Turbulence Driven by Reflected Internal Tides in a Supercritical Submarine Canyon

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ABSTRACT: The La Jolla Canyon System (LJCS) is a small, steep, shelf-incising canyon offshore of San Diego, California. Observations conducted in the fall of 2016 capture the dynamics of internal tides and turbulence patterns. Semidiurnal (D2) energy flux was oriented up-canyon; 62% ± 20% of the signal was contained in mode 1 at the offshore mooring. The observed mode-1 D2 tide was partly standing based on the ratio of group speed times energy cgE and energy flux F. Enhanced dissipation occurred near the canyon head at middepths associated with elevated strain arising from the standing wave pattern. Modes 2–5 were progressive, and energy fluxes associated with these modes were oriented down-canyon, suggesting that incident mode-1 waves were back-reflected and scattered. Flux integrated over all modes across a given canyon cross section was always onshore and generally decreased moving shoreward (from 240 ± 15 to 5 ± 0.3 kW), with a 50-kW increase in flux occurring on a section inshore of the canyon’s major bend, possibly due to reflection of incident waves from the supercritical sidewalls of the bend. Flux convergence from canyon mouth to head was balanced by the volume-integrated dissipation observed. By comparing energy budgets from a global compendium of canyons with sufficient observations (six in total), a similar balance was found. One exception was Juan de Fuca Canyon, where such a balance was not found, likely due to its nontidal flows. These results suggest that internal tides incident at the mouth of a canyon system are dissipated therein rather than leaking over the sidewalls or siphoning energy to other wave frequencies.

KEYWORDS: Continental shelf/slope; Energy transport; Internal waves; Turbulence; Tides; In situ oceanic observations

1. Introduction

Submarine canyons are common along continental margins worldwide (Allen and Durrieu de Madron 2009; Harris and Whiteway 2011). They exhibit an array of geometries and configurations—some are straight and narrow, others are wider and meandering; some extend to and incise the continental slope region (slope canyons), others are confined to the shelf (shelf canyons). Wherever they occur, canyons act as conduits that connect the shallower shelf to deeper waters offshore. Generally more narrow than the adjacent shelf is wide, canyons tend to cause ageostrophic flows that enhance cross-shelf exchange and strengthen wind-driven up- and downwelling flows (Klinck 1989; Hickey 1997; Kämpf 2009; Allen and Hickey 2010). Breaking of focused internal waves, hydraulic control over ridges and bends, and enhanced ageostrophic flows all lead to small-scale turbulent processes such that canyons are known to be hotspots of enhanced turbulent dissipation and diapycnal diffusivity (Carter and Gregg 2002; Lee et al. 2009; Gregg et al. 2011; Alford and MacCready 2014; Waterhouse et al. 2017; Alberty et al. 2017; Hall et al. 2017).

The classical theory of turbulence driven by internal waves in canyons is presented by Hotchkiss and Wunsch (1982). They theorize that internal waves incident at a canyon’s offshore extent (mouth) reflect downward off of steep sidewalls and toward its nearshore terminus (head) over an axis (thalweg) that slopes gently in the along-canyon direction. Wave focusing leads to energy convergence, high velocities, and shear near the canyon floor and head that generates turbulence there. Waves that encounter an axis slope close to their own angle of propagation (critical) are also expected to scatter and break near the bottom (Nazarian and Legg 2017). In the handful of canyons in which turbulence has been observed to date, the expected patterns of enhanced turbulence near the bottom and head of a canyon do occur (Petruccio et al. 1998; Kunze et al. 2002; Carter and Gregg 2002; Lee et al. 2009; Kunze et al. 2012; Wain et al. 2013; Hall et al. 2014, 2017), but so do patterns of turbulent enhancement at middepths and in other regions of the canyon (Gregg et al. 2011; Zhao et al. 2012; Hall et al. 2014; Waterhouse et al. 2017; Aslam et al. 2018; Alberty et al. 2017).

Several studies of internal waves in canyons have reported patterns of energy partitioning consistent with partly standing waveforms in canyon systems (Petruccio et al. 1998; Zhao et al. 2012; Hall et al. 2014; Waterhouse et al. 2017; Alberty et al. 2017). Many canyons steepen approaching their nearshore terminus or have sharp bends that can reflect incident waveforms. As suggested by observations made by Alberty et al. (2017) and ray tracing models by Nazarian and Legg (2017), the superposition of incident and reflected waves may set the stage for elevated shear, strain, and breaking events at middepths.

Studies of internal wave reflection in canyons are limited, but studies of reflection from steep continental slopes show that incident mode-1 waves are accompanied by back reflection and scattering to modes 2 and higher. Globally, mode-1 internal tides encountering the continental slope transmit 20% of their energy onto the shelf, while 40% scatters to higher modes and 40% reflects back to the ocean interior (Kelly et al. 2013). Locally, these percentages vary depending on stratification...
and the geometry of the slope. Thorpe (2001) presents an analytical treatment of reflection from a rough slope, and finds that the shear in scattered waves scales with steepness and roughness of the slope. Observations by Nash et al. (2004) over a corrugated continental slope in the mid-Atlantic Bight show that the convergence of low-mode semidiurnal onshore energy flux is balanced by high-mode offshore flux; they postulate that enhanced near-bottom mixing offshore of the steep topography and enhanced diffusivity extending over the bottom ~1000 m are a direct consequence of the reflection process.

