Flow and Mixing around a Small Seamount on Kaena Ridge, Hawaii

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ABSTRACT

Microstructure observations over a small seamount on the Kaena Ridge, Hawaii, showed asymmetry in the along- and across-ridge directions. The ~400-m-high seamount is on the southern edge of the ridge (centered at 21°43′49″N, 158°38′48″W), 42 km northwest of Oahu. A 1-km-resolution numerical simulation shows that the flow within the depth range of the seamount tends to be accelerated around the seamount rather than going up and over it. The flow patterns, however, are more complicated than for an isolated seamount because of the influence of the ~3000-m-high Kaena Ridge. Comparison with the numerical simulations indicates that the across-ridge asymmetry, in which dissipation on the north-northeastern side of the seamount was higher and more concentrated toward the bed than on the south-southwestern side, is consistent with an $M_{2}$ tidal beam generated at the northern edge of the ridge. The along-ridge asymmetry, with higher dissipation on the east-southeastern flank than on the west-northwestern flank, is in qualitative agreement with $M_{2}$ shear variance from the model simulation. The average observed dissipation rate over the seamount was $\tau = 6.2 \times 10^{-8}$ W kg$^{-1}$, and diapycnal diffusivity was $K_{\rho} = 1.3 \times 10^{-3}$ m$^2$ s$^{-1}$. Dissipation measurements following the 1000-m isobath south-southwest of the seamount suggest along-ridge internal tide generation caused by topographic steering that creates an along-ridge current over critical topography northwest of the seamount.

1. Introduction

Seamounts are the dominant topographic feature of the Pacific Ocean. There may be close to one million seamounts on the Pacific Plate, with from 30 000 to 70 000 being taller than 1 km (Smith and Jordan 1988; Wessel and Lyons 1997). The majority of these large seamounts occur in the western Pacific. Seamounts are sites of enhanced turbulence (Lueck and Mudge 1997; Toole et al. 1997; Kunze and Toole 1997; Lavelle et al. 2004), internal tide generation (Noble et al. 1988; Holloway and Merrifield 1999), and internal wave scattering/reflection (Kunze and Sanford 1986; Eriksen 1998; Johnston and Merrifield 2003; Johnston et al. 2003). The presence of seamount chains affects large-scale circulation features, such as the Gulf Stream (Roden 1987; Ezer 1994). When the summit is close to the surface, seamounts often support large nektonic (aquatic organisms that swim) communities and are an important fishing resource (Rogers 1994; Dower and Mackas 1996). It has been hypothesized that this is due to increased abundance of planktonic species through increased primary production with upwelling at seamounts or the trapping of plankton during diurnal migration (Rogers 1994). Acceleration of currents around seamounts and varied habitat regions lead to rich benthic communities (Genin et al. 1986; Rogers 1994).

Topographic features can interact with the barotropic (surface) tide to produce baroclinic (internal) tides (e.g., Prinsenberg et al. 1974; Baines 1982; Holloway and Merrifield 1999; Egbert and Ray 2000; Simmons et al. 2004). The energy removed from the barotropic tide then cascades through the internal wave spectrum while propagating through the ocean, before dissipating...
as heat (Rudnick et al. 2003). Internal tide generation at abrupt features, such as seamounts and ridges, is a potential source of the mixing that maintains the abyssal stratification, although exact pathways remain unclear.

Numerical simulations show that the nearly 1:1 horizontal aspect ratios of seamounts makes them less efficient than ridges at extracting energy from the barotropic tide (Holloway and Merrifield 1999) because more of the flow can go around a seamount, resulting in less across-isobath flow. Johnston and Merrifield (2003), however, report that when considering scattering of a mode-1 internal tide to higher modes the difference between seamounts and ridges is small. Although most of the mode-1 energy is transmitted past the topographic feature, the height of the feature affects the proportion of energy that is reflected. The slope and width (relative to the internal tide wavelength) determine the formation of beams and higher modes. As with many internal wave–topography interactions, a key parameter is the ratio of bottom slope $b$ to the slope of internal wave energy propagation

$$\gamma = \frac{k_h}{k_z} = \left(\frac{\omega^2 - f^2}{N^2 - \omega^2}\right)^{1/2} \tag{1}$$

where $\omega$ is the internal wave frequency, $f$ is the Coriolis frequency, $N$ is the buoyancy frequency, and $k$ is the wavenumber (the $z$ and $h$ subscripts refer to vertical and horizontal, respectively). An incident mode-1 internal tide, which can be thought of as the superposition of an upward- and downward-propagating ray, is focused into a beam if the topographic feature is near critical ($b/\gamma \approx 1$). For a subcritical ($b/\gamma < 1$) feature, the mode-1 internal tide propagates away largely unaltered. In the supercritical case ($b/\gamma > 1$), energy is reflected back from the flanks. Topographic scattering of energy from defined beams is not understood well and may be very different from barotropic or mode-1 situations.

Lueck and Mudge (1997) report that dissipation rates over the upper flanks of Cobb Seamount were $10^2$–$10^4$ times open-ocean values, decaying to background levels by 14 km from the summit. Fieberling Guyot, a relatively isolated seamount in the eastern Pacific, has been studied using a combination of vertical profilers and moorings (Kunze and Toole 1997; Toole et al. 1997; Eriksen 1998). Kunze and Toole (1997) observed a 200-m-thick anticyclonic vortex and $\pm 0.15$ m s$^{-1}$ diurnal fluctuations over the $\sim 10$-km-diameter summit plateau. Diapycnal diffusivities within the vortex were $K_p \sim 3 \times 10^{-4}$ m$^2$ s$^{-1}$, 100 times the open-ocean values observed 10 km away. A 500-m-thick stratified layer with $K_p = (1-5) \times 10^{-4}$ m$^2$ s$^{-1}$ was observed over the flanks of Fieberling Guyot (Toole et al. 1997). These observations covered approximately three-quarters of the circumference and indicated that the mixing was axisymmetric. Using mooring data from the flanks of Fieberling Guyot, Eriksen (1998) found dramatic departures from the Garrett–Munk spectra near the local critical frequency extending $\sim 750$ m above the bed, indicating internal wave scattering.

