ABSTRACT: An array of moorings deployed off the coast of Palau is used to characterize submesoscale vorticity generated by broadband upper-ocean flows around the island. Palau is a steep-sided archipelago lying in the path of strong zonal geostrophic currents, but tides and inertial oscillations are energetic as well. Vorticity is correspondingly broadband, with both mean and variance $O(f)$ in a surface and subsurface layer (where $f$ is the local Coriolis frequency). However, while subinertial vorticity is linearly related to the incident subinertial current, the relationship between superinertial velocity and superinertial vorticity is weak. Instead, there is a strong nonlinear relationship between subinertial velocity and superinertial vorticity. A key observation of this study is that during periods of strong westward flow, vorticity in the tidal bands increases by an order of magnitude. Empirical orthogonal functions (EOFs) of velocity show this nonstationary, superinertial vorticity variance is due to eddy motion at the scale of the array. Comparison of kinetic energy and vorticity time series suggest that lateral shear against the island varies with the subinertial flow, while tidal currents lead to flow reversals inshore of the recirculating wake and possibly eddy shedding. This is a departure from the idealized analog typically drawn on in island wake studies: a cylinder in a steady flow. In that case, eddy formation occurs at a frequency dependent on the scale of the obstacle and strength of the flow alone. The observed tidal formation frequency likely modulates the strength of submesoscale wake eddies and thus their dynamic relationship to the mesoscale wake downstream of Palau.

KEYWORDS: Ageostrophic circulations; Eddies; Nonlinear dynamics; Small scale processes; Topographic effects

1. Introduction

Vorticity wakes are often generated by energetic ocean currents flowing past islands. These wakes are of interest to the oceanographic community because they can extract momentum from the incident flow and modify water masses on short time and spatial scales. For sufficiently strong currents, eddies form and separate from the island (Heywood et al. 1990). Energetic, unstable wake eddies may dissipate locally and increase mixing close to the island (Chang et al. 2013). More stable eddies can transport isolated water masses hundreds to thousands of kilometers downstream (Caldeira et al. 2014). This variability is not well constrained in part due to nonlinearities associated with the production of vorticity. The broadband frequency content of geophysical flows introduces additional variability and presents a challenge in characterizing island wakes. In this study, a year of moored current data is used to calculate submesoscale vorticity near the northern end of Palau, where flow separation has been observed (MacKinnon et al. 2019; Wijesekera et al. 2020). The high temporal resolution of the data enables quantification of vorticity on time scales of hours to months, and insight into the physical processes of wake generation due to broadband flows.

Corresponding author: Kristin L. Zeiden, kfitzmorris@ucsd.edu

The classic paradigm for vorticity wakes is a circular cylinder in steady, unidirectional flow (Kundu and Cohen 1990). Vorticity is generated by a no-slip boundary condition and flow separation leads to recirculation in the wake. Eddies are shed when oppositely signed shear layers across the wake interact, interrupting the supply of vorticity (Bearman 1984). Downstream the wake is nonlinear and analytic solutions are not available. However, two characteristics are remarkably predictable: the frequency of eddy shedding and their downstream configuration. The former is given by the Strouhal number, $St = f_d(U/L)$, a nondimensional parameter with an empirically determined dependence on the Reynolds number, $Re = UL/\nu$ (Roshko 1954). Here $f_d$ is the (nonangular) eddy shedding frequency, $U$ is the unperturbed upstream velocity, $L$ is the diameter of the cylinder and $\nu$ is the kinematic viscosity. The configuration of wake eddies satisfies a constant spacing ratio similar to a von Kármán vortex street (VKVS), (von Karman 1912). These characteristics persist even for turbulent wakes with high Reynolds numbers (Bearman 1969).

Observational studies have routinely found eddies in the wakes of islands resembling idealized VKVS wakes (e.g., Barkley 1972; White 1973; Heywood et al. 1990; Bowman et al. 1996; Herández-León 1991; Aristegui et al. 1994; Zheng et al. 2008; Chang et al. 2013; Caldeira et al. 2014). This similarity has motivated oceanographers to use the cylinder paradigm to
extrapolate observations with limited temporal and spatial resolution to a more complete picture of the wake (e.g., Heywood et al. 1990; Teague et al. 2002). However, the broadband content of real geophysical flows challenges the utility of this comparison. Basin-scale geostrophic flows are often considered quasi-steady, but typically advect mesoscale features with significantly shorter time scales. Tides and inertial currents can also be energetic, especially in shallow coastal regions. Many studies examining oscillatory flows past small-scale headlands have shown that the relative strengths of different flow components, as well as the geometry of the headland, dictate the dynamics in the wake (Pawlak et al. 2003; Pattiaratchi et al. 1987; Signell and Geyer 1991). How these finescale, high-frequency processes relate to larger-scale island wakes is an outstanding question in the field.

This study is part of a large field campaign designed to address this gap by observing the wake generated by an island over a wide range of spatial and temporal scales. The Office of Naval Research department research initiative Flow Encountering Abrupt Topography (FLEAT) combines current observations from ship surveys, moorings, autonomous vehicles and remote sensing to characterize the wake generated by Palau, a steep sided, meridionally elongated mesoscale archipelago lying in the path of two major geostrophic currents in the tropical North Pacific. Here “mesoscale” means on the order of the Rossby radius of deformation, which is $O(100)$ km in the region (Chelton et al. 1998). A panoply of synoptic observations has captured wake eddies around the island from topographic to mesoscales. Rapid repeat ship surveys around the northern and southernmost ends of Palau observed flow separation around small-scale headlands and the subsequent generation of energetic eddies $O(1)$ km during strong tidal flows with $Ro$ up to $O(10^5)$ (MacKinnon et al. 2019; Johnston et al. 2019). Here and throughout this manuscript, $Ro$ is the Rossby number, a nondimensional ratio of relative to planetary vorticity ($Ro = \xi/f$). At $O(10)$ km, repeat box surveys around the northern end of Palau sampled larger-scale recirculation with $Ro$ $O(1)$ during strong northwestward flow (Wijesekera et al. 2020). In another study, ship surveys revealed the presence of eddies $20-40$ km in scale and with $Ro \sim O(1)$ downstream of the island concurrent with strong eastward flow (Rudnick et al. 2019). Additional larger-scale wake eddies were observed with HF radar to the east of Palau with similar vorticity during a the island concurrent with strong eastward flow (Rudnick et al. 2019). Finally, repeat glider sections over 2.5 years show persistent $O(100)$ km recirculation to the west of the island with $Ro \sim O(0.1-1)$, which develops in response to westward low-frequency geostrophic flow (Zeiden et al. 2019).

