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Can Tidal Perturbations Associated with Sea Level Variations in the Western Pacific Ocean be used to Understand Future Effects of Tidal Evolution?

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21

Abstract

This study examines connections between mean sea level (MSL) variability and diurnal and semidiurnal tidal constituent variations at 17 open-ocean and 9 continental shelf tide gauges in the western tropical Pacific Ocean, a region showing anomalous rise in MSL over the last 20 years and strong interannual variability. Detrended MSL fluctuations are correlated with detrended tidal amplitude and phase fluctuations, defined as tidal anomaly trends (TATs), to quantify the response of tidal properties to MSL variation. About 20 significant amplitude and phase TATs are found for each of the two strongest tidal constituents, K_1 (diurnal) and M_2 (semidiurnal). Lesser constituents (O_1 and S_2) show trends at nearly half of all gauges. Fluctuations in MSL shift amplitudes and phases; both positive and negative responses occur. Changing overtimes suggest TATs are influenced by changing shallow water friction over the equatorial Western Pacific and the eastern coast of Australia (especially near the Great Barrier Reef). There is a strong connection between semidiurnal TATs at stations around the Solomon Islands and changes in thermocline depth, overtide generation, and the El Niño Southern Oscillation (ENSO). TATs for O_1 , K_1 and M_2 are related to each other in a manner that suggests transfer of energy from M_2 to the two diurnals via resonant triad interactions; these cause major tidal variability on sub-decadal time scales, especially for M_2 . The response of tides to MSL variability is not only spatially complex, it is frequency dependent; therefore, short-term responses may not predict long-term behavior.

1.Introduction

This study examines the connections between mean sea level (MSL) rise and variations in major diurnal and semidiurnal tidal constituent properties at open-ocean and continental shelf tide gauges in the Western Tropical Pacific Ocean, a region showing anomalous MSL rise over the last 20 years (Merrifield, 2011). Additionally, this domain exhibits large excursions of interannual sea level and tidal variability, especially during El Niño and La Niña events. We attempt to answer the question posed in the title, can we use the relations of the observed short-term MSL fluctuations to the short-term tidal variations to understand the future effects of tidal evolution, and to predict changes in total water level (MSL + tides) under higher MSL conditions? MSL rise rates have increased since ca. 1993 at 15 of 26 gauges used in this study, and the rate has increased at both deep ocean island stations and at continental shelf gauges (Figure 1; Tables 1a and 1b). This rapid increase in Western Pacific MSL rate of change provides an opportunity to seek connections between MSL rise and tidal evolution. If MSL rise is a primary driver of changes in the tides, as suggested by Arbic and Garrett (2010), then this area should exhibit unusually large rates of tidal evolution over the same time period, with a relatively simple spatial pattern. Alternatively, if changes in internal wave propagation (as in Hawaii, Colosi and Munk, 2006) are dominant, then smaller-scale, less spatially coherent tidal evolution would be expected. Shallow water frictional processes can also alter tidal constituents and may be altered by MSL variations (Amin, 1983), but this local process would not likely drive regional trends. Finally, the existence of multiple mechanisms could lead to complex patterns of tidal evolution, and this is, in fact, what our examination of

Pacific Island gauges suggests. Furthermore, the results of our analyses suggest that different processes are active in different parts of the frequency spectrum, and so the behavior of short-term fluctuations do not, in general, allow a clear prediction of what may be expected for long-term trends of tidal properties.

We examine the interactions of detrended tidal anomalies—the deviations from long-term trends in tidal constituent properties—in relation to detrended MSL anomalies (deviations from long-term MSL trends) as a means of diagnosing dynamical factors that might influence the longer-term evolution of tides. Specifically, we examine MSL anomalies and tidal anomalies at 26 tide gauges in the Western Pacific region, at both island gauges, and at nearby continental shelves. We focus on four constituents, the two largest semidiurnal (twice daily) constituents, M_2 (twice daily lunar tide) and S_2 (twice daily solar tide), and the two largest diurnal (once daily) components, K_1 (lunisolar diurnal) and O_1 (lunar diurnal). In addition, a subset of gauges with a minimum length-of-record (LOR) ≥ 40 yr is used to determine whether any changes in tidal trends are associated with the onset of anomalous regional MSL rise in recent decades. We will compare the long-term linear trends of both MSL and tidal properties (amplitude and phase), as well as the shorter-term fluctuations of MSL anomalies and tidal property anomalies from these long term trends (LTTs). Under rapidly rising and warming sea level scenarios, tidal amphidromes may shift as MSL rises and stratification increases (Pugh, 2004, Müller 2010). Amphidromes may shift center position, amplify, and/or rotate. For all four tides, we will compare the changes in amplitudes and phases with changes in MSL in the context of possible amphidromic migrations.

We will also attempt to identify the mechanisms behind the observed trends in tidal properties in several ways. First, we consider non-linear, shallow-water overtides as a tool for understanding changing friction over continental shelves and in shallow water regions in relation to changing tides. Then we will compare the variations of tidal trends at select stations to fluctuations in regional thermocline depth. Finally, we will consider the possibility of a resonant triad interaction between the K , O_1 , and M_2 tides.

2. Background

2.1. Changing sea level and changing tides

Ocean tides are usually considered stationary in time because of their close relationship to astronomical forcing. However, recent work has shown that tides are evolving at diverse rates in different parts of the ocean without any apparent relationship to astronomical forcing (Woodworth, 2010). Changes in major tidal constituents such as M_2 , S_2 , K_1 and O_1 are observed in the Eastern Pacific (Jay, 2009), the Gulf of Maine (Ray, 2006), and the North Atlantic (Ray, 2009; Müller et al., 2011). Changes in tidal range and datum levels along many parts of the US coast line (Flick et al., 2003) also indicate evolution of constituents. While several mechanisms have been suggested that are independent of mean sea level (MSL) rise (Jay 2009), MSL may influence or be correlated with tidal evolution in a variety of ways, both locally and on amphidromic scales, as well as over a multitude of frequency bands, from seasonal to multi-decadal. One possibility is that changes in continental shelf depth may influence tides on a basin scale (Arbic and Garrett, 2010, Arbic et al., 2009), through changing bed friction in shallow water. Rising MSL also alters tidal

wavelength and wave speed in shallow water areas, and reduces energy dissipation due to the reduced effect of bottom friction (Pugh, 1987). Alternatively, MSL rise may coincide with changes in stratification and/or thermocline depth. These internal changes may alter the surface manifestation of internal tides in ways that are detectable at tide gauges, e.g., at Honolulu (Colosi and Munk, 2006). Finally, fluctuations in multiple constituents may be related via resonant triads (cf. Lamb, 2007). These factors, individually or in combination, could lead to a migration of the tidal amphidromes that would appear as a change in tidal properties at fixed observation stations.

The long-term global average trend in MSL is 1.7 ± 0.3 mm yr^{-1} , as determined from tide gauge records (Church and White, 2006; 2011). Satellite observations show that the global MSL rise since 1993 has been 3.3 ± 0.4 mm yr^{-1} (Nicholls and Cavenaze, 2010), suggesting a recent acceleration. However, MSL rise is spatially non-uniform. Five-year running means suggest that the MSL trend in the western Tropical Pacific has approached $+10$ mm yr^{-1} at some locations since 1993 (Merrifield, 2011). In contrast, MSL rise rates in the Northeastern Pacific have been below the global average over the same time period (National Research Council Ocean Science Board, 2012), with some places along the West Coast of the US having constant or slightly negative rates since 1980, partly due to prevailing wind patterns. While it is unlikely that the extreme rise rate in the western tropical Pacific will persist, and Bromirski et al. (2011) have suggested that a trend reversal along the US West Coast is imminent, such decadal-scale variations in MSL rise rates offer an opportunity to examine the connections between different parts of the water level spectrum, in this case, between tidal evolution and MSL.

MSL also exhibits fluctuations related to climate cycles. The El-Nino Southern Oscillation (ENSO) is a strong contributing factor to sea surface anomalies in the Pacific (Kohl et al. 2007; Lombard et al. 2009; Timmermann et al. 2010), and local, short-term MSL anomalies associated with ENSO are often much larger than long-term trends. The 1982-1983 and the 1997-1998 El Nino events gave rise to widespread MSL fluctuations, with some coastal localities showing yearly averaged MSL rises or falls of 200mm or more (Nerem et al, 1999; White et al, 2001). ENSO-related sea surface height (SSH) trends are, however, an order of magnitude too small (over a ~ 15 yr time period) and too transient to account for the observed MSL rise for the western tropical Pacific region (Merrifield, 2011). Instead, Merrifield argues that the unusually rapid MSL rise in the western tropical Pacific is correlated with changing wind stress patterns in the region from after 1993, corresponding roughly to the period of satellite observations. ENSO related sea level variability is difficult to remove entirely from MSL time series, due to its quasi-periodic behavior, and large geographical extent that the signal affects. Hamlington, et al. (2011) present a detailed study on the capture and removal of the ENSO signal and of modulated annual variability from global sea level records for 1950 through 2009, using cyclostationary empirical orthogonal function techniques, based on the methods of Church et al. (2004).

Predictions of future inundation due to extreme events must account for both MSL rise and changes in tidal properties. For example, Haigh et al. (2010) found that MSL rise rates of $+0.8$ - 2.3 mm yr^{-1} in the English Channel were accompanied by smaller increases in high water elevation of $+0.1$ - 0.3 mm yr^{-1} (relative to MSL). A numerical model from Pickering et al. (2012) found that a 2m increase in MSL increased the spring tidal range from -0.49 m to

+0.35m, depending on location. Hence, changing tidal processes can either mitigate or exacerbate changes to extreme water levels caused by MSL rise (Horsburgh and Wilson, 2007).

Changing internal tides are a likely mechanism connecting MSL change and the tidal evolution, at least between the critical latitudes for the major tidal species, $\pm 30^\circ$ latitude for K_1 and, and $\pm 75^\circ$ for M_2 . Ray and Mitchum (1997) report that time-series of the monthly variations of M_2 and MSL were coherent at annual, inter-annual and intra-annual bands, and Mitchum and Chiswell (2000) note that higher amplitudes of M_2 are associated with times of a deeper thermocline. Colosi and Munk (2006) conclude that the observed increase in the total M_2 amplitude from 161mm to 169mm at Honolulu over the past century occurred because the internal M_2 wave moved more closely into phase with the surface tide. This conclusion is dependent on the assumption that rising sea level is a proxy for a deeper thermocline, which then alters the phase and/or the amplitudes of the internal tide.

2.2 Resonant Triads-- Theory

Resonant triad interactions are a possible mechanism for coherent fluctuations of multiple tidal constituents. Resonant (and near-resonant) triad interactions, were first described in the context of light waves by Armstrong et al (1962), discussed generally by Bretherton (1964), and solutions for water waves are detailed by Craik (1985). Resonant internal wave triads arise from the (nonlinear) convective acceleration terms and are essential to energy transfer across the oceanic tidal spectrum (Hibiya et al 2002; Gerkama et al, 2006), including shoaling surface gravity waves in shallow water (Frelich and Guza, 1984), and are involved in parametric subharmonic instabilities (PSI) at or near critical latitudes (McKinnon and Winters, 2005).

A resonant triad is a pathway for an energy conserving, nonlinear energy exchange between multiple wave modes. The three (or more) complex plane waves involved in a triad have the form:

$$Z_j(t) = A_j(t) \exp[i\theta_j(t)] \quad (1)$$

$$\text{with : } \theta_j = (\mathbf{k}_j \cdot \mathbf{x} - \omega_j t) = (k_j x + l_j y + m_j z - \omega_j t)$$

The ω_j 's are the frequencies of each wave; $\mathbf{k}_j = (k_j, l_j, m_j)$ are the wavevectors of each wave, and each wave has a dispersion relation given by $\omega_j(\mathbf{k}_j)$. For three interacting waves, the main resonance conditions are (Craik, 1985; Simmons, 1969) that the waves are phase locked (i.e., the sum of the phases is a constant), and that the frequencies and wavenumbers are also matched via a summation. Thus:

$$\theta_1 + \theta_2 + \theta_3 = C, \quad (2a)$$

$$\begin{aligned} \omega_1 \pm \omega_2 \pm \omega_3 &= \Delta\omega, \\ \mathbf{k}_1 \pm \mathbf{k}_2 \pm \mathbf{k}_3 &= \Delta\mathbf{k} \text{ [for : } \mathbf{k} = (k, l, m)] \end{aligned} \quad (2b, c)$$

Small frequency and wavenumber mismatches ($\Delta\omega$ and $\Delta\mathbf{k}$) allow for the possibility of near-resonant triads (Craik, 1985); exact triads have $\Delta\omega=0$ and $\Delta\mathbf{k}=0$. Near-resonant triads are possible in the ocean for internal tide generation at or near topographic features (Lamb, 2007). These relationships hold for a single triad of three waves, but are applicable to multiple triads of three waves, or even groups of four waves (tetrads). While Eq. (2b) is satisfied exactly ($\Delta\omega=0$) by a number of tidal constituent combinations, Eq. (2c) is geometry-dependent. The dispersion relationship, $\omega_i(\mathbf{k}_i)$, for planar and linear internal waves in the ocean is given by (Gill, 1982):

$$\omega_j(\mathbf{k}_j)^2 = \frac{N^2(k_j^2 + l_j^2) + f^2 m_j^2}{k_j^2 + l_j^2 + m_j^2} \quad (3)$$

Where N indicates the buoyancy frequency, and f is the Coriolis frequency; f is a constant for each gauge, while N is dependent on the local depth profile of temperature and salinity, and may vary seasonally.