With steep, supercritical sidewalls and thalwegs that generally steepen approaching the canyon head, submarine canyons are prime locations in which reflection and scattering of low-mode incident waves may give rise to dissipative dynamics. If the mechanism by which incident waves dissipate their energy in a reflective canyon is via instability during middepth strain events rather than critical near-bottom breaking, the time-averaged vertical profile of dissipation rate $\epsilon$ would be enhanced throughout the water column rather than increasing toward the bottom. Given the prevalence of canyons along the continental margins, this vertical distribution of dissipation would alter the distribution of heat and other tracers on global scales (Melet et al. 2016). Furthermore, due to their near-ubiquitous presence, canyons may be responsible disproportionately in the dissipation of global mode-1 internal tides, and the global assessment of mode-1 internal tide dissipation offered by Kelly et al. (2013) might be improved accordingly by including these effects.

The observations presented here were collected in the La Jolla Canyon System (LJCS), a shelf-incising canyon located offshore of Scripps Institution of Oceanography in La Jolla, California, United States. Unlike previously studied systems like Monterey Canyon (MC) and Eel Canyon (EC) that incise the continental slope, the LJCS is largely confined to the shelf region. Where previously studied “slope canyons” often contain, connect to, and foster mixing along deep isopycnals relevant for the overturning circulation of the deep ocean (Kunze et al. 2012), “shelf canyons” like the LJCS are impactful for regional dynamics and interactions with the inner shelf and nearshore regions. Turnover rates, circulation patterns, and sediment transport in nearshore waters are affected by nearby canyon topography and the surface wave refraction associated with it (Shepard and Inman 1950; Hickey et al. 1986); connection to deep, nutrient rich waters influences biological productivity within canyons and in canyon-adjacent shallows (Vetter 1994; Vetter and Dayton 1999; Allen et al. 2001; Vetter et al. 2010; Alford and MacCready 2014; Kavanaugh et al. 2015); sediments along the shelf adjacent to canyon systems have a tendency to flow toward canyon incisions and offshore (Liu et al. 2002; Puig et al. 2003); the presence of shelf valleys can enhance cross-shore transport driven by alongshore winds (Zhang and Lentz 2017, 2018).

Understanding the dynamics at play within shelf-incising canyon systems both informs our ability to parse out these complicated multidisciplinary puzzles and expands our understanding of canyon physics in general. Comprehensive measurements can be made in these shallower and more easily accessible systems to examine the effects of variable stratification and forcing, and the physical intuition can be applied and generalized to submarine canyons throughout the ocean.

2. Experiment layout

The La Jolla Canyon System is a shelf-incising canyon that terminates in the La Jolla Cove offshore in La Jolla, California (Fig. 1). The greater LJCS is 24 km long, 3 km wide, and 1000 m deep at its farthest offshore extent. Between 8 and 14 km from its head the canyon meanders to the south. Two kilometers from its head, the LJCS bifurcates into two branches: the La Jolla Canyon (LJC) extends to the south parallel to the direction of the main canyon axis while the Scripps Canyon (SC) branches off to the north in a direction nearly perpendicular to LJC. Located 1.6 km away, the headland of Point La Jolla sits south of the canyon head blocking strong alongshore flows in the vicinity. The LJCS is situated in a portion of the shelf break where it and much of the surrounding bathymetry is steep and supercritical to the D$_2$ internal tide ($\alpha = z_{bottom}/z_{wave} > 1$; red areas in Fig. 1, inset). Although myriad internal waves are generated at complicated bathymetry in the southern California Bight, some portion of the internal waves observed in LJCS likely propagate up from a steep escarpment to the southwest of the LJCS (Ponte and Cornuelle 2013).

The La Jolla Internal Tide Experiment 2 (LaJIT2) was carried out in the LJCS in September 2016 [Fig. 1; see Alberty et al. (2017) for LaJIT1 experiment details]. The experiment consisted of 7 days of shipboard surveys and 3 months of moored data collected from three locations. On 10 September, three moorings were deployed along the axis of the canyon. The deepest mooring (MP1) was located at 415-m depth onshore of the meander and consisted of a downward-looking 75-kHz Workhorse acoustic Doppler current profiler (ADCP) and a McLane moored profiler. The McLane profiler carried a SeaBird (SBE) 52 CTD and a Falmouth Scientific acoustic current meter (ACM) and made full-depth profiles every 15 min. Onshore of the bifurcation point, mooring T1 reached 240-m depth and consisted of 36 SBE56 thermistors and two SBE37 CTDs spaced 3–8 m apart (denser spacing near the thermocline) and a downward-looking 75-kHz Workhorse ADCP. At the head of the canyon in 105-m depth, station WW consisted of a Wirewalker wave-powered profiler (Rainville and Pinkel 2001; Pinkel et al. 2011) that carried a sideways-looking Aquadopp ACM and temperature, conductivity, pressure, oxygen, chlorophyll fluorescence, and backscatter sensors. Moorings MP1 and T1 were recovered on 14 December 2016 while WW remained in place transmitting data from its surface float in real time through September of 2017. The moored profiler on MP1 stopped profiling after 5 weeks, and the ACM on WW experienced intermittent failures due to power supply issues.