Our observations over the Kaena Ridge showed that dissipations were highest in hydraulically controlled regions shoreward of the 500-m isobath (Gregg and Carter 2006, unpublished manuscript). Here we concentrate on the second-most dissipative region observed over the Kaena Ridge, an $\sim 400$ m-high seamount atop the southern edge of the ridge. A combination of shipboard observations and numerical simulations is used in this analysis. The experimental context is presented in section 2. The complex flow field around the seamount is explored in section 3. Dissipation measurements are presented in section 4. Results are summarized in section 5.

2. Experiment

a. Location

The Hawaiian Ridge lies almost perpendicular to the direction of propagation for the $M_2$ barotropic tide, making it an important location for $M_2$ internal tide generation. Using an inverse regional model that assimilates altimetry data, Zaron and Egbert (2006) estimate that nearly 25 GW of energy are removed from the barotropic tide over the Hawaiian Ridge. Of this, 19 GW are from the $M_2$ constituent, 3.1 GW are from $S_2$, and 1.4 GW are from $K_1$. Numerical simulations (Merrifield and Holloway 2002; Zaron and Egbert 2006) and observations (Rudnick et al. 2003; Lee et al. 2006) show that the Kauai Channel is one of the strongest generation sites along the ridge. The Kauai Channel and, in particular, the Kaena Ridge, which extends 75–100 km northwest from Oahu, were the location of the Nearfield component of the Hawaii Ocean Mixing Experiment (HOME; Pinkel et al. 2000).

Numerical simulations (Merrifield and Holloway 2002) and observations (Martin et al. 2006; Nash et al. 2006; Rainville and Pinkel 2006) show an $M_2$ internal tide emanating from both flanks of the Kaena Ridge. Approximately one-third of the energy flux observed off the ridge originated on the opposite flank (Nash et al. 2006), which results in energy fluxes having opposite signs at different depths or canceling out, depending on across-ridge location (Nash et al. 2006; Rainville and Pinkel 2006). Dissipation rates were elevated over the
ridge crest (Klymak et al. 2006), although it appears that much of the baroclinic $M_2$ energy radiated away as a low-mode $M_2$ internal tide (Merrifield and Holloway 2002; Lee et al. 2006) or was transferred to other frequencies before radiating off the ridge (Carter and Gregg 2006; Rainville and Pinkel 2006).

A small (unnamed$^1$) seamount, centered at $21°37.2′-21°48′$N, $158°45′-158°27′$W, rises $\sim 400$ m above the southern edge of the ridge, 42 km from the westernmost tip of Oahu (Fig. 1). The seamount is reasonably Gaussian in shape, although its base is elongated in the along-ridge direction (11.6 km, vs 7.9 km in the across-ridge direction). Bottom slopes on the flanks range from 0.05 to 0.3. The seamount is surrounded on three sides by a locally relatively flat plain with depths between 1000 and 1100 m. South of the seamount, the ridge drops away to abyssal depths, with slopes of more than 0.2. Slopes in excess of 0.7 are found near the 1500-m isobath. A small ridge, between two concave scars, runs down the southern face of the Kaena Ridge and is almost in line with the center of the seamount (Fig. 1a).

**b. Observations**

During September of 2002, microstructure and velocity observations were made over the Kaena Ridge as part of HOME. The microstructure data were collected with the loosely tethered deep advanced microstructure profiler (AMP), which has been operated as deep as 1100 m. The AMP, which measures the turbulent ki-

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$^1$ Investigators in the HOME project started referring to this seamount as Munk Mound in honor of Walter and Judith Munk.
netic energy dissipation rate \( e \) (Osborn and Crawford 1980; Wesson and Gregg 1994), temperature, and salinity, was yo-yo cycled from the stern as the ship maintained steerage [ship speeds of at least 1 kt (\( \sim 0.5 \) m s\(^{-1}\))]. Following Osborn (1980), the diapycnal diffusivity can be estimated as

\[
K(z) = \frac{\gamma e(z)}{\langle N^2(z) \rangle},
\]

where the mixing efficiency \( \gamma \) is taken to be \( \leq 0.2 \), and \( \langle N^2(z) \rangle \) is the buoyancy frequency averaged to represent the local background stratification. Altimeters were used to terminate the profiles \( \sim 20 \) m above the rough volcanic bed.

Velocity observations were made with the hydrographic Doppler sonar system mounted on the R/V Revelle. This system, built by the Scripps Institution of Oceanography, consists of two four-transducer Doppler current profilers: one a deep-profiling 50-kHz sonar and the other a high-resolution 140-kHz sonar. Both sonars have transducers that are aligned 30° from the vertical direction, resulting in sidetone interference in the lower 15% of the water column. Here, we consider data from the 50-kHz sonar, which gives velocity measurements sampled every 8.6 m between \( \sim 80 \) and 800 m. Each estimate is the average from an overlapping trapezoidal window having a base length of 25.8 m. The raw data were filtered in time to give overlapping 4-min averages output every 30 s.

Measurements were made along two survey lines across the seamount (Fig. 1), each occupied for 14–16 h, or approximately one semidiurnal tidal period. The cross-ridge line (Cr9; Fig. 1) was occupied from 0905 UTC 20 September to 0945 UTC 21 September 2002, with 22 profiles. Twenty-four profiles were taken in the along-ridge line (Sww; Fig. 1) between 2026 UTC 22 September and 1230 UTC 23 September 2002. A line approximating the 1000-m isobath (Ew2; Fig. 1) was occupied for 12.8 h (15 profiles) and allows for the identification of some of the larger-scale influences of the seamount.

An insert in Fig. 1b defines the along- and across-ridge directions, which are rotated \(-26.24°\). The locations of observations within the two across-seamount surveys (Cr9, Sww) are given as along- and across-ridge distances relative to the seamount summit. Directions throughout the paper are relative to true north. For example, the across-ridge survey line runs north-northeast (NNE) to south-southwest (SSW), and the along-ridge line runs west-northwest (WNW) to east-southeast (ESE). When distinguishing between the sides of the Kaena Ridge, we simply use the descriptors “northern” and “southern.”

**c. Numerical simulations**

Near-bottom and local flow patterns are important to understanding the mixing signals at the seamount. Numerical simulations using the Princeton Ocean Model (POM) were found to agree well with the Doppler sonar measurements and so were used in the analysis. POM is a nonlinear three-dimensional, hydrostatic, sigma-coordinate, primitive equation model (Blumberg and Mellor 1987) and has been used previously to examine internal tide generation at idealized seamount and ridge topography (Holloway and Merrifield 1999), at the Hawaiian Ridge using realistic topography (Merrifield et al. 2001; Merrifield and Holloway 2002; Holloway and Merrifield 2003) as well as internal tide scattering (Johnston and Merrifield 2003; Johnston et al. 2003).