There are more than 450 tidal frequencies in the ocean (Cartwright and Tayler, 1971); many of these frequencies could in theory form triads that satisfy Eq. (2b), but most such triads are based on minor tidal constituents and are unlikely to be energetic enough to be measurable. The triad most likely to be detected consists of K_1 , O_1 and M_2 . The K_1 and O_1 tidal frequencies are close to each other, being equidistant (plus or minus ~2%) from half the M_2 frequency; in terms of M_2 , they are:

$$\omega_{m2} = \omega_{k1} + \omega_{o1} \quad (4)$$

$$(\omega_{k1} = 0.52\omega_{m2}); (\omega_{o1} = 0.48\omega_{m2})$$

We note that shallow water (frictional) overtides satisfy Eqs. (2a) and (2b), and may satisfy Eq. (2c).

Ball (1964) details how triad interactions can occur at the interface of a two-layer fluid, e.g., at the thermocline. The strength of the interactions is dependent on the depths of the layers as well as the stratification and the angles of incidence of each wave relative to the thermocline interface. Recent upper-layer warming in the Western Pacific (Domingues et al, 2008) may be leading to a deepening thermocline and stronger stratification (Müller, 2012), which makes the ocean effectively two-layered, and enhanced triad interactions may occur at the interface. MSL rise may also cause an extension in a basin's areal extent, as new areas are inundated. The close frequencies of K_1 and O_1 will yield similar wavenumber vectors for any given mode. In addition, because they are both nearly half the M_2 frequency, internal modes of K_1 and O_1 may energy exchange via M_2 . Alternatively, remote changes in barotropic M_2 may alter low-mode diurnal internal tides via parametric subharmonic instability (MacKinnon and Winters, 2005). Both possibilities are explored below.

3. Methods

3.1. Station Selection and Data

Twenty-two of the 26 tidal records used in this study were selected from the University of Hawaii Sea Level Center (UHSLC) archives. The Australian stations of Cairns, Gladstone, and Williamstown were provided by the Australian National Tidal Center and Auckland, New Zealand was provided by Land Information New Zealand (LINZ). The factors that guided our station selection were:

1) Location: all stations are in or near the area of anomalous MSL rise.

2) Temporal coverage: all stations used have an LOR (length of record) greater than one nodal cycle (18.6yrs); LOR ranged from 19 to 107yrs.

3) Completeness: the data were more than 80% complete over the record.

There were 17 island and 9 coastal stations that met these criteria, allowing comparison of open-ocean and continental shelf dynamics. Locations are shown in Figure 1, along with bathymetry. Figure 2 shows yearly average MSL time series at 7 long-term gauges (Guam, Pago Pago, Kwaljein, Malakal, Yap, Honiara, Kanton). These suggest a break in sea level trend at or around 1993 (broken vertical line). All of these 7 gauges and more than half of the 26 total gauges of our study show an increase in MSL rate after this point up to the present. The longer-period gauges used will be subject to larger timing errors in the earlier years of the records, but these clock errors have improved in the recent decades with the advent of precise GPS clocks. A detailed discussion of the effect of clock errors in the Western Pacific can be found in: (Zaron and Jay, 2014).

Table 1a shows the gauges used (location name and country), start and end year of the complete length of record (LOR), and a descriptor indicating the environment where the gauge is located. For islands, delineation is made between low-lying atolls, steep islands with active volcanism, and non-volcanic mountainous terrain. Locations where a coral reef surrounds one or both sides of the island are also indicated. For coastal locations, we note whether the gauge location involves a river estuary, or a coastal enclosed bay. The last three columns of Table 1a show the MSL rates, expressed in mmyr^{-1} , shown first for the entire record, and then for the years before 1993, and after 1993. Rates are only reported for the early years if there exists at least a full nodal period of data (~ 19 years), otherwise, they are not calculated, since these determinations may be inaccurate.

3.2. Harmonic Analysis and Tidal Admittance

To account for nodal cycle variability, we investigate tidal trends through the use of a tidal admittance. Admittance is a unitless, complex ratio of an observed tidal constituent (m) to its corresponding astronomical equilibrium amplitude, V/g (m); tidal potential V is in m^2s^{-2} and g is the acceleration due to gravity, in ms^{-2} . Because nodal and other low-frequency astronomical variability is present with similar (relative) strength in both the observed tidal record and in V/g , it is largely absent in admittance time series.

For all relevant stations, overlapping yearly tidal harmonic analyses (at monthly time steps) were carried out on hourly records and hourly V/g, using the tidal harmonic analysis package in MATLAB developed by Pawlowicz (2002), and modified to include robust fitting by Leffler and Jay (2009). The tidal potential is determined using a program provided by Richard Ray (National Atmospheric and Space Administration, personal communication) that is based on the methods of Cartwright and Tayler (1971), and Cartwright and Edden (1973). Results from a single harmonic analysis determine an amplitude, A , and phase, θ , at the central time of the analysis window for each tidal constituent with error estimates for both. Use of a moving analysis window produces time-series of amplitude and phase; both monthly (767hr) and annual analyses were used. From amplitude $A(t)$ and phase $\theta(t)$ time series one can construct complex amplitudes $\mathbf{Z}(t)$,

$$\mathbf{Z}(t) = A(t)e^{-i\theta(t)}. \quad (5)$$

We form time-series of tidal admittance ratio (AR) and phase difference (PD) for any constituent using Eqs. (6) and (7),

$$\mathbf{AR}(t) = \frac{A_{obs}(t)}{A_{pot}(t)} \quad (6)$$

$$\mathbf{PD}(t) = \theta_{obs}(t) - \theta_{pot}(t) \quad (7)$$

Where the subscripts ‘obs’ or ‘pot’ denote the the observed data or the equilibrium amplitude, respectively. Both the admittance ratio in Eq. (6) and the phase difference in Eq. (7) largely remove the nodal-cycle variability, allowing easier examination of the non-tidal signals.

3.3. Mean Sea Level and Anomaly Trends

The 26 gauges used have a variable time coverage; some have less than 40 years of data, others have 100 years or more. While the long-term trends are an important concern that must still be considered in the overall picture of the ocean, our primary focus here is the interannual variations of MSL and the tides. MSL time-series are generated by taking overlapping yearly averages of the original hourly water level data at monthly intervals, without corrections for local ground motion or inverted barometer effects. AR and PD time series are generated by overlapping yearly harmonic analyses. We remove trends from the yearly AR, PD, and MSL time-series for each location and constituent over the entire record. The removed residual trends are fit by linear regression, reported in Table 1a (MSL) and 1b (ARs and PDs).

The removal of the long-term trends leaves us with time-series of residual variations in AR, PD, and MSL, which are compared with one another (and with other ancillary time-series like thermocline depth), using scatter plots, cross-correlations, and regression to determine a tidal anomaly trend (TAT) and related statistics (Huber, 1981). The mean value of the tidal potential at each location is used as a scale factor to convert the non-dimensional TAT into a

dimensional amplitude trend (A-TAT, expressed as millimeter change per meter sea level rise (mmm^{-1}). The same approach is used with the phase difference time-series to provide phase anomaly trends relative to the potential (P-TAT), with the trends expressed as degree change per meter sea level rise (degm^{-1}).

Sample results (Figure 3) show the A-TATs (top 4 panes, (a) - (d)) and P-TATs (bottom 4 panes, (e) - (h)) for the diurnal (K_1 and O_1) and semidiurnal (M_2 and S_2) tides vs. MSL at Honiara, Solomon Islands (9.4167 S, 159.950 E). The scatter plots at Honiara exhibit some of the most coherent trends in our data set, for M_2 , with $r^2 > 0.9$. While the M_2 tide amplitude is relatively small at this location ($\sim 50\text{mm}$), the anomaly trend is large, $+65.6 \pm 3.3 \text{ m mm}^{-1}$ (132% of the local M_2 amplitude per meter of MSL rise). The S_2 trend in amplitude is not as coherent, but is still significant, as are the K_1 and O_1 A-TATs (Table 2; plots in supplementary materials). Trends in M_2 and S_2 phase are smaller but still significant (Table 3). The diurnal P-TATs are not significant. A-TAT and P-TAT plots for semidiurnal and diurnal tides at all 26 stations are provided in supplementary materials (Figures S1-1, through S1-26; Online Resource 1).

3.4. Friction and Overtides

Overtide generation is superficially similar to a resonant triad interaction, since both may satisfy a similar frequency condition to Eq. (2b), (i.e., $\omega_1 + \omega_2 = \omega_3$). Overtides can be linked to bottom topography, but may also results from internal waves of large amplitude (Legg and Klymak, 2008), or from strong flow curvature (Parker, 1991). The largest overtide at most locations is M_4 , twice the frequency of M_2 ; M_4 (ω_3) results from the non-linear interaction of M_2 tidal currents with itself ($\omega_1 + \omega_2$). However, the wavenumber condition Eq. (2c) will only be satisfied in certain cases (Kukulka and Jay, 2003), and is not a requirement. In addition, the driving non-linearity (quadratic bed friction) is not the same as that for resonant triads (convective accelerations). Still, a frictional triad may interact with a resonant triad, as can a mean flow or an eddy field (Lelong and Kunze, 2013). It is, therefore, reasonable to suppose constituent interactions to be part frictional and part resonant. If overtides change substantially over time, it may indicate changes in non-linear frictional interaction (Parker, 1991). Because M_4 is not represented in the astronomical potential, it is normalized by an appropriate overtide ratio (OR) to its forcing constituent, M_2 ,

$$OR_{m4}(t) = \frac{A_{m4}(t)}{[A_{m2}(t)]^2} \quad . \quad (8)$$

Here, OR_{m4} indicates the overtide ratio of M_4 . This ratio is analogous to an admittance calculation and will remove known long-period variations such as the nodal cycle. Additional overtides investigated are: S_4 (interaction of S_2 with itself), MK_3 (interaction of M_2 and K_1), M_6 (third harmonic of M_2), MS_4 (interaction of M_2 and S_2), and MO_3 (interaction of O_1 and M_2). In each case, the overtide “ratio” is that between the overtide amplitude and the product of the amplitudes of the forcing waves. Overtides are small and may not be resolvable against background noise; we consider only overtides with a signal-to-noise ratio greater than 2.0. Temporal trends are normalized as the percentage change (per year) from local means of the overtide ratios (ORs). Additionally, we compare time-series of M_4 and M_6

overtide ratios at Honiara and Rabaul, two stations where the local overtimes are anomalously large, to local MSL variations, and to the El Nino Southern Oscillation (ENSO) through the use of the Multivariate El Nino Index (MEI), as defined by the National Oceanic and Atmospheric Administration (NOAA) climate data center: <http://www.ncdc.noaa.gov>. Note that the Rabaul station was decommissioned in 1997.

3.5 Thermocline Depth and EEMD Analysis

Observed tidal properties are compared to estimated thermocline depths to diagnose the possible role of internal processes in tidal anomaly trends. Thermocline depths have been synthesized from the NOAA Tropical Atmosphere Ocean project (TAO) buoys: (<http://www.pmel.noaa.gov/tao/>). The TAO array, established in the early 1990s, is a gridded network of 70 buoys extending from 8°N to 8°S and from 137° E to 95° W. These buoys record sea surface temperature (SST), wind speed, and air pressure plus water temperature at increments of 25m down to 250m, and also at 300m and 500m. From this profile, the depth of the 20°C isotherm, D_{20} , is estimated and used as a proxy for thermocline depth. D_{20} time-series are compared to tidal property (AR and PD) time-series for gauges at Honiara, Pago Pago, and Kapingamarangi, where some of the largest TATs (relative to local mean tides) are seen for the semidiurnal tides (M_2 and S_2). Monthly averaged D_{20} data are compared to monthly M_2 and S_2 admittance data, calculated using 767 hour analysis windows in *R_T_Tide*, allowing examination of seasonal, as well as interannual, variations.

