Observations of Flow Separation and Mixing around the Northern Palau Island/Ridge

HEMANTHA W. WIJESEKEIRA, JOEL C. WESSON, AND DAVID W. WANG
Naval Research Laboratory, Stennis Space Center, Mississippi

WILLIAM J. TEAGUE AND Z. R. HALLOCK
NVision Solutions, Inc., Bay St. Louis, Mississippi

(Manuscript received 22 November 2019, in final form 23 June 2020)

ABSTRACT

Turbulent mixing adjacent to the Velasco Reef and Kyushu–Palau Ridge, off northern Palau in the western equatorial Pacific Ocean, is examined using shipboard and moored observations. The study focuses on a 9-day-long, ship-based microstructure and velocity survey, conducted in November–December 2016. Several sections (9–15 km in length) of microstructure, hydrographic, and velocity fields were acquired over and around the reef, where water depths ranged from 50 to 3000 m. Microstructure profiles were collected while steaming slowly either toward or away from the reef, and underway current surveys were conducted along quasi-rectangular boxes with side lengths of 5–10 km. Near the reef, both tidal and subtidal motions were important, while subtidal motions were stronger away from the reef. Vertical shears of currents and mixing were stronger on the northern and eastern flanks of the reef than on the western flanks. High turbulent kinetic energy dissipation rates, $10^{-6} - 10^{-4}$ W kg$^{-1}$, and large values of eddy diffusivities, $10^{-4} - 10^{-2}$ m$^2$ s$^{-1}$, with strong turbulent heat fluxes, 100–500 W m$^{-2}$, were found. Currents flowing along the eastern side separated at the northern tip of the reef and generated submesoscale cyclonic vorticity of about 2–4 times the planetary vorticity. The analysis suggests that a torque, imparted by the turbulent bottom stress, generated the cyclonic vorticity at the northern boundary. The northern reef is associated with high vertical transports resulting from both submesoscale flow convergences and energetic mixing. Even though the area around Palau represents a small footprint of the ocean, vertical velocities and mixing rates are several orders magnitude larger than in the open ocean.

1. Introduction

Flow interactions over space and time with abrupt topography, such as with small-scale islands and deep ocean ridges, can significantly impact regional to large-scale ocean circulation through intense turbulent mixing, formation of mesoscale wakes and eddy shedding behind islands, submesoscale upwelling and downwelling, and generation of lee internal waves (Baines 1995). Along with ocean variability due to topographic features, islands and atolls also modify the atmospheric boundary layer structure by generating island wind-wakes extending several hundreds of kilometers in the downstream direction (Xie et al. 2001). The upper-ocean current interactions and the modified wind stress curl in the wake of the island are associated with biologically productive upwelling zones.

Eddies and island wakes have received more attention from observational programs due to their local impact on biology and chemistry (Barton et al. 2000; Basterretxea et al. 2002). There are numerous numerical and observational studies of flow-topography interactions in shallow seas and over the continental shelf (Pattiaratchi et al. 1987; Wolanski and Hammer 1988; Dietrich et al. 1996; Nash and Moum 2001; Dong and McWilliams 2007; Jarosz et al. 2014). The primary vorticity-generation mechanism in shallow water is bottom drag (Pingree and Maddock 1980; Wolanski et al. 1984; Denniss and Middleton 1994; Dong and McWilliams 2007), and it can be modified by tilting of lateral vorticity also generated by bottom drag (Smolarkiewicz and Rutunno 1989). Numerical studies have suggested many processes for vorticity formation around islands and headlands, which include bottom drag on a sloping bottom with a
turbulent-shear boundary layer and a large vertical eddy viscosity, form drag due to pressure distribution around the topographic feature, and submesoscale instabilities (e.g., Canals et al. 2009; Molemaker et al. 2015; Gula et al. 2015). Recent measurements show that wake/wave-induced form drag over coastal banks is an important flow retardation mechanism for both tidal and low-frequency motions (Edwards et al. 2004; Warner et al. 2013; Wijesekera et al. 2014).

Studies of flow–topography interactions with ocean islands are limited. In deep water, the lateral boundary stress is the dominant driving mechanism of eddies. Existing observational and numerical studies show generation of wakes and eddy shedding (Heywood et al. 1996; Flament et al. 2001; Coutis and Middleton 2002; Sangrà et al. 2005; Dong et al. 2007). Wave drag over ridges acts on the entire water column and leads to the generation of mixing, eddies, and internal waves away from the bottom (Polzin et al. 1997; Jayne and St. Laurent 2001; Garrett and Kunze 2007). In the vicinity of deep ocean ridges, especially in the western Pacific Ocean, small-scale bathymetry is complex and in most places, isolated deep-ocean islands are connected with shallow reefs/banks and submarine ridges.

Ocean-model predictions are currently challenged in accurately representing flow fields in the vicinity of isolated deep-ocean islands and across ridges due to the poor representation of abrupt topographical features in the bathymetry and the consequent inadequate parameterizations of small-scale topographic effects on the large-scale flow (Johnston et al. 2019; Simmons et al. 2019). Furthermore, assimilation, to correct predictions of these flows, is limited by the poor knowledge of the appropriate horizontal and vertical scales in and around these complex regions.

a. FLEAT program

The overall objective of the U.S. Office of Naval Research (ONR) sponsored study “Flow Encountering Abrupt Topography” (FLEAT) is to determine temporal and spatial scales of ocean processes and the appropriate spatial scales of bathymetry for accurately capturing flow around an island/ridge complex (Johnston et al. 2019). FLEAT is a multi-investigator program, including scientists from U.S. universities, the U.S. Naval Research Laboratory (NRL), and the Coral Reef Research Foundation (CRRF) in Palau. Specific objectives of NRL are to observe temperature, salinity, currents, and turbulent mixing fields to characterize the flow when a large-scale current encounters Northern Palau, Velasco Reef and the Kyushu–Palau Ridge in the western Pacific Ocean (Fig. 1a), and to obtain a quantitative understanding of processes generating wakes/eddies and turbulent mixing around the complex topography. To achieve these specific objectives, the NRL deployed nine moorings, consisting of acoustic Doppler current profilers (ADCPs), and numerous temperature and conductivity sensors on and around the Velasco Reef (Fig. 1b), from
May 2016 through March 2017. Velocity profiles were measured in the upper 600 m of the water column at five deep water moorings. Profiles for nearly the full water column were measured at four moorings on the reef (Table 1). In addition, shipboard ADCP transects and microstructure measurements, utilizing a vertical microstructure profiler, were conducted during the middle cruise (using the R/V Roger Revelle from 26 November to 4 December 2016), which are the main focus of this study. Here, we examine turbulent-mixing and submesoscale processes around Velasco Reef (Fig. 1).

### Palau geography and large-scale currents

Palau is a chain of about 200 volcanic islands located in the southwestern Pacific Ocean at latitude 7.5°N, about 650 km southeast of the Philippines. Its topography ranges from mountains to low coral islands that are surrounded by large barrier reefs. Velasco Reef is about 65 km north of Babelthuap, the largest island of the eight principle islands of Palau (Fig. 1). Collectively, Palau has a footprint of about 466 km². The surrounding sea-floor, over 4000 m deep, rises abruptly to depths approaching 10 m at the outer edges of the reef. The reef is oval in shape and is oriented north–south, about 30 km × 15 km in size.

Velasco Reef is located between the westward-flowing North Equatorial Current (NEC) and the eastward-flowing North Equatorial Counter Current (NECC). The North Equatorial Undercurrent (NEUC) flows eastward beneath the NEC (Qiu et al. 2013). Using Argo float data (Roemmich et al. 2009), zonal geostrophic velocities were computed between 130° and 135°E and between 2° and 30°N by Qiu et al. (2015). They found the NEC between 7.5° and 25°N and the NECC between 2.5° and 7.5°N. Beneath the NEC, they also observed three eastward-flowing subsurface jets (NEUC) centered around 9°N, 13°N, and 18°N (Qiu et al. 2013; Schönau and Rudnick 2015).

The paper is organized as follows. Background winds and currents are described in section 2. Shipboard current and microstructure observations are in section 3. Discussions of vorticity generation and vertical transports are in section 4. Summary and conclusions of this study are presented in section 5.

### 2. Background atmospheric and oceanographic settings

#### a. Atmospheric conditions

Meteorological observations from an autonomous XMET weather station on the Kayangel Atoll (8°04’N, 134°42’E; Fig. 1a), at the northern end of the Palau Islands (Fig. 2), indicate that winds were relatively moderate to weak, with pulses of high winds associated with passage of cold fronts/low pressure events. Typical winds were slightly above 5 m s⁻¹, westerly to northwesterly during the shipboard survey between 26 November and 4 December 2016. A pulse of westerly–northwesterly winds of magnitude ~12 m s⁻¹ occurred about three weeks prior to the ship survey. The wind event lasted nearly 3 days. Similar events occurred in early July and the middle of August. Pulses of high-wind events can spin off near-inertial waves, and the resulting vertical shear and mixing in the thermocline can persist for several weeks after the passage of the wind events (Gregg et al. 1986).

#### b. Oceanographic conditions

Background currents (u, v) and temperature (T), salinity (S), and pressure (P) were collected from five deep subsurface moorings (M1–M5) and four bottom moorings.