From 10 to 17 September, the R/V Gordon Spraul repeatedly occupied eight cross-canyon lines (SL1–8) for 25–36 h towing the Shallow Water Integrated Mapping System (SWIMS) to resolve semidiurnal (D$_2$) tidal signals. The sections are detailed in Table 1. SWIMS is a 300-kg towed body that is winched up and down at 1–2 m s$^{-1}$ capable of resolving measurements with 0.5-m spacing in the vertical and 10–610 m horizontally depending on water depth and towing speed (details in Table 1). SWIMS houses
up- and downward looking 300-kHz ADCPs that resolve velocities near the bottom near steep canyon bathymetry. It also houses temperature, conductivity, pressure, oxygen, chlorophyll fluorescence, and turbidity sensors (Alford et al. 2006; Wain et al. 2013). The R/V Sproul continuously operated a 300-kHz narrowband hull-mounted ADCP measuring currents in the upper ~150 m of the water column. Sidelobe reflections from the walls of the canyon preclude the use of traditional shipboard ADCP measurements to estimate velocities within the canyon walls.

In addition to the moorings deployed specifically for LaJIT2, data from nearby long-term monitoring stations were used to provide regional context for the canyon measurements. The Del Mar mooring (DM), located 2 km north of the canyon on the 100-m isobath, provides measurements at 10-min intervals from nine CTDs, three dissolved oxygen sensors, two chlorophyll fluorescence sensors, and an upward-looking 300-kHz ADCP (Send and Nam 2012). These data were available until November 16 when it was recovered for servicing. We use hourly sea surface height data from NOAA (2020) and wind data from the Coastal Data Information Program collected from the pier at Scripps Institution of Oceanography (CDIP 2020).

3. Analysis methods

a. Notation conventions

In what follows, the following conventions are used: \( \langle X \rangle \) is the tidal average of variable \( X \); \( \overline{X} \) is the depth average of variable \( X \); \( X_n \) is the mode-\( n \) component of variable \( X \), where \( n \) is mode number (see below); \( X_t \) is the space and time average of variable \( X \); and \( X_{\text{int}} \) is the depth-integrated value of variable \( X \).
Table 1. SWIMS lines SL1–SL8 details. From left to right, the line number, start time, duration, length, maximum depth, number of occupations, maximum spacing between profiles in space (Δx_max) and time (Δt_avg), and energy and flux adjustment factor (R) are shown for each line. Note that R(t) is computed based on the maximum value of E observed at MP1 over the full 5 weeks for which data were available, and that R reported here is the average value of R(t) over the time the line was occupied. Note also that lines SL1–SL4 have the same start and end date—these were occupied in series using a figure-eight pattern.

<table>
<thead>
<tr>
<th>Line</th>
<th>Start time</th>
<th>Duration (h)</th>
<th>Length (km)</th>
<th>H_max (m)</th>
<th>N</th>
<th>Δx_max (m)</th>
<th>Δt_avg (h)</th>
<th>R</th>
</tr>
</thead>
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<tr>
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<td>25.25</td>
<td>1.5</td>
<td>200</td>
<td>16</td>
<td>120</td>
<td>1.55</td>
<td>0.62</td>
</tr>
<tr>
<td>SL2</td>
<td>2300 UTC 13 Sep 2016</td>
<td>25.25</td>
<td>1.0</td>
<td>300</td>
<td>17</td>
<td>175</td>
<td>1.55</td>
<td>0.62</td>
</tr>
<tr>
<td>SL3</td>
<td>2300 UTC 13 Sep 2016</td>
<td>25.25</td>
<td>0.7</td>
<td>225</td>
<td>16</td>
<td>100</td>
<td>1.55</td>
<td>0.62</td>
</tr>
<tr>
<td>SL4</td>
<td>2300 UTC 13 Sep 2016</td>
<td>25.25</td>
<td>0.7</td>
<td>225</td>
<td>16</td>
<td>105</td>
<td>1.55</td>
<td>0.62</td>
</tr>
<tr>
<td>SL5</td>
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<td>38.37</td>
<td>2.5</td>
<td>300</td>
<td>38</td>
<td>270</td>
<td>1.01</td>
<td>0.67</td>
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<tr>
<td>SL6</td>
<td>0410 UTC 16 Sep 2016</td>
<td>25.33</td>
<td>6.8</td>
<td>400</td>
<td>21</td>
<td>375</td>
<td>1.21</td>
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<tr>
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<td>0630 UTC 11 Sep 2016</td>
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<td>7.1</td>
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<td>415</td>
<td>1.82</td>
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<tr>
<td>SL8</td>
<td>0200 UTC 15 Sep 2016</td>
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<td>8.1</td>
<td>525</td>
<td>15</td>
<td>610</td>
<td>1.73</td>
<td>0.53</td>
</tr>
</tbody>
</table>

b. Isopycnal displacement

After converting measurements of temperature, conductivity, and pressure into salinity and density, isopycnal displacement η is computed at each station and mooring. The background density profile (ρ(z)) is computed as an average over two D2 tidal cycles. For longer time series from moorings, a slowly varying background state (ρ(z, t)) is computed as a running mean over two D2 tidal cycles. At each time step, η(z, t) is computed as the difference between the isopycnal at each depth in the instantaneous profile and its corresponding depth in the time-mean profile (ρ(z, t)).

c. Isolating the semidiurnal internal tide

Shipboard surveys across the canyon (SL1–8, Fig. 1; see Fig. 2 for example data) are used to diagnose the spatial characteristics and energetics of the semidiurnal internal tide. Each transect is divided into a set of stations corresponding to the variable spacing (between 100 and 600 m) expected for a single SWIMS profile to be made at that water depth given an average low speed of 3 kt (1 kt = 0.51 m s⁻¹) and fall rate of 1.5 m s⁻¹. For each station, all profiles falling within half the distance between the adjacent stations are concatenated, and semidiurnal motions are isolated from the reconstructed time series using harmonic analysis. Semidiurnal components of eastward and northward velocity (u, v) and η are then determined by fitting the data A(ξ, z, t) to the model

\[ A(\xi, z, t) = A_0(x, z) \sin(\omega_D t + \phi(x, z)), \]

where x is the along-track distance, ω_D is the D2 tidal frequency, A_0 is amplitude, φ is phase relative to a common reference time, and A represents u, v, or η.