The D_{20} and monthly A-TAT time-series are separated into quasi-orthogonal scales using an EEMD (Ensemble Empirical Mode Decomposition) analysis (Huang, et al., 1998; Huang and Wu, 2008). EEMD uses a sifting process to split signals into approximately dyadic modes—a signal of length 2^n yields $\sim n$ modes. We selectively recombine modes to preserve the desired scales of variation (from about annual to decadal scale), while removing higher frequency variability and the long-term trend. In the language of EEMD, the decomposition of the variables are expressed as intrinsic mode frequencies, or IMFs. For all variables considered, we will use a decomposition of 10 IMFs, with IMF #10 being the extracted long-term trend of the time series.

3.6 Analysis of Resonant Triads

Resonant triad interactions are nonlinear and occur through the convective accelerations. They are described by coupled evolution equations for the “envelopes” of the three interacting waves, assumed to vary on a “slow” timescale, much longer than the actual wave periods. These non-dimensional evolution equations relate the rate of change of the complex amplitude of one wave to the product of the other two waves’ complex amplitudes (Craig, 1985),

$$(\partial_t + \mathbf{c}_j \cdot \nabla) \mathbf{Z}_j = i \delta_j (\mathbf{Z}_k^* \mathbf{Z}_l^*) \quad (9)$$

$l \neq j \neq i$

Where $\mathbf{c}_i$ is the group velocity, $\mathbf{Z}$ is the complex amplitude, $\mathbf{Z}^*$ is the complex conjugate of $\mathbf{Z}$, and δ_j the interaction coefficient of the j^{th} mode. The δ_j describe the strength of the wave interactions; each is the sum of a number of bilinear correlations between waves. We neglect the spatial gradient terms (a “local” assumption), assume that amplitudes and phases will be slowly-varying, and also assume that only one triad—the one linking the M_2 , K_1 and O_1 tides—is active. Note that the triad relationships involve the slow timescale (interannual or decadal) modulations of the tidal harmonics, not the fast timescale of the tidal frequencies themselves (approximately once and twice daily), and our yearly admittance ratios and phase difference have already removed the fast-scale variations of the tides. Thus, our time-series for AR and PD can be used to represent slow-scale variation time series of amplitudes and phases, respectively. We rewrite Eq. (1) to show the form of the slowly varying complex triad modes,

$$Z_j(\tau) = AR_j(\tau) \exp[-i(\omega_j \tau + PD_j(\tau))] . \quad (10a)$$

Both the AR’s and the PD’s are time-dependent, and the AR’s are real. Thus, taking the time derivative of the LHS under the local assumption of Eq. (9) yields three terms,

$$\frac{d}{d\tau}(Z_j(\tau)) = \left[\frac{dZ_j(\tau)}{d\tau} \right] Z_j(\tau) = \left[\frac{1}{AR_j(\tau)} \frac{d}{d\tau} AR_j(\tau) - i\omega_j - i \frac{d}{d\tau} PD_j(\tau) \right] Z_j(\tau) . \quad (10b)$$

Equating this expression to the RHS, combining complex exponential terms, and separating the real and imaginary terms (via the Euler formula) into two equations leads to,

$$\left[\frac{1}{AR_j(\tau)} \frac{d}{d\tau} AR_j(\tau) \right] AR_j(\tau) = \delta_j (AR_k(\tau) AR_l(\tau)) \sin(\Phi(\tau) + \Delta\omega\tau) \quad (11a)$$

$$\left[\omega + \frac{d}{d\tau} PD_j(\tau) \right] AR_j(\tau) = \delta_j (AR_k(\tau) AR_l(\tau)) \cos(\Phi(\tau) + \Delta\omega\tau) . \quad (11b)$$

Here, the $\Delta\omega$ term is the sum of the frequencies as in Eq. (2b): $\Delta\omega = \omega_1 + \omega_2 + \omega_3$; for the $M_2+K_1+O_1$ triad, $\Delta\omega = 0$. The $\Phi(\tau)$ term is the sum of the three phase differences: $\Phi(\tau) = PD_1(\tau) + PD_2(\tau) + PD_3(\tau)$.

Sets of equations like Eq. (11a; b) can, in principle, be solved using inverse scattering methods (Zakharov and Manakov, 1973; Kaup, 1980), but sufficient environmental information to do so is rarely present in an oceanographic context. We manipulate Eq. (11a; b), therefore, to obtain a tool for diagnosis of triad interactions; Eq. (11a) and Eq. (11b) can be rewritten in terms of M_2 , K_1 , and O_1 as:

$$C_{\text{Re}}[\text{Re}\{AR_{M_2}(\tau)\}] = \Gamma_{\text{Re}}[\text{Re}\{AR_{K_1}(\tau) AR_{O_1}(\tau)\}] \sin(\Phi(\tau)) \quad (12a)$$

$$C_{\text{Im}}[\text{Im}\{AR_{M_2}(\tau)\}] = \Gamma_{\text{Im}}[\text{Im}\{AR_{K_1}(\tau) AR_{O_1}(\tau)\}] \cos(\Phi(\tau)) \quad (12b)$$

C_{Re} and C_{Im} represent the terms in brackets in Eqs. (11a,b), and $\Gamma = \Gamma_{Re} + i^*\Gamma_{Im}$ is a complex interaction coefficient related to the phase speeds and dispersion relations of the interaction. Finally, we can examine the absolute variation by combining both expressions of Eqs. (11a) and (11b) or (12a) and (12b) into a single equation,

$$AR_{M_2}(\tau) = \frac{\Gamma_{Abs}}{C_{Abs}} AR_{K_1}(\tau) AR_{O_1}(\tau) . \quad (13)$$

The salient features of Eq. (13) are that: a) AR_{M_2} is proportional to the product on the RHS, and b) $\Gamma_{Abs}/C_{abs} = \beta < 0$, so that energy lost from M_2 is taken up by the two diurnals (at least if the two diurnals behave in a similar manner). A complete triad problem requires two more equations similar to Eq. (13), but we shall only focus on the one that uses M_2 as the primary. Here, we use Eq. (13) as a diagnostic tool to identify and describe qualitatively the M_2 - K_1 - O_1 triad. Eq (2) does not use phase information. But from Eq. (2a) we have a “phase-lock” constraint,

$$\Phi(\tau) = PD_1(\tau) + PD_2(\tau) + PD_3(\tau) = const. \quad (14)$$

For the PD variations, we convert variations from degrees to minutes. Also, β in Eq. (13) can be estimated by regressing AR_{M_2} against $AR_{K_1} \times AR_{O_1}$. Finally, triad interactions occur on slow, and possibly multiple, timescales. Thus, it is logical to compare variations on multiple scales, separated via EEMD. Trends were extracted, and modes of variability separated via EEMD, as with the thermocline data. Applying this approach, we compare the two sides of Eq. (13) using IMFs 4-10 of all quantities. To illustrate the possible importance of triad interactions, we examine below the coupled variations of M_2 , K_1 , and O_1 at Honiara and Rabaul, stations at which constituent variations are particularly strong and coherent with MSL.

Because the data we analyze are surface water levels and not internal wave amplitudes at the thermocline level, we cannot use Eqs. (11ab, to 13) without additional assumptions. We assume that, for nearly linear internal waves, the induced surface displacement is governed by the depth of the interface and the ratio of the restoring forces between surface and internal waves. If the changes in interfacial depth and stratification are small, then the surface displacements measured at tide gauges will be proportional to the interfacial amplitudes (and their variations) described by Eqs. (12-13). This approach allows us to analyze interactions between baroclinic and surface tides, because the surface displacements of both involve similar potential energies. The method outlined here also assumes that the wavenumber vectors for each wave are constant, with amplitudes and phases that evolve slowly. Rising MSL renders a constant wavenumber assumption questionable, especially for the vertical wavenumbers (m_i). Still, it is of interest to identify wavenumbers for which triads are possible, assuming stationary wavenumbers. For the triad of interest (K_1 , O_1 , and M_2), Eq. (2b) is exact, but Eqs. (2a), (2c) and (3) must still be satisfied. To simplify finding wavenumber vectors that satisfy the triad restraints, the dispersion relation, Eq. (3), can be recast as an aspect ratio or slope factor (Gill, 1982),

$$\sqrt{\frac{(\omega_i^2 - N^2)}{(f^2 - \omega_i^2)}} = \frac{m_i}{\kappa_i} \quad (15)$$

using a general horizontal wavenumber $\kappa_i^2 = (k_i^2 + l_i^2)$. The Coriolis frequency, f , will be dependent only on latitude, but N^2 will depend on temporally variable temperature and salinity profiles. However, because all three triad waves are present at the same place, all three will have the same N at any time. Choosing one vertical wavenumber, (m_1) , κ_1 is determined from Eq. (15). We then select $(m_2, \kappa_2, m_3, \text{ and } \kappa_3)$ combinations that are consistent with Eq. (15) and then test their consistency with Eq. (2). Because higher modes can be involved in a resonant triad, there could be several wavenumber solutions that satisfy Eqs. (2a, c) and (3); however, higher modes are usually less energetic. We therefore only consider (vertical) modes 1 to 3, along with the zero-mode (barotropic) waves.

We use the Honiara and Rabaul records to illustrate the role of triad interactions in tidal evolution, because both stations show strong fluctuations in M_2 , K_1 and O_1 that are relatively coherent with MSL variations. Application of Eqs. (12-15) then requires determination of a locally appropriate density profile. N^2 profiles are estimated from temperature-salinity (T-S) profiles obtained from the World Ocean Atlas 2009: (Locarnini, et al., 2010), using the TEOS-10 (Thermodynamics Equations of Seawater) standard (Millero, et al., 2008). For convenience, we express all frequencies non-dimensionally, relative to the M_2 forcing frequency. Thus, the M_2 frequency = 1; K_1 = 0.52; O_1 = 0.48. The Coriolis frequency, f , at Honiara (9.5° S), is ~ 0.17 ; that at Rabaul is ~ 0.07 . For both stations, non-dimensional N in the thermocline ranges from about 10 to 100. With these values of N , Eq. (15) can be simplified, for conceptual purposes, to:

$$\frac{N}{\omega_i} \cong \frac{m_i}{\kappa_i} \quad (16)$$

This form emphasizes the fact that K_1 and O_1 will have similar wavenumbers for any given internal wave mode.

Because we consider only the barotropic waves and the first three internal modes; there are only four choices for each m_i ; $m=0, 1, 2, 3$, when non-dimensionalized by depth. Inserting these values of m_i into m_i/κ_i ratio (Eqs. 15; 16) gives values for the κ_i , which may be examined for conformity with wavenumber resonance condition Eq. (2c) for each mode. The vertical to horizontal aspect ratio of the ocean is small, so the $m_i \gg \kappa_i$. Because $|M_2|$ is quite small in the Solomon Sea relative to the diurnal constituents, it seems likely that the most easily detected triad is the barotropic M_2 wave interacting with a first-mode internal K_1 , and an oppositely propagating first-mode internal O_1 . This is a form of parametric sub-harmonic instability that has been analyzed in the context of a resonant triad (but not specifically the M_2 - K_1 - O_1 triad) in the South China Sea (Xie, 2008) and also near the Hawaiian Ridge (Carter and Gregg, 2006).

4. Results

4.1 Anomaly Trend Maps

The spatial distribution of anomaly trends (A-TATs and P-TATs) may help identify connections between MSL rise and tidal evolution on a basin-wide, or amphidromic scale. Figures 4 to 7 show the A-TATs for each constituent. P-TATs are not shown on the plots, but are discussed below in the context of possible amphidromic migration. Exact values and error bands are reported in Tables 2 (A-TATs) and 3 (P-TATs). The magnitude of the A-TATs is shown by the color intensity of the dots, with the color indicating positive (red) or negative (blue). For a gauge with an insignificant A-TAT, the dot is white.

These plots provide a picture of the tidal response to a 1m MSL rise and an indication of the spatial coherence of tidal evolution. For a gauge influenced by only one amphidrome, a positive (negative) A-TAT and a zero P-TAT indicates that the amphidrome is moving away (towards) the gauge. Gauges that show significant P-TATs suggest a rotational component to the movement of the amphidrome. If changes in tides (as represented by A-TATs and P-TATs) are coherent on an amphidromic scale, they should suggest consistent amphidromic movement or rotation. If statistically significant TATs do not suggest consistent amphidromic change, then local or regional process are likely more important.