### Table 1. Summary of moorings deployed during the FLEAT program. M1, M2, M3, M4, and M5 are deep subsurface ADCP moorings and B1, B2, B3, and B4 are shallow, bottom-mounted ADCP moorings. In M1–M5, 300- and 75-kHz ADCPs were mounted in a subsurface buoy at a depth of about Zn. The 300-kHz ADCPs were upward looking, and 75-kHz ADCPs were downward looking. In B1–B4, the bottom mounted 300-kHz ADCPs were upward looking. Note that H is the water depth, Δt is the sampling period of ADCPs, and Δz is the bin size; Zn is the mean position of the ADCP buoy from the surface for the M moorings. The depths of the first and last bins of a given ADCP varied due to the movement of the mooring, and the ADCP velocity profiles were interpolated between Z1 and Zn depth levels. The ADCPs at B1, B2, B3, and B4 were located respectively at 4.2, 4.2, 0.5, and 5.6 m, above the bottom. All depths are in meters. The site locations of moorings are marked in filled red circles in Fig. 1.

<table>
<thead>
<tr>
<th>Site</th>
<th>Start date</th>
<th>End date</th>
<th>Lat (°N)</th>
<th>Lon (°E)</th>
<th>H (m)</th>
<th>Δt (min)</th>
<th>Δz (m)</th>
<th>Zn (m)</th>
<th>Z1 (m)</th>
<th>Zn (m)</th>
<th>Type (kHz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>24 May 2016</td>
<td>4 Apr 2017</td>
<td>8.438</td>
<td>134.431</td>
<td>2600</td>
<td>60, 20</td>
<td>8, 4</td>
<td>64</td>
<td>8</td>
<td>600</td>
<td>75, 300</td>
</tr>
<tr>
<td>M2</td>
<td>21 May 2016</td>
<td>31 Mar 2017</td>
<td>8.258</td>
<td>134.794</td>
<td>2610</td>
<td>60, 20</td>
<td>8, 4</td>
<td>80</td>
<td>8</td>
<td>600</td>
<td>75, 300</td>
</tr>
<tr>
<td>M3</td>
<td>23 May 2016</td>
<td>3 Apr 2017</td>
<td>8.618</td>
<td>134.487</td>
<td>2846</td>
<td>60, 20</td>
<td>8, 4</td>
<td>66</td>
<td>8</td>
<td>600</td>
<td>75, 300</td>
</tr>
<tr>
<td>M4</td>
<td>22 May 2016</td>
<td>2 Apr 2017</td>
<td>8.605</td>
<td>134.602</td>
<td>1350</td>
<td>60, 20</td>
<td>8, 4</td>
<td>91</td>
<td>8</td>
<td>600</td>
<td>75, 300</td>
</tr>
<tr>
<td>M5</td>
<td>21 May 2016</td>
<td>1 Apr 2017</td>
<td>8.523</td>
<td>134.871</td>
<td>3610</td>
<td>60, 20</td>
<td>8, 4</td>
<td>73</td>
<td>8</td>
<td>600</td>
<td>75, 300</td>
</tr>
<tr>
<td>B1</td>
<td>25 May 2016</td>
<td>5 Apr 2016</td>
<td>8.378</td>
<td>134.545</td>
<td>110</td>
<td>15</td>
<td>4</td>
<td>8</td>
<td>100</td>
<td>300</td>
<td></td>
</tr>
<tr>
<td>B2</td>
<td>25 May 2016</td>
<td>5 Apr 2017</td>
<td>8.400</td>
<td>134.666</td>
<td>135</td>
<td>15</td>
<td>4</td>
<td>8</td>
<td>115</td>
<td>300</td>
<td></td>
</tr>
<tr>
<td>B3</td>
<td>26 May 2016</td>
<td>6 Apr 2017</td>
<td>8.472</td>
<td>134.623</td>
<td>53</td>
<td>15</td>
<td>4</td>
<td>8</td>
<td>45</td>
<td>300</td>
<td></td>
</tr>
<tr>
<td>B4</td>
<td>26 May 2016</td>
<td>7 Apr 2017</td>
<td>8.479</td>
<td>134.636</td>
<td>78</td>
<td>10</td>
<td>4</td>
<td>8</td>
<td>68</td>
<td>300</td>
<td></td>
</tr>
</tbody>
</table>
(B1–B4) with acoustic Doppler current profilers (ADCPs) around Velasco Reef (Fig. 1b), from May 2016 through March 2017 (Schönau et al. 2019; Teague et al. 2020). Table 1 provides positions, measurement depths, measurement times, sampling intervals, water depths, and instrument types used in the moorings. Here we use a subset of the moored currents that encompass the ship survey period. A detailed analysis of currents from these moored observations are out of the scope of the present study, but a description of the spatial distribution of background currents is necessary to provide a context for short-term observations. In the following, a brief description of background currents is presented before discussing specific details of the velocity field encompassing the microstructure observations of mixing around Velasco Reef.

1) TIDES AND MEAN CURRENTS

Geographical variability of currents around the Velasco Reef were examined by computing current ellipses for tidal and subtidal bands from the 10-month-long ADCP records. Tidal ellipse parameters were computed using harmonic analysis, for eight primary constituents, consisting of four diurnal (K1, O1, P1, Q1), and four semidiurnal (N2, M2, S2, K2) constituents. M2 and S2 were generally the dominant constituents and their geographical variabilities are shown in Fig. 3a. Significant variability is evident among the moorings. Tidal ellipses were largest at M4 in the deep water, likely due to shallower depths on the ridge, and at B4 on the steep slope of the reef. The ratios of minor to major axes of the M2 and S2 ellipses are small, indicating that the ellipses are nearly rectilinear. Figures A1 and A2 in the appendix display vertical profiles of ellipse orientations and magnitudes for M2
and S2. The tides on the west side of the reef at B1 were nearly barotropic (\(\sim 3-5 \text{ cm s}^{-1}\)). Tides on the east side of the reef at B2 changed their vertical structure and orientation near 40-m water depth (Fig. A1). Semidiurnal tidal components had a baroclinic structure in the deep water where the vertical structure changed near 200-m water depth (Fig. A2). Significant tidal currents near Palau were expected since tidal-height differences between low and high tides from observations at Palau tide stations can be as large as 2 m (https://coralreefpalau.org/research/oceanography/weather/ocean-observations/).

Geographical pictures of the mean currents and their variability are shown by mapping the depth-averaged currents and their standard deviation at each of the moorings (Fig. 3b). Here we used 40-h low-passed ADCP currents to generate a geographical picture of mean subtidal currents around the reef. Near-inertial waves (inertial periods \(\sim 80-81 \text{ h}\)) and intraseasonal oscillations (periods \(\sim 30 \text{ days}\)) dominated the low-frequency variability (Schönaug et al. 2019; Teague et al. 2020). The center of the standard-deviation ellipse is at the tail of the mean current arrow and reflects the area that is within one standard deviation of the mean. Mean currents are well determined when the mean vectors are not encompassed by their respective deviation ellipses. High variability, indicated by the major axes of the standard deviation ellipses, did not always appear in the direction of the mean flows. The variability was largest at B4 with the major axis oriented in the northwest direction. Orientations of the ellipses at B2 and B3 were north-northwest and northeast, respectively. Ellipse orientations at M4 ranged from northwest at M4 on top of the ridge to north-northwest at M2 and M5 located east of the reef/ridge; ellipse orientation at the two moorings located west of the reef/ridge were northeastward at M1 and M3. Generally, all of the depth-averaged flows had a northward component except for B1.

2) TIDAL AND SUBTIDAL VARIABILITY DURING THE SHIP SURVEY, NOVEMBER–DECEMBER 2016

In the following, we describe the time–depth structure of current variability from moored observations, which are relevant to the interpretation of turbulent mixing processes and shipboard ADCP currents for the period between 26 November and 4 December 2016. We divided the flow field into two frequency bands: tides to higher-frequency flows [frequency greater than 0.5 cycles per day (cpd)], and subtidal flows (frequency less than 0.5 cpd) including near-inertial waves and low-frequency motions. A pulse of strong current, with strong contributions in the tidal, near-inertial, sub-inertial bands, lasting about a month, occurred during the microstructure observations. In general, currents were strongest on the northern and eastern sides of the reef.

The flow was highly energetic, especially on the northern and eastern sides of Velasco Reef (Figs. 4 and 6). Near-inertial wave packets with strong vertical shears were found around the reef at B2 and B4 and in deep water at M2–M5, while currents at B1 were the weakest (Figs. 4 and 5). The subtidal currents in the upper 50 m at B2 and B4 were oriented nearly northward and northwest, respectively, with magnitudes as large as 0.5 m s\(^{-1}\). The subtidal flow at B3 was relatively small compared to that at B4, although these two stations were separated only by 1.6 km. B3 was located on the northern end of the reef at 53-m water depth, and B4 was located on the edge of the northern slope at 78-m water depth (Table 1). The semidiurnal tide was strongest at B2 and B4, where depth-averaged currents were as large as 0.75 m s\(^{-1}\) (Fig. 6). The tidal flow dominated in a nearly east–west direction at the northern edge of the reef at B3 and B4, and in a north–south direction at B2 (Fig. 6), indicating the tidal flow was aligned approximately with the bathymetry. In general the orientation of tides at B2–B4 during November–December 2016 are consistent with long-term statistics (Fig. 3a, Fig. A1). Currents were weak at the western edge of the reef (B1), where \(u\) and \(v\) components were about 0.1 m s\(^{-1}\). Downward propagating packets of near-inertial waves were found at moorings deployed in the deep water around the reef (M1, M2, M3, and M5) and over the ridge (M4) (Fig. 5).