Semidiurnal signals from each of the moorings are isolated by standard bandpassing methods. A fourth-order Butterworth filter with zero-phase response, a center frequency ω = 12.4 h⁻¹ and quarter-power points at 10 and 14.8 h⁻¹ is designed to minimize filter ringing. Profile data from the moored profiler are linearly interpolated onto a grid with a constant time step of 10 min and 1-m depth intervals before filtering.

d. Energetics

Available potential energy (APE, J m⁻³) and horizontal kinetic energy (HKE, J m⁻³) are computed for each station and mooring as

\[ \text{APE}(z, t) = \frac{1}{2} \rho N^2(z, t) \eta^2(z, t), \]  
and

\[ \text{HKE}(z, t) = \frac{1}{2} \rho \left| \mathbf{u}^2(z, t) \right|, \]

where ρ is density, u are the horizontal components of velocity, and N²(z) = -(g/ρ) [δρ(δz)/δz]. Total energy E is taken as the sum of HKE and APE at any given location and time.

Baroclinic pressure perturbation is calculated as

\[ p'(z) = \int_{-z}^{0} \eta(z') N^2(z') dz' - p_o, \]

where p_o is the hydrostatic pressure and H is water depth.

Semidiurnal velocity perturbation u'(z, t) and energy flux F(z, t) = u' + p' are computed by standard methods following Kunze et al. (2002) and Nash et al. (2005). Uncertainty in these calculations that arises from sparse temporal sampling and/or incomplete coverage of the water column is computed following Nash et al. (2005). Wave-averaged quantities (APE), (HKE), and (F) are also computed for each location and mooring. For SWIMS stations, a single vertical profile of each quantity is computed from available data spanning two full semidiurnal tidal cycles. For moorings, (APE)(z, t), (HKE)(z, t), and (F)(z, t) are computed from the running mean spanning two tidal cycles at each depth.

e. Modal decomposition and reflection

The vertical structure of internal tides can be described as vertical modes that depend only on N(z). These modes are the solutions to the equation

\[ \frac{\partial^2}{\partial z^2} \eta(z) + \frac{N^2(z)}{c_n^2} \eta(z) = 0, \]

with boundary conditions η(0) = η(H) = 0, and where n is mode number, c_n is eigenspeed, and H is water depth. The eigenspeed c_n is the geometric mean of the phase speed c_p and group speed c_g, c_n^2 = c_p c_g (Alford and Zhao 2007b). This method is applied to obtain the first five vertical modes for
\( u, v, \) and \( \eta, \) after which HKE, APE, and \( F \) are computed for each mode (Gill 1982).

It should be noted that this formulation of modal structure assumes that vertical motion cannot occur at the surface or bottom and is valid rigorously only over a flat bottom (Wunsch 1968; Haney 1991; LaCasce 2017). Over bathymetric features (like a canyon), the bottom boundary condition should allow for vertical and horizontal velocities that are coupled over the slanted bottom. The resulting velocity modes, then, are surface intensified with weaker horizontal (cross-isobath) flows near the bottom (Rhines 1970; Charney and Flierl 1981). In the limit of very steep topography (where horizontal flows are assumed to be zero at the bottom), the gravest mode is not a depth-invariant horizontal flow, but a bottom-intensified topographic wave. With these issues in mind, we proceed cautiously with the above decomposition as a diagnostic tool.

For a progressive wave, the expected ratio of HKE to APE for all modes is \( \frac{\text{HKE}}{\text{APE}} = \frac{\omega^2 + f^2}{\alpha^2 - f^2} = 1.92 \) at this latitude (Gill 1982). If the wave is reflected (or partly reflected), the interference between incident and reflected waves alters this ratio such that (depending on the distance from the reflection point and the fraction of a wavelength that distance represents) HKE/APE is not constant but can deviate above or below this value. Maxima and minima of HKE and APE alternate at quarter-wavelength intervals.

Here we assess HKE/APE for all computed modes in order to diagnose reflection, by mode. Martini et al. (2007) use a similar by mode diagnostic of reflection, but use the ratio \( \frac{c_g E}{F} \) as an indicator of progressive wave behavior for each mode (as suggested by Alford and Zhao 2007b). For a progressive wave, the ratio of the observed energy flux \( F \) to the theoretical group speed times observed energy \( c_g E \) should equal 1; for a partly standing wave, that ratio is less than 1 (Alford et al. 2006;
Alford and Zhao (2007a). Here, the two diagnostics produce qualitatively similar results.

f. Quantifying and qualifying turbulence

The dissipation rate $\epsilon$ of turbulent kinetic energy (TKE) is estimated using overturns in density profiles measured by SWIMS and at the moored profiler following the Thorpe scale method (Thorpe 1977; Dillon 1982). Although no microstructure sensors were used during LaJIT2, measurements were made in the LJCS during a subsequent field campaign in February 2017. These measurements are used to determine that the Thorpe scale method is valid in this region (see the appendix). Error bars for profiles of $\epsilon$ are computed as 95% bootstrap confidence intervals.