4.1.1 Diurnal Constituents

K_1 has its largest amplitudes along the western Pacific equator, with much lower amplitudes eastward in the open-ocean (Figure 4). Based on the tidal potential, diurnal tides are expected to be small near the equator, and the large amplitudes seen (4-5 times that predicted by astronomical forcing) around Indonesia and Papua New Guinea may represent a resonance. Diurnal tides are larger within the Coral Sea, but are very small around New Zealand, where semidiurnal tides prevail.

There are statistically significant ($SNR > 2.0$) positive K_1 amplitude and phase tidal anomaly trends at 18 of 26 gauges. Large positive A-TATs are seen at both island and shelf stations, while significant negative A-TATs are only seen at island gauges. There is a concentration of positive A-TATs at Australian gauges. All negative A-TATs are north of $\sim 10^\circ$ S. For the K_1 P-TATs, 18 of 26 stations are significant (though not the same 18 as the A-TATs). The strongest positive P-TAT is seen at Guam, and strong negative trends in phase are seen at the islands near the center of the map.

The amplitude and phase distributions and amphidromic patterns for O_1 are similar to those for K_1 , though O_1 amplitudes are generally smaller (Figure 5). Like K_1 , O_1 is resonant near Papua New Guinea and the Indonesian archipelago. Amphidromic points are again marked by “A” and “B”. Significant positive A-TATs are also found at most Australian shelf stations. In total, there are 14 significant A-TATs, and 20 significant P-TATs.

4.1.2 Semidiurnal constituents

M_2 amplitudes are generally larger than diurnal amplitudes in the study area, with the largest amplitudes occurring around New Zealand and the northern coast of Australia (Figure 6). An amphidromic point is located just to the east of Pago Pago (marked by “B”), with an anti-node (high amplitudes) along the equator west of the date line. Very low M_2 amplitudes are seen in and around the Solomon Sea.

There are significant M_2 A-TATs at 17 of 26 stations and significant M_2 P-TATs at 19 of 26 stations. Positive and negative A-TATs are observed at both island gauges and shelf gauges. Honiara and Rabaul, which are near the M_2 amphidrome in the Solomon Sea, have small mean M_2 amplitudes (~50mm), but display large relative trends; (+132% for Honiara and +240% for Rabaul), though the strong phase trends at these two gauges have opposite magnitudes. With the exception of Auckland, all significant positive A-TATs are south of the equator, and all negative A-TATs are north of the equator.

S_2 is much smaller than M_2 at most locations, but with a similar (large) amplification on the northwest coast of Australia (Figure 7). Also similar is the location of the amphidromic points, with “A” indicating another line-shaped amphidrome (which more closely follows the Mariana trench), and an amphidromic point at “B” very near Pago Pago.

4.1.3 Combined Tidal Trends

We show the combined A-TATs in Figure 8; values and confidence limits are tabulated in Table 2. The constituent TATs were combined as complex numbers (using A-TAT and P-TAT determinations for each of the four constituents examined), and the real part of the result is plotted. Conceptually, a large and positive combined A-TAT describes locations in which the tidal range increases strongly during periods with a positive sea-level anomaly, whereas a negative combined A-TAT implies decreased tidal range for a positive MSL anomaly. Therefore, shelf-locations such as Brisbane, or island stations such as Pago Pago or Noumea could see large tidal range increases for a positive sea-level anomaly. By contrast, stations such as Legaspi, Guam, and Johnston could see decreased tidal range for a positive sea-level anomaly. Large positive tidal-range responses are all south of 9°N, while all but one of the large negative changes are north of the latitude.

4.2 Overtides

Figure 9(a-d) summarizes the spatial distribution of overtide ratios (ORs) for: M_4 , M_6 , S_4 , and MK_3 ; significant long-term trends are shown as colored dots. MS_4 and MO_3 plots are shown in Figures S2 and S3 (Online Resource 1). Significant changes over time for these four ORs are mainly seen at coastal shelf stations, with a few islands showing some moderate change over time. The largest M_4 and M_6 ratio means are in the Solomon and Coral Seas, and in general, regions of high M_4 and M_6 ratios correspond to areas where lower M_2 amplitude are found (Figure 6). Mean S_4 ratios are generally smaller than those for M_4 and M_6 , and fewer gauges show changes over time; those that do are decreasing. Again, the S_4 ratio is large in areas where S_2 amplitudes are small (Figure 7). MK_3 is also generally smaller than other overtides, but shows larger relative changes at many stations. Examination of the numerators and denominators of the ORs shows that in almost all cases the trend is driven by changes in the observed

overtides (numerator), not by changes in the forcing (denominator). This suggests a mechanism of changing ORs in the area being driven by frictional processes rather than by secular increases or spatial shifts of the gravitationally forced tides.

We next consider the fluctuations in ORs at Honiara and Rabaul (which show some of the largest OR values), as well as the simultaneous changes in forcing tides, in MSL, and in the Multivariate El Niño Index (MEI). Figure 10 shows the time-series of the ORs for M_4 and M_6 ; AR amplitudes for M_2 , K_1 and O_1 ; MSL; and the MEI. These two gauges show the largest ORs in the region, and very large fluctuations are associated with high values of MEI (El Niño events). M_2 amplitudes are much smaller than predicted by astronomy at these gauges (~2% of gravitational potential prediction), suggesting the possibility of increased relative error. However, the coherence between the two records at several frequencies suggests that this is unlikely— M_2 at these two gauges show an $r^2 = 0.8$ correlation to each other, M_4 overtide ratios at Honiara and Rabaul show an $r^2=0.4$ correlation to each other, and M_6 ratios have a very strong $r^2=0.85$ correlation. MSL signals at the two gauges are also well correlated ($r^2=0.80$). M_4 ratios are fairly well correlated to the MEI ($r^2=0.4$ for Honiara and $r^2=0.64$ for Rabaul). M_6 at Honiara shows an $r^2 = 0.7$ correlation to MEI, while Rabaul shows a correlation of $r^2=0.65$ for M_6 to MEI. Both gauges show very strong correlations ($r^2=0.85$ for Honiara; $r^2 = 0.75$ for Rabaul) of local MSL to the MEI.

Figure 9 suggests that there is a threshold relationship between the occurrence of large ORs, local MSL, and the MEI. As shown by the yellow boxes, significant increases in ORs by a factor of $3\times$ to $10\times$ occur when the MEI is high and MSL is low, under El Niño conditions. Increased OR values during El Niño periods of low MSL could result from several factors, including increased friction over reefs and in shallow lagoonal areas. Most of Indonesia and Australia receive less rain during El Niño events (Roplewski and Halpert, 1987, Figure 21). Honiara and Rabaul are on the edge of this ENSO-induced drought region, and MSL rapidly decreases at both these gauges during El Niño events. M_4 and M_6 ORs have started to decrease in recent years at Honiara, while M_2 has shown a strong increase. This is likely due to the fact that sea level has risen rapidly in recent years, reducing bed friction.

4.3 Thermocline depth and tidal variations

Thermocline depth, MSL and tidal properties are closely linked at some, but not all, stations. Figure 11 shows EEMD decomposed time-series (seasonal scales and longer) of D_{20} from the TAO buoys (as detailed in section 3.5), along with the M_2 and S_2 tidal amplitude variations (determined from monthly admittance amplitudes); for all relevant time series, we only use IMFs #4-9 so the long-term trends are not included. This analysis is performed at Honiara, Kapingamarangi, and Pago Pago; buoys and gauge locations are shown in Figure 11a. The D_{20} variations at the three TAO buoys nearest Honiara are mutually coherent; we take, therefore, a mean of these three equidistant buoys (Figure 11b). For the other two gauges, only one TAO buoy is used to compare thermocline depths. Figure 11c and 11d show this D_{20} time-series with M_2 and S_2 variations at Honiara and Kapingamarangi; both gauges show positive correlations of thermocline depth fluctuations to M_2 and S_2 fluctuations, with r^2 being between + 0.69 and + 0.89 in all cases. For Pago Pago, the coherence between the time-series is insignificant ($r^2 < 0.1$). It is unclear whether this is the result of

the distance between buoy 5 and Pago Pago, or whether different processes are at work at this station. The thermocline, as represented by D_{20} , shoaled during the 1997-1998 ENSO event. At the same time, there was drop in tidal admittance for M_2 and S_2 at Honiara, Kapingamarangi, and Pago Pago, but available data do not provide any apparent connection of the tides to changes in thermocline depth at Pago Pago.

4.4 Spatial Summary

The long-term trends of the M_2 tide are mainly positive, while diurnals are mixed, but generally negative, in Australian and New Zealand waters. For the A-TATs, M_2 is mainly negative, while the diurnal A-TATs are all positive. Overtides, even M_4 , are generally growing, despite extensive harbor development at some locations. These changes suggest regionally driven tidal evolution involving a combination of frictional and triad interaction, perhaps tied to the reef systems in the Coral Sea. The waters around the Solomon Islands show M_2 A-TAT to be increasing, while K_1 , O_1 and the main overtide ratios are decreasing, with all variations coherent with, and likely driven by, variations in the depth of the thermocline.

4.5 Anomaly Trends Over Time

The majority of the Western Pacific stations used in this study show larger yearly MSL rise rates after 1993 than before (Table 1a; Figure 2). This rather abrupt increase in MSL rise in the southwest Pacific after 1993 provides an opportunity to test the relationship between MSL rise and tidal evolution. Thus, if TATs are sensitive to absolute MSL or rate of change of MSL, we might expect to see differences in TATs between these two epochs at gauges that show MSL acceleration. If TATs are the same in both time periods for such gauges, then they are likely independent of MSL. Thus, analyses were performed to evaluate differences in TATs before and after 1993. Twenty out of 26 stations have enough data both before and after 1993 to cover a nodal period and allow meaningful comparison. Results of the “before and after” 1993 analyses for these 20 gauges are summarized in Tables S1 (K_1) and S2 (M_2) (Online Resource 1). To illustrate spatial trends in these shifts, figures are provided for K_1 (Figure 12) and M_2 (Figure 13), analogous to Figure 4 through 7, but now the color intensity of the dots represent the *difference* in A-TATs between the two time eras.

There are strong positive K_1 A-TAT temporal shifts at three shelf gauges, and at two islands; there is a negative shifts at only one shelf gauge. M_2 A-TAT shifts tend to be generally larger than K_1 . Most shelf gauges show significant M_2 shifts, positive at four, and negative at two stations. Two island gauges show positive shifts, with only one negative result. In general, shelf gauges show larger magnitude shifts than island gauges, where it is harder to distinguish the two periods (pre and post-1993). Similar plots and tables are provided for O_1 (Figure S4, Table S1, Online Resource 1) and S_2 (Figure S5, Table S2, Online Resource 1). O_1 A-TATs shifts are also mainly positive after 1993, but S_2 A-TATs and P-TATs have shifted only slightly.

4.6 Triad Results

The Honiara and Rabaul tidal records show striking examples of what appears to be M_2 - K_1 - O_1 triad interactions (Figures 14 and 15). Figure 14 (a) shows $K_1 O_1$ vs. M_2 at Honiara; Figure 14(b) shows Rabaul. M_2 amplitudes are small, 51mm at Honiara and 40 mm at Rabaul, and their temporal variability is high (Figures 10 and 11). Diurnal tides are larger; K_1 amplitudes are 229mm and 240mm at Honiara and Rabaul, while O_1 amplitudes are 112mm and 126mm, respectively. Since we further multiply the K_1 and O_1 ARs together, the magnitude of the RHS of Eq. (13) will be much larger than the M_2 AR appearing on the LHS, meaning that the interaction coefficients, β , must be small as well as negative (Figure 14). As noted above, β can be estimated by regression, using Eq. (13); see Figure 15. If triad resonance is occurring, the variations of the summed variables (LHS and RHS sides of Eq. (13)) should be less than the variations in the LHS and RHS individually, a condition that is generally fulfilled for both stations in Figure 14, though some deviations are seen during El Niño. Also, the r^2 values of 0.59 (Honiara) and 0.72 (Rabaul) in Figure 15 suggest that resonant triad dynamics explain a substantial fraction, though not all, of the variations in the major constituents at these stations. A fraction of this energy may be lost to overtide generation during El Niño events, but direct comparison of M_4 and M_2^2 fluctuations does not show a significant correlation. Finally, the variations in AR_{M_2} are impressive: $\pm 30\%$. Thus for a small constituent like M_2 , triad interactions can lead to very large time variations, in relative terms.

