so rapidly that the depressions fill. More specifically, for a given sediment generation rate on the reef, there may be an equilibrium spacing and groove depth such that the flux of sediment flowing into the grooves equals the rate of offshore transport of the sediment through the grooves. I am currently exploring the theoretical influence of these feedbacks in a morphodynamic model of fore reef geomorphic evolution.

In Chapter 4 I quantified variability in the abrasion rate due to sediment load and wave orbital velocity. However, abrasion rate is likely also a function of sediment grain characteristics, such as density and diameter/size. The higher the sediment density, the higher the settling velocity and the kinetic energy of saltating grains. Greater diameter or grain size also implies a faster settling velocity. Thus, sediment with greater density or size than the sediment I used in my experiment would induce higher abrasion rates, assuming it was still mobilized by a given flow. I did not observe a dropoff in the power law relationship at low orbital velocities (Fig. 4-5), which implies that once sediment is mobilized, its abrasion rate does not depend on the mobilization threshold. Thus, I hypothesize that changes in sediment characteristics would mainly affect the constant K in Eq. 4.4, and negligibly change the slope of the relationship in log-log space (i.e. n). The dependence of the abrasion rate on sediment composition and size is worth exploring further, especially because sediment in coral reef environments generally has a lower density than that I used in my experiments.

Finally, previous researchers have argued that the importance of abrasion for the downwearing of rocky shore platforms is limited by the lack of any accumulated sediment on many shore platforms and the localization of accumulated sediment to the landward margins of those shore platforms that do have such deposits (Stephenson, 2014). However, the tools effect in my experiments was apparent as a steep increase in the abrasion rate for very low sediment loads. Thus, if abrasion is an important

process for downwearing where sediment has accumulated as beaches on the shore platform (implying the cover effect is dominant), it may still be important on the more seaward ends of the platform where sediment is transitory and much more sparse (implying the tools effect is dominant). Testing this hypothesis would entail isolating abrasion from other erosive processes on shore platforms with sparse sediment.

6.2.3 Paleotempestology

Through analysis of climate model experimental results in Chapter 5, I identified decreased vertical wind shear in the western North Pacific as a potential driver of enhanced tropical cyclogenesis during the Little Ice Age. I further traced this reduction in vertical wind shear to anomalously negative mean values of the Pacific Meridional Mode (PMM), a climate mode, and a westward shift in the Pacific Walker Circulation (PWC). However, before this causal chain is accepted as the explanation for centennial variability in cyclogenesis over the past millennium, it should be confirmed with paleoclimate proxy evidence, both in the form of precipitation proxies confirming the modeled PMM and PWC behavior and in the form of additional tropical cyclone reconstructions. These additional reconstructions can a) confirm anomalous cyclogenesis during the Little Ice Age; b) confirm the spatial extent of anomalous tropical cyclone activity, and whether it conforms to what would be expected from anomalous PMM and PWC; and c) identify other systematic factors affecting tropical cyclone activity in reconstructions of the last millennium, such as shifts in dominant storm tracks.

As I mentioned in Chapter 5, the climate models whose results I analyzed do not replicate the magnitude of climate variability during the last millennium that we would expect to see given variability in our paleoclimate proxy reconstructions of precipitation and sea surface temperatures in the Pacific. Without clear explanation of the discrepancy, we cannot assume that the lack of variability is uniform throughout the climate system. In particular, it's possible that the models do not replicate pre-historic variability in the extent and mean location of the Intertropical Convergence Zone (ITCZ) and Hadley circulation cells as well as variability in PWC and PMM.

Anthropogenic climate change has widened the tropics over the past few decades, and this expansion has shifted tropical cyclogenesis poleward in the North Atlantic and western North Pacific (Sharmila and Walsh, 2018; Studholme and Gulev, 2018). Paleoclimate proxy evidence indicates the tropics, or at least the seasonal geographic variability of the ITCZ, shrank during the Little Ice Age (Yan et al., 2015b). It's possible that this contraction in the ITCZ/tropics shifted tropical cyclogenesis equatorward, opposite the effect of current tropical expansion. This shift in cyclogenesis could occur due to enhancement of low-level vorticity at low latitudes, as the ITCZ is historically associated with a band of high vorticity (Wu et al., 2012). Establishing the driver of centennial variability in tropical cyclogenesis over the Late Holocene with any certainty requires additional paleoclimate reconstructions to verify or rule out as drivers potential large-scale shifts in climate.

The work presented in this thesis sheds light on three very different processes: cross-shore sediment transport, wave-driven abrasion, and paleoclimatic forcing of tropical cyclogenesis. Additionally, this thesis focuses on the expression of these processes in coral atoll environments, which are difficult to study as a consequence of their remoteness. It is my sincere hope that future research builds on that presented here and that my research inspires curiosity and adoration of these other-worldly landscapes. Where my work has fallen short of my aspirations, I hope it enables the attainment of others'. As always, there is still much work to be done.

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