Although the comparison shown in the appendix indicates agreement between Thorpe scale estimates and microstructure measurements, it is known that the Thorpe scale method is not valid in regions with insufficient stratification to detect density overturns. These estimates cannot capture turbulence in the well-mixed region near the bottom, for example, so some care must be taken in interpreting the results. Most of the profiles measured in this experiment did not come within 10 m of the bottom due to the steepness of the topography and the need to turn the profiler around rapidly. We, therefore, do not discuss bottom boundary layer dynamics and turbulence in great detail here.

Quantities useful in diagnosing the processes that give rise to elevated dissipation are computed at all stations using unfiltered data. Four-meter vertical squared shear is computed from velocity profiles as $S^2 = \langle \partial \omega/\partial z \rangle^2 + \langle \partial \omega/\partial z \rangle^2$. The Richardson number is computed from 4-m shear and stratification as $Ri = N^2/S^2$ (Miles and Howard 1964). Four-meter Eulerian strain is computed as $\langle N \rangle^2 p(z, t)/N^2(z, t)$, where, following Alford and Pinkel (2000), the numerator indicates the time-mean average of $N^2$ along the isopycnal, and the denominator is the local value of $N^2$.

g. Along-canyon energy budget

Data collected along transects SL2–8 are used to compute a canyon-wide energy budget. Following Gregg et al. (2011), the energy budget for tidally averaged quantities is given by the equation

$$\frac{dE_A}{dt} = \frac{dF_A}{dx} + E_A + D,$$

(7)

where $x$ is along-canyon distance, $E$ is a production term, $D$ is a dissipation term which includes both dissipation of turbulent kinetic energy and energy transfers to different frequencies integrated over the volume between lines, and the subscript $A$ indicates that each quantity is integrated over the cross-width area of the canyon at each line. For energy flux $F$, only the component parallel to the axis of the canyon is considered. The terms on the right-hand side of Eq. (7) can be computed directly from the measurements made by SWIMS. The production term $E$ is computed at each location following the methods of Kelly et al. (2012) as

$$E = \left( \frac{dh}{dx} U_{BT} + \frac{dh}{dy} V_{BT} \right) p,$$

(8)

where $dh/dx$ and $dh/dy$ are east–west and north–south bottom slopes and $U_{BT}$ and $V_{BT}$ are barotropic velocities in the east and north directions computed as the depth average of the measured SWIMS velocity at each station, and $p$ is pressure perturbation [as defined in Eq. (4)] at the bottom, $z = H$. The values at each station are integrated to estimate $C_A$.

The left-hand side of Eq. (7) accounts for unsteady variability in the energy of the system which, in this experiment, is affected by the transition from neap to spring tide that occurs during the period of shipboard observations. To account for this variability, time series of $E$ and $F$ in the semidiurnal band observed at MPI are used following the methods of Alford et al. (2011) and Wain et al. (2013). For each time step, a scale factor $R$ for energy and flux is computed as

$$R(t) = \frac{E_{\text{max}} - E_{\text{lp}}(t)}{E_{\text{max}}},$$

(9)

where $E_{\text{max}}$ is the maximum value observed during the period over which observations from MPI are available and $E_{\text{lp}}$ is the 3-day low-passed time series of $E$. The same is done for $F$, and the relevant scale factor is applied to measurements of $E$ and $F$ along SL1–8 for the corresponding time. These adjusted values are used to compute the along-canyon energy budget and to examine patterns in turbulence and energetics (see Table 1).

4. Results

a. Low-frequency currents and tidal flow structure

In the two days preceding the LaJIT2 field campaign, sustained winds blow from the northwest at $\sim 5 \text{ m s}^{-1}$ over the study region, and a surface-intensified southward current is observed at the DM mooring (Fig. 3). After making a $180^\circ$ reversal and blowing from the southwest between 15 and 17 September, winds from the northwest begin to exhibit a diurnal cycle and weaken to an average of $\sim 2.5 \text{ m s}^{-1}$. The direction of the barotropic velocity observed at DM also transitions from southward to northward midcruise. Correlations between wind forcing and current direction is minimal, consistent with prior observations over the southern California Shelf (Lentz and Winant 1986). Measurements of surface tides from the Scripps Institution of Oceanography pier 1.5 km north of LJJC indicate that the 7-day shipboard experiment spans a transition from neap to spring tide.

At DM, the observed velocity signals are typically dominated by either a barotropic or mode-1 structure (Fig. 3a) and exhibit a range of mesoscale variability. This is qualitatively different from velocity signals observed at MPI (located within the canyon walls) where barotropic currents are relatively weak, the amplitude of tidal currents exceeds the subtidal flow, and the currents exhibit multiple zero crossings throughout the water column (Fig. 3).