The strength of mixing in the water column depends on the kinetic energy, velocity shear, and strain fields associated with internal tides, near-inertial waves, and low-frequency currents. Therefore we subdivided the moored observations into tidal and subtidal bands to examine the impacts of multiple-scale forcing on mixing around the reef. Figure 7 shows profiles of time-averaged kinetic energy (KE) and squared vertical shear around the reef for tidal and subtidal bands. Two-day high-passed and two-day low-passed filtered currents were used to obtain tidal and subtidal bands, respectively (Fig. 7). Components of vertical shear (\(\partial u/\partial z\) and \(\partial v/\partial z\)) were estimated by finite differencing of vertical \(u\) and \(v\) profiles at 8-m scales, and a shear-squared profile was constructed by averaging \((\partial u/\partial z)^2 + (\partial v/\partial z)^2\) between 15 November and 15 December. KE in the tidal band was largest at B4 and smallest at B1. Shear-squared estimates around the reef (B1, B2, B3, and B4) were within a factor of 2, but higher shear estimates were found along the eastern side (B2) and northern edges of the reef (Figs. 7a–d). For the subtidal band, KEs at B2 and B4 were two orders of magnitude larger than at B1, and the largest subtidal shear was found at B4, between 40 and 60 m. Both tidal
and subtidal shears were comparable along the northern and eastern sides of the reef indicating that they both contribute to mixing. The asymmetry in KE and shear on the east and west sides of the reef can impact on the strength of turbulent mixing during the period encompassing the ship observations.

Away from the Velasco Reef, at the deep-water moorings, KE and squared shear in the tidal and subtidal bands were computed (Figs. 7e–h). The tidal component of squared shear, about $10^{-2} \text{s}^{-2}$ near 50 m below the surface, gradually increased to about $10^{-4} \text{s}^{-2}$ at 100 m, and then dropped to about $10^{-5} \text{s}^{-2}$ below 400 m. KE in the subtidal band varied with depth but was primarily dependent on location. KE and shear were high on the eastern side of the ridge in the subtidal band. Away from the reef, the subtidal energy was larger than energy in the tidal band. Far-field shears at deep water sites were comparable but dropped from $10^{-4} \text{s}^{-2}$ in the upper 100 m to $2 \times 10^{-6} \text{s}^{-2}$ below 400 m. On average, shear squared in the subtidal band was close to $4 \times 10^{-5} \text{s}^{-2}$ near the surface with a maxima of about $10^{-4} \text{s}^{-2}$ at 100–120 m, and dropped to about $2 \times 10^{-5} \text{s}^{-2}$ below 400 m. Subtidal shears were largest in the upper 120 m, and gradually decreased with depth to about $10^{-6} \text{s}^{-2}$ at 400 m. In the upper 200 m, both tidal and subtidal shears were comparable, but tidal shear was one order magnitude larger than subtidal shear below 300 m. In this manuscript we discuss mixing in the upper 250 m, where both tidal and subtidal shears are equally important.

3. Ship surveys

A 9-day-long ship-based microstructure and velocity survey was conducted between 26 November and 4 December 2016. The microstructure survey, using a vertical microstructure profiler (VMP) (Jarosz et al. 2014), sampled in the upper 250 m of the water column around the reef and over the ridge. Eight VMP sections (5.5–10 km in length) were collected over the Velasco Reef and slope, where water depths ranged from 50 to 3000 m, and over one 12-km-long east–west section (134.561°–134.670°E).
over the Kyushu–Palau Ridge along the 8.525°N meridian (Fig. 1, Table 2). We also collected a day-long time series of microstructure profiles at a fixed position over the ridge at 8.525°N, 134.636°E (Fig. 1, Table 2). The VMP profiles around the reef were collected while the ship moved slowly toward or away from the reef. Vertical profiles of horizontal currents in the upper 600 m along the VMP transects were collected from shipboard 150- and 75-kHz ADCPs.

Mean hydrographic profiles of temperature, salinity, potential-density anomaly, and buoyancy frequency near Velasco Reef are illustrated in Fig. 8. The mean profiles were computed by taking an average of all profiles collected during VMP transects. Buoyancy frequency $N$ was estimated at 8-m vertical scales from 4-m averaged potential-density anomaly $\sigma_0$ for each VMP profile, where $N^2 = -g/\rho_0 (d\sigma_0/dz)$, with $g$ gravitational acceleration, $z$ depth in meters, and $\rho_0$ is the reference density. In general, temperature was weakly stratified in the upper 50 m, but salinity stratification generated a density stratification near the surface. Salinity has a maximum near 100-m water depth, where $N$ was strongest (Fig. 8d), with an average $N$ of about 0.024 s$^{-2}$ or 13.7 cycles per hour (cph).

a. Flow separation and generation of vorticity and divergence

Several ADCP surveys were conducted around box-shaped tracks, for evaluating vorticity and divergence

Fig. 5. Two-day low-pass filtered currents (cm s$^{-1}$) at M1–M5.
around the reef (Table 3, Fig. 1b). Box sides ranged from 8 to 18 km and were significantly less than the Rossby radius of deformation, $U/f \approx 100 \text{ km}$ for $f = 2 \times 10^{-5} \text{ s}^{-1}$ and $U = 0.5 \text{ m s}^{-1}$. Six ADCP box surveys were conducted over the northern slope (boxes 1, 2, and 5–6), one survey over the eastern slope (box 3), and one survey over the western slope (box 4) (Fig. 1b). We have noted that tidal and subtidal currents are equally important around the reef (Figs. 4–6). Therefore currents along box survey patterns contain tidal to subtidal bands, but separation of these motions from short time–space records of a box survey is not a trivial task. For example, a change in tidal phase between two opposite sides of a box can cause errors in lateral velocity gradients. Therefore, we examine tidal phases at B3 and B4 that highlight the time interval for which each box was surveyed. Figures 9a–d shows time series of 10–14-h band-passed $u$ and $v$ at B3 and B4 for boxes 1, 2, 5, and 6, respectively. Variability of tidal currents at B3 and B4 indicate that the velocity change during box 1 and box 6 surveys could be about 10 cm s$^{-1}$ (Figs. 9a,d). The smallest tidal variability was observed during the box 5 survey period (Fig. 9c). The box 2 survey, conducted over a full semidiurnal tidal cycle (Fig. 9b), had tidal velocities of about 20 cm s$^{-1}$. However, the ADCP survey around box 2 was repeated four times within the semidiurnal tidal cycle, and therefore averaging of these four consecutive loops can minimize the tidal signal from the velocity record.

Figures 10a and 10b show current vectors based on depth-averaged $u$ and $v$ for 17–105 m and 105–200 m, respectively. North of the reef, the flow entered from the east, accelerated, and moved northwestward. Currents were strongest in the upper 100 m (Fig. 10a), similar to the ADCP moored velocity records around the reef (Fig. 4). The observations showed that on the northern edge of the reef, there was an eastward flow of about 0.1 m s$^{-1}$ at the western end, and a northwestward flow of about 1 m s$^{-1}$ at the eastern end, thus forming a strong velocity front or convergence across the northern edge of the reef, connecting to the Kyushu–Palau Ridge. About 10 km away from the
northern boundary, the flow in the upper 100 m was toward the northwest at about 0.5 m s\(^{-1}\). Similar flow convergence can be seen in the upper 200 m, though currents were smaller (Fig. 10b).

Flow separation and formation of cyclonic vorticity off the northern end of the reef are shown in Fig. 11. Here we used all available current observations off the northern reef, including box surveys and other transects during the survey period. The vertical component of the relative vorticity \(\zeta\), estimated as a function of depth, was computed using east–west (E-W) and north–south (N-S) transects of current data from boxes 1, 2, and 5, where

\[
\zeta = \nabla \times \mathbf{V}_H = \frac{\delta v}{\delta x} \frac{\delta u}{\delta y}. \tag{1a}
\]

\(\mathbf{V}_H\) is the horizontal velocity vector, \(\delta x\) is the E-W separation and \(\delta y\) is the N-S separation of box surveys. First, velocity gradients were computed as,
\[
\delta u/\delta x = \frac{[u_E(x_E, y, z, t) - u_W(x_W, y, z, t)]}{(x_E - x_W)}, \quad (1b)
\]

and

\[
\delta u/\delta y = \frac{[u_N(x, y_N, z, t) - u_S(x, y_S, z, t)]}{(y_N - y_S)}, \quad (1c)
\]

where the subscripts E, W, N, and S denote east, west, north, and south sides of a given box. Average velocity gradients \( \delta u/\delta x \) and \( \delta u/\delta y \) were calculated by averaging individual estimates along N-S and E-W sides, respectively. N-S and E-W transects were separated in time approximately by 2.8 h (22\% of M2 tidal period) for box 1 and 1.6 h (12\% of M2 tidal period) for boxes 2 and 5 (Fig. 9, Table 3). Smaller spatial scales (5–18 km) and shorter separation times of opposing N-S and E-W segments can minimize significant tidal space–time contamination due to phase