The simulation used in this analysis was made with the latest version of POM (POM2k), had 1-km horizontal grid spacing, 51 sigma levels in the vertical direction, and a density structure from the Hawaii Ocean Time Series (HOT) experiment (22°45′N, 158°00′W) and is discussed further in M. Merrifield et al. (2006, unpublished manuscript). Although the HOT site is \( \sim 100 \) km north of Oahu, the density and stratification are in good agreement with the average of the 46 AMP profiles over the seamount (Fig. 2). Density and stratification also remain nearly constant over the length of the Hawaiian Ridge (Klymak et al. 2006). Horizontal velocities and vertical displacements for \( M_2 \) were simulated by forcing the model boundaries with \( M_2 \) frequency surface elevations obtained from a global inverse model (“TXPO”; Egbert and Ray 2001; Egbert...
and Erofeeva 2002). After a 7-day simulation, by which time energy growth within the model was found to be small (M. Merrifield et al. 2006, unpublished manuscript), harmonic analysis was used to obtain barotropic (depth averaged) and baroclinic (total minus depth averaged) amplitude and phases. Time series can then be constructed for any period.

Our observations were taken from a constantly moving ship; phase errors in the model could lead to significant differences when mapped to the ship’s time/location. To address this concern, a comparison was made with the 4-min-averaged shipboard Doppler sonar. Previous 6-day simulations using POM97, the same resolution, and the same density structure were available for $M_2$, $S_2$, $K_1$, and $O_1$. Energy analysis was not performed on these simulations, but a comparison of the two $M_2$ simulations showed that the general flow patterns agreed, although details, such as shear, differed—which suggests that the 6-day simulation was not as stable as the 7-day simulation. For the comparison with the measured velocity, the benefit of using multiple constituents outweighs the increased noise in the details, and hence we use a superposition of $M_2$, $S_2$, $K_1$, and $O_1$ simulations. This approach also assumes linearity and the absence of background flow. Holloway and Merrifield (2003) report that velocities calculated from the linear superposition of separate $M_2$ and $S_2$ model runs agreed with the results of a model with combined $M_2$ and $S_2$ forcing—suggesting that at their 4-km spacing the tidal currents were approximately linear.

Figure 3 shows the comparison of the measured velocities and the superposition of the four model constituents at 30-s intervals along the ship track during the Sww survey (across the seamount in the along-ridge direction). It indicates good agreement, although the model does not include any background flow or mesoscale features. There appears to be a small phase offset (~1 h) in the 200–500-m-averaged northward velocity component (Fig. 3f) but no phase offset for the eastward velocity component (Fig. 3e). For both components the amplitudes are in good agreement. A similar
phase offset in northward velocity was observed at all the survey lines we occupied over Kaena Ridge [see Carter and Gregg (2006) for the location of the other survey lines], which suggests that spatial comparisons within the model should be internally consistent.

Rainville (2004) compared the $M_2$ POM97 output with the semidiurnal currents measured from R/P Floating Instrument Platform (FLIP) over the ridge crest ($21^\circ40'48"$N, $158^\circ37'45"$W) and found good agreement in amplitude and phase for both barotropic and baroclinic velocities. The barotropic ellipses were fairly rectilinear (the major axis was about 4 times the minor axis), and the modeled and observed orientations agreed to within $\sim15^\circ$. The average velocity in the semidiurnal band from the R/P FLIP deployment (32 days) showed similar amplitudes to the model at all depths, but the difference between observed and modeled phase varied with depth. It was not possible for Rainville (2004) to separate the $S_2$ constituent from the $M_2$, although averaging over 1 month should have minimized the errors from this source.

Across-ridge energy fluxes from a POM simulation with 4-km horizontal grid spacing (Merrifield and Holloway 2002) agreed to within a factor of 2 with absolute velocity profiler (AVP) observations along the 3000-m isobath of the Hawaiian Ridge (Lee et al. 2006). Lee et al. (2006) report, however, that along-ridge energy fluxes and horizontal kinetic energy tended to be underestimated in the model relative to observational values.

Overall, it is justifiable to use the model to interpret our observations. Only the $M_2$ constituent is considered here because it dominates the flow (e.g., Levine and Boyd 2006; Nash et al. 2006). Energy analysis shows that the 7-day simulation is stable (M. Merrifield et al. 2006, unpublished manuscript). Use of a single constituent simplifies the discussion.

3. $M_2$ flow around the seamount

In agreement with simulations of idealized isolated seamounts (Holloway and Merrifield 1999), the modeled $M_2$ flow encountering the seamount as a barrier tends to go around, rather than over, the seamount. This path results in accelerated flows over the flanks of the seamount. In many ways the flow patterns are more complicated than for an isolated seamount because of the location on the $\sim3000$-m-high Kaena Ridge. However, the small size of the seamount means that it does not support the rotationally trapped waves seen at many larger seamounts and that the flow above the seamount summit tends to be largely unaffected by the presence of the seamount. The mode-1 internal radius of deformation ($NH\pi^{-1}f^{-1}$, where $H$ is the water depth) calculated from the stratification over the seamount ($N = 8.2 \times 10^{-3}$ s$^{-1}$) is $\sim40$ km, significantly larger than the $\sim5$-km radius of the seamount. The thickness of a seamount-trapped vortex can be estimated as $O(\sqrt{L/N})$ (Kunze and Toole 1997). Using $L = 2$ km for the radius of the summit plateau gives a potential vortex thickness of $\sim15$ m, below the termination depth of our profiles ($\sim20$ m above the bed).

Although currents over the Hawaiian Ridge are dominated by the $M_2$ tide, the magnitude is modulated by other constituents. The $S_2$ and $K_1$ barotropic across-ridge currents are about 40% of $M_2$, Holloway and Merrifield (2003) report that, at 4-km grid spacing, modeled baroclinic currents are 50% weaker during neap tide than during spring tide. Observed semidiurnal energy fluxes over the Kaena Ridge crest varied from negligible at neap tide up to 3 kW m$^{-1}$ at spring tide (Rainville and Pinkel 2006). Klymak et al. (2006) estimate spring–neap variability in dissipation to be a factor of 2. Around the seamount, the strongest diurnal currents tend to correspond to regions of large $M_2$ currents—for example, over the ESE, WNW, and NNE flanks. This spatial correspondence of $M_2$ and diurnal currents leads to modulation through both constructive and destructive interference, which is not addressed in the following analysis.

a. Topographic steering

The $M_2$ model output shows that flow within the depth range of the seamount tends to be accelerated around the seamount as an along-isobath current, resulting in no significant enhancement of across-isobath flow. The location of the seamount, on top of the Kaena Ridge, adds another level of complication to the flow. In contrast to an isolated seamount, the velocity vectors are not parallel prior to encountering the seamount.