Depth-averaged velocities are weak everywhere in the canyon except at the offshore extent (SL8) where a northward surface current is observed (Fig. 2a). Barotropic tides are also weak ($\sim 1 \text{ cm s}^{-1}$) over the canyon axis but strengthen to $\sim 10 \text{ cm s}^{-1}$ over the canyon rim and walls—particularly on the southern flank (not shown). Internal tides visibly dominate the velocity signal at depths below the canyon rim where downward phase propagation
of oscillations of semidiurnal (12.42 h) period is clear (Figs. 2b,d). At all stations, the baroclinic velocity structure has two or more zero crossings, indicating mode-1 structure (Figs. 2b,d).

b. Energy flux

Depth-integrated energy fluxes (Fig. 4a) in the LJCS are, in general, relatively weak [O(100) W m\(^{-2}\)] compared to MC and other canyons (Petruncio et al. 1998; Kunze et al. 2002; Lee et al. 2009; Gregg et al. 2011; Wain et al. 2013; Waterhouse et al. 2017). Depth-integrated energy flux is oriented toward the canyon head (up-canyon) with a maximum near the canyon axis. Very small energy flux (<5 W m\(^{-1}\)) oriented toward the canyon mouth (down-canyon) is observed over the sidewalls on the north side of the canyon (Fig. 4a). Energy flux is elevated in one or more middepth swaths on all transect lines (Fig. 5); up-canyon flux on lines SL2, SL3, and SL5 is concentrated in a middepth band below the canyon rim. We expect the vertical structure of energy flux in mode-1 waves to exhibit minimal energy flux in the middle of the water column and maxima at the top and bottom, but that is not what we observe here (Fig. 5). Nevertheless, computing energy flux by mode reveals that mode 1 contains the majority (68%) of the flux, with an exception near the bifurcation point where the energy flux in mode 2 exceeds that in mode 1 (not shown).

c. Diagnosing reflection: Energy and energy partitioning

To diagnose patterns in the reflectivity of the internal tide in LJCS we first consider quantities computed from the total measurement at each location. APE is elevated along the canyon axis while HKE is higher over the sidewalls (not shown). Total energy \(E\) increases toward the canyon head (Fig. 4b) due to a corresponding increase in APE. The increase in energy toward the canyon head is expected as the depth and canyon width both decrease moving up-canyon.

Because the ratio of HKE/APE for a partially reflected wave is not constant along its wavelength, we must consider the observed HKE/APE ratios in the context of the horizontal wavelength of the mode in question. In the region offshore of the LJCS, the wavelength of the mode-1 D\(_2\) internal tide is \(~27\) km, which is only 3 km longer than the length of the canyon itself (24 km; see Table 3). If this wavelength were not altered by changes in stratification and bottom depth moving up-canyon, energy ratios from a fully reflected mode-1 wave would exhibit a maximum in HKE at the canyon head, a maximum in APE one quarter wavelength down-canyon, another APE maximum halfway down and an HKE maximum three quarters of a wavelength away. Rather, we observe a relative increase in APE toward the canyon head that lowers the ratio of HKE/APE such that offshore of the meander (beyond halfway down the canyon, on lines SL7–8) a maximum in HKE occurs, while onshore (lines SL2–3; SL5–6) APE is higher (HKE/APE < 1.92, indicating reflection). The discrepancy is likely due to partial reflection, and the pattern in HKE versus APE is characteristically consistent with the theoretical model of Nash et al. (2004) modified by Waterhouse et al. (2017).
their Fig. 14c) in which an incident wave loses 50% of its energy to dissipation before reflecting from the canyon head and losing its remaining energy on the way out.

Because the diagnosis of HKE/APE ratios depends on the location of reflection points and the wavelength of the corresponding mode and quickly becomes very complex, we also examine the ratio of $c_g E/F$ where values greater than 1 indicate reflection (Alford et al. 2006; Alford and Zhao 2007a) to assess internal tide reflection by mode. For mode 1 (Fig. 4d), this metric shows a mix of incident waves throughout the canyon system; line SL6 exhibits reflection everywhere, while SL5 shows primarily progressive waves (except on the southern canyon wall). Moving up-canyon from SL5, $c_g E/F$ increases and approaches unity. If conversion from barotropic to baroclinic motions within the canyon is negligible (as is shown below), this pattern suggests that an incident mode-1 internal tide is partly reflected at the meander between SL6 and SL5, and the portion not reflected continues up-canyon where it undergoes a second reflection off of the canyon head. This pattern was also observed at the continental shelf break by Hamann et al. (2018).

Evaluating these energetic quantities by mode (Table 2) shows that on offshore lines SL6–SL8 APE is dominated by mode 1, while inshore APE is contained primarily in mode 2. Everywhere (except on line SL8) mode-1 APE exceeds HKE indicating some amount of reflection. Examining the higher
modes, more complicated patterns are apparent. Compared to APE, a greater proportion of HKE is contained in modes 2 and 3, especially on nearshore lines SL2–5, suggesting that incident mode-1 signals may scatter to higher modes inshore. For modes 2–5, depth-integrated energy tends to increase toward the canyon head. The $c_{gE}/F$ ratios for each mode indicate reflection for modes 1 (Fig. 4d) and 2 (except on lines SL5 and SL8, where mode 1 waves are progressive), while for the higher modes $c_{gE}/F$ is closer to unity throughout the system suggesting that high-mode waves may be more progressive in character than their low-mode counterparts.

At MP1, which is located at the meander offshore of line SL6 where $c_{gE}/F > 1$ indicates a partly standing wave (Fig. 4d), the diagnostics for reflection suggest a pattern similar to that discovered in the modal analysis of SL1–8. During the time period corresponding to shipboard observations, $c_{gE}/F < 1$ for mode 1, in mode 2 $c_{gE}/F$ transitions in the middle of the experiment from being close to 1 to being less than 1, and in the high modes $c_{gE}/F$ remains around 1 (Fig. 6). That is, the lowest modes exhibit the characteristics of a partly standing wave, while high-mode waves are largely progressive.