The goal of this study is to expand upon these synoptic observations by quantifying the temporal variability of vorticity close to the generation point at the northern end of Palau. While a few of the studies cited above included time series of vorticity, these have predominantly been estimates over limited time periods concurrent with the surveys. In this study, 11 months of sustained current observations from moorings enables characterization of vorticity on supertidal to subinertial time scales. The moorings were deployed at variable distances within 40 km of the northern end of the island. The full array is used to characterize the flow field, while a $\sim 7$-km triangular array close to the topographic separation point gives the vorticity at that scale. A key observation of this study is that periods of strong low-frequency flow coincide with elevated high-frequency vorticity variance, particularly in the tidal bands. This tidal band vorticity is not simply related to tidal band kinetic energy, which suggests a degree of nonlinearity in the vorticity generation process.

This paper is organized as follows: section 2 provides details on the oceanographic context, data collection, and analysis methods in this study; section 3 presents results characterizing first the flow field around the island, followed by vorticity; section 4 discusses the vorticity generation process and implications for the wake; and section 5 provides a summary and conclusions.

2. Data and methods
   a. Oceanographic setting

Palau is a meridionally elongated archipelago in the western North Pacific. The archipelago is composed of over 300 small islands which are connected by a submerged reef extending $\sim 200$ km from its northern to southern end. In this manuscript, when we refer to the “island” of Palau we mean this submerged reef. Palau lies in the path of two opposing geostrophic currents. At the northern end, the North Equatorial Current (NEC) flows between 8° and 17°N with maximum speeds of 0.5 m s$^{-1}$ (Qiu and Lukas 1996; Schönau and Rudnick 2015). On average, the NEC is westward and surface intensified, but extends down to the pycnocline centered at about 150-m depth close to Palau (Schönau and Rudnick 2015). Below the pycnocline there are weak eastward undercurrents, the North Equatorial Undercurrents (NEUCs), with average current speeds less than 0.1 m s$^{-1}$ (Qiu et al. 2013). Transport in these undercurrents is correlated with transport of the NEC (Schönau and Rudnick 2015), although this relationship weakens downstream of Palau (Zeiden et al. 2019). At the southern end, the North Equatorial Countercurrent (NECC) flows eastward at variable latitude between the equator and 7°N with maximum currents that can exceed 1 m s$^{-1}$ (Hsin and Qiu 2012).

Data presented in this manuscript were collected at the northern end of Palau, where on average the NEC and southernmost NEUC are incident on the archipelago. Here the steep-sided island boasts bathymetric gradients up to 45°, culminating in a 40-km-wide subsurface reef called Velasco. The reef rises to just under 20 m around the rim and blocks the incident flow almost completely to the surface. A concurrent FLEAT study quantified the mean mesoscale currents around Velasco and their variability using repeat glider surveys (Zeiden et al. 2019). Mean incident currents were westward and surface intensified (0.2 m s$^{-1}$ within the upper 200 m) with eastward return flow in the lee of Velasco. Significant variability on subinertial time scales was likely due to the mesoscale eddy field advected westward by the NEC (Schönau and Rudnick 2015; Hsin and Qiu 2012). Strong northward currents occurred on occasion due to recirculation of the NECC, likely spurred by westward propagating Rossby waves (Heron et al. 2006).

b. Dataset description

Fourteen moorings were deployed around Velasco over an 11 month period from April 2016 to May 2017 (Fig. 1). The
array was designed to sample high-frequency currents with adequate spatial resolution to resolve their submesoscale structure around the island. A subselection of these moorings was used in this study. Four far-field (greater than 10 km from the reef) deep-water moorings were used to characterize the ambient flow field (F1, M1, M2, and M5 deployed in 3390, 2600, 2610, and 3690 m, respectively). Three near-field (within 10 km) moorings were used to calculate vorticity, F2, F3, and F4. These moorings were deployed in 1515, 1666, and 880 m, respectively. Each had two ADCPs mounted around 60-m depth; an upward-looking 300-kHz RDI Work Horse Monitor and a downward-looking 75-kHz Work Horse Long Ranger. F2 had an additional upward-looking Long Ranger deployed at 770 m and a downward looking Work Horse at 1470 m. F3 had a downward looking Long Ranger at 760 m, and a downward-looking Work Horse at 1570 m. F4 had one additional upward looking Long Ranger at 880 m. These moorings form an equilateral triangle ~7-km scale and centered 5 km from the northernmost promontory of Velasco (Fig. 1c). Moorings were each equipped with a series of thermistors, CTDs and ADCPs. In this study we examine the ADCP data. F moorings each have ADCP coverage of over 80% of the water column at their respective depths and M moorings have coverage from the surface to 700 m. For all deep-water moorings, upward looking ADCPs deployed near the surface achieved a vertical resolution of 4 m and a minimum temporal resolution of 16 min, enough to resolve tidal motions. Velocities were smoothed and interpolated to a 10 m x 1 h grid.

To examine the depth dependence of vorticity around Palau we perform a layer-average in the surface (0–100 m) and subpycnocline (hereafter subsurface, 300–400 m). The surface layer was chosen to capture the variability of the NEC, which extends through the base of the pycnocline but decays substantially below 100 m (Schönau and Rudnick 2015). The subsurface layer similarly corresponds to the mean depth of the peak NEUCs velocities at this latitude, well below the pycnocline. An example of zonal and meridional velocity in the upper 1400 m during a period of westward surface flow at F3 gives an idea of the temporal and vertical scales included in these averages (Fig. 2). Vertically sheared geostrophic currents have length scales of O(100) m in the surface (NEC) and hundreds of meters in the subsurface (NEUCs). Energetic tides can lead to flow reversals despite the strong geostrophic westward flow. There is likely additional vertical variability on finer scales which has been
filtered by our gridding process. The surface layer encompasses the mixed layer and the upper half of the pycnocline, although ADCP data in the surface 10 m were often discarded due to sidelobe contamination. The mean mixed layer depth (MLD) during the observation period was 40 m, with a standard deviation of 12 m (obtained with CTD data from concurrent glider surveys). Correlation values between velocity at any two depths across the MLD are no less than 0.5, and mostly greater than 0.8, despite a potential dynamic shift across the base of the mixed layer. Correlation values are highest in the semidiurnal band ($R \geq 0.7$), but also strong in the diurnal band ($R \geq 0.6$). Correlation values are lowest in the inertial band, most likely due to inertial oscillations confined to the mixed layer ($R = 0.5$). Nonetheless, the main results of our analysis are not sensitive to varying the base of the surface layer from 50 to 150 m. The subsurface layer is well below the pycnocline and above the bottom boundary. Correlation values within this layer are above 0.8 in all bands due to the large vertical length scales of the NEUC and tide, and absence of energetic mixed layer currents. Correlation values between velocity at any depth level in the surface and subsurface layer are less than 0.3, which supports treating them independently.