<table>
<thead>
<tr>
<th>VMP line</th>
<th>Start time</th>
<th>End time</th>
<th>Duration (h)</th>
<th>Distance (km)</th>
<th>No. of VMP profiles</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1737 UTC 29 Nov 2016</td>
<td>0043 UTC 30 Nov 2016</td>
<td>7.10</td>
<td>6.4</td>
<td>39</td>
</tr>
<tr>
<td>2</td>
<td>0422 UTC 29 Nov 2016</td>
<td>1256 UTC 29 Nov 2016</td>
<td>8.57</td>
<td>7.7</td>
<td>48</td>
</tr>
<tr>
<td>3</td>
<td>1802 UTC 28 Nov 2016</td>
<td>0142 UTC 29 Nov 2016</td>
<td>7.67</td>
<td>5.4</td>
<td>28</td>
</tr>
<tr>
<td>4</td>
<td>0852 UTC 28 Nov 2016</td>
<td>1306 UTC 28 Nov 2016</td>
<td>4.23</td>
<td>7.7</td>
<td>16</td>
</tr>
<tr>
<td>5</td>
<td>0809 UTC 30 Nov 2016</td>
<td>1155 UTC 30 Nov 2016</td>
<td>3.77</td>
<td>9.3</td>
<td>16</td>
</tr>
<tr>
<td>6</td>
<td>1734 UTC 1 Dec 2016</td>
<td>0055 UTC 2 Dec 2016</td>
<td>7.35</td>
<td>10.4</td>
<td>30</td>
</tr>
<tr>
<td>7</td>
<td>0405 UTC 1 Dec 2016</td>
<td>1236 UTC 1 Dec 2016</td>
<td>8.52</td>
<td>9.0</td>
<td>33</td>
</tr>
<tr>
<td>8</td>
<td>1733 UTC 30 Nov 2016</td>
<td>0300 UTC 1 Dec 2016</td>
<td>7.50</td>
<td>8.2</td>
<td>41</td>
</tr>
<tr>
<td>9</td>
<td>0317 UTC 2 Dec 2016</td>
<td>1304 UTC 2 Dec 2016</td>
<td>9.78</td>
<td>12.0</td>
<td>39</td>
</tr>
<tr>
<td>Time series</td>
<td>1933 UTC 2 Dec 2016</td>
<td>0315 UTC 3 Dec 2016</td>
<td>7.70</td>
<td>—</td>
<td>25</td>
</tr>
<tr>
<td>0524 UTC 3 Dec 2016</td>
<td>0712 UTC 3 Dec 2016</td>
<td>1.80</td>
<td>—</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>0754 UTC 3 Dec 2016</td>
<td>1253 UTC 3 Dec 2016</td>
<td>4.98</td>
<td>—</td>
<td>15</td>
<td></td>
</tr>
<tr>
<td>1735 UTC 3 Dec 2016</td>
<td>0051 UTC 4 Dec 2016</td>
<td>7.27</td>
<td>—</td>
<td>30</td>
<td></td>
</tr>
</tbody>
</table>

**Fig. 8.** The cruise-mean profiles of temperature \( T \), salinity \( S \), potential density anomaly \( \sigma_\theta \), and buoyancy frequency squared \( N^2 \), from VMP data collected during the cruise. Dashed lines are the 95\% confidence level based on standard error estimates.
changes. Normalized vorticity $\zeta/f$ profiles (where $f = 2 \times 10^{-5}$ s$^{-1}$) computed as a function of depth for box 1, box 2, and box 5, are given in Fig. 11b. Vorticity estimates for boxes are within a factor of 2 of each other. Box 1 with the largest length scales produced smallest vorticity. The vorticity estimated from box 2 has the minimum tidal contamination due to temporal averaging of four estimates representing four different phases of the $M_2$ tidal cycle. The normalized vorticity $\zeta/f$ was about 4 between 25 and 50 m, and decreased with depth to about 1 at 200 m. The Rossby number,

<table>
<thead>
<tr>
<th>ADCP box around the reef</th>
<th>Start time</th>
<th>End time</th>
<th>Duration (h)</th>
<th>Complete cycles</th>
<th>Box size (width x length) (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (north)</td>
<td>1500 UTC 26 Nov 2016</td>
<td>2038 UTC 26 Nov 2016</td>
<td>5.6</td>
<td>1</td>
<td>$14.4 \times 17.6$</td>
</tr>
<tr>
<td>2 (north)</td>
<td>0625 UTC 27 Nov 2016</td>
<td>1908 UTC 27 Nov 2016</td>
<td>12.7</td>
<td>4</td>
<td>$7.8 \times 8.8$</td>
</tr>
<tr>
<td>3 (east)</td>
<td>1317 UTC 29 Nov 2016</td>
<td>1640 UTC 29 Nov 2016</td>
<td>3.4</td>
<td>1</td>
<td>$9.1 \times 12.4$</td>
</tr>
<tr>
<td>4 (west)</td>
<td>1056 UTC 1 Dec 2016</td>
<td>1642 UTC 1 Dec 2016</td>
<td>5.8</td>
<td>1</td>
<td>$10.4 \times 12.5$</td>
</tr>
<tr>
<td>5 (north)</td>
<td>1305 UTC 2 Dec 2016</td>
<td>1623 UTC 2 Dec 2016</td>
<td>3.3</td>
<td>1</td>
<td>$10 \times 12$</td>
</tr>
<tr>
<td>6 (north)</td>
<td>1110 UTC 3 Dec 2016</td>
<td>1624 UTC 3 Dec 2016</td>
<td>4.2</td>
<td>1</td>
<td>$10 \times 12$</td>
</tr>
</tbody>
</table>
is about 4, where $U$ is the velocity scale, and $L$ is the length scale. Earth’s rotation is a smaller factor than vorticity for 5–10-km scale motions, especially in the upper 100 m off the northern reef. We address a plausible vorticity generation mechanism in section 4.

The ship surveys provided a snapshot of vorticity north of Velasco Reef. To see the temporal variability of vorticity away from the reef, $\zeta$ was computed from ADCP current observations at M1–M5. Here, 2-day low-passed $u$ and $v$ were used to compute the subtidal vorticity at $O(40)$ km scale. Note that the distances between M1–M3, M3–M5, and M2–M5 are 21, 44, and 31 km, respectively. For a given time and depth, $\partial u/\partial x$ was estimated using M3 and M5 moorings, and similarly $\partial u/\partial y$ was estimated using both M1–M3 and M2–M5 mooring combinations. Figure 12 displays a time–depth section of normalized vorticity $\zeta/f$ for the period encompassing the ship survey. Strong positive (cyclonic) vorticity occurs during the cruise period. Some high values of $\zeta/f$ were found in the deep water column, but the strongest vorticity was limited to the upper 100 m (Fig. 12). The vorticity was positive between the middle of November and December, as observed during the ship surveys, and was about $(0.5–1)f$ in the upper 100 m during 26–27 November (Fig. 12) when near-inertial waves were strong (Figs. 4 and 5). Our analysis indicates that $O(40)$ km vorticity is a factor of 2–4 times smaller than vorticity estimated at 10-km spatial scale (Fig. 11b) reflecting scale dependence in vorticity as we can expect. However, such an analysis, along with modulation of vorticity at different frequency bands is out of the scope of this study.

Deceleration of flow near the boundary, and flow reversals off the northern reef edge, generate vorticity and divergence. Velocity divergence of lateral velocities $\nabla \cdot \mathbf{V_H}$ was computed for box 2, where

$$\nabla \cdot \mathbf{V_H} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \quad (3)$$

Similar to vorticity estimates, we averaged divergence estimates from four consecutive loops (Fig. 13a). By integrating velocity divergence [Eq. (3)] vertically from the surface to a given depth with $W = 0$ at the surface, a profile of vertical velocity off the northern reef (Fig. 13b) is estimated. Error bars in the vertical velocity increase due to the vertical integration of the divergence of the velocity from the surface to a given depth. Here, error in the divergence was estimated by using the standard deviation of the ADCP currents of about 0.5 cm s$^{-1}$, and therefore a total uncertainty resulting from velocity components associated with velocity differences ($\delta u$ and $\delta v$) is about 2 cm s$^{-1}$ (i.e., $4 \times 0.5$ cm s$^{-1}$). A standard error of velocity differences for a given depth was based on the number of samples or velocity differences averaged along sides of a box. The standard error of $W$ at a given depth is the summation of errors from the surface to that depth. The vertical velocity shows downwelling in the upper 125 m with a maximum of about 50 m day$^{-1}$ at 75 m, and upwelling below 150 m with a maximum of about 50 m day$^{-1}$ at 200 m. These vertical velocity estimates off the northern reef edge are significantly larger than estimates in the main thermocline in the open ocean.
indicating the major influence of steep topography on regional scale circulation. Strong downwelling on the top of the thermocline, and upwelling below the thermocline generate a stronger thermocline and high vertical stratification between 50 and 175 m.