Also, it is evident from Figure 16 and Table 4 that the phase-lock condition Eq. (2a) is generally approximately met for IMF#s 4-10. While $\Phi(\tau)$ is not constant, particularly during the ENSO events, it is less variable than the individual phases. For most of the record, the O_1 and K_1 PDs are negatively correlated, with M_2 making up the difference between the two. During El Niño events, in contrast, there are large M_2 phase excursions that perturb the phase lock. The sum of the variances is 2-13 $\times$ as large as the variance Φ , and the phase lock is more closely met after 1993 than before, despite very large fluctuations in the O_1 PD. Interestingly, the 1998 El Niño does not cause large excursions in the M_2 PD, even though large M_4 amplitudes are seen at Honiara.

To confirm that the variations shown in Figures 14-16 indeed represent a triad interaction, we should verify that the resonance conditions of Eq. (2) and Eq. (3) are satisfied. For M_2 , K_1 , and O_1 , the frequency condition, Eq. (2b), is satisfied by definition, and the phase-lock condition, Eq. (2a), is shown in Figure 15. There are, however, only a few wave combinations that will satisfy the resonance condition, Eq. (2c), that $\Sigma \mathbf{k}_i = 0$. For simplicity, we consider only cases in which Eq. (2c) can be satisfied in two dimensions (2D). We exclude 3D solutions on the grounds that they would require a rather special geometry and that the matching conditions would likely be transient -- small changes in thermocline depth could cause large changes in the angles between the matching wavenumber vectors. Given fixed geometry, it seems unlikely that the resonance would endure. There are only two possible 2D interactions between waves of zero or low vertical wavenumber. The most likely situation is a parametric sub-harmonic instability (PSI): a barotropic M_2 wave interacting with first-mode K_1 and O_1 internal waves. The other possibility is a second mode M_2 wave interacting with first and third mode diurnal waves, the case examined numerically by Lamb (2007).

For the PSI case, the vertical wavenumber for M_2 is $m_{M_2}=0$; and κ_{M_2} will be small relative to the κ_i for the diurnals, because barotropic tidal wavelengths are much longer than internal wavelengths. As detailed above, we find

values of the m_i/κ_i ratio for the internal waves from Eq. (11). Table 5a shows these ratios for a range of values of N of 10-100. The m_i for K_1 and O_1 will be equal and of opposite sign, while their κ_i will differ by $\sim 8\%$. An exact triad would occur when the small difference between the two diurnal κ_i matches the κ_{M_2} . For the barotropic forcing wave, M_2 , we find κ from the shallow-water relation: $\kappa = \omega/\sqrt{gH}$, where g is the gravitational acceleration, ω is the frequency of M_2 , and H is the bottom depth, taken to be $\sim 500\text{m}$ based on the average depth near Honiara. These values yield a $\kappa = 2.0 \times 10^{-6} \text{ m}^{-1}$. However, the depth in this area is highly variable and water depths of up to 5km are found within a wavelength of Honiara. If, on the other hand, the triad interaction is localized near the gauge, the relevant depth could be as little as $100\text{-}200\text{m}$. Thus, a plausible range of κ values for barotropic M_2 is ~ 0.7 to $4 \times 10^{-6} \text{ m}^{-1}$. For the K_1 and O_1 first-mode internal waves, the internal wave relation gives $\kappa = \omega/c_{ph}$, where c_{ph} is the phase speed of each wave. We use the N profile near Honiara (with $10 < N < 100$) to find a typical first-mode diurnal $c_{ph} \sim 2.7 \pm 0.1 \text{ ms}^{-1}$ for both K_1 and O_1 (Rainville and Pinkel, 2006). The resulting first-mode wavenumbers are: $\kappa \sim 2.7$ and $2.5 \times 10^{-5} \text{ m}^{-1}$ for K_1 and O_1 , respectively. The difference between these two κ values, $\sim 2 \times 10^{-6} \text{ m}^{-1}$, gives an exact resonant triad ($\Delta\kappa = 0$) for a barotropic M_2 wave corresponding to depth of 500m , as well as near-resonant triads for depths ranging from 5000 to 100m (Table 5b). Given the uncertainty in and temporal variability of N , an exact resonant triad is likely present only on a transient basis, and a near-resonant triad is probably the usual situation. It is encouraging, however, that such a simple configuration is possible in 2D, and Table 5b indicates that $\Delta\kappa$ remains small for a range of possible values of N .

There is only one 2D, low-mode M_2 - K_1 - O_1 triad that does not involve a barotropic wave: 1st ($m = \pm 1$) and 2nd mode ($m = \pm 2$) waves of the same sign, coupled with a 3rd mode wave ($m = \mp 3$) of opposite sign. To satisfy the wavenumber condition Eq. (2c), the second mode wave must be M_2 , the case studied by Lamb (2007). There are two "flavors" of this triad, because there are two diurnals (K_1 and O_1), one third mode and one first mode. Which diurnal should be chosen as the first mode depends on the fit to Eq. (2c) -- the correct choice will minimize $\Delta\kappa$. Table 5c lists the κ_i and $\Delta\kappa$ for these two possibilities. $\Delta\kappa$ is minimized by factor of 5 for a K_1 mode 3 and an O_1 mode 1 over the reverse situation, but both show a very small $\Delta\kappa$. While both the Lamb analysis and Table 5c suggest that this triad is possible, measurable surface deflections seem unlikely for an M_2 wave with $m = \pm 2$, given the small barotropic M_2 in the Solomon Sea. Thus, it seems a less likely candidate to account for the triad features seen in Figures 14-16.

5. Discussion

We observed above that fluctuations in observed tidal properties are associated with sea-level anomalies. At three stations (Honiara, Kapingamarangi, and Rabaul), semidiurnal tidal properties (influenced by both local and amphidromic scale processes) and overtide generation (a local, frictional process) are strongly correlated with anomalies in sea level and stratification. These observations leave open, however, the question of whether constituent variability is primarily related to local processes, to basin-scale phenomena, or to some combination thereof. Furthermore, it is yet to be determined if the observed tidal variability is influenced only at the at the interannual time

scale, or if there is also a combination of dynamics across multiple frequency bands, including the long-term trends of MSL. In the following discussion we attempt to identify the mechanisms behind the observed TATs.

5.1 Spatial Patterns

Here we summarize and interpret spatial patterns in selected areas, focusing on two regions that exhibit reasonably coherent changes; a) the eastern coast of Australia and the Great Barrier Reef; and b) the Solomon Sea. We will also consider not just the anomaly trends (TATs), but also the long term trends (LTTs) in tidal constituents.

5.1.1 Australian Shelf and the Great Barrier Reef

Anomaly trends for the Australian stations are relatively coherent. There is a positive trend in the A-TATs for K_1 and O_1 along the eastern Australian coast, and (with the exception of Brisbane) a negative trend to M_2 A-TATs. Stations north of Brisbane lie on the semi-enclosed Coral Sea behind the Great Barrier Reef (GBF); the regional bathymetry is highly variable, from deep ($>4000\text{m}$) to inter-tidal. The pattern of TATs suggests larger diurnal and smaller semidiurnal tides in this region as MSL rises. The large trends in A-TATs at Auckland show a heightened sensitivity of tidal dynamics to water level, though New Zealand has its own M_2 amphidrome, the dynamics are still likely connected to the Australian waters across the Coral Sea. Auckland shows a negative M_2 A-TAT and a negative LTT in M_2 , but for diurnals the A-TAT is positive and the LTT negative.

Overtide ratios (Figure 9) for M_4 and MK_3 are increasing for most Australian stations north of Brisbane, with some of the largest increases seen at Cairns ($+3\%\text{yr}^{-1}$ for M_4 , and $+4\%\text{yr}^{-1}$ for MK_3). Decreases are seen only at Townsville for S_4 and MK_3 ($-2\%\text{yr}^{-1}$) and at Bundaberg and Gladstone for M_4 ($-2\%\text{yr}^{-1}$). Overtides are typically driven by frictional interactions, and friction at harbor stations is usually reduced, not increased, by navigational development (e.g., Jay et al., 2011), which may help explain the decreasing trends at Townsville, Bundaberg, and Gladstone. These harbors are amongst the busiest in Australia, and have had an extensive history of development over the last 50 years. This would not, however, explain the overtide increases seen at other nearby Australian gauges. It is likely that the regional overtide changes are related to larger scale processes, perhaps a coupling of resonant triads and overtides. Tidal evolution at the Australian gauges on the Coral Sea may be a response in part to MSL rise over the complex topography of the Great Barrier and other reef systems (Hughes, et al, 2010; Hughes et al, 2003; Wolanski, 1994).

5.1.2. Solomon Islands: Overtide Ratios and Thermocline Depth Variations

The Honiara and Rabaul tide gauges are located in an area of water shallower ($\sim 300\text{-}700\text{m}$) than at nearby island stations. To the direct west of the island chain is the deep Solomon Sea basin, which is connected to the waters of Honiara and Rabaul via a semi-enclosed basin directly east of Papua New Guinea known as “The Slot”. Much deeper water lies both north and south of Honiara and Rabaul, but the only open-water connection to the north is a through a narrow deep channel east of the Honiara gauge. The connecting waters south of Honiara and Rabaul have very strong currents due to the motion of the South Equatorial Current (SEC) that runs through the area, both on the surface, and in the deeper thermocline layer. The local bathymetry (Figure 1) is irregular and steep, with relatively

narrow deep-water connections between neighboring deep-water basins. This region also has some of the strongest ORs for M_4 and M_6 , an order of magnitude greater than neighboring stations. The connections detailed above between the temporal fluctuations in the ORs, MSL and the El Niño (Figure 10) highlight the heightened sensitivity of tides to water level in this area. Thus, changes in friction, related to changes in lagoonal inundation or “sill depth” that connect the deep and interconnecting basins in the complex bathymetric region may be tied to changing ORs, since the large MSL fluctuations due to ENSO represent a larger relative change in water level. The M_2 A-TAT is positive at Honiara and Rabaul, while the both the diurnals show negative A-TATs. Long-term trends are negative at Honiara for M_2 , K_1 , and O_1 . At Rabaul, the long-term trends are negative in M_2 and positive for the diurnals, but this trend is only estimated up to 1997 when the Rabaul gauge was de-commissioned so a direct comparisons of LTTs will not be accurate.

The similarity of behavior between TATs and D_{20} variations (Figure 11) suggests that the semidiurnal tidal properties in this region may be dependent on local thermocline depth. These results are (at least partially) consistent with the processes inferred for Hawaii; i.e., that increased M_2 amplitude is correlated with times of deeper thermocline, due to changing internal wave phases (Mitchum and Chiswell, 2000; Colosi and Munk, 2006). In principle, changing thermocline depth could change ORs by altering internal tidal wave steepness. However, the observation (above) that changes in ORs are driven by changes in the overtides (not the parent constituents) suggests a frictional mechanism. That is, it seems unlikely that the M_2 wave would become significantly steeper at several stations without modifying M_2 amplitudes at any of them. Moreover, the deeper thermocline during El Niño does not suggest internal tide steepening. Finally, scaling relationships suggest that changes in friction associated with lowered sea levels are a more likely cause of increased ORs than internal wave steepening. In any event, evidence for a direct relationship between overtides, the thermocline, local MSL and the forcing tides can only so far be provided for the gauges near the Solomon Islands (Rabaul, Honiara, Kapingamarangi), presumably due in part to the variable topography and shallower relative water level amplifying both the overtide and thermocline mechanisms. However, lack of correlation between ORs and thermocline depth may also be attributable to the distance of the TAO buoy datasets from other gauges in the study area, e.g., near Pago Pago.

5.2 Temporal changes

Analyses of TATs before and after 1993 show that most stations on the Australian shelf exhibit larger A-TATs before than after 1993 for M_2 , and also for K_1 (Figure 12 and 13; Tables S1 and S2; Online Resource 1). No other region in this study showed such spatial coherence in the shifting of TATs. Overall, these results are locally interesting, but do not conclusively show that the rate of MSL during different epochs is directly influencing the response of the tides to MSL variability, since the majority of all gauges analyzed show more or less unchanged TATs before 1993 and after. Gauges that do show a significant shift in TATs are likely due to local dynamical factors mentioned above (overtide and thermocline variability), as well as the fact that the post-1993 period captures the 1997-1998 El Niño event, the strongest such event in the modern record, which produced the largest interannual variability in both MSL and the tidal dynamics, particularly in this region. Therefore, the comparison of the tidal variability

between these two time periods with different MSL rates is either not able to isolate the effect of long-term sea level change, or it is not the mechanism for the observed tidal variability at this time scale.