c. Vorticity calculation

The spatial gradients necessary to compute the vertical component of relative vorticity at the center of a triangular array can be approximated by fitting a plane to zonal and meridional velocity at each time step and calculating their slopes (i.e., $\zeta = v_z - u_x$). Equilateral triangles provide the lowest error estimate of spatial gradients via this method (Davies-Jones 1993). This is intuitive because acute/obtuse triangles sample spatial gradients over different length scales in each direction and therefore give an over/underestimate of the magnitude of vorticity at the average scale of the array. The triangle formed by F2, F3, and F4 is nearly equilateral, with side lengths of 5.87, 5.82, and 5.87 km (in the clockwise direction starting at F2, Fig. 1). Together they inscribe a circle 6.8 km in diameter.

The implicit assumption when using a plane-fit to calculate vorticity is that the observed velocity variability is due to spatial gradients which are near the scale of the array. The error associated with this assumption is dependent on the relative energy at spatial scales smaller than the triangle. Without knowing the actual velocity field, it is difficult to estimate this error (Davies-Jones 1993). While KE spectra in the open ocean are typically red, it is unclear how proximity to Palau may impact this distribution. For example, MacKinnon et al. (2019) observed the generation of tidal eddies ~1 km in scale during ship surveys around the northern tip of Velasco. If these eddies were to propagate away from their generation site and cross a single mooring in the array, it would be aliased as vorticity variance at the scale sampled by the triangle. To better understand the spatial structure of currents included in our vorticity estimate, we perform an empirical orthogonal function (EOF) analysis of the velocity time series in each layer.

d. EOF analysis

Previous studies have used EOFs to identify the statistical modes with maximum covariance across an array (Rudnick and Davis 1988; Rudnick and Weller 1993; Fischer et al. 2002). In this study we calculate EOFs of velocity across the triangular array. The EOFs are orthogonal eigenvectors of the data covariance matrix. Each describe a certain percentage of the
total variance, given by their eigenvalues. The projection of each EOF back onto the data gives its time-dependent amplitude. While EOFs should not automatically be interpreted as real physical modes, there is no more efficient way to describe the velocity variance. Typically, velocity time series are reconstructed using only the most energetic EOFs. In this study we are specifically concerned with calculating vorticity, so we also consider the Okubo–Weiss value (OW) of each EOF. OW is a measure of the relative magnitude of strain and vorticity and is given by \( s^2 - \xi^2 \), where \( s^2 \) is the magnitude of horizontal strain given by \( (u_x - v_y)^2 + (v_x + u_y)^2 \) and \( \xi^2 \) is the enstrophy (Okubo 1970; Weiss 1991). This value springs from approximating the horizontal flow field around a singular point as a power series, where the first-order terms are a function of local linear velocity gradients. The sign of OW determines whether the flow field follows spiral trajectories (negative OW) or nodal/saddle trajectories (positive OW). Regions of negative OW are frequently used to identify coherent vortices in modeling and satellite data (e.g., D’Addezio et al. 2020).

To minimize the impact of finescale features on the vorticity calculation, we select for currents with spatial scales close to that of the array. First, we identify the EOFs in each layer which describe either currents in phase across the array (i.e., in the same direction), or with vertical structure near the array scale (i.e., currents out of phase, but with a clear sense of rotation characterized by negative OW). We then “reconstruct” the velocity time series by multiplying these EOFs by their respective amplitudes and adding those time series. We find that the subsequent reconstructed velocity time series retain 94% and 89% of the total velocity variance in the surface and subsurface layers, respectively. We then proceed to analyze the vorticity obtained via plane-fit to these velocity time series with confidence that our estimates are not significantly aliased by finescale motions. Further details of the EOFs are examined in the following section, as well as their individual contributions to the vorticity.

3. Results

We begin by describing the broadband flow around Velasco using the full array of moorings. We follow with a detailed analysis of the vorticity generated at the northernmost promontory where flow separation has been both predicted and observed (Simmons et al. 2019; MacKinnon et al. 2019; Wijesekera et al. 2020). The focus of this analysis is a ~7-km triangular array. We examine the vorticity time series and compare them with the velocity time series. We then examine the EOFs in an effort to understand the structure of currents which contribute to the vorticity. Finally, we present scalograms of vorticity to quantify its time-varying spectral content.

a. Broadband currents around Palau

Surface currents around Velasco are broadband in frequency, as seen in time series of surface velocity (0–100 m) from a selection of the moorings counterclockwise around the reef: M2, M5, F3, F1, and M1 (Fig. 3). In the mean, M2 is upstream of Velasco and M1 is downstream. Four strong subinertial current events dominate the variance and show topographic blocking of large-scale flows: strong eastward flow in August (absent downstream at M2, marked with red boxes) and strong northwestward flow in September, December, and March/April (absent downstream at M1, marked with blue boxes). Strong northward flow to the east of Velasco precedes the fourth westward event, appearing first at M2 and M5 before transitioning to north-westward flow. Significant high-frequency variability frequently leads to flow reversals.
west of Palau, Zeiden et al. (2019) found the time-mean incident westward flow diverges at ~8.2°N. The current above that latitude gains a strong northward component as it circumvents the island. This effect can be seen clearly in average surface velocity vectors during the westward flow event in December (Fig. 1b). A fourth event in early March begins as strong northward flow on the eastern side of the island (at M2 and M5) and then transitions to westward flow. Subinertial velocities are well correlated across the array (e.g., $R = 0.6$ between F3 and the other moorings), suggesting they have spatial scales of at least 60 km. This is consistent with glider observations which indicate that subinertial currents often have spatial scales exceeding 100 km. Weak downstream currents are further evidence of topographic blocking. For example, the strong eastward current in July appears in velocity data at M1, F1, F3, and M5 but does not appear downstream of Velasco at M2 (orange shaded boxes in Fig. 3). Similarly, currents are negligible downstream at M1 during westward flow (blue shaded boxes in Fig. 3). Thus, topographic blocking reduces the strength of large-scale, subinertial currents by almost an order of magnitude.