### b. Microstructure observations

Estimates of eddy viscosities, diffusivities, and turbulent fluxes of heat and momentum are computed using measurements of turbulent kinetic energy (TKE) dissipation rate $\varepsilon$ (e.g., Gregg 1987; Osborn 1980). These mixing parameters and background flow variables across Velasco Reef and over the Kyushu–Palau Ridge were calculated by combining VMP and shipboard ADCP measurements. Eight time–depth transects were conducted around the reef, one transect was made over the northern ridge, and a day-long time series was measured at a single position on the ridge (Table 2). The VMP was free falling at a speed of about 0.6–0.8 m s$^{-1}$. The sampling rates for turbulent velocity and temperature sensors were 512 Hz. The TKE dissipation rate is calculated from measurements of high wavenumber vertical shear (for vertical scales of approximately 0.01–1 m). Note that turbulent velocity shear spectra were computed in the frequency domain, and the conversion from frequency to wavenumber was made using Taylor’s frozen field hypothesis (Tennekes and Lumley 1972). Bad data buffers infrequently occurred during our sampling, and were removed before computing turbulent-shear variances (Lueck 2016). The TKE dissipation rate was estimated by evaluating turbulent-shear variance from a spectrum computed for 512 data points corresponding to a 1-s data record, where

$$
\varepsilon = \frac{15}{2} \nu \left\langle \left( \frac{\partial u}{\partial z} \right)^2 \right\rangle,
$$

and $\left\langle \left( \frac{\partial u}{\partial z} \right)^2 \right\rangle$ is the variance of turbulent shear (angle brackets indicate suitable ensemble averages). Estimates of $\varepsilon$ were averaged into 8-m vertical bins to match the vertical resolution of the ADCP velocity-shear measurements. In regions where dissipation rates are large, shear probes

---

**FIG. 11.** (a) Showing the flow separation at the northern edge of the Velasco Reef based on ADCP velocities in the upper 105 m from multiple cruise tracks between 26 Nov and 4 Dec. (b) Normalized relative vorticity $\zeta/f$ based on ADCP surveys along box 1, box 2, and box 5.

**FIG. 12.** Normalized relative vorticity $\zeta/f$, based on 2-day low-pass filtered currents from the deep moorings (M1–M5) around the Velasco Reef for the period encompassing the shipboard ADCP survey, 26 Nov–4 Dec 2016 (thin dashed lines).
Wolk et al. (2002) do not have adequate wavenumber resolution to resolve the velocity-shear spectrum in order to evaluate \( \varepsilon \), therefore the theoretical spectrum for the inertial subrange was fitted to the observed shear spectrum to estimate the unresolved variance (Wang and Wijesekera 2018).

Vertical shear of horizontal currents (\( \text{sh}^2 \)) was computed by finite differencing of \( u \) and \( v \), where \( \text{sh}^2 = [(\partial u / \partial z)^2 + (\partial v / \partial z)^2] \) for 8-m vertical resolution. Note that we used 8-m vertically averaged profiles of \( u \) and \( v \) from the 150-kHz ADCP measurements. ADCP currents were recorded at 2-min time intervals, and therefore squared shear was interpolated to the VMP time frame before computing the Richardson number (\( \text{Ri} \)) at 8-m vertical scales, where \( \text{Ri} = N^2 / \text{sh}^2 \).

The eddy diffusivity \( K_D \) was estimated by assuming the turbulent buoyancy flux, \( (g/\rho)\langle \rho'w' \rangle \) is about 20% of the TKE dissipation rate (i.e., the coefficient of mixing efficiency = 0.2) (Osborn 1980), where

\[
K_D = -\frac{\langle \rho'w' \rangle}{\frac{dp}{dz}} \approx \frac{0.2\varepsilon}{N^2}. \tag{5}
\]

and primes denotes turbulent fluctuations. Although the coefficient of mixing efficiency for a given realization can have large fluctuations due to variations in ocean processes, a constant value of 0.2 has been established as appropriate for describing large ensemble averages of mixing events (Ijichi and Hibiya 2018; Gregg et al. 2018).

The eddy viscosity was estimated by assuming a steady-state balance among shear production terms, \( \langle u'w' \rangle \langle \partial u / \partial z \rangle \), \( \langle v'w' \rangle \langle \partial v / \partial z \rangle \), buoyancy flux, and dissipation rate in the TKE equation. Here turbulent momentum-flux terms are expressed as downgradient momentum fluxes, \( \langle u'w' \rangle = -K_M \langle \partial u / \partial z \rangle \) and \( \langle v'w' \rangle = -K_M \langle \partial v / \partial z \rangle \), and then the eddy viscosity can be written as (Gregg 1987),

\[
K_M = \frac{(g/\rho)\rho'w' + \varepsilon}{\text{sh}^2} \approx \frac{1.2\varepsilon}{\text{sh}^2}. \tag{6}
\]

The majority of these data records show strong velocity shears and energetic mixing in the water column. Before discussing individual transects in detail, we examined depth-averaged \( \varepsilon \) and \( \text{sh}^2 \) around the reef. Parameters \( \varepsilon \) and \( \text{sh}^2 \) were examined for a weakly stratified layer atop of the thermocline between 20 and 50 m, and a strongly stratified thermocline below 50 m.
Here $\varepsilon$ in the upper 20 m was discarded to exclude contamination of ship-induced turbulence. The scatterplots of $\varepsilon$ (Figs. 14a,b) show high dissipation rates in the weakly stratified layer around the reef, and to the east and north in the thermocline, where background vertical shear was also strongest (Figs. 14c,d).

To study spatial variability in mixing around the reef, we examined several transects representing east, west, north, and over the reef (Figs. 15–19) and a time series over the reef (Fig. 20). As mentioned above, background flow contained tides, near-inertial waves, and subseasonal and seasonal currents (Figs. 3–6), but the tidal flow may have a bigger impact on short-term, space–time shipboard observations. It is difficult to separate tides and their impacts on turbulence from these short space–time records, and therefore we examine tidal variability around the reef that occurs in the time interval for which each microstructure transect was conducted.

As noted above, mixing was intense on the eastern flank as observed along transects 1 and 2. Figure 15 shows distance–depth sections of $u$, $v$, $T$, $S$, $sh^2$, $N^2$, ($sh^2 - 4N^2$), $\varepsilon$, and $K_D$ along the eastern transect. Here, distance is referenced from the 50-m isobath. Semidiurnal tidal variability representing 10–14-h band-passed filtered currents at B2 is shown Fig. 15a. Note that B2 is the nearest mooring to transects 1 and 2. Currents as large as 0.5 m s$^{-1}$ were found in the upper
100 m, and the strongest northward moving subsurface flow was found near the eastern boundary. In general the currents in the upper 100 m (Figs. 15c,d) are consistent with observations at the B2 and M2 mooring sites (Figs. 4 and 5). Depth-averaged tidal currents (~0.1 m s$^{-1}$) at B2 (Fig. 6) were relatively small compared to northward-flowing subtidal currents (~0.5 m s$^{-1}$) at B2 (Fig. 4) and M2 (Fig. 5). Mixing properties along the transect 1 (Figs. 15j,k) do not follow the tidal variability (Fig. 15a). Current observations revealed a layered structure of...
velocity, which in turn generated thin ribbons of high vertical shear with a maximum of $10^{-3} \text{s}^{-2}$. The stratification in the water column was strongest between 75 and 125 m, where the local buoyancy frequency is $0.03 \text{s}^{-1}$ or 18 cph (Fig. 15h), which is larger than the average $N$ of about $0.02 \text{s}^{-1}$ or 12 cph (Fig. 8). Regions with high shear and low density stratification generated patches and thin layers of low $\text{Ri}$ (<0.25) or regions of $(\text{sh}^2 - 4N^2) > 0$ (Fig. 15i), where dissipation was large and eddy diffusivity $K_D$ was as high as $10^{-3}$ to $10^{-2} \text{m}^2\text{s}^{-1}$ (Figs. 15j,k). Note that the variable $[\text{sign}(\text{sh}^2 - 4N^2)\log_{10}((\text{sh}^2 - 4N^2) \times 10^3)]$ was plotted instead of $\text{Ri}$ to show clear separation between dynamically stable and unstable regions in the thermocline, where $[\text{sign}(\text{sh}^2 - 4N^2)]$ is $+1$ for $(\text{sh}^2 - 4N^2) > 0$ and $[\text{sign}(\text{sh}^2 - 4N^2)]$ is $-1$ for $(\text{sh}^2 - 4N^2) < 0$.

Highly sheared currents and energetic mixing were found along transects (3, 4, 5, 6, and 9) off the northern end of Velasco Reef and over the Kyushu–Palau Ridge (Figs. 16–18). Flow patterns show complex velocity structures over the steep slopes, but away from the

**FIG. 16.** VMP and ADCP survey along line 4 representing the northeastern side of the reef. (a) Time series of depth-averaged semidiurnal tidal currents at B3 for the duration of the transect beginning at 0852 UTC 28 Nov (Fig. 1b, Table 2). (b) Water depth $H$. Depth-distance sections of survey variables as described in Fig. 15.
slopes the flow fields exhibited layered structures (Figs. 16 and 17), as found on the eastern side (Fig. 15). As on the eastern side, there was a layer of subsurface salinity maxima between 50 and 100 m, with $N^2 = 10^{-4}$–$10^{-3}$ s$^{-2}$. Mixing was energetic and $\varepsilon$ varied from $10^{-6}$ to $10^{-4}$ W kg$^{-1}$, where shear was strong, $N^2$ was relatively weak; and $(sh^2 - 4N^2) > 0$ or $Ri$ fell below the critical value of 0.25. Energetic mixing ($\varepsilon > 10^{-5}$ W kg$^{-1}$) and large diffusivities were found both near and away from the reef (Figs. 16j,k and 17j,k).