Depth-integrated modeled $M_2$ horizontal kinetic energies [$HKE = \frac{1}{2}(u^2 + v^2)$], calculated using barotropic plus baroclinic velocities] are highest on the along-ridge flanks of the seamount (Fig. 4a). This result suggests that the dominantly across-ridge flow is accelerated around the seamount. The largest flow intensification is confined to near the bed (section 3c). The greater extent for the high HKE northwest of the seamount relative to that southeast of the seamount is likely due to another seamount farther northwest along the ridge (Fig. 1b).

The across-isobath component of a barotropic flow over sloping topography will generate a vertical velocity [$w = (z/H)(u \cdot \nabla H)$; Baines 1982] and hence isopycnal displacements $\eta$. The available potential energy (APE = $\frac{1}{2}N^2\eta^2$) can, therefore, be used as an indicator
of cross-isobath flow. Recall, of course, that baroclinic flows generate isopycnal displacements independent of topographic forcing. Although the barotropic ellipses only partially follow the topography (not shown) and therefore have a cross-isobath component, the APE magnitude is similar over both the seamount flanks and the flat portions of the ridge (Fig. 4b). Across-ridge sections show that the highest APE values tend to be distinct from the bottom (Fig. 5) and consistent with internal tidal beams (section 3b). This result suggests little topographic enhancement of isopycnal displacements at the seamount. The highest modeled APEs are found over the steep southern flank of the Kaena Ridge, in particular in the gully east of the small ridge (Fig. 1). Because the topography is less steep there than to the west of the small ridge, stronger across-isobath flows are required in this region, which are reflected in the HKE (Fig. 4a).

Streamlines provide a useful way to visualize fluid flows. They are defined as a curve that is everywhere tangential to the instantaneous velocity (e.g., Acheson 1990; Kundu 1990). In a steady flow the streamlines are identical to the particle paths; in an unsteady flow, however, the two can be very different. Figure 4c shows selected streamlines through the modeled $M_2$ velocity field at the time when the $M_2$ across-ridge (NNE) barotropic current is strongest. (Because amplitudes and phases from a single constituent were used in the integration, the streamlines also describe the instantaneous flow at the time of maximum SSW current.) The
streamlines were chosen to pass through points 100 m above the topography at distances of ±2, ±2.5, ±3, ±3.5, ±4, ±4.5, ±5, ±6, ±7, and ±8 km along ridge from the seamount center. Streamlines above the seamount summit are mainly across ridge but are not plotted because they obscure the view of the topographically steered streamlines over the flanks.

The flow field near the seamount is obviously complex, with considerable small-scale variability away from the flanks. The streamlines often cross at sharp angles (up to 90°) as a result of markedly different flow fields at different depths. Near the seamount, the shallower streamlines lead the deeper ones in the clockwise rotation during a tidal cycle. Streamlines that pass over the WNW flank closer than ~4 km to the summit flow across the SSW flank of the seamount before turning north to cross the WNW flank. On the ESE flank, the deeper streamlines turn sharply east at the northeastern
edge of the seamount. South of the seamount, 13 of the 20 plotted streamlines cross the southern edge of the Kaena Ridge above the gully east of the small ridge, the location of the largest APE.

It seems unlikely that currents encountering such a pronounced feature as the southern edge of the ridge would be symmetric in the positive and negative across-ridge directions, as described by a single constituent ellipse. More topographic steering would be expected for currents flowing toward the ridge than toward the open ocean. Examination of the residuals currents left after the harmonic analysis may reveal marked asymmetries.

The streamlines in Fig. 4c are colored by the horizontal current speed \(\left(\sqrt{u^2 + v^2}\right)\) that a particle moving through the instantaneous velocity field would experience. The acceleration around the seamount is clear, with speed reaching 0.4 m s\(^{-1}\) over the flanks as compared with background flows of \(\sim 0.05\) m s\(^{-1}\).

### b. Tidal beams

Two internal tidal sheets,\(^2\) one from either side of the ridge, cross above the Kaena Ridge. These sheets are horizontally coherent in the along-ridge direction and intersect the seamount. Local generation of internal tides also occurs at the seamount.

St. Laurent and Garrett (2002) suggest that internal tides are generated by cross-isobath flow, but critical topography is required to generate high vertical modes and hence tidal beams. Both criteria are met at the Kaena Ridge, and internal tidal beams have been observed in measured energy fluxes (Nash et al. 2006; Rainville and Pinkel 2006) and in tidally averaged HKE (Martin et al. 2006). Numerical simulations and observations found generation on both the northern and southern edges of the ridge, with approximately one-third of the energy flux observed away from the ridge having propagated over the ridge crest after being generated on the far edge (section 2a). Nash et al. (2006) found that the magnitudes of the energy flux crossing the ridge from generation sites on the northern and southern sides of the ridge were similar but of opposite sign; hence, the total observed energy flux was nearly zero over the ridge crest. Likewise, the APE was reduced over the ridge crest, but the HKE was enhanced where the two beams crossed. The energetics of this horizontally standing, vertically propagating internal tide over the ridge crest are examined in detail by Nash et al. (2006).

The internal tidal structure in the vicinity of the seamount can be inferred from a series of modeled \(M_2\) horizontal kinetic energy, available potential energy, and across-ridge energy flux sections (Fig. 5). The across-ridge energy fluxes are calculated, following Kunze et al. (2002), as

\[
F_E = (v_{\text{across}}' p')_\phi
\]

where \(v_{\text{across}}'\) is the vertically de-meaned across-ridge velocity, \(p'\) is the vertically de-meaned pressure perturbation:

\[
p' = \int_0^H \sqrt{2} c_\phi \eta(\xi) \, d\xi - \frac{1}{H} \int_0^H \sqrt{2} c_\phi \eta(\xi) \, d\xi \, dz,
\]

\(H\) is the water depth, and \((\cdot)_\phi\) indicates an average over a tidal cycle.