5.3 Amphidromic movements

The observed TATs for K_1 (Figure 4) are consistent with a westward, slightly northwest-ward movement and counter-clockwise rotation of amphidrome “A”, and an eastward movement and clockwise rotation of amphidrome “B” associated with positive MSL fluctuations. These combined motions would also allow an increase in the K_1 amplitudes within the Coral Sea, as observed in TATs for the area. Though diurnal A-TATs are decreasing at nearby Honiara and Rabaul, these gauges are within a semi-enclosed basin, and are essentially separated from the Coral Sea and the open Pacific. The O_1 tidal field shows a similar movement of amphidromic points as K_1 (Figure 5) Unlike K_1 , there is an additional amphidromic point off the New Zealand coast. A westward movement of this point would explain trends observed at the Australian shelf and Auckland. Thus, we conclude that changes in diurnal tides within amphidromes “A” and “B” are amphidromic scale, though modulated by other local factors at some stations.

For M_2 , (Figure 6) an eastward movement of amphidrome “B”, along with a counterclockwise rotation that expands anti-node “C” westward and southwest-ward would agree with most of the observed M_2 TATs. The long, narrow amphidrome, “A” (more or less aligned with the Mariana Trench) is not changing in an obvious pattern. The S_2 patterns (Figure 7) of A-TATs and P-TATs show some similarities to those for M_2 (e.g., at Yap, Kapingamarangi, and Noumea), but also many differences (e.g., Legaspi, Rabaul and Pago Pago). Honiara has a similar M_2 A-TAT to other regional gauges, but an opposite P-TAT. The overall pattern of change for S_2 may echo M_2 , with an amphidromic point east of Pago Pago moving westward, and a central anti-node rotating counter-clockwise. It is not surprising that the M_2 and S_2 behavior show some difference, because S_2 has a significant radiational component that is absent for M_2 (Godin, 1986; Ray, 2001; Arbic, 2005). Still there is a strong suggestion of regional coherence for the semidiurnals, as with the diurnals.

If these suggestions of amphidromic-scale tidal evolution are real, they require an explanation relevant to such a scale. The two most likely candidates are the barotropic “back-effect” mechanism suggested by Arbic and Garrett (2010) and altered stratification (Müller, 2012). The first requires continental shelf areas over which friction or depth are changing, which seems more relevant to amphidrome B, for both the diurnals and the semidiurnals. Altered stratification may have occurred over large areas in both amphidromes A and B, and there is a clear connection between stratification and tidal properties at Honiara and Kapingamarangi. At least the latter is affected by amphidrome A for both diurnals and semidiurnals. Thus altered stratification seems the more likely agent of tidal evolution in amphidrome A. We note that amphidromic movements are not a cause of the observed TATs, but are rather symptoms of tidal evolution that represent a re-distribution the tidal fields driven by other factors.

5.4 Triad Interactions

Honiara and Rabaul in the Solomon Sea show resonant triad behavior at sub-decadal and longer timescales, with some fluctuations during ENSO events. Estimates of possible wave vectors show there are multiple situations that can satisfy the resonance conditions, Eqs. (2) and (3). Triads that involve higher-mode internal tides are, however, unlikely to be observed at a coastal tide gauge. Thus, the triad interaction that we do see is possibly a parametric subharmonic instability (PSI) interaction of barotropic M_2 with first-mode internal K_1 and O_1 .

There are also frictional triad interactions in the Solomon Sea that are correlated with MSL fluctuations, likely due to decreased lagoonal depths during periods of lowered MSL (Figure 10). Thus, frictionally-driven overtides M_4 and M_6 strongly vary with MSL (and thermocline depth) even though MO_3 and MK_3 do not. This likely reflects the fact that M_2 is decreased and the diurnals (K_1 and O_1) increased by a deeper thermocline. These opposite fluctuations compensate each other in MO_3 and MK_3 , while M_4 and M_6 variations are proportional to the square and cube (respectively) of M_2 . Interestingly, the strong ENSO events, which caused large fluctuations in MSL, O_1 , K_1 , M_2 , M_4 , and M_6 individually, do not destroy the triad conservation relation Eq. (14), because of the manner in which the oscillations in individual constituents are coupled.

Resonant triad behavior at Honiara shows a difference in behavior before and after 1993, with larger excursions from stability after 1993. These excursions are caused by decreases in K_1 and O_1 amplitudes that are not fully compensated by changes in M_2 . These may be caused by the rising MSL in the area. Also, the ENSO response in 2009 and 2010 was less dramatic than in the four previous El Niños, perhaps because the most recent El Niño was weaker. Determining whether MSL rise has fundamentally altered Solomon Sea triad behavior awaits the next major El Niño.

6. Conclusions

Tidal anomalies are related to MSL anomalies via tidal anomaly trends for both amplitude and phase (A-TATs and P-TATs) over much of the western Tropical Pacific, at both coastal and open-ocean stations. Comparison of periods before and after 1993 at stations with records >40 yrs long shows considerable temporal variability. However, complex spatial patterns, differences between constituents and differences between TATs and trends make it difficult to draw conclusions regarding causes of tidal evolution. Observed trends so far suggest the existence of multiple mechanisms with spatial scales from strictly local (e.g., lagoonal depth and harbor modification) to amphidromal, as evidenced in the apparent movement and rotation of both diurnal and semidiurnal amphidromes. Comparison of the long-term trends (LTTs) in MSL and tidal properties with the short-term variability of MSL and tides, as exhibited by the A-TATs and P-TATs, do not show ubiquitous overlap of behavior. At some locations, the LTTs and TATs are correlated, but at others, anti-correlated. This suggests that multiple dynamics are active in different parts of the frequency spectrum, with temporal scales from interannual to decadal, and as of yet, no significant prediction about the future behavior of tidal constituents may be inferred from the analysis of interannual tidal and MSL fluctuations. Thus, the question posed in the title must be answered in the negative, at least for now, though further analysis with longer records from a different part of the world ocean might suggest a different answer.

Two regions were studied in detail; the Solomon Sea, and the Coral Sea area including the east coast of Australia and New Zealand. Both areas show regionally coherent changes, but with different manifestations. In both cases, there is considerable variability between stations related to local processes. To the extent that the reefs of the Coral Sea are a factor, changes in the Australia-New Zealand area may be unusual – regionally coherent but not likely applicable in other settings.

To summarize the available evidence:

1. Internal tides: The surface manifestation of internal tides, generated across steep topography, can affect major tidal constituents over the entire region investigated here, which is between the critical latitudes for both diurnal and semidiurnal tides. There has been a deepening and strengthening of the thermocline in the area (documented for 1993-2003 by Church et al., 2011), which necessarily changes internal tides and may affect surface tides (Müller, 2012). The M_2 and S_2 constituents are strongly (positively) correlated with nearby thermocline depth in the region surrounding the Solomon Islands as well as MSL, and diurnal tides show a negative (decreasing) response to increasing MSL and deepening thermocline. Thus, changing stratification appears to be the most likely agent of large-scale tidal evolution in the northern part of the study area (amphidrome A for both diurnals and semidiurnal). It may affect amphidrome B in the southern part of the study area, but we have no definite evidence of this.
2. Changing friction: Frictional interactions are most prominent for overtides but affect the main tidal species as well. A-TATs, and LTTs of the major tides sometimes reinforce each other, but also show opposing behavior at some locations. These two calculations of trends have different time scales, and are likely due to unrelated mechanisms. The similarity in trends between coastal and island stations gives some support to the hypothesis that changes in amphidrome B may be driven by continental shelf processes, which can have a back-effect upon open-ocean trends (Arbic and Garrett, 2010). Comparison of TATs for the pre- and post-1993 periods shows the trends have changed for many Australian shelf stations, including sign reversals. Changes in ORs over time also suggest a frictional component to changing tides in the region of study. Particularly over the Australian shelf, this appears to have a regional rather than local origin.
3. Resonant triad interactions: Triad interactions of the diurnal and semidiurnal tides (K_1 , O_1 , and M_2), may both be mediated by and enhanced by variations in thermocline properties (Ball, 1964). The strength of these interactions can be modulated by changing water depth (MSL), which changes bottom friction, and also the deepening of the thermocline, which changes both surface-layer depth and stratification. Triads can involve a transfer of tidal energy across multiple frequencies, and is therefore a likely candidate to connect the other variability explored in this work, as illustrated by analysis of the Honiara and Rabual in the Solomon Sea region. Whether triad interactions are relevant on a regional basis, or only locally important, remains unclear.

It is likely the above mechanisms are not altogether independent, and the situation in the Solomon Sea suggests that there is a connection between triad and frictional interactions, with M_2 energy feeding both the resonant

and frictional triads. This results in large ($\pm 30\%$) interannual fluctuations in M_2 amplitude. Variations in thermocline depth may be tied to changes in shallow water frictional interactions, because a shallower total depth makes fluctuations in the upper-layer thickness larger, in percentage terms, than at deep water stations. The amphidromic movements associated with MSL will likely have secondary regional effects as the basin-wide water levels and tidal admittances adjust to changes in sea level and in thermocline profile. Resonant triads may also help redistribute the amphidromes of each wave involved in the triad. Further work is underway to test the contribution of each of these mechanisms through modeling in coordination with analyses of satellite altimetry data, as well as investigations focusing on sub-annual (seasonal) dynamics based on complete monthly analyses.

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933 **Table 1a** Station information for tidal records used in this study; locations are shown in Figure 1^a

Station	Country	Start Year	End Year	Environment	MSL Rates ^a	Pre-1993 ^a	Post-1993 ^a
Pohnpei	Micronesia	1974	2012	volcanic	2.22 ± 0.01	-2.20 ± 0.28	4.71 ± 0.27
Majuro	Marshall Is.	1974	2012	atoll	3.91 ± 0.15	2.66 ± 0.45	7.02 ± 0.41
Malakal	Belau	1974	2012	mountainous/reef	2.98 ± 0.15	-1.56 ± 0.44	10.36 ± 0.42
Yap	Micronesia	1974	2012	mountainous/reef	1.99 ± 0.12	-2.91 ± 0.34	7.30 ± 0.34
Honiara	Solomon Is.	1975	2012	volcanic	0.88 ± 0.08	-5.15 ± 0.24	8.51 ± 0.20
Rabaul	Pap. N. Gui.	1966	1997	volcanic	-3.16 ± 0.09	-5.94 ± 0.10	N/A
Kanton	Kiribati	1949	2012	atoll	0.89 ± 0.05	0.12 ± 0.08	4.69 ± 0.29
Noumea	France	1967	2012	reef	1.37 ± 0.09	0.53 ± 0.20	5.93 ± 0.34
Saipan	N. Mar. Is.	1978	2012	mountainous	2.27 ± 0.10	N/A	N/A
Kapinga	Micronesia	1978	2009	atoll	2.82 ± .12	N/A	N/A
Lautoka	Fiji	1992	2012	volcanic	5.58 ± 0.10	N/A	N/A
Midway	USA	1947	2012	atoll	0.86 ± 0.02	-0.25 ± 0.04	5.40 ± 0.17
Wake	USA	1950	2012	atoll	1.99 ± 0.05	1.42 ± 0.07	1.12 ± 0.36
Johnston	USA	1947	2012	atoll	0.85 ± 0.05	0.51 ± 0.06	3.04 ± 0.59
Guam	USA	1948	2012	mountainous	1.32 ± 0.04	-0.89 ± 0.06	9.32 ± 0.31
Kwajalein	Marshall Is.	1946	2012	atoll	1.72 ± 0.07	0.71 ± 0.10	8.07 ± 0.50
Pagopago	USA	1948	2012	volcanic	2.13 ± 0.05	1.45 ± 0.08	6.44 ± 0.37
Brisbane	Australia	1984	2012	estuary	2.75 ± 0.10	N/A	N/A
Bundaberg	Australia	1984	2012	estuary	1.90 ± 0.08	N/A	N/A
Ft. Denison	Australia	1914	2012	estuary/ria	0.92 ± 0.12	1.06 ± 0.22	2.84 ± 0.56
Townsville	Australia	1985	2012	coastal bay	0.81 ± 0.16	N/A	N/A
Legaspi	Philippines	1984	2007	coastal bay	8.97 ± 0.42	N/A	N/A
Cairns	Australia	1985	2010	coastal inlet	2.14 ± 0.33	N/A	N/A
Gladstone	Australia	1978	2010	coastal bay	2.47 ± 0.58	N/A	N/A
Williamstown	Australia	1966	2010	coastal bay	1.39 ± 0.07	0.09 ± 0.17	2.48 ± 0.25
Auckland	New Zealand	1904	2010	coastal bay	1.48 ± 0.09	1.33 ± 0.07	3.75 ± 0.18

934 ^a -Expressed in mm yr⁻¹. N/A indicates there was not a full nodal cycle (~18.6 years) of data before 1993 available.