The northern end of the headland where the Kyushu–Palau Ridge and Velasco Reef connect is characterized by a highly sheared complex flow field with energetic mixing (Fig. 18). As discussed above, TKE dissipation rates and eddy diffusivities were large in regions/layers where $Ri$ was small (Figs. 18i–k). Across the ridge along the 8.525°N meridian, flow was eastward on the western side of the ridge, while less than 10 km away, the flow was westward on the eastern side. The flow converged due to opposing east–west currents (where $\partial u/\partial x < 0$;
Fig. 18c), while the north–south flow across the ridge was likely to generate a positive vorticity ($\partial u / \partial x > 0$; Fig. 18d). Doming of isopycnals including both isotherms and isohalines were found over the ridge (Figs. 18e,f), where $\partial \rho / \partial x < 0$ on the eastern side and $\partial \rho / \partial x > 0$ on the western side.

On the western side of the reef (Fig. 19), TKE dissipation rates and eddy diffusivities in the upper 50 m were comparable to those on the eastern and northern sides, while mixing in the thermocline away from the sloping boundary was significantly weaker (Fig. 19j). However, mixing was strong near sloping boundaries, where $\varepsilon \sim 10^{-6} - 10^{-5}$ W kg$^{-1}$ and $K_D \sim 10^{-3} - 10^{-2}$ m$^2$ s$^{-1}$ and comparable to eastern and northern slopes. Thin layers of shear and a strongly stratified thermocline with salinity maxima around 100 m were consistent with similar features on both sides of the reef.

A day long time series of currents, hydrographic, and mixing properties were collected over the ridge off the Velasco Reef (Fig. 20). The time series had significant
data gaps, nevertheless we used it to explore the temporal variability of mixing over a tidal period. Currents in the upper 100 m were as large as 0.5 m s$^{-1}$ (Figs. 20b,c), while squared shear exhibited thin layered structure (Fig. 20f). Velocities show tidal variability below 100 m (Figs. 20b,c) and are qualitatively consistent with the depth-averaged tides observed at B4 (Fig. 20a). Dissipation rates were large ($\varepsilon \sim 10^{-6}$–$10^{-4}$ W kg$^{-1}$) and were similar to the observations made along transect 9. Mixing parameters such as $\varepsilon$, $K_D$, and $\varepsilon$, and $K_D$, and $\varepsilon$ (Figs. 20h–j) do not display clear correlation with the M$_2$ tidal cycle, and perhaps mixing is driven by a combination of tides and subtidal motions (Fig. 7).

We further examined east–west asymmetry in mixing in the thermocline as a function of depth by averaging profiles of $\varepsilon$, $K_D$, and turbulent heat flux $F_H$ along east, west, and north segments. Figure 21 shows the vertical structure of $\varepsilon$, $K_D$, and diapycnal heat flux $F_H$, representing east, west, and north segments off of the Velasco Reef, where
\[ F_H = \rho C_p (\bar{w}'T') = -\rho C_p K_D \left( \frac{\partial T}{\partial z} \right), \]

where \( C_p \) is specific heat of seawater, and \( \rho \) is seawater density. East transects are 1 and 2, west transects are 7 and 8, and north transects are 3, 4, 5 and 6. In the upper 50 m, \( \varepsilon, K_D \), and \( F_H \) were comparable around the reef (Fig. 21), but there was a marked contrast in mixing in the thermocline between west and east/north. Eddy diffusivities in the east and north \((-10^{-2} \text{ m}^2 \text{s}^{-1})\), were one order of magnitude larger than those in the west.

Diapycnal heat flux in the thermocline was about 100–500 W m\(^{-2}\) for northern and eastern slopes, but it was only 20 W m\(^{-2}\) for the western slopes. Note: negative turbulent flux \((\rho C_p w' T' > 0)\) indicates cold water moves upward (Fig. 21).

Spatial variations of mixing in the thermocline were further examined as a function of distance from Velasco Reef. Depth-averaged \( \varepsilon, K_D \), and \( K_M \), and the slope angle of bathymetry along VMP transects are plotted versus distance from the 50-m isobath (Fig. 22).
Bathymetry is steep, with slopes as large as 60° near the reef, especially on the eastern side of Velasco Reef, and reduces to half within 2 km distance. Turbulent mixing was strongest near the boundary, where $\epsilon \sim 10^{-6}$ W kg$^{-1}$, $K_D \sim 10^{-2}$ m$^2$ s$^{-1}$, and $K_M \sim 10^{-1}$ m$^2$ s$^{-1}$. Dissipation rates and diffusivities at the boundary were highest and fell by an order of magnitude 2 km away from the reef. Away from the reef, $\epsilon$, $K_D$, and $K_M$ over the western slopes were one order of magnitude smaller than that over the eastern and northern slopes. On average, for the far field (about 10 km away from boundary), $\epsilon$, $K_D$, and $K_M$ were reduced; $\epsilon \sim 10^{-7}$ W kg$^{-1}$, $K_D \sim 10^{-3}$ m$^2$ s$^{-1}$, and $K_M \sim 10^{-4}$ m$^2$ s$^{-1}$ (Fig. 22).

4. Discussion

The flow off the northern edge of Velasco Reef can be characterized as a high Rossby Number flow, where $Ro$ is about 2–4 (Fig. 11b), and is scaled as $U/Lf$ [Eq. (2)]. The corresponding length scale $L$ becomes 5–10 km for $U = 0.5$ m s$^{-1}$, and $Ro = 2–4$; $L$ of about 10 km is similar to the width of Velasco Reef. The nature of a flow behind an obstacle can be characterized from the Reynolds number (Batchelor 1967),

$$Re = UL/\nu,$$  \(8a\)

where $\nu$ is the molecular kinematic viscosity. $Re$ is $O(10^0)$ for $\nu = 1.2 \times 10^{-6}$ m$^2$ s$^{-1}$. By replacing $\nu$ from turbulent eddy viscosity $K_M$, we can obtain a representative Reynolds number, $Re_T$ which is $O(10^3)$ for $K_M = 10^{-4}$ m$^2$ s$^{-1}$. For $Re_T > 10^3$, the flow separates behind an obstacle and becomes highly turbulent and irregular in time, and this type of flow is expected off the northern Velasco Reef. Typically wakes off steep topographic features in deep water are modeled as flows around a vertical cylinder (e.g., Dong et al. 2007) where lateral friction plays an important role for generating vorticity while the bottom frictional torque is neglected, and the relevant Reynolds number,

$$Re_H = UL/\nu_H,$$  \(8b\)

where $\nu_H$ is the horizontal eddy viscosity, which is several orders of magnitude larger than $K_M$. Numerical simulations require $\nu_H$ of about 100 m$^2$ s$^{-1}$ to obtain
realistic mesoscale wake structure behind Islands in deep water (e.g., Pattiaratchi et al. 1987; Heywood et al. 1996; Chang et al. 2013). However, several investigators (Pingree and Maddock 1980; Pattiaratchi et al. 1987; Tomczak 1988; Denniss and Middleton 1994) noted that for shallow water regions, Re based on horizontal eddy viscosity [Eq. (8b)] is not adequate to describe the flow variability. They suggested a nondimensional parameter or an equivalent Reynolds number $R$ representing bottom frictional drag, where

$$ R = \frac{D}{C_D L}, $$

where $C_D$ is the frictional drag coefficient near the bottom, and $D$ is the water depth. Magnitudes of $Re_H$ and $R$ can be similar, but similar values of $R$ and $Re_H$ can generate vastly different flow fields (Denniss and Middleton 1994). Around the Velasco Reef, $D$ varies from 50 to 1000 m within a short distance from the top of the reef (Fig. 1), and therefore both bottom frictional and lateral diffusivities can be important in describing the flow variability. As described above, intense mixing occurred on the sloping boundaries, and therefore the bottom frictional drag may be an important factor in the vorticity dynamics. Recent numerical studies show that formation of eddies on wall boundaries can be modeled
by coupling frictional drag associated with the sloping bottom boundary layer (e.g., Molemaker et al. 2015; Gula et al. 2015). MacKinnon et al. (2019) observed intense mixing around the reef with \( \varepsilon \) as large as \( 10^{-5} \text{W kg}^{-1} \) (Fig. 3 in MacKinnon et al. 2019), similar to the dissipation rates discussed here, and suggested that the bottom frictional stress is an important factor for generating wake eddies. They further suggested that small-scale vorticity, generated off the edge of Velasco Reef advects downstream, which in turn helps to generate effective eddy viscosity, typically used in regional-scale numerical models. In the following we explore the impacts of bottom frictional drag on the generation of submesoscale vorticity and also discuss a plausible role of submesoscale vortices on effective lateral eddy viscosity applicable to mesoscale flow.

### a. Bottom boundary layer and generation of vorticity

The Velasco Reef is quite steep, with bathymetry slope angles of 30°–60° that are associated with energetic turbulence and large eddy viscosity. Even though slopes are large, the frictional forces resulting from the bottom boundary layer turbulence and strongly sheared lateral currents can slow down the flow around the reef, which in turn can lead to flow separation and formation of vorticity. By assuming the turbulent stress is nearly uniform in the bottom boundary layer and away from the viscous sublayer, bottom stress components can be approximated as

\[
\tau_{sb} = -\langle u'w' \rangle_b = K_M \left( \frac{\partial u}{\partial z} \right)_b, \quad (9a)
\]

and

\[
\tau_{yb} = -\langle u'w' \rangle_b = K_M \left( \frac{\partial v}{\partial z} \right)_b, \quad (9b)
\]

where \( \tau_{sb} \) and \( \tau_{yb} \) are east–west and north–south components, and the bottom stress is

\[
|\tau_b| = (\tau_{sb}^2 + \tau_{yb}^2)^{1/2} = K_M (sh)^{1/2}, \quad (9c)
\]

for \( K_M \sim 10^{-1} \text{m}^2 \text{s}^{-1} \) (Fig. 22), \( (sh)^{1/2} \sim 2 \times 10^{-2} \text{s}^{-1} \) (Fig. 7), then \( \tau_b \sim 2 \times 10^{-3} \text{m}^2 \text{s}^{-2} \).