Subinertial currents are weaker in the subsurface layer (300–400 m), with strong events reaching ~0.25 m s$^{-1}$ (Fig. 4). One sustained westward flow event spanning August and September 2016 stands out in the time series of M2 and M5 (orange shaded boxes in Fig. 4), but otherwise high-frequency currents are the dominant component of the variance. Note that although the northward current is actually stronger than the zonal component here, concurrent glider observations show the incident large-scale subinertial flow is westward during this time (not shown). None of the five subinertial current events observed in the surface layer have a signature at depth, consistent with a weakened relationship between the NEC and NEUC close to the island (Zeiden et al. 2019). Correlation in space is divided between moorings to the east and west of Velasco (Fig. 4). For example, subinertial zonal currents at M5 are strongly correlated with those at M2 (both east of Velasco, $R = 0.7$) but are uncorrelated with those at F1 (west of Velasco). Interestingly, subinertial subsurface velocity at the tip of the reef is correlated with currents both to the east and west of Velasco ($R > 0.6$ between F3 and all other moorings).

In both layers, high-frequency currents are energetic enough to induce flow reversals except during the strongest subinertial current events. Their contribution to the total velocity variance is quantified with surface and subsurface kinetic energy spectra at the moorings in the triangular array (Figs. 5a,b). Subsurface KE variance is higher at F4 because it is shallower and therefore has stronger depth-mean stratification. Spectra are red with broad peaks around the local inertial ($f$) frequency, and sharp peaks at the diurnal (D1) and semidiurnal (D2) frequencies. These bands account for 10%, 9%, and 12% of the total variance in the surface layer, and 12%, 14%, and 32% in the subsurface layer. The greater percentage of KE variance contained in these bands at depth (>50%) reflects the weak subinertial currents there.

At the surface, tidal currents are highly irregular. A scalogram of array-mean surface KE shows sharp increases in tidal variance that occurs on time scales of about a week (Fig. 5d). The most energetic tidal event occurs in late July, coincident with strong subinertial eastward flow. In the subsurface, tidal KE variance is comparatively stationary (Fig. 5f). A strong spring–neap cycle is apparent in the semidiurnal band, and there is weak intermittency in the diurnal band. This spring–neap cycle reflects the barotropic tide (i.e., depth average, not shown). It is likely internal waves contribute more to the irregularity in surface tidal KE. In another FLEAT study, Voet et al. (2020) observe the formation of lee waves from a combination of subinertial and tidal flows over the ridge north of Velasco. However, their observations suggest it is unlikely these waves contribute appreciably to upper-ocean velocity. In an earlier study of Palau, Wolanski et al. (2004) observed isotherm shoaling at tidal periods. These internal tides were
synchronous the local spring–neap cycle, suggesting they were locally generated. In our observations, strong peaks in tidal surface KE are not synchronous with the local spring tide (seen in Fig. 5f). Changes in the local stratification could cause detuning between the barotropic tide and locally generated internal tides, but remotely generated internal tides may better explain the observed variance. To the west of Palau, the Luzon Strait is a known generation site of strong internal tide beams which may encounter Palau during favorable current conditions (Zhao et al. 2013). Propagation speeds of mode-1 internal waves in the tropical western North Pacific are \( \sim 3 \text{ m s}^{-1} \) (Rainville and Pinkel 2006), which corresponds to a lag between the Luzon Strait and Palau on the order of 1 week. The arrival of such remote internal tides would not likely have a regular pattern that corresponds with the local spring–neap cycle.

There are similar irregular peaks in the inertial band at the surface, most notably twice in November and once in early March. This intermittency is consistent with wind forced inertial oscillations, which are sporadic by nature. In another FLEAT study, Voet et al. (2020) characterize the inertial currents around Palau using moorings from both the northern and southern ends of the island. Although only one inertial event is easily explained using local wind data, they argue that the large decorrelation length scales of the other inertial events suggest they are wind generated. At depth, near-inertial internal waves likely contribute to the variance as well. In early November there is a weak peak around \( \sim 2 \text{ days} \), as well as a

FIG. 5. (a) Surface and (b) subsurface kinetic energy spectra at F2 (green), F3 (red), and F4 (blue) are red, with strong peaks in the inertial, diurnal, and semidiurnal bands. The dotted gray line in (a) and (b) gives the Garrett–Munk internal wave spectrum for reference. (c) Time series of array-averaged surface velocity and (d) a corresponding scalogram of kinetic energy reveal that energy in these bands is nonstationary in the energetic surface layer. Conversely, (e) subsurface array-averaged velocity and (f) corresponding kinetic energy scalogram have steady tidal energy, modulated only by spring–neap variability.
smattering of energy in the superinertial, subtidal band in September and October. Andres et al. (2020) observed a similar signal in acoustic travel time records from pressure-sensor-equipped inverted echo sounders (PIES) around the island. This signal propagated through their array in the southeast direction, and they concluded near-inertial internal waves likely propagate into the region from higher latitudes.

b. Submesoscale vorticity time series

Here we present time series of submesoscale (~7 km) vorticity obtained via plane fit to the EOF filtered velocity time series from the triangular array (see section 2 for methodology). Because the reconstructed time series retain over 94% (89%) of the total velocity variance in the surface (subsurface) layer, we present the time series first and then the EOFs to help interpret the signals in them. Here and throughout the manuscript we often refer to “shear,” by which we mean lateral shear. When discussing vertically sheared currents, we use “vertical shear.”