Similar to findings by Nash et al. (2004) over the continental shelf, at this location energy flux in mode 1 is directed up-canyon, while higher-mode fluxes are directed down-canyon (Fig. 6, rose plots). This pattern is consistent throughout the full 5-week span of measurements at MP1 and is indicative of reflection of low-mode waves and scattering into high modes.

d. Turbulence patterns and mechanisms

Over the LJCS as a whole, Thorpe-inferred turbulent dissipation rate has a time- and space-averaged value $\epsilon_0 = 3.3 \times 10^{-3}$ W kg$^{-1}$ for all data below the canyon rim. The spatial distribution of time-averaged, depth-integrated dissipation ($\langle \epsilon \rangle_{\text{int}}$; Fig. 4c) shows that $\langle \epsilon \rangle_{\text{int}}$ increases toward the canyon head and is enhanced near the mouth of the Scripps canyon where the LJCS bifurcates and over the canyon axis at the head of the system (Fig. 4c). Enhancement of $\langle \epsilon \rangle_{\text{int}}$ on lines SL2 and SL3 qualitatively aligns with regions where $c_{gE}/F$ for the semidiurnal internal tide exceed unity, indicating that enhanced dissipation may be found in concert with reflected waveforms.

Examining the vertical structure of tidally averaged dissipation (Figs. 2 and 5), two major patterns are apparent. On the offshore lines (SL6–8), $\langle \epsilon \rangle$ is enhanced near the bottom and over the sidewalls of the canyon (Figs. 2c and 5c), and enhanced dissipation occurs preferentially with up-canyon flows (Fig. 7d). Enhanced values of dissipation are not observed during the occupation of line SL7, which is sampled at the beginning of the experiment during neap tide and likely represents a lower bound for tidally driven turbulence at that location. Because our profiles did not come closer than 10 m from the sea floor and because the Thorpe scale method does
not capture turbulence in insufficient stratification, there may be additional enhanced dissipation occurring in the well-mixed bottom boundary layer that cannot be estimated here.

In the following sections we discuss several processes that may explain the turbulence patterns apparent in Figs. 5 and 2c,e.

1) INTERNAL TIDE BREAKING NEAR LOCAL GENERATION

Depth-averaged $\langle \epsilon \rangle$ is also enhanced over the sidewalls of the canyon, particularly near the depth of the canyon rim. On lines SL5 and SL6 that sidewall enhancement is skewed with stronger dissipation over the southern rim than on the northern; on line SL6, depth-averaged dissipation over the southern flank (2.5 ± 0.12 × 10⁻⁷ W kg⁻¹) is an order of magnitude greater than over the northern flank (3.2 ± 1.1 × 10⁻⁷ W kg⁻¹). At locations above the canyon rim, the highest values of $\langle \epsilon \rangle$ near the bottom occur in conjunction with velocities oriented in the up-slope direction and consistent with conditions for conversion from surface to internal tides (Llewellyn-Smith and Young 2002).

2) INTERNAL WAVE SHEAR AND STRAIN

The vertical structure of dissipation on nearshore lines (SL2, SL3, and SL5) exhibits enhanced dissipation $\langle \epsilon \rangle$ throughout most of the water column (Fig. 5). In time series from SWIMS, we observe downward phase propagation of the semidiurnal tide and signals of middepth enhancement in dissipation that increase in depth in conjunction with up-canyon flows in a middepth layer. A relationship between velocity shear and enhanced dissipation is visually apparent in the raw time series data (Figs. 2d,e).

The association between dissipation and strain is demonstrated by conditional probability density functions (as in Alford and Pinkel 2000, their Fig. 15) that are skewed toward higher strain values and up-canyon velocity (e.g., Figs. 7b,d) when considering overturning regions and values of $\epsilon$ that exceed 10⁻⁷ W kg⁻¹. On all lines wherever 4-m Ri falls below 0.25 (the threshold value below which shear instability is possible), $\epsilon$ is 2–5 times greater than the average value on that line. As expected, values of Ri falling below 0.25 are more probable when considering overturning regions and high values of $\epsilon$ than for the dataset as a whole (Fig. 7a). The PDF of shear is also skewed slightly toward higher values of shear when considering only overturning regions (Fig. 7c) indicating isopycnal strain as well as shear are responsible for creating conditions for enhanced dissipation.

e. Canyon energy budget

After applying the computed energy and flux scale factors $R(t)$ to adjust terms 2–4 of Eq. (7) for spring–neap changes in overall energy in the system, the along-canyon energy budget is assessed. In general, area-integrated semidiurnal energy flux decreases from off- to onshore (Fig. 8a), except for an increase in up-canyon energy flux of 71 W m⁻¹ from line SL6 to SL5. Conversion is computed both from the observed bottom pressure perturbations $p_b(y)$ and barotropic velocities $u_{BT}$ at each station, and over the whole region shown in Fig. 1 by assuming an average $p_b$ and $u_{BT}$ everywhere. For both methods, conversion from barotropic to baroclinic motions is ≤5 kW between any given line [in agreement with values computed by Ponte and Cornuelle (2013)] and small relative to dissipation and flux terms. Except for the region surrounding the increase in energy flux on line SL5, the observed dissipation integrated in the volume between lines balances the observed flux convergence within error bars (Fig. 8b). Because we assume that the dissipation observed on one line extends between lines, we make two estimates of volume-integrated $\langle \epsilon \rangle$—one integrating in the space toward the canyon head and one toward the canyon mouth. The difference between the two methods contributes to the computed error. Considering the system as a whole, the convergence of semidiurnal energy flux from mouth (SL8) to head (SL2) is balanced by dissipation integrated over the canyon volume within error bars (Fig. 8b, stars), putting an upper bound on...
energy siphoned to frequencies not captured by the harmonic analysis.