As we discussed above, when flow passes over the northern edge of Velasco Reef and the connected Kyushu–Palau Ridge from the southeast, strong lateral currents with high vertical shears, and frictional stresses, were generated on the steep bottom slopes. The ADCP surveys indicated a generation of positive (cyclonic) vorticity, which is \( -2-4f \) atop of the thermocline and comparable to \( f \) at 200 m (Fig. 11b). If the vorticity is generated by the bottom stress on the sloping bathymetry, then the vorticity balance in the water column can be simplified as (e.g., Dong and McWilliams 2007)

\[
\frac{\partial \zeta}{\partial t} = -\nabla \times \tau_b = -\frac{1}{H} \left( \frac{\partial \tau_{yb}}{\partial x} - \frac{\partial \tau_{sb}}{\partial y} \right), \quad (10)
\]

where \( H \) is the height of vortex stretching or the height of the Ekman boundary layer. Figure 4 shows surface intensified westward and northward currents and strong vertical shears, where \( \partial \omega/\partial z < 0 \) and \( \partial \omega/\partial z > 0 \) and \( \tau_{sb} < 0 \) [Eq. (9a)] and \( \tau_{yb} > 0 \) [Eq. (9b)]. Since eddy viscosity decays rapidly away from the reef (Fig. 22d), a similar decrease in turbulent stresses away from the boundary can be expected, and therefore we can approximate \( \frac{\partial \tau_{yb}}{\partial x} - \frac{\partial \tau_{sb}}{\partial y} \) and \( (\partial \tau_{yb}/\partial x) - (\partial \tau_{sb}/\partial y) < 0 \) where \( l \) is the length scale associated with the high turbulent stress region, which is about 2 km (Fig. 22d). The negative bottom-stress curl supports generation of positive (cyclonic) vorticity [Eq. (10)]. Assuming the relevant time scale is similar to the advective time of the flow, \( H/U \), we can approximate \( \zeta = |\tau_b|/UH \), where \( U \) is the velocity scale associated with the high stress region. For \( U \sim 0.2–0.5 \text{m s}^{-1} \), \( H = (2K_Mf)^{1/2} = 100 \text{m} \) as \( K_M \sim 0.1 \text{m}^2 \text{s}^{-1} \) and \( \tau_b \sim 2 \times 10^{-3} \text{m}^2 \text{s}^{-2} \) then \( \zeta = (4 - 10) \times 10^{-5} \text{s}^{-1} \), which is about \( 2-5f \), and is similar to the observed vorticity (Fig. 11b).

### b. Effective lateral eddy viscosity

Our analysis reveals that the submesoscale (\( \sim 5–10 \text{km} \)) to mesoscale (\( \sim 20–40 \text{km} \)) vorticities on subtidal time scales were (2–4)\( f \) (Fig. 11b) and (0.5–1)\( f \) (Fig. 12), respectively. MacKinnon et al. (2019) reported formation of \( O(1) \text{-km-scale wake eddies on tidal time scales with vertical components of relative vorticity of about \( 20–30 \text{f} \), and Rudnick et al. (2019) reported, based on glider observations on the northwestern side of the northern end of the Velasco Reef, that the magnitude of relative vorticity for spatial scales of 40 km was about 0.3\( f \). This low-frequency, large-scale vorticity away from the reef is two orders of magnitude smaller than the 1-km-scale estimate near the reef. Vorticity is scale dependent, as noted by differing estimates of vorticity on 1-km and 5–40-km scales off the northern end of the Velasco Reef.

As discussed in section 4a, the frictional drag in the sloping bottom boundary layer can generate small-scale to submesoscale vorticity around the Velasco Reef. The vorticity was generated by multiple-timescale motions when tides, near-inertial currents, and subseasonal to seasonal mean currents interacted with the reef. The injection of energy and vorticity at small scale to submesoscale around the reef can be advected downstream while transforming the wake structure, perhaps through...
nonlinear interactions and flow instabilities. From a two-dimensional (2D) turbulence point of view (Salmon 1998), the kinetic energy $E$ cascades toward larger scales while the enstrophy (which is described as the integral of the square of the vorticity) cascades toward smaller scales. Here, we define a spatially averaged enstrophy as $\text{Ens} = \langle \zeta^2 \rangle$. The time scale of enstrophy transfer $T_\lambda$ from one cascade step to the next cascade step can be taken as an inverse strain rate, where $T_\lambda = C\text{Ens}^{-1/2} = C\langle \xi^2 \rangle^{-1/2}$, where $C$ is a constant, say, $C \sim 1$. The scaled enstrophy transfer rate becomes $\text{Ens}/T_\lambda = C^{-1}\text{Ens}^{3/2}$. By taking the time rate of change in enstrophy as a diffusive process, we can write $C^{-1}\text{Ens}^{3/2} = \nu_H \nabla^2 \text{Ens}$, where $\nu_H$ is an effective lateral eddy diffusivity. Suppose $\lambda$ is the dissipation scale of enstrophy, then the scaling arguments suggest that the lateral eddy diffusivity can be approximated as

$$\nu_H = C^{-1} \lambda^2 \text{Ens}^{3/2} = C^{-1} \lambda^2 \langle \xi^2 \rangle^{1/2}.$$  \hspace{1cm} (11)

The representative Reynolds number for the mesoscale flow around the Velasco Reef of width $L$ (8b) becomes

$$\text{Re}_H = \frac{UL}{\nu_H} = \frac{\text{Ro}}{\text{Ek}_H} \left( \frac{L}{\lambda} \right)^2 = \left( \frac{Cf}{\langle \xi^2 \rangle^{1/2}} \right) \left( \frac{U}{fL} \right) \left( \frac{L}{\lambda} \right)^2,$$  \hspace{1cm} (12)

where $\text{Ek}_H = \nu_H / f \lambda^2$ is a kind of horizontal Ekman number defined for scale $\lambda$, and $L/\lambda$ is the aspect ratio. In this context, $\lambda$ can be treated as a characteristic length scale of the lateral boundary layer due to diffusion of momentum around the Velasco Reef. As shown in Fig. 22d, the decay scale of $K_M$ is about 2 km. For $C = 1$, $\lambda = 2 \text{ km}$, and $\xi = (2-4)f \text{ s}^{-1}$, the effective lateral eddy viscosity $\nu_H$ is about 160–320 m$^2$ s$^{-1}$. The representative $\text{Re}_H$ varies from 6 to 31 for $U = 0.2–0.5 \text{ m s}^{-1}$, $L = 10 \text{ km}$, and $\nu_H = 160–320 \text{ m}^2 \text{ s}^{-1}$. For the $5 < \text{Re}_H < 40$ range, the flow separation and recirculation occurs prior to the onset of vortex shedding, which appears to be consistent with the observations discussed here (Fig. 11).

c. Form drag

Apart from bottom frictional effects, flow separation, eddies, and lee-wave disturbances downstream of an obstacle can be generated by form drag, which is related to dynamic pressure differences across an obstacle arising from upstream blocking (e.g., Baines 1995). This drag occurs on multiple time scales. Form drag resulting from tides over a sloping ridge in Puget Sound, Washington, was examined using both observational and model evaluations (Edwards et al. 2004; McCabe et al. 2006). They found that the drag resulted from both isopycnal deformations (“internal drag”) and change in surface elevation (“external drag”). Warner et al. (2013) measured form drag directly over this headland in Puget Sound and found that the drag was consistent with linear theory. Analysis of a 6-month-long time series of pressure estimates over an isolated submarine bank in the northern Gulf of Mexico revealed that form drag was generated by multiple timescale processes and deviated from the linear theory (Wijesekera et al. 2014), while the background flow was in a high-drag state. Similar processes can occur during flow encounters with the steep bathymetry around the northern Velasco Reef, but we lack direct pressure measurements to evaluate the magnitude of the form drag.

d. Mixing, shear, and upwelling/downwelling

Our analysis shows that turbulent eddy diffusivities were as large as $10^{-3} \text{ m}^2 \text{ s}^{-1}$ near the reef and dropped rapidly by two orders of magnitude about 10 km from the reef. There is an asymmetry in mixing in the thermocline around the reef, with the strongest mixing occurring on the eastern and northern slopes and the weakest mixing occurring on the western slopes. The large eddy diffusivities generate downward turbulent heat fluxes of about 100–500 W m$^{-2}$ in the thermocline.