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937 **Table 1b** Long-term (linear) trends, with 95% confidence limits, for tidal amplitudes and phases^a

Tidal Comp:	K ₁ A-LTT	K ₁ P-LTT	O ₁ A-LTT	O ₁ P-LTT	M ₂ A-LTT	M ₂ P-LTT	S ₂ A-LTT	S ₂ P-LTT
Station	(mmcy ⁻¹)	(degcy ⁻¹)	(mmcy ⁻¹)	(degcy ⁻¹)	(mmcy ⁻¹)	(degcy ⁻¹)	(mmcy ⁻¹)	(degcy ⁻¹)
Pohnpei	-7.5 ± 2.5	-7.5 ± 1.7	4.4 ± 1.3	5.7 ± 1.1	22.0 ± 6.7	8.0 ± 1.8	18.2 ± 4.4	1.8 ± 2.0
Majuro	0.4 ± 2.6	17.7 ± 2.7	3.7 ± 1.5	23.2 ± 2.4	44.9 ± 8.3	0.8 ± 1.2	35.8 ± 3.8	4.6 ± 0.9
Malakal	15.7 ± 4.8	2.6 ± 1.3	-1.0 ± 3.0	-5.5 ± 2.8	62.3 ± 5.8	0.1 ± 1.2	29.5 ± 2.3	-0.7 ± 1.4
Yap	14.0 ± 3.2	4.1 ± 1.3	21.8 ± 4.6	16.3 ± 4.7	41.3 ± 6.1	10.5 ± 1.8	16.2 ± .31	10.6 ± 2.2
Honiara	-2.1 ± 4.9	15.6 ± 1.9	-4.8 ± 4.5	22.5 ± 9.8	-6.2 ± 5.4	17.8 ± 2.8	4.0 ± 2.7	6.8 ± 1.7
Rabaul	23.6 ± 3.9	1.0 ± 1.3	5.4 ± 3.1	0.0 ± 1.4	-13.1 ± 13.1	-0.8 ± 12.8	14.4 ± 5.6	0.8 ± 2.5
Kanton	3.0 ± 1.0	-2.4 ± 2.6	4.0 ± 0.9	6.4 ± 2.4	27.8 ± 4.8	-5.0 ± 0.8	15.3 ± 1.3	-1.5 ± 0.6
Noumea	6.9 ± 1.2	0.6 ± 0.5	0.3 ± 0.9	0.9 ± 1.2	25.3 ± 6.1	-2.5 ± 0.7	19.3 ± 2.5	-4.6 ± 0.8
Saipan	-11.2 ± 4.9	0.7 ± 1.7	1.3 ± 2.3	-0.2 ± 2.0	33.2 ± 6.7	7.6 ± 2.6	-15.8 ± 3.9	-13.4 ± 5.2
Kapinga	19.8 ± 6.3	1.7 ± 3.1	-1.5 ± 12.5	7.2 ± 5.9	49.7 ± 9.7	7.9 ± 5.7	35.1 ± 6.6	10.3 ± 5.7
Lautoka	1.8 ± 2.2	-3.5 ± 15.1	2.8 ± 2.3	-8.1 ± 8.4	36.7 ± 7.1	-5.2 ± 0.9	12.4 ± 6.0	14.1 ± 5.2
Midway	1.5 ± 0.7	0.2 ± 0.4	1.6 ± 0.5	-2.3 ± 0.5	7.4 ± 1.2	-5.5 ± 0.7	-0.3 ± 0.5	-17.2 ± 1.7
Wake	-3.4 ± 0.6	-0.7 ± 0.6	0.9 ± 0.6	2.0 ± 0.7	-11.9 ± 2.7	2.4 ± 0.5	5.7 ± 1.1	2.1 ± 0.5
Johnston	-3.9 ± 0.8	-7.0 ± 0.5	-2.0 ± 0.4	1.8 ± 0.6	2.6 ± 1.7	-4.1 ± 0.7	8.1 ± 1.4	-11.3 ± 0.7
Guam	4.7 ± 1.9	4.3 ± 0.9	-3.4 ± 1.6	3.5 ± 0.7	-0.5 ± 2.4	1.6 ± 0.6	0.1 ± 0.9	7.7 ± 1.2
Kwajalein	1.6 ± 0.6	0.4 ± 0.5	-1.0 ± 0.5	0.2 ± 0.4	-7.3 ± 1.7	0.4 ± 0.2	-0.1 ± 1.4	-1.8 ± 0.2
Pagopago	2.2 ± 0.9	0.3 ± 0.7	2.2 ± 0.9	3.8 ± 0.9	10.1 ± 2.5	-0.2 ± 0.5	1.0 ± 0.8	-1.8 ± 0.2
Brisbane	-12.9 ± 3.7	2.5 ± 1.1	-4.9 ± 3.3	6.2 ± 1.5	15.6 ± 13.6	13.9 ± 1.1	12.3 ± 7.2	9.1 ± 1.5
Bundaberg	-8.7 ± 3.1	0.3 ± 1.1	-8.3 ± 2.4	0.5 ± 1.1	2.3 ± 5.6	1.9 ± 1.2	7.9 ± 4.4	2.5 ± 1.4
Ft Denison	-1.1 ± 0.4	-0.4 ± 0.2	-0.2 ± 0.4	-0.9 ± 0.2	-27.8 ± 10.0	-2.0 ± 0.2	-5.1 ± 0.4	-3.5 ± 0.2
Townsville	3.3 ± 1.7	-1.6 ± 0.3	2.6 ± 1.2	1.1 ± 0.5	66.8 ± 3.2	-3.0 ± 0.4	9.9 ± 2.0	1.9 ± 0.4
Legaspi	-14.8 ± 6.9	0.1 ± 2.7	-28.7 ± 7.2	9.4 ± 3.8	-69.3 ± 27.6	-4.1 ± 4.1	-13.1 ± 15.1	-0.1 ± 3.4
Cairns	8.8 ± 5.7	-0.1 ± 0.8	10.2 ± 3.1	0.4 ± 0.9	49.5 ± 12.5	-1.4 ± 1.4	15.4 ± 6.0	-3.1 ± 1.7
Gladstone	-2.2 ± 3.3	0.1 ± 0.7	-8.9 ± 2.3	1.7 ± 1.1	14.1 ± 8.2	1.6 ± 0.7	57.3 ± 6.7	2.3 ± 0.8
Wllmstwn	18.2 ± 1.6	5.1 ± 1.3	7.8 ± 1.6	9.6 ± 1.7	34.8 ± 2.1	11.7 ± 1.2	6.6 ± 1.5	18.9 ± 1.8
Auckland	-1.4 ± 0.4	1.2 ± 0.3	-0.8 ± 0.3	19.5 ± 1.3	-31.7 ± 2.0	0.6 ± 0.2	-3.6 ± 0.7	1.6 ± 1.5

938 ^a Expressed as mm century⁻¹ (mmcy⁻¹) for amplitudes, or degrees century⁻¹ (degcy⁻¹) for phases. Significant values
939 are shaded grey; significant values have SNR > 2, and an absolute magnitude of >10 mmcy⁻¹ or degcy⁻¹

Table 2 Amplitude tidal anomaly trends (A-TATs) with 95% confidence limits and combined A-TATs for major diurnal and semidiurnal constituents^a.

Tidal Comp: (± 95% CI)	K ₁ A-TAT	O ₁ A-TAT	M ₂ A-TAT	S ₂ A-TAT	Comb. A-TATs
Station	(mm ⁻¹)	(mm ⁻¹)	(mm ⁻¹)	(mm ⁻¹)	(mm ⁻¹)
Pohnpei	22.9 ± 3.6	0.2 ± 2.3	-4.1 ± 12.2	-5.6 ± 3.2	13.5 ± 13.3
Majuro	7.6 ± 6.8	-17.7 ± 4.2	-21.6 ± 23.3	-0.9 ± 5.0	-32.6 ± 25.2
Malakal	46.4 ± 3.8	6.3 ± 3.4	-28.5 ± 7.2	-2.2 ± 1.5	21.9 ± 9.0
Yap	19.6 ± 4.0	3.6 ± 5.8	-39.6 ± 7.9	-9.7 ± 1.5	-26.1 ± 10.7
Honiara	-21.2 ± 6.3	-27.6 ± 4.5	65.6 ± 3.3	15.9 ± 0.8	32.7 ± 8.4
Rabaul	-24.3 ± 3.1	-8.4 ± 2.9	91.1 ± 7.5	-11.7 ± 1.8	46.7 ± 8.8
Kanton	-14.0 ± 3.6	-2.4 ± 3.7	33.0 ± 18.3	8.7 ± 2.2	25.3 ± 19.1
Noumea	23.6 ± 4.2	2.9 ± 3.3	70.9 ± 19.0	12.9 ± 3.3	110.4 ± 20.0
Saipan	-10.6 ± 8.2	8.9 ± 3.8	17.0 ± 11.1	-8.1 ± 2.6	7.2 ± 14.6
Kapinga	-18.3 ± 5.8	-20.7 ± 12.5	47.6 ± 7.7	17.4 ± 1.6	26.1 ± 15.8
Lautoka	5.4 ± 3.3	12.3 ± 2.9	21.9 ± 10.6	6.4 ± 3.8	46.0 ± 12.1
Midway	7.4 ± 3.1	-7.7 ± 2.4	6.6 ± 6.0	2.9 ± 1.1	9.1 ± 7.2
Wake	-9.0 ± 2.6	4.1 ± 2.6	-43.6 ± 11.8	-2.4 ± 2.1	-50.9 ± 12.6
Johnston	-26.1 ± 3.4	2.1 ± 2.1	-38.4 ± 8.9	-21.9 ± 2.6	-84.3 ± 10.1
Guam	-29.0 ± 4.2	-35.9 ± 2.9	-23.3 ± 5.8	1.3 ± 0.9	-86.9 ± 7.7
Kwajalein	-1.1 ± 2.6	8.2 ± 2.0	-2.7 ± 6.6	9.1 ± 2.0	13.4 ± 7.6
Pagopago	20.3 ± 3.1	2.7 ± 1.5	82.7 ± 7.7	-9.9 ± 1.1	95.8 ± 8.5
Brisbane	19.4 ± 10.2	30.3 ± 8.4	177.6 ± 28.8	15.2 ± 8.4	242.4 ± 32.8
Bundaberg	50.8 ± 10.6	1.9 ± 6.8	-32.2 ± 15.4	-9.3 ± 4.9	11.2 ± 20.5
Ft. Denison	18.1 ± 4.8	10.2 ± 4.1	-47.9 ± 11.0	-0.5 ± 1.8	-20.1 ± 12.8
Townsville	10.4 ± 11.6	1.0 ± 7.2	-4.7 ± 19.6	-6.0 ± 4.4	0.7 ± 24.3
Legaspi	-11.2 ± 7.1	-29.1 ± 6.2	-152.0 ± 21.2	26.0 ± 4.5	-166.2 ± 23.6
Cairns	15.8 ± 16.4	27.1 ± 8.0	35.5 ± 36.0	1.7 ± 6.9	80.1 ± 40.9
Gladstone	38.3 ± 10.5	6.8 ± 7.7	-34.1 ± 26.8	-23.1 ± 8.5	-12.1 ± 31.0
Wllmstown	1.0 ± 6.3	-0.1 ± 6.2	9.2 ± 8.2	9.2 ± 2.2	19.3 ± 12.3
Auckland	6.0 ± 3.0	-0.8 ± 2.3	-58.2 ± 17.2	3.4 ± 2.5	-49.5 ± 17.8

^a -All values are expressed as millimeter change in tide per meter rise in MSL (mm m⁻¹). Significant values are shaded grey, based on a SNR > 2, and an absolute magnitude of > 10 mm m⁻¹

Table 3 Phase anomaly trends (P-TATs) with 95% confidence limits for major diurnal and semidiurnal constituents^a.