Over the saddle ESE of the seamount (Figs. 5a–f), tidal beams from both edges of the ridge form two continuous along-ridge tidal sheets. In agreement with Nash et al. (2006), the model output shows that the energy flux (Fig. 5, right column) and APE (Fig. 5, middle column) decrease over the ridge crest where the two tidal sheets intersect. These sections, however, show that the SSW-directed tidal sheet generated on the north edge is stronger, resulting in a net negative flux over the center of the ridge. Flux divergence between the upward–SSW beam and the downward–NNE beam indicates that generation occurs between 1200- and 1600-m depth on the northern side of the ridge in all six sections. The flux divergence on the southern side of the ridge tends to occur at the break in slope between the ridge crest and the steep southern side of the ridge, although it is not as pronounced because the energy flux crossing the ridge is swamped by the SSW energy flux within \(\sim 2\) km of the generation site. The exception to the generation occurring at the top of the southern flank of the ridge is at the seamount, where it appears that generation occurs at the summit (Fig. 5i) or over the WNW flank (Fig. 5o). It is possible that reflection of the SSW tidal sheet occurs at the seamount; however, the role of this mechanism is hard to assess in such a complex environment.

Tidal beams are identifiable in the HKE, APE, and energy flux, but the beam structure is clearest in the HKE plots (Fig. 5, left column) because HKE is enhanced, rather than decreased, by the nearly horizontal standing wave structure over the ridge crest. The vertical location of the HKE beams within the water column is remarkably consistent over the 20 km represented in Fig. 5. At 25 km across ridge, all of the \(\sim 200\)-m-thick upward and NNE beams lie between 200- and 500-m depth, with a tendency for the beams to be shal-

\(^2\) The term “tidal sheet” rather than “tidal beam” is used, so as to emphasize its three-dimensional nature.
lower toward the ESE. Likewise, at ~15 km the beams all lie between ~150 and 500 m. Carter and Gregg (2006) found that averaged HKE from a series of along-ridge Doppler sonar runs was consistent with tidal sheets crossing the ridge. All of these findings strongly suggest that the $M_2$ internal tide is along-ridge coherent and should be considered as three-dimensional sheets rather than as a series of distinct beams.

Katsumata (2006) derived the barotropic and baroclinic energy terms for a POM simulation on the northwest Australian shelf; M. Merrifield et al. (2006, unpublished manuscript) evaluated these energy terms for Kaena Ridge and found that 90% of the barotropic flux divergence over Kaena Ridge is converted into baroclinic tidal energy. In Fig. 6 we present the barotropic conversion term (based on $u \cdot \nabla H$; Baines 1982) for the seamount area. The topography and region of barotropic conversion on the northern side of the ridge is nearly linear, suggesting that the southward tidal sheet crossing the seamount can be considered as having a line source. The southern edge of the ridge consists of a number of landslide scars (Fig. 1b), which results in the horizontal structure of the barotropic conversion being less linear and more intermittent. This implies that a series of “point” sources may be a more appropriate description of the generation on the southern side.

Figure 6 shows that barotropic conversion occurs on the NNE flank of the seamount, although it is weaker than at either edge of the ridge (~1.2 W m$^{-2}$ as compared with >2 W m$^{-2}$). The along-ridge coherence of the northward tidal sheet suggests that generation at the southern edge of the ridge (~1000 m depth) and generation at the seamount (~600 m depth) both contribute to the same tidal sheet, reinforcing the three-dimensional nature of the internal tide.

For the current analysis, the key elements are that the northward and southward tidal sheets intersect the seamount and that they produce a quasi-stationary-wave structure in the horizontal plane. This standing-wave structure results in reduced energy flux but increased HKE over the ridge crest, which in turn makes HKE a better tracer for the tidal sheets over the ridge crest.

c. Flow evolution

Flow in the region of the seamount is strongly baroclinic and is accelerated around the seamount, and horizontally coherent tidal sheets are generated at both edges of the ridge. Now the evolution of the flow over an $M_2$ tidal cycle is considered. Figure 7 shows the modeled $M_2$ HKE corresponding to the along- and across-ridge survey lines. Only one-half of the tidal cycle is displayed, because the square of a sinusoidal function (i.e., a single constituent) is symmetric over one-half of the original period. Figure 7a shows the structure as the modeled barotropic flow turns from SSW to NNE. Each subsequent panel is offset by one-twelfth of an $M_2$ period.

The evolution is easiest to follow starting with Fig. 7b, where the flow is weakest. Horizontal kinetic energies of approximately less than 0.025 m$^2$ s$^{-2}$ are confined mainly to the tidal sheets described in section 3b, and, as discussed in that section, the intersection of the two beams over the NNE flank of the seamount results in elevated HKE. In the across-ridge direction, this tidal sheet appears to be distorted upward by the presence of the seamount. As the across-ridge flow increases, this layer migrates down and widens, with the HKE reaching ~0.05 m$^2$ s$^{-2}$. As the across-ridge barotropic current approaches its maximum (Fig. 7c), near-bottom intensification of currents occurs over the seamount summit. This intensification grows and spreads to the flanks (Fig. 7d). As the across-ridge flow ebbs, the near-bottom intensified regions contain HKEs in excess of 0.09 m$^2$ s$^{-2}$ that propagate down the flanks of the seamount (Figs. 7a,e,f).

The maximum modeled currents for $S_2$, $K_2$, and $O_1$ are approximately less than 0.2 m s$^{-1}$, as compared with 0.5 m s$^{-1}$ for $M_2$. The $S_2$ HKE structure is similar to
that presented in Fig. 7, although the near-bottom values are no larger than those in the mid-water-column layer. The diurnal constituents \( (K_1, O_1) \) primarily show bottom-intensified HKE over the flanks and off the seamount. This diurnal intensified layer is thickest (reaching 200 m) over the ESE flank of the seamount.