Vorticity in the surface layer has a clear relationship to the subinertial flow. Magnitudes are $O(1)$, reaching up to $Ro \sim 6$ during westward flow events (Figs. 6d and 7d). This is consistent with synoptic observations in the same region obtained by Wijesekera et al. (2020) during the period of strong westward flow in early December. Over the course of a week they conducted repeat ship surveys in box patterns $O(10)$ km around the northern end of Velasco. These surveys captured the flow accelerating and separating near F4, with strong recirculation immediately to the west. The recirculation was $\sim 5$–10 km in scale and had vorticity $O(1)$ (see their Figs. 10 and 11). This suggests that vorticity here also reflects flow separation and recirculation at the scale of the array. We will show this interpretation is supported by the EOFs (section 2c).

Subinertial vorticity is highly correlated with subinertial velocity, especially meridional velocity at F3 ($R = 0.9$). During the periods of energetic westward flow, vorticity increases from zero to $Ro \sim 5$. An $\sim 0.1 \text{ m s}^{-1}$ increase in velocity corresponds to an increase in vorticity of $Ro \sim 1$. It is noteworthy that the flow event with greatest total kinetic energy does not have a strong vorticity signature (eastward flow in late July). The reason for this is evident in the relative lack of array-scale shear, which distinguishes it from the other events.

Surface vorticity is highly broadband, and a novel observation of this study is that high-frequency vorticity variance increases substantially during strong westward flow. This can be seen by comparing the range of high-frequency vorticity prior to and during the westward event in December (Fig. 6d). Initially, values range from $|Ro| \leq 1$. At maximum flow, this increases to $|Ro| \leq 4$. We define a low-frequency envelope for this variance by 10-day low-passing the magnitude of superinertial vorticity (not shown). This envelope is strongly correlated with subinertial velocity, especially meridional velocity at F3 and F4 ($R = 0.8$). The relationship between high-frequency vorticity and corresponding low-frequency envelopes of high-frequency velocity is weaker, with $R \leq 0.5$ at all moorings other than F3, where $R = 0.6$.

Similar relationships between the subinertial current and vorticity exist in the subsurface layer as well (Fig. 7). Subinertial vorticity is strongest during the westward flow event in September and October, and high-frequency vorticity variance is nonstationary. Recall that close to the topography, currents become rectilinear and the meridional component dominates during westward flow (Fig. 4). For most of the time series, subinertial vorticity is negligible and high-frequency vorticity is $|Ro| \leq 1$. During the westward flow event this increases to $|Ro| \leq 4$. The correlation between high-frequency vorticity and subinertial velocity is weaker than at the surface. The strongest correlation is with meridional velocity at F4 ($R = 0.7$), but otherwise less than 0.5. It is clear that the strength of the current is not the distinguishing factor leading to elevated vorticity. Instead, strong shear in the meridional component across the array separates this period from the rest. As with the surface layer, the period with greatest kinetic energy (here January) does not have a vorticity signature due to a lack of array-scale shear.

![Fig. 6. Time series of array-mean surface (a) zonal velocity, (b) meridional velocity, (c) kinetic energy, and (d) vorticity. Velocity time series were reconstructed using EOFs (section 3b). Bold lines are 10-day low passes of the respective time series, except for (a) and (b), which are 10-day low passes at each of the three moorings in the array. Subinertial vorticity varies linearly with subinertial velocity during westward flow. A key result of this study is that high-frequency vorticity variance also increases during strong westward flow. Eastward flow does not have a strong vorticity signature.](image-url)
c. Dominant statistical modes

The EOFs of velocity used to filter out unresolved spatial variability give the horizontal structure of currents around Velasco and their contributions to the vorticity (Fig. 8). At the surface, the most energetic EOF accounts for 72% of the total velocity variance and reflects large-scale flow around the island (Fig. 8a). Current vectors are in phase with similar magnitude. Velocities are strongest at F4 where the flow is forced to accelerate around the promontory, and weakest in the lee at F2 where it must decelerate. Because its vorticity is the result of shear rather than a coherent vortex, the OW value of the first EOF is weakly positive (Fig. 8b). The second most energetic EOF contains 17% of the total velocity variance and is similar in character to the first EOF. Together these EOFs describe large-scale currents forced to circumvent the island from either direction, producing shear at the array scale. This shear contains 36% and 13% of the total vorticity variance, respectively. The third most energetic EOF contains just 5% of the total velocity variance, but nearly half (45%) of the total vorticity variance. Current vectors rotate counterclockwise around the center of the triangle with similar magnitudes, and the OW value is strongly negative. This suggests the third EOF describes vortex motion at the scale of the array, which likely reflects a combination of attached recirculating flow and fully separated eddies.

The EOFs of subsurface velocity have a similar variance distribution and similar structures (Figs. 8e–h). The two most energetic EOFs account for 64% and 18% of the total velocity variance and describe large-scale flow around the island. The third EOF accounts for only 7% of the velocity variance and describes eddying motions at the scale of the triangle. This EOF is responsible for a bulk of the vorticity variance. While the first two EOFs account for only 5% and 4% of the total vorticity variance, the third EOF accounts for 75% of the total vorticity variance. An important difference between the surface and subsurface EOFs is that vectors of the first subsurface EOF follow the topography. Additionally, the second EOF describes flow from the east but has a westward component at F4 and therefore significant convergence. These two dissimilarities suggest that the large-scale subsurface flow often does not separate, reflected in the negligible vorticity of the first two EOFs.

Both in the surface and subsurface layers, the three least energetic EOFs are dominated by strain and divergence. We interpret these EOFs as currents which are not properly resolved by the triangle and they were not included in the reconstruction of velocity as a means of scale filtering.

d. Nonstationary vorticity variance

The EOFs enable us to decompose vorticity into contributions from 1) currents in phase across the array and 2) eddy motion at the scale of the array (Fig. 9). Our goal in this section is to gain physical insight into the processes leading to vorticity generation around Palau, i.e., to understand which motions are responsible for the observed nonstationary vorticity variance. For both layers, we define a “shear” component of vorticity as due to the first two EOFs (in phase) and an “eddy” component of vorticity as due to the third EOF (eddy motion). However, it is important to keep in mind that these are statistical modes, not true physical modes, and the real currents are a linear superposition of them (plus a mean).