Tidal Comp: (± 95% CI)	K ₁ P-TAT	O ₁ P-TAT	M ₂ P-TAT	S ₂ P-TAT
Station	(degm ⁻¹)	(degm ⁻¹)	(degm ⁻¹)	(degm ⁻¹)
Pohnpei	-18.9 ± 2.1	8.7 ± 1.6	12.1 ± 2.9	-13.9 ± 3.1
Majuro	-26.6 ± 6.4	22.0 ± 6.0	-0.3 ± 2.9	-0.2 ± 2.2
Malakal	6.7 ± 1.4	-31.3 ± 1.3	-3.7 ± 1.3	-6.5 ± 1.4
Yap	11.0 ± 1.4	57.6 ± 2.6	17.4 ± 1.8	23.0 ± 2.1
Honiara	1.8 ± 0.6	2.3 ± 7.9	28.2 ± 2.8	-16.9 ± 1.6
Rabaul	-2.0 ± 1.2	4.2 ± 1.5	-39.6 ± 10.8	3.6 ± 2.4
Kanton	-2.8 ± 10.5	32.0 ± 10.9	-7.4 ± 3.6	4.9 ± 2.6
Noumea	2.8 ± 1.9	-3.3 ± 4.3	1.0 ± 2.5	1.2 ± 3.1
Saipan	-1.6 ± 3.0	-12.9 ± 2.9	3.7 ± 4.3	-28.2 ± 9.1
Kapinga	-13.3 ± 2.8	-27.2 ± 5.4	-32.6 ± 4.5	-37.0 ± 4.0
Lautoka	23.8 ± 39.2	-10.0 ± 12.8	-2.2 ± 1.3	14.5 ± 3.4
Midway	0.8 ± 2.1	-1.1 ± 2.2	-12.3 ± 3.1	-49.8 ± 7.2
Wake	0.6 ± 2.7	10.8 ± 3.2	17.6 ± 2.1	7.8 ± 2.3
Johnston	-3.4 ± 2.9	6.0 ± 3.4	15.8 ± 3.7	-3.0 ± 3.9
Guam	18.9 ± 1.6	-4.3 ± 1.6	-1.9 ± 1.4	15.9 ± 2.8
Kwajalein	7.6 ± 1.8	-3.6 ± 1.7	-1.6 ± 0.7	-3.2 ± 0.9
Pagopago	2.9 ± 2.5	23.6 ± 3.3	17.8 ± 1.5	-6.7 ± 3.0
Brisbane	-6.9 ± 2.8	-15.1 ± 3.3	-11.9 ± 2.9	-20.7 ± 3.8
Bundaberg	-4.4 ± 4.2	0.6 ± 3.3	-3.9 ± 3.6	-6.3 ± 4.0
Ft. Denison	10.9 ± 1.9	2.2 ± 2.4	13.2 ± 2.6	17.4 ± 2.3
Townsville	6.3 ± 1.7	-0.5 ± 2.8	-0.1 ± 1.8	2.7 ± 1.7
Legaspi	3.2 ± 2.9	16.9 ± 2.7	-17.3 ± 3.5	-7.9 ± 3.2
Cairns	1.7 ± 2.3	-1.9 ± 2.6	11.9 ± 3.8	-2.3 ± 4.8
Gladstone	8.2 ± 2.2	8.7 ± 3.4	-0.9 ± 2.6	-2.1 ± 3.1
Wllmstown	-2.3 ± 5.2	-10.1 ± 6.6	-10.2 ± 5.2	-32.0 ± 6.7
Auckland	1.9 ± 2.9	22.4 ± 11.6	1.2 ± 1.4	60.3 ± 11.6

^a -All values are expressed as degree change in tide per meter rise in MSL (deg/m). Significant values are shaded grey, based on a SNR >2, and an absolute magnitude of >10 degm⁻¹. Combined trends have not been calculated for phase anomaly trends, since each forcing frequency has a different period; therefore, cumulative trends may not physical meaningful

Table 4 Phase variances and phase sum variances (of the IMF 4-10 combination) for Honiara and Rabaul. For Honiara, we calculate the variances for the entire time record as well as splitting the record about the year 1993^a.

<u>Honiara</u>	K₁ phase	O₁ phase	M₂ phase	Sum of	$\Phi(\tau)$
	variance	variance	variance	variances	variance
(1974-2012)	0.49	0.62	0.71	1.06	0.42
(1974-1993)	0.11	0.12	0.89	0.91	0.48
(1993-2012)	0.47	0.54	0.51	0.88	0.31
<u>Rabaul</u>					
(1966-1997)	0.15	0.20	1.56	1.58	0.68

^a- All variances are expressed as a percentage of a full cycle, with Φ being relative to K_1 . Values shaded grey in the last column indicate that the variance is less in the combined quantity than in the sum of the individual quantities.

978 **Table 5a** Slopes (m/κ) of the dispersion relations of M_2 , K_1 , and O_1 at for Honiara for typical values of non-
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N/ω_{m2}	(m/κ) <i>for M_2</i>	(m/κ) <i>for K_1</i>	(m/κ) <i>for O_1</i>
10	10.0	20.2	22.1
23	27.6	45.5	49.8
32	32.1	64.0	70.1
71	71.8	143.8	157.5
100	101.4	203.6	222.9

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994 **Table 5b** PSI triad wavenumber estimates for three assumed depths^a.

Depth, (m)	<i>(k,m) for M₂</i>	<i>(k,m) for K₁</i>	<i>(k,m) for O₁</i>	Δk
500	$(2.0 \times 10^{-6} \text{m}^{-1}, \sim 0)$	$(-2.7 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$(+2.5 \times 10^{-5} \text{m}^{-1}, \pm 1)$	~ 0
5000	$(0.7 \times 10^{-6} \text{m}^{-1}, \sim 0)$	$(-2.7 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$(+2.5 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$1.3 \times 10^{-6} \text{m}^{-1}$
100	$(4.0 \times 10^{-6} \text{m}^{-1}, \sim 0)$	$(-2.7 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$(+2.5 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$2.0 \times 10^{-6} \text{m}^{-1}$

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996 ^a- Calculations of wavenumbers for the case of a barotropic M₂, and a first mode baroclinic K₁, and O₁ at Honiara, Solomon
997 Islands. Three cases are shown for three depths typical of the region; the mean depth of 500m gives an exact triad (Δk). For
998 much deeper (5000m) and much shallower (100m) water, a near-resonant triad is possible Horizontal wavenumbers (k 's, and Δk)
999 are dimensional (meters) and mode numbers are non-dimensional.

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1012 **Table 5c** Baroclinic triad wavenumber estimates.

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<i>(k,m) for M₂</i>	<i>(k,m) for K₁</i>	<i>(k,m) for O₁</i>	$\Delta \mathbf{k}$
$(\pm 1.1 \times 10^{-4} \text{m}^{-1}, \pm 2)$	$(\pm 8.4 \times 10^{-5} \text{m}^{-1}, \pm 3)$	$(\pm 2.5 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$1.0 \times 10^{-6} \text{m}^{-1}$
$(\pm 1.1 \times 10^{-4} \text{m}^{-1}, \pm 2)$	$(\pm 2.7 \times 10^{-5} \text{m}^{-1}, \pm 1)$	$(\pm 7.8 \times 10^{-5} \text{m}^{-1}, \pm 3)$	$5.0 \times 10^{-6} \text{m}^{-1}$

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1015 ^a - As in Table 5b, this time for the 3-2-1 triad of Lamb (2007) (with a mode 2 M₂ wave), at Honiara, Solomon Islands. The first of
1016 the two estimates assumes that K₁ is mode 3 and O₁ is mode 1; the second assumes that K₁ is mode 1 and O₁ is mode 3.

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Figure Captions

Figure 1 Bathymetry of the western Pacific Ocean, with tide gauge stations used in this study shown as white dots and bold text for island gauges, and yellow dots and italicized text for coastal gauges; the depth scale is shown at right, in meters

Figure 2 Time series of yearly average MSL (in meters) at Guam, Pago Pago, Kwaljein, Malakal, Yap, Honiara, and Kanton, showing differences in sea level evolution before and after 1993 (indicated by the dashed vertical line)

Figure 3 Amplitude anomaly trends (A-TATs) for Honiara in the Solomon Islands: (a) K_1 ; (b) O_1 ; (c) M_2 ; (d) S_2 ; and phase anomaly trend (P-TATs) for (e) K_1 ; (f) O_1 ; (g) M_2 ; and (h) S_2 . The red bars show 95% confidence limits on each annual estimate. The green line is the robust linear regression trend, in mmm^{-1} or degm^{-1} , as shown as text, with 95% confidence limits on the anomaly trend. Significance is indicated by correlation (r_0^2) values within each subplot

Figure 4 K_1 A-TAT map, showing changes in amplitude (per m MSL rise). Map backgrounds show mean tidal amplitudes (meters) over the entire time record (color scale) and phases (solid lines), from the ocean tidal model of TPXO7.2, (Egbert and Erofeeva, 2002, 2010). Red and blue colored markers show positive and negative A-TATs, respectively. The magnitudes are indicated by color intensity, as shown by legend at the bottom, in units of mm of tidal change per meter of sea level rise (mmm^{-1}). To avoid potentially spurious results due to large percentage changes in small constituents, A-TATs are only plotted if the ratio of the 95% confidence limit of the trends has an $\text{SNR} > 2$. Statistically insignificant values are indicated by white circles

Figure 5 O_1 A-TAT map, showing changes in amplitude anomaly trends (for a 1 meter MSL rise); symbols are as in Figure 4, and units are mmm^{-1}

Figure 6 M_2 A-TAT map, showing changes in amplitude anomaly trends (for a 1 meter MSL rise); symbols are as in Figure 4, and units are mmm^{-1}

Figure 7 S_2 A-TAT map, showing changes in amplitude anomaly trends (for a 1 meter MSL rise); symbols are as in Figure 4, and units are mmm^{-1}

Figure 8 Combined A-TAT map, showing changes in amplitude anomaly trends (for a 1 meter MSL rise); symbols are as in Figure 4, and units are mmm^{-1} . Tidal magnitudes were added together as complex quantities, and the real part of the result is what is plotted

Figure 9 Overtide ratios (ORs) for: (a) M_4/M_2^2 ; (b) M_6/M_2^3 ; (c) S_4/S_2^2 ; and (d) $MK_3/(K_1 \times M_2)$; the background color scale represents the mean OR on a logarithmic scale. For stations that show a significant change over time in this ratio, numbers and colored markers indicate change in the ORs, in units of $1/\text{m}$, (or $1/\text{m}^2$ for M_6) expressed as percentage change per year

Figure 10 Time-series for Honiara (blue line with red error bars) and Rabaul (black line with cyan error bars) of: (a) M_4 and (b) M_6 overtide ratios; and detrended (c) M_2 amplitude; (d) K_1 amplitude; and (e) O_1 amplitude; and (f) MSL at Honiara (blue) and Rabaul (black); and (g) the Multivariate El Nino Index (MEI). Horizontal lines in (f) indicate approximate threshold values of MSL at the two gauges, vertical lines indicate times of ENSO events (yellow shading), when MSL is forced below this threshold. To standardize to a common water level datum, MSL is expressed as difference from the Revised Local Reference (RLR) datum (for 1983), as defined by the Permanent Service for Mean Sea Level (PSMSL), <http://www.psmsl.org/>; MSL is 6.232m on RLR at Honiara and 6.401m at Rabaul

Figure 11 (a) Map of Solomon Islands region, with locations of tide gauges and TAO buoys used for thermocline analyses; (b) time series of D_{20} at buoys 1-3 in blue, red and green, each about equidistant to Honiara, and the composite average of all three (black dotted line); (c) time series of composite D_{20} (blue), M_2 (green) and S_2 (red) amplitude variations at Honiara; (d) time series of D_{20} (blue) at buoy 4, M_2 (green) and S_2 (red) amplitude variations at Kapingamarangi; and (e) D_{20} for Pago Pago (blue) and M_2 amplitude (green) at buoy 5. All datasets represent an IMF decomposition of monthly time series, using IMFs #4-9 out of 10 so that the long term trends are not apparent

Figure 12 K_1 TAT map, showing differences in A-TATs and P-TATs between years before 1993 and after 1993 (for a 1m MSL rise); symbols are as in Figure 4-7; values are tabulated in Table 4

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Figure 14 Resonant triad interactions for Honiara (a) and Rabaul (b) in the Solomon Sea region, represented in terms of the IMF decomposition (IMFs #4-10) of the LHS (blue, $|M_2|$) and RHS (red, $\beta|O_1||K_1|$) of (13). As in Figure 9, stronger ENSO-related fluctuations are highlighted in yellow

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Figure 16 Phase-lock condition at Honiara, as in (2b), showing the IMF decomposition (IMFs #4-10) of the individual PD variations of M_2 (blue), K_1 (red), and O_1 (green), along with errors. The sum of the phase differences, Φ , is shown in black

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