Although the formation of the near-bottom intensified regions of HKE corresponds to the thickening and deepening of the tidal sheet intersecting the seamount, these two effects do not appear to be directly related. Flow acceleration caused by topographic steering appears to be a more likely explanation. For the HKE

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**Fig. 7**. The \( M_2 \) HKE from the POM across the seamount in the along- and across-ridge directions. These lines correspond to extensions of the Sww and Cr9 survey lines. The tick marks along the top of each HKE slice are spaced every 2 km in the along- and across-ridge directions. Only one-half of the tidal cycle is displayed, because HKE is symmetric over this period for a single constituent. Panels (a)–(f) are separated by 30° (12.42/12 h). The thick black lines in the circles indicate the tidal phase for that panel, and the gray lines indicate the phase one-half of an \( M_2 \) cycle later with the same HKE distribution. The upper hemisphere of the phase indicator corresponds to positive across-ridge barotropic (northeastward) flow.
intensification to be caused directly by focusing of the across-ridge internal tide on to the along-ridge flanks, the azimuthal reflection would need to be across isobath. Linear theory applied to an upward-propagating beam intersecting topography, such as one generated on the northern edge of the ridge intersecting the seamount, does not allow this orientation.

Eriksen (1982) presents a linear theory for internal waves with a nonzero azimuthal angle reflecting off a sloping planar bottom in a semi-infinite domain. It predicts that upon reflection the wave azimuthal angle will be more across the isobaths (i.e., along ridge) than the incident wave. However, implicit in this formulation is that the incident wave is isobaths (i.e., along ridge) than the incident wave. Because of the formulation of Eriksen (1982), there is no topographic variation in the along-isobath direction, and consequently all changes to the horizontal wavenumber must be in $k_{across	ext{ isobaths}}$. Decreasing the across-isobath scale (increasing $k_{across	ext{ isobaths}}$) makes the reflected wave more normal to the isobaths. This magnitude of the turning increases as the wave characteristic slope tends toward the bottom slope.

The upward propagation of the tidal beam emanating from the northern ridge break that intersects the seamount results in the opposite azimuthal turning: the vertical scales increase upon reflection (Fig. 8d), $k_{across	ext{ isobaths}}$ decreases, and the reflected wave is more parallel to the isobaths (i.e., across ridge) than the incident wave. Therefore, azimuthal reflection of the internal tide does not appear to be responsible for the near-bottom intensification; rather, the acceleration as the flow is steered around the seamount seems the likely mechanism. Linear theory may be inappropriate for reflection from such a complex region, but it seems unlikely that the nonlinear effects would change the sense of the azimuthal rotation and hence our conclusion.

4. Observed mixing over the seamount

Turbulent kinetic energy dissipation rates observed over the seamount ranged from $10^{-10}$ to $2 \times 10^{-8}$ W kg$^{-1}$. The average over the 46 profiles was $\bar{\epsilon} = 6.2 \times 10^{-8}$ W kg$^{-1}$, with a 95% bootstrapped confidence interval from $5.8 \times 10^{-8}$ to $6.7 \times 10^{-8}$ W kg$^{-1}$. The along-(Sw) and across-ridge (Cr) surveys both had similar average observed dissipation ($\bar{\epsilon} = 6.2 \times 10^{-8}$-$6.3 \times 10^{-8}$ W kg$^{-1}$). However, dividing the microstructure profiles into along-track bins showed differences in binned averaged dissipation $\bar{\epsilon}$ with distance from the seamount summit and with orientation relative to the Hawaiian Ridge (Fig. 9, Table 1). The asymmetry in the across-ridge direction seems robust, whereas the asymmetry in the along-ridge direction may be due to undersampling the flow.

The 2–3.6-km-long bins divide the surveys into two regions over the summit as well as regions over the flanks. Most bins adequately span the range of tidal phases (circles in Fig. 9). The effective tidal phase, as defined by Levine and Boyd (2006), varies linearly between consecutive zero crossings of across-ridge barotropic velocity. Given that the total velocity from POM agrees with the Doppler sonar observations (Fig. 3), we calculate the effective tidal phase using four-constituent ($M_2$, $S_2$, $K_1$, and $O_1$) POM barotropic across-ridge velocity at the seamount summit. Because of the combination of constituents, the period for a $2\pi$ change in effective phase is variable, but the dominant $M_2$ constituent means that the average is 12.42 h. The standard deviation over 1 yr is 39 min. For the tidal cycles sampled over the seamount, the range was 12.10–12.52 h.

Most of the dissipation occurs in the lower half of the

3 The overbar is used to denote an average using all 46 AMP profiles over the seamount, whereas () denotes an average over a 2–3.6-km-long along-track bin.

4 Levine and Boyd (2006) use the global TPXO.5 model, which has 2° horizontal resolution, to obtain the barotropic velocities. However, over the seamount summit the TPXO across-ridge velocity lags the POM across-ridge barotropic velocity by 50°–70° (53.5° for $M_2$ only). Because this is a region of rapidly changing topography, the differences between models with different topographic representations are not surprising.
water column, with the largest \( \langle \varepsilon \rangle \) observed over the ESE flank of the seamount. The tendency for the flanks to have higher dissipations differs from a number of previous observations at isolated seamounts. At Fieberg and Toole’s 1997) was greater than dissipation over the flanks (2 \( \times \) 10^{-9} W kg^{-1}; Toole et al. 1997). Lavelle et al. (2004) report a column of elevated mixing \( \mathcal{K}_p = 4 \times 10^{-4} \text{ m}^2 \text{s}^{-1} \) above the 200-m-deep summit of Irving Seamount. This difference is probably because this seamount does not appear to support a trapped anticyclonic vortex over its summit.

Fig. 9. Turbulent kinetic energy dissipation rate \( \varepsilon \) averaged into along-track bins: (a) the along-ridge section (Sww), and (b) the across-ridge section (Cr9). The circles above each panel indicate the tidal phase (relative to the across-ridge barotropic current; see text) of profiles in each bin. Upper hemispheres of the circles indicate positive across-ridge (NNE) velocity. The numbers in the circles give the number of profiles in each average. In both panels, the four regions are defined as less than \(-2\) km, from \(-2\) to \(0\) km, from \(0\) to \(2\) km, and greater than \(2\) km.