In the surface layer, shear and eddy components contribute 49% and 45% to the total vorticity variance, respectively. This equitable distribution is reflected in their individual time series (Figs. 9a,b). Both components range up to $|Ro| \leq 4$. However, their frequency content differs substantially. Shear vorticity contains greater relative subinertial variance, while eddy vorticity has greater relative superinertial variance. This is evident from the time series but is quantified in their corresponding power spectral densities (Fig. 9e). The spectrum of shear vorticity closely resembles kinetic energy, with strong inertial and tidal peaks (gray line in Fig. 9e). Eddy vorticity has similarly strong tidal peaks, but they are less prominent due to the broadband elevation in high-frequency variance. It is clear from the
surface time series that nonstationary high-frequency variance is primarily due to the eddy component of vorticity (Fig. 9a). The range of values increases abruptly from $|Ro| \approx 1$ up to $\approx 4$ at the end of November, concurrent with the onset of strong westward flow. There is a corresponding low-frequency increase in the shear component of vorticity. High-frequency variance in the shear component is comparatively steady.

In the subsurface layer, the shear component contributes only 9% to the total vorticity variance while the eddy component contributes 75%. As a result, the eddy vorticity time series closely resembles the full vorticity time series (Fig. 7d). The eddy component ranges up to $|Ro| \approx 1$ for most of the record but increases up to $|Ro| \approx 4$ in September and October during westward flow. Shear component vorticity is characterized by a weak spring–neap cycle and strong peaks in the tidal bands, with close adherence to the spectrum of KE. As in the surface layer, eddy vorticity has greater relative variance at superinertial frequencies and as a result its tidal peaks are less prominent.

As with KE, we examine the spectral distribution of total vorticity as a function of time using scalograms (Figs. 10 and 11). We do not show scalograms of each component because our goal is to quantify the temporal variability of the true vorticity around Palau. However, we note that the scalogram of

![FIG. 8. Empirical orthogonal functions of velocity in the (a) surface layer (0–100 m) and (e) subsurface layer (300–400 m). Colored contours in (a) and (e) are 100-m interval isobaths from 0 to 1000 m, and the black contour is the depth of the EOFs. (b),(f) Corresponding normalized magnitudes of vorticity, strain, divergence, and OW value. (c),(d),(g),(h) The percentage of the total variance described by each EOF. In both layers the two most energetic EOFs (red and blue lines and vectors, respectively) describe currents in phase with similar magnitude. The third EOF (magenta lines and vectors) describes vortex motion at the scale of the triangle. The three least energetic EOFs are characterized by strong relative divergence and strain. Subsequent vorticity time series are calculated using only the first three EOFs.](image-url)
shear vorticity nearly mirrors KE. Differences between the scalograms of total vorticity and KE reflect nonstationary variance in the eddy component (as noted above). At the surface, the scalogram reveals that elevated variance during westward flow is the result of strong peaks in both tidal bands, especially the diurnal band in December and March (Fig. 10c). Energy increases at supertidal frequencies as well. Peaks in near-inertial variance appear to loosely coincide with the waxing and waning of the subinertial current. At depth, there are clear peaks in the tidal and inertial bands, but the increase in high-frequency variance is more broadband than at the surface. This is predominantly due to elevated variance in the superinertial, subtidal band, especially a strong peak at a period of ~2 days in early November. As mentioned in section 3a, this may be related to superinertial waves observed by Andres et al. (2020).

These vorticity scalograms differ substantially from corresponding KE scalograms (described in section 3a, see Fig. 5). Surface tidal KE is irregular, and its low-frequency modulation only loosely coincides with westward flow. For example, KE in both bands increases in mid-October, over a month before the...
onset of westward flow in late November. In contrast, tidal vorticity variance only increases during strong westward flow. Also, peaks in tidal band KE mostly do not coincide with peaks in tidal band vorticity. At depth these differences are starker.

Superinertial KE is dominated by a steady spring–neap cycle in the semidiurnal band, but vorticity is entirely event driven. These differences reflect the asymmetric distribution of velocity and vorticity variance between the shear and eddy components.

FIG. 10. Time series of surface (a) subinertial velocity at all three moorings and (b) vorticity, along with (c) the scalogram of vorticity gives time variability of its frequency content. Bold lines are 10-day low passes of each time series. Scalograms are normalized by the total variance contained at frequencies higher than (10 days)\(^{-1}\). High-frequency vorticity variance is highly episodic and strongly correlated with the subinertial flow. Increases in surface tidal variance occur during strong subinertial westward flow in December and March. Kinetic energy (Fig. 5d) in the tidal bands is comparatively stationary, and the strongest peaks do not coincide with those in vorticity.

FIG. 11. As in Fig. 10, but for the subsurface layer. High-frequency vorticity variance increases dramatically in September and October during a subinertial current event characterized by strong shear. This is dissimilar to kinetic energy at depth, which has a strong spring neap cycle (Fig. 5f).
The statistical modes which are responsible for most of the KE variance contain vorticity in the form of lateral shear. Therefore, the relationship between shear vorticity and KE is approximately linear across all frequencies. However, most of the vorticity is due to eddy motions, which increase nonlinearly in the tidal bands during strong westward flow events.

4. Discussion

In this study we have used velocity data from moored ADCPs to characterize the broadband current as well as submesoscale vorticity (~7-km scale) around the northern end of Palau. We have examined properties of the flow in both the shallow surface layer (0–100 m) and a subsurface layer (300–400 m). We find that flow in the surface layer is broad band, composed of large-scale subinertial currents and strong tidal and inertial currents (Figs. 3 and 5). Subinertial currents are topographically blocked and downstream currents are weak. Oscillatory flows are highly intermittent, likely in part due to internal tides as well as wind-forced near inertial currents. At depth, subinertial currents are weaker and a single northward flow event dominates the variance (Figs. 4 and 5). For most of the record, however, high-frequency currents are stronger than the subinertial flow. Tidal kinetic energy is steady, dominated by a clear spring–neap cycle in the semidiurnal band.