Microstructure measurements collected during the last several decades indicate that mixing in the stratified ocean is closely related to background shear, strain, and energy levels of the internal-gravity wave field. The common assumption in the most popular models is that nonlinear wave–wave interactions transfer energy from larger scales to smaller scales, and eventually mixing is generated by the Kelvin–Helmholtz (KH) instability (where the Richardson number becomes less than 0.25; Miles 1961). In the presence of highly sheared currents as observed here, turbulent mixing was likely generated by just such KH instabilities. The flow field near the reef consisted of flows with comparable tidal and subtidal (near-inertial and low-frequency motions) bands. Away from the reef, subtidal flows dominated in both KE and shear variances. The flow field around Velasco Reef deviates substantially from the open ocean environment. We noted that the dissipation rate was high when shear was strong, $N^2$ was weak, and $\text{Ri}$ was low or below the critical value of 0.25. Detailed exploration of functional forms of $\varepsilon$ as a function of background parameters are out of the scope of this manuscript, but we can examine the parameter dependence of the dissipation rate by averaging $\varepsilon$ into $\text{sh}^2$ and $\text{Ri}$ bins (Fig. 23). The dissipation rate is highly correlated with the squared shear, with a linear relationship between $\log_{10}(\text{sh}^2)$ and $\log_{10}(\varepsilon)$ (Fig. 23), where $\varepsilon \sim (\text{sh}^2)^n$ with $n \approx 0.5$, which is closer to the functional form suggested by MacKinnon and Gregg (2005), but the predicted estimates are at least one order of magnitude smaller.
than the observations. We also find a close correlation between Ri and $\varepsilon$. A rapid increase in $\varepsilon$ occurred as Ri fell below the critical limit of 0.25.

As discussed above, currents off Velasco Reef and over the Kyushu–Palau Ridge generated flow divergence, $\nabla \cdot \mathbf{V}_H$, which in turn generated vertical motions. Vertical velocity $W$ was computed by vertically integrating $\nabla \cdot \mathbf{V}_H$ from the surface to a given depth, with $W = 0$ at the surface (Fig. 13). The vertical velocity at 75 m indicates that downwelling atop of the thermocline and upwelling at 200 m can be as large as 50 m day$^{-1}$ (Fig. 13b). Large downwelling motions in the upper 75 m can advect heat and lateral momentum downward and similarly large upwelling motions at 200 m can advect heat and lateral momentum upward, thus generating strong convergences of heat, mass, and momentum in the thermocline. Converging vertical motions can enhance the thermocline stratification and push subsurface flows off of Velasco Reef laterally. The doming thermohaline structure across the ridge was evident in Figs. 18e and 18f and the rising isotherms and isohalines in the thermocline qualitatively suggest upwelling in the thermocline. The intense mixing in the water column can diffuse heat and momentum vertically with a diapycnal velocity $w_d$ of about 1 m day$^{-1}$, where $w_d \sim K_D/h$, for $K_D = 10^{-3}$ m$^2$ s$^{-1}$, and $h = 100$ m. Both diapycnal advection associated with turbulent motions and vertical advection associated with the mesoscale flow field are significantly larger than open ocean estimates, indicating that the Palau–Velasco Reef, with the associated complex steep topography, is a hot spot of vertical transfer processes in the western equatorial Pacific region.

5. Summary

Flows encountering complex topographic features connected with Velasco Reef and Kyushu–Palau Ridge off northern Palau in the western equatorial Pacific Ocean, were examined as part of the Office of Naval Research Initiative, Flow Encountering Abrupt Topography (FLEAT). The main objectives of this FLEAT study are to understand and quantify multiscale processes when a large-scale current system interacts with the abrupt topography in the vicinity of Velasco Reef and Kyushu–Palau Ridge. The Naval Research Laboratory deployed five ADCP moorings in the deep water around the reef, and four bottom-mounted ADCP moorings in the shallower water near the reef, from May 2016 to March 2017. Each deep mooring had a 300-kHz upward-looking ADCP and a 75-kHz downward-looking ADCP, and several temperature, conductivity, and depth sensors, for measuring current profiles in the upper 600 m and hydrographic fields below. In addition to shipboard ADCP transects, temperature and salinity were measured and TKE dissipation rates were estimated using a vertical microstructure profiler during a 9-day cruise in November–December 2016. In this study we focused on turbulent mixing and flow separation around Velasco

Fig. 23. Bin-averaged TKE dissipation rate plotted (a) as a function shear squared and (b) as a function of Ri. Dashed lines are the 95% confidence levels based on lognormal statistics of $\varepsilon$. The dotted line in (a) is $\langle \varepsilon \rangle = 10^{-4}(sh^{3/2})$ and is not a fit to the data.
Reef from the 9-day shipboard survey and moored velocity measurements during this time period.

The Velasco Reef is an oval-shaped submarine bank that is roughly 10 km wide (east–west) and 30 km long (north–south) in reference to the 50-m isobath. The shallowest point at the top of the reef is only 20 m deep. Bathymetric slope angles around the reef vary from 30° to 60°, with the steepest slopes on the eastern side. Several sections (9–15 km in length) of microstructure, hydrography, and velocity around Velasco Reef were collected, where water depths ranged from 50 to 3000 m. Microstructure profiles were obtained while progressing slowly either toward or away from the reef. Underway ADCP surveys were conducted along quasi-rectangular boxes with side lengths of 5–10 km. Findings of this study are:

- Asymmetric structure of currents, vertical shears, turbulent kinetic energy dissipation rate, and eddy diffusivities were apparent around the reef. For example, currents were as high as 0.75 m s⁻¹ and stronger on the northern and eastern sides than on the western side, where currents were about 0.1 m s⁻¹.
- Around the reef both tidal and subtidal (inertial waves to low frequency) motions were important, while away from the reef, subtidal motions were stronger.
- Highly sheared currents, energetic TKE dissipation rates, from 10⁻² to 10⁻¹ W kg⁻¹, large eddy diffusivities, from 10⁻⁴ to 10⁻² m² s⁻¹, and downward turbulent heat fluxes of 100–500 W m⁻² in the thermocline, were found around the reef.
- Dissipation rate is highly correlated with vertical shear of lateral currents; there is a linear relation between sh² and $\varepsilon$ (Fig. 23). There is a close correlation between Ri and $\varepsilon$. A rapid increase in $\varepsilon$ occurred as Ri fell below the critical limit of 0.25 indicating that KH type instabilities drive mixing.
• The strongest eddy diffusivities and eddy viscosities occurred near the reef and decayed rapidly away from reef; $e$, $K_D$, and $K_M$ fell by an order magnitude within 2 km from the reef boundary.

• The turbulent fluxes estimated in the thermocline were significantly larger than the net surface heating, indicating that turbulent mixing moves cold water up from below at a cross-isopycnal speed ($K_D/h$) of about 0.5–1 m day$^{-1}$. 

**FIG. A2.** Vertical structure of major (blue) and minor (red) axes for M$_2$ (solid) and S$_2$ (dashed) tidal ellipses and orientation at M1–M5 mooring locations. Triangle and diamond symbols are M$_2$ and S$_2$ tidal axes and orientations from RCM9 current meters at deeper depths.
• Currents flowing along the eastern side of the reef separated at the northern edge, and generated cyclonic vorticity at 5–10-km scales of about (2–4)\( f \). Our analysis suggests that cyclonic vorticity is generated by the negative bottom-stress curl resulting from energetic turbulent mixing around the reef.
• Flow divergences in the water column caused strong downwelling (50 m day\(^{-1}\)) and upwelling (50 m day\(^{-1}\)) above and below the thermocline, respectively. Vertical advection resulting from large downwelling and upwelling velocities above and below the main thermocline (at 75–150 m) can generate strong convergences of heat, mass, and momentum in the thermocline, thus enhancing thermocline stratification, while high mixing rates tend to reduce the stratification.
• Rough estimates of effective lateral eddy diffusivity and a representative Reynolds number were estimated based on scaling arguments of the conceptual idea that small-scale to submesoscale eddies generated by the bottom frictional drag around the northern Velasco Reef advect downstream while cascading enstrophy toward small scales. The estimated effective lateral eddy diffusivity was \( \sim 160–320 \text{ m}^2 \text{s}^{-1} \). The representative Reynolds number has a range of 6–31, where flow separation and recirculation occur prior to the onset of vortex shedding.

In summary, abrupt topographies like Velasco Reef and the connected submarine ridge are regions of flow separation where vorticity is generated at multiple space–time scales, most likely resulting from the torque imparted by the bottom stress over the steep sloping bottom boundary through intense turbulent mixing. Furthermore, the region is associated with high vertical transports resulting from flow convergences and diapycnal transports due to energetic mixing in the thermocline. These small-scale to mesoscale vertical transports are 100–1000 times larger than for the typical open-ocean conditions.

Acknowledgments. This work was sponsored by the ONR Grant N0001416WX01186. Special thanks to Mr. Andrew Quaid and Mr. Ian Martens for their efforts and their technical support. We thank Dr. Eric Terrill and Mr. Travis Schramek at Scripps Institution of Oceanography for providing XMET meteorological products. We also thank Drs. Patrick Colin and Lori Colin of Coral Reef Research Foundation in Palau for their assistance, and the Captain and crew of the R/V Roger Revelle. We thank two anonymous reviewers for their helpful comments and suggestions.

APPENDIX

Tidal Analysis

Tidal harmonic analysis was conducted on eastward, \( u(z, t) \) and northward, \( v(z, t) \) current series using T_TIDE (Pawlowicz et al. 2002). Four diurnal (\( K_1, O_1, P_1, O_1 \)) and four semidiurnal (\( N_2, M_2, S_2, K_2 \)) constituents were computed. \( M_2 \) and \( S_2 \) were the most dominant constituents. Profiles of \( M_2 \) and \( S_2 \) tidal ellipse parameters are shown in Fig. A1 for shallow water mooring (B1–B4) and in Fig. A2 for deep water moorings (M1–M5).

REFERENCES


Unauthenticated | Downloaded 07/30/22 03:58 AM UTC