<table>
<thead>
<tr>
<th>Region</th>
<th>( 10^6 \times \langle \varepsilon \rangle ) (W kg^{-1})</th>
<th>( 10^4 \times \langle \mathcal{K}_p \rangle ) (m^2 s^{-1})</th>
<th>( 10^5 \times \langle \mathcal{N}^2 \rangle ) (s^{-2})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Along-ridge direction (Sww)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>2.7 (2.4–3.0)</td>
<td>6.2 (5.0–7.4)</td>
<td>5.2 (4.7–5.8)</td>
</tr>
<tr>
<td>2</td>
<td>4.5 (3.4–5.7)</td>
<td>4.1 (3.2–5.1)</td>
<td>6.6 (6.0–7.2)</td>
</tr>
<tr>
<td>3</td>
<td>2.0 (1.8–2.2)</td>
<td>1.3 (1.1–1.5)</td>
<td>7.1 (6.4–7.8)</td>
</tr>
<tr>
<td>4</td>
<td>13.6 (11.9–15.5)</td>
<td>39.9 (32.4–48.4)</td>
<td>5.1 (4.6–5.6)</td>
</tr>
<tr>
<td>Across-ridge direction (Cr9)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>5.1 (4.5–5.8)</td>
<td>13.5 (11.2–16.2)</td>
<td>4.8 (4.4–5.2)</td>
</tr>
<tr>
<td>2</td>
<td>3.4 (2.9–3.8)</td>
<td>2.2 (1.9–2.6)</td>
<td>6.7 (6.1–7.3)</td>
</tr>
<tr>
<td>3</td>
<td>6.7 (4.8–9.0)</td>
<td>6.9 (4.8–9.1)</td>
<td>6.7 (6.1–7.3)</td>
</tr>
<tr>
<td>4</td>
<td>8.9 (7.8–10.0)</td>
<td>21.9 (18.9–26.0)</td>
<td>4.8 (4.4–5.3)</td>
</tr>
</tbody>
</table>

The diapycnal diffusivity [Eq. (2)] is calculated with \( \langle \mathcal{N}^2(z) \rangle \), defined as the average buoyancy frequency profile over the 4–7 AMP profiles in each along-track bin. This calculation gives an average over the summit of \( \mathcal{K}_p = 1.3 \times 10^{-3} \text{ m}^2 \text{s}^{-1} \), with a 95% confidence interval from 1.2 \( \times \) 10^{-3} to 1.5 \( \times \) 10^{-3} m^2 s^{-1}. Although the pattern of \( \langle \mathcal{K}_p \rangle \) (not shown) is similar to that shown in Fig. 9, decreasing stratification with depth causes, for a fixed \( \varepsilon \) value, larger \( \mathcal{K}_p \) values above the flanks than over the summit. This, in turn, results in the differences between dissipation and diapycnal diffusivity shown in Table 1; for example, in the across-ridge survey, \( \langle \varepsilon \rangle \) in bin 3 was 30% larger than in bin 1 but \( \langle \mathcal{K}_p \rangle \) in bin 3 was one-half of that in bin 1 (Table 1).

**a. Dissipation higher on ESE flank than WNW**

The dissipation measurements from the along-ridge line (Fig. 9a) showed a substantial asymmetry, with the highest values observed over the ESE flank of the seamount. Modeled \( \mathcal{M}_2 \) shear variance also shows an asymmetry, with more shear on the ESE flank than the WNW.

An almost fivefold difference in dissipation was observed between the two along-ridge flanks (Table 1). The biggest difference was below 775 m, where \( \langle \varepsilon \rangle \) in bin 4 (ESE flank) ranged from 1.1 \( \times \) 10^{-7} to 6.3 \( \times \) 10^{-6} W kg^{-1}, as compared with from 1.1 \( \times \) 10^{-10} to 3.1 \( \times \) 10^{-7} W kg^{-1} over the WNW flank (bin 1). Sixty-four percent of the observed dissipation over the ESE flank was from below 775 m in a single profile, which had an along-ridge distance of 4.0 km.

The dissipation rate is dynamically related to vertical shear. The stability of the 7-day simulation allows an estimation of first-difference \( \mathcal{M}_2 \) shear variance \( [\mathcal{S}^2 = (\Delta u/\Delta z)^2 + (\Delta v/\Delta z)^2] \). Regions of high shear variance
Fig. 10: Modeled $M_2$ first-differenced shear variance $S_e'$ averaged over a tidal cycle for (a) the along-ridge survey and (b) the across-ridge survey. The bin-averaged dissipation rates are overplotted. The across-ridge bins are the same as in Fig. 9.

(Fig. 10) tend to follow the regions of high HKE (Fig. 7) because of the small vertical scales in the tidal sheets and the near bed where HKE is large. Figure 10a shows that there is more modeled near-bed shear variance on the ESE flank than on the WNW flank. The WNW flank has higher modeled shear variance in the mid-water-column than is found ESE of the seamount; the majority of this shear is at distances of less than $4\,\text{km}$ and hence was not sampled by the microstructure survey.

Although the modeled shear variance pattern agrees with the observed dissipation pattern, the fivefold difference in magnitude may be a result of undersampling. The five profiles in bin 1 map onto the flow structures described in Figs. 7a–c, which miss the major bottom-intensified currents and in all likelihood the highest dissipations on the WNW flank, whereas the majority of dissipation observed on the ESE flank came from a single drop that appears to have coincided with the high near-bed HKE region described in Fig. 7e.

b. Across-ridge $\varepsilon$ consistent with tidal beam

The observed dissipation over the NNE half of the seamount was elevated and more concentrated toward the bed relative to that on the SSW side. This pattern is consistent with the internal tidal sheets described in section 3b, and, therefore, the asymmetry appears to be robust.

Average dissipation rates from the across-ridge survey were higher over the NNE side of the seamount than over the SSW side (Table 1, Fig. 9b). In bin 4, $78\%$ of the dissipation occurred below $750\,\text{m}$—a depth range of $250\,\text{m}$. The majority ($86\%$) of the dissipation in bin 3 was within a $105\,\text{m}$-thick layer adjacent to the summit of the seamount. Over the southern side of the seamount the high dissipation tended to be removed from the topography. In bin 2 dissipations greater than $1.8 \times 10^{-7}\,\text{W kg}^{-1}$ were mainly between $445$ and $530\,\text{m}$, and bin 1 had three distinct regions of high $\varepsilon$: $425-480, 670-740$, and $936-1000\,\text{m}$.

Figure 10b shows the bin-averaged dissipation rates overplotted on the tidally averaged vertical shear variance. The tidal beam structure from Fig. 5m is visible in the shear as well, although there is clearly shear not associated with the tidal beams.