A novel observation of this study is that vorticity variance increases by up to an order of magnitude in the tidal bands during periods of strong westward flow (Figs. 10 and 11). Intermittency in tidal band KE does not easily explain this variance, pointing to nonlinearity in the wake generation process. Empirical orthogonal functions (EOFs) of velocity reveal that although 80%–90% of the velocity variance is due to large-scale currents in phase as they flow past the island, 45%–75% of the vorticity variance is contained in eddying motions which contribute only minimally to the total velocity variance (~4%–7%). Spectra of vorticity due to these eddy motions are characterized by strong peaks in the tidal bands, and scalograms show this variance increases during strong westward flow (Figs. 10c and 11c). These observations suggest that eddies are formed (and possibly shed) at tidal frequencies, but that the strength of their vorticity varies with the subinertial current.

In this section we discuss the physical mechanism likely responsible for this nonlinearity. We argue that the broadband content of the flow modulates the strength of vorticity as well as the rate at which eddies are formed. In light of significant intermittency of tidal KE at the surface, we also consider the potential contribution of internal waves to the observed vorticity. Finally, we speculate on the relationship between submesoscale eddies and the island scale wake.

a. Nonlinear broadband eddy formation

The observations detailed above are consistent with wake eddies generated by a combination of oscillatory and steady flows past a headland. The source of vorticity in the recirculating wake is shear across the upstream boundary layer, which is a function of total incident flow strength. We see this dependence reflected in the EOFs with shear in the cross-shore direction (Fig. 8). These EOFs contain most of the KE variance and thus shear vorticity is strongly correlated with the total velocity ($R = 0.7$ in both layers, Figs. 9a,c). This attached recirculating flow forms eddies if the flow reverses and generates an opposing shear layer inshore of the wake which disrupts the upstream supply of vorticity (Signell and Geyer 1991; Pawlak et al. 2003). In our observations, tidal currents are often strong enough to either arrest or reverse the low-frequency flow (Figs. 3 and 4). Thus oscillatory currents force the formation of eddies, while low-frequency currents modulate their vorticity. An example snapshot of surface velocity vectors at both flood and ebb tide during strong westward flow encapsulates this process (Fig. 12). The velocity is decomposed into the shear and eddy components as discussed in section 3d. At flood tide, shear is generated by the superposition of strong subinertial and tidal currents. Later at ebb tide, the zonal component of velocity close to the reef at F2 has reversed, introducing oppositely signed shear and generating an eddy near the scale of the array. We note that elevated tidal vorticity variance occurs in this study when the free-stream flow does not fully reverse, but is substantially weakened by the opposing tide (Fig. 6). Weakened or slack currents may be sufficient to allow an inshore shear layer to grow due to friction between the attached recirculation and the headland (Davies et al. 1990).

Two recent synoptic observational studies off the coast of Taiwan provide further support for this interpretation. Using a combination of drifter and satellite data Cheng et al. (2020) captured the evolution of three successive submesoscale eddies generated by strong northward flow past the southern cape of Taiwan. These eddies appeared to be generated at a diurnal frequency with similar scales and vorticity to those observed in this study. Using a realistic model of the region, the authors show that eddies are indeed likely shed at tidal frequencies with vorticity modulated by the tide and subinertial current strength (see their Fig. 6). At smaller scales, synoptic surveys Chang et al. (2019) captured the generation of eddies by the flow of the Kuroshio past Green Island, a 7-km island northeast of the cape. These synoptic observations suggest eddies shed close to a diurnal period, and they observed strong tidal peaks in measurements of velocity from two downstream moorings. However, they estimate that the natural shedding frequency of the island (i.e., the Strouhal period) was also close to a tidal period, and so there was a high degree of uncertainty in their estimate of the wake period.

In our case, the observations are inconsistent with eddy shedding at a natural frequency. Although difficult to predict due to its dependence on the Reynolds number, we can estimate an upper limit for this frequency. Maximum values of St are 0.45 for Re $> 10^5$ (Bearman 1969). Mean length scales in the wake on these longer time scales reflect the width of Velasco, so $L \sim 70\text{km}$ (Zeiden et al. 2019). Peak velocities observed during westward flow are 0.5 m s$^{-1}$ (Fig. 3). This gives a maximum shedding period of 86 h, many times either the diurnal or semidiurnal tidal period. We also consider the possibility that eddy–eddy interactions could modulate the near-field vorticity, as in Callendar et al. (2011). However, this
modulation would likely reflect a spring–neap cycle (see their Fig. 6). In our observations tidal vorticity is not modulated by a spring–neap cycle, even at depth where the spring–neap forcing is strong.

b. Internal waves and vorticity variance

In the simple case of regular tides (i.e., barotropic) plus subinertial currents, the latter is responsible for modulating the eddy strength. At depth this explains most of the vorticity and velocity variance. Subsurface semi-diurnal tidal currents dominate tidal KE variance with a clear spring–neap cycle (Fig. 5f).

Tidal band vorticity is weak and steady, except during a period of subinertial westward flow which generates array scale shear (Fig. 7). This simple relationship is reflected in the scalogram of subsurface vorticity (Fig. 10e). At the surface, however, there is significant intermittency in tidal band KE (Fig. 5d). This is likely due to internal waves, which are not steady in time. Here we consider how internal waves may contribute to non-stationary tidal vorticity variance, whether directly via aliasing or indirectly via modulation.

First, we consider remotely generated internal tides which may propagate into the region. The horizontal length scales of
freely propagating mode-1 internal tides in the tropical North Pacific are $O(100)$ km (Zhao et al. 2013). At these scales remote internal tides would likely appear at the array as currents in phase with almost uniform magnitude. Our EOF analysis has shown that tidal band vorticity is mostly due to rotational eddy motion across the scale of this triangle. Therefore, remote internal tides with these characteristics do not likely contribute directly to eddy vorticity. However, they may modulate the strength of boundary layer shear in a manner analogous to the subinertial current. There are several instances which match this description for internal tides. For example, in late December/early January KE variance increases sharply in the diurnal band (Fig. 5d). Tidal velocities increase simultaneously at far-field moorings F1 and M5, suggesting large length scales (Fig. 3). This is one of the few times when tidal band vorticity variance increases in both the shear and eddy component (not shown). The diurnal KE event in mid-March has similar characteristics, and notably both events occurred during strong westward flow. Two additional tidal KE events fit the description of an internal tide (abrupt increases in KE variance, large length scales), but neither have strong vorticity signatures. The first occurred in July during strong eastward flow, and the second in November when the subinertial current was weak. This may suggest the combination of tidal and westward flow is essential to generating eddies, or this may reflect the location of the array just west of the separation point (Fig. 1c).

Locally generated, high-mode internal waves with small horizontal scales have the potential to alias directly into the observed eddy vorticity. These waves might be perceived to contain vorticity, but only if they have a wavelength close to the scale of the array. Another FLEAT study observed the generation of internal waves over the ridge extending north of Velasco, at the same latitude as F3 (Voet et al. 2020). These waves were dependent on the combined strength of the subinertial current and the semidiurnal tide, which is the dominant component of the local barotropic tide (Fig. 5f). However, the observations of Voet et al. (2020) suggest it is unlikely the waves propagate through the strong vertical shear at the base of the surface layer. Further, high-mode internal waves do not typically contain much energy (Pinkel 1984). Finally, as previously noted, the observed variance in tidal band KE is weakly modulated on subinertial time scales but does not align closely with the local flow. Thus, it seems unlikely that local internal waves contribute significantly to the observed eddy vorticity.

c. Implications for the Palau wake

Here we speculate on implications for the Palau wake. Submesoscale eddies observed in this and other FLEAT studies are $O(10)$ km. This is consistent with advective length scales estimated from typical velocities over a tidal period during strong westward flow (Fig. 5c). Given $U \sim 0.25$ m s$^{-1}$, eddy length scales are $L \sim 10$ and 20 km for the semidiurnal and diurnal tidal period, respectively. If vorticity is similarly approximated as $\zeta \sim U/L$, these eddies have $Ro \sim O(1)$. This is consistent with vorticity observed in this and the synoptic study by Wijesekera et al. (2020). In the case of steady flow, wake eddies are proportional to the island scale (Heywood et al. 1990; Dong et al. 2007). Were eddies to develop on subinertial time scales to this scale [$O(100)$ km] they would have $Ro \sim O(10^{-1})$. Thus, when tidal currents control the formation rate they likely reduce the scale and increase the vorticity of wake eddies by an order of magnitude. FLEAT observations suggest both an island scale wake and submesoscale eddies are generated around Palau. Mean currents downstream of Velasco show island scale recirculation with $Ro \sim O(10^{-1})$ at the surface (Zeiden et al. 2019). This wake pattern is often visible on time scales as short as a few days, but there is significant variance. Wake vorticity occasionally reaches $Ro \sim O(1)$ with 10-km scales and smaller, possibly the result of smoothing over many smaller eddies.

Submesoscale eddies likely relate dynamically to the island scale wake in one of two ways. First, they may act as agents of momentum transfer. This is the process which motivates the use of an “eddy” viscosity in lieu of molecular viscosity when comparing island wakes to idealized wakes (e.g., Teague et al. 2005). For example, Delandmeter et al. (2017) observed the development of finescale eddies in the wakes of shallow islands in the Great Barrier Reef. They argued these eddies control momentum transfer between the outer flow and the wake by enhancing the viscosity of the shear layer. Second, submesoscale eddies may behave like two-dimensional turbulence and merge, undergoing an inverse cascade to larger scales. A numerical study by Molemaker et al. (2015) depicts submesoscale anticyclones in the California Undercurrent coalescing until stabilizing near the local Rossby radius of deformation. In another study by Callendar et al. (2011), eddies generated by tidal flows coalesce on spring–neap time scales to larger scales. In these scenarios, submesoscale eddies are a direct source of island-scale wake vorticity. To what degree either (or both) of these processes influences the development of an island scale wake behind Palau is an open question.

5. Conclusions

In this study we have used 11 months of velocity observations from moorings deployed around a steep-sided mesoscale island, Palau, to quantify submesoscale vorticity generated by broadband currents circumventing the topography. In the energetic surface layer there is commensurate energy in tidal, inertial and low-frequency flows. Vorticity as measured by the Rossby number is $O(1)$ and similarly broadband, but its relationship to the flow is nonlinear. During strong westward flow vorticity increases by almost an order of magnitude in both the subinertial and tidal bands. Empirical orthogonal functions (EOFs) of velocity reveal that this increase in tidal vorticity during strong low-frequency flows is due to eddying motions at the array scale ($\sim 7$ km). In contrast, lateral shear across the array is linearly related to velocity. We argue this reflects the linear superposition of currents which generates shear in the boundary layer. This shear is the source of eddy vorticity in the wake, but high-frequency oscillatory currents control the rate of eddy formation. Thus, eddies likely increase in strength during strong subinertial flow, but are formed at a tidal frequency and so contribute to vorticity variance in the tidal band. This is unlike the case of steady flow, in which eddies are
formed at a frequency dependent on the island scale and current strength alone.

The results of this study contribute to a growing body of research aimed at understanding how the broadband content of real geophysical flows influences the development of wakes. We have shown that energetic oscillatory currents modulate the formation of submesoscale eddies, but it is unclear how this impacts the island scale wake. We have discussed momentum transfer and an inverse cascade as two possible mechanisms by which submesoscale eddies may modulate or contribute directly to large-scale vorticity. These processes require eddies to be stable downstream of the separation point. However, eddies with $|Ro| > 1$ and strong baroclinicity may decay in the near field. Anticyclonic eddies with $|Ro| > 1$ are susceptible to centrifugal instability and may diffuse significantly within a few turnover time scales (Dong et al. 2007). Tilted eddies generated around headlands with sufficient vertical shear may overturn and generate strong vertical mixing (Pawlak et al. 2003; Farmer et al. 2002). The submesoscale eddies around Palau are preferentially cyclonic due to the mean westward flow, but anticyclonic eddies can be generated if oscillatory flow reversals are strong enough (Fig. 9). Although steep compared to typical topography, the 45° slope around Velasco has the potential to generate tilted eddies. Strong vertical shear in the upper few hundred meters likely enhances any baroclinicity (Fig. 2). Finally, finescale eddies may influence the life cycle of submesoscale eddies. In another FLEAT study, MacKinnon et al. (2019) observed the generation of 1-km-wide eddies on tidal time scales immediately in the lee of the northernmost promontory of Velasco (Fig. 1c). A full understanding of the Palau wake will hinge on understanding how eddies across this wide range of scales interact.


Herández-León, S., 1991: Accumulation of mesozooplankton in a wake area as a causative mechanism of the “island-mass