Over the NNE flank, where the two tidal sheets intersect, the high measured dissipation corresponds to regions of high modeled shear. Not all regions of high shear correspond to observed dissipation. This, however, is not surprising, because the shear is averaged over an entire tidal cycle whereas the dissipation was measured at discrete times; also, differences in vertical resolution result in such differences even between concurrent observations (e.g., Carter and Gregg 2002). Over the SSW side of the seamount the tidally averaged shear (Fig. 10b) follows the upper edge of the HKE beam (Fig. 5m) and is higher in the water column than the observed dissipation. The fact that the dissipation in bins 1–3 tends to have the same shape as the tidal beam shear but is just offset, along with vertical migration of the tidal beam within a tidal cycle (section 3b), suggests that the dissipation is probably related to the tidal beams. The two deeper dissipation regions in bin 1 fall within the depth range of this bin and may likely be due to boundary layer processes.

c. Along-ridge tidal beam

Dissipation measurements following the 1000-m isobath SSW of the seamount (EW2; Fig. 1), adjacent to the edge of the ridge, are suggestive of along-ridge internal tide generation caused by topographic steering of currents around the seamount.

The four AMP profiles of the second run along the 1000-m isobath showed $\sim100\,\text{m}$-thick patches of high dissipation, which sloped upward across isopycnals with distance southeastward (Fig. 11a). In the last three profiles (18347–18349), these dissipation patches were in
midwater, distinct from the bottom boundary layer. In fact, elevated boundary layers were often observed beneath them (Figs. 11a,b). The elevated HKE associated with the tidal sheet generated on the northern side of the ridge lies between 200- and 600-m depth (maximum at ∼300 m, Figs. 5j,m) as it crosses the plane of this along-ridge survey line; hence, the observed dissipations are not related to this tidal sheet. Although this survey line ran along the base of the 5-km-radius seamount, the dissipation patches are unrelated to side boundary processes. A comparison of the location of the profiles with the across-ridge flow direction (Fig. 11c) shows that profile 18348 is most in line with the summit and therefore would have dissipation highest in the water column.

The location of this dissipation within the water column is in remarkable agreement with an along-ridge $M_2$ characteristic (Fig. 11b). As noted (section 3b), the dominant internal tide generation is across ridge, but near-bottom $M_2$ currents were steered around the seamount (section 3a), which results in enhanced along-ridge currents. These four profiles were taken when the barotropic across-ridge current at the seamount was positive and increasing toward its maximum value (Fig. 11c). This is a period for which the model suggests a strong along-ridge component to the flow south of the seamount. Figure 11d shows that the instantaneous flow at 1532 UTC 21 September 2002, the time corresponding to the average time of the last three profiles (7–9). The red stars give the location of the AMP profiles.

**Fig. 11.** (a) Three-dimensional view of the seamount showing the dissipation measured by the four AMP profiles (18346–18349, labeled using the last digit) from the second transect along the 1000-m isobath SSW of the seamount. The transect was run WNW to ESE. The light purple grid is the same as in Fig. 1a. (b) Dissipation rates $\varepsilon$ in comparison with an $M_2$ characteristic (black line). (c) Barotropic across-ridge velocity, showing the time and duration of the AMP profiles. (d) Map showing the criticality of the 75-m-resolution topography at $M_2$ frequency and the measured $N^2$ profile. White indicates subcritical slopes, and gray indicates critical or supercritical slopes. Streamlines show how sharply the flow is steered around the seamount. The streamlines are $M_2$ only and are calculated at a phase corresponding to the average time of the last three profiles (7–9). The red stars give the location of the AMP profiles.
survey line contains regions of both sub- and supercritical slopes. The topographic steering of the tide around the seamount appears to create an along-ridge flow over this region of critical topography and hence along-ridge internal tide generation that is consistent with our dissipation observations.

5. Summary

Microstructure and velocity observations were made over an ∼400-m-high seamount on the Kaena Ridge as part of the Hawaii Ocean Mixing Experiment. Observations and numerical simulations both find Kaena Ridge to be a site of strong $M_2$ barotropic-to-baroclinic conversion, with internal tides generated on both flanks.

Velocities from a linear combination of $M_2$, $S_2$, $K_1$, and $O_1$ 1-km-resolution Princeton Ocean Model simulations were in good agreement with observed Doppler sonar velocities. The $M_2$ tidal constituent dominates the velocities around Hawaii, and $M_2$ simulations were used to interpret the microstructure observations.

The $M_2$ simulations show that most of the flow encountering the seamount as a barrier goes around, rather than over, the seamount. This pattern results in accelerated along-isobath flow over the flanks. The flow patterns are more complicated than for an isolated seamount, because of the presence of the ∼3000-m-high Kaena Ridge. However, the seamount does not support rotationally trapped waves because of its small size.

Two internal tidal sheets, which are coherent in the along-ridge direction, cross above the Kaena Ridge. These tidal sheets intersect the seamount, where local generation also occurs. Over the ridge crest the tidal sheets form a quasi-across-ridge standing-wave structure, because they travel in opposite directions from generation at either side of the ridge. Energy flux and available potential energy are decreased by the superposition of the two waves, whereas horizontal kinetic energy is increased.

Two survey lines over the seamount, one in the along-ridge direction and one in the across-ridge direction, were occupied for a semidiurnal period. The average observed dissipation rate was $\overline{\varepsilon} = 6.2 \times 10^{-5}$ W kg$^{-1}$, and the average diapycnal diffusivity was $K_n = 1.3 \times 10^{-3}$ m$^2$ s$^{-1}$. This was the second most dissipative environment measured on the Kaena Ridge, after hydraulic flows near Oahu. Average dissipation rates over the seamount were approximately 3 times those over the adjacent saddle.

Bin-averaging along the survey lines showed asymmetries across the seamount. The highest dissipations ($\varepsilon > 10^{-5}$ W kg$^{-1}$) were observed over the east-southeastern flank of the seamount; at similar depth, bin-averaged dissipations over the west-norwestern flank were less than $10^{-9}$ W kg$^{-1}$. The $M_2$ shear variance calculated from the 1-km model output shows a similar asymmetry. Observed dissipation was higher and more concentrated toward the bed on the north-northeastern side of the seamount than on the south-southwestern side. This dissipation pattern was consistent with the across-ridge tidal beam structure.

Dissipation measurements following the 1000-m-isobath south-southwest of the seamount suggest along-ridge internal tide generation. The location of ∼100-m-thick high-dissipation regions is consistent with an $M_2$ characteristic parallel with the edge of the ridge. This situation appears to be caused by topographic steering that creates an along-ridge current over a region of critical topography northwest of the seamount.

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