5. Discussion

The classical theory of internal waves in canyons describes waves that focus toward the canyon head and floor where they eventually break near the bottom (Hotchkiss and Wunsch 1982). Our analysis of the LJCS, on the contrary and in alignment with results from Waterhouse et al. (2017), suggests that steep topography along the sidewalls of a meander and at the head of the canyon system leads to reflection of incident low-mode internal waves and scattering to higher modes, giving way to enhanced dissipation at middepths. Unifying patterns in the mechanisms driving turbulence may be revealed by examining similar observations across multiple canyons.

a. Intercanyon comparison: Monterey, Eel, and La Jolla

Monterey Canyon, Eel Canyon, and the La Jolla Canyon System are all fairly well-sampled canyons located on the West Coast of the United States. The three have distinct geometry, and the physical processes observed in each differ. A detailed comparison of the three is offered here.

One primary distinction for this set of canyons is that both MC and EC extend to and incise the continental slope, while the LJCS is confined to the shallower shelf and does not interact with isopycnals in the deep ocean. MC is in all dimensions nearly quadruple the size of the relatively small LJCS (Table 3); EC is only 9 km longer than LJCS, but is as deep as MC (∼1500 m) at its offshore extent. Both MC and LJC are about as long as the length of the mode-1 \( D_2 \) internal tide offshore of their mouths, while EC is only half the length of the local mode-1 \( D_2 \) internal tide. Where LJC has only one large
bend in its axis, MC and EC exhibit several meanders, and MC is the trunk off of which multiple branch canyons stem. Despite the size discrepancy, MC, EC, and LJCS share some key characteristics: all are shelf-incising canyons that lie poleward of the turning latitude of the diurnal frequency band. Waves of semidiurnal frequency, which are free to propagate at these latitudes, are expected to dominate the baroclinic tidal signal in all three.

The differing bathymetry exhibited by the canyons, accordingly, causes different expectations for how semidiurnal internal tides should behave within their walls. All three canyons have steep, supercritical sidewalls that will reflect $D_2$ waves toward the canyon floor. With typical near-bottom stratification ($10^{-5} \text{s}^{-2} < N^2 < 10^{-4.5} \text{s}^{-2}$), the canyons’ thalwegs are all subcritical to the semidiurnal internal tide at their offshore extent. In MC, the thalweg remains subcritical for the majority of the canyon’s great length. Because of its subcritical geometry, $D_2$ internal waves in MC are expected to progress forward to the head of the canyon where they may later break at critical bathymetry. Changes in stratification within the canyon, though, can shift the criticality of the topography enough to make the same slopes reflective (Petruncio et al. 1998; Zhao et al. 2012). Observations made by Wain et al. (2013) align well with this expectation – turbulent hotspots in their study of the upper MC are primarily in the bottom 300m near critical reflection points and near topographic features.
that induce hydraulically controlled flows. HKE/APE ratios in MC are closer to that of freely propagating $D_2$ waves at that latitude and actually increase from offshore to onshore as baroclinic velocities strengthen while isopycnal displacements remain constant; the opposite is true for the LJCS.

On the other hand, in EC and LJCS where the length of the supercritical thalweg slopes near the canyon head is longer relative to the overall canyon length, $D_2$ waves are expected to partly reflect and generate partly standing wave patterns. In LJCS, such reflection is observed by Albery et al. (2017) as well as the observations reported here. The observations suggest that the $D_2$ internal tide is a standing wave. In conjunction with steepening waves, the party standing pattern leads to middepth turbulence near the head of the canyon, accounting for a portion of the dissipation observed in the canyon system as a whole. The enhancement of dissipation near the bottom is much weaker in LJJC than in MC, suggesting that mid-water-column processes can be equally if not more dissipative than the near-bottom bores and breaking that are commonly reported. Patterns of turbulent enhancement at middepths and semidiurnal frequency, similar to those in the LJCS, are also reported near the head of EC (Waterhouse et al. 2017).

In addition to differences in reflectivity and turbulence patterns, the presence of high modes in the vertical structure of velocity toward the head of the LJCS is not reported by Wain et al. (2013) near the head of MC. Time series presented by Petruncio et al. (1998, see Figs. 9 and 10 therein) made in 400-m water depth near the head of MC, however, do qualitatively demonstrate high-mode velocity structure. Such structure is not expected in a paradigm where incident mode-1 waves simply enter and dissipate at a critical breaking point at a location up-canyon, but is consistent with a reflected wave that scatters into higher modes. In velocity time series made nearest the head of Eel canyon by Waterhouse et al. (2017, see Fig. 6 therein), multiple flow reversals indicating high-mode velocity structure are apparent, as were periods of mid-depth turbulence. From along-canyon SWIMS surveys conducted by Gregg et al. (2011) in Ascension Canyon (a canyon with a supercritical thalweg slope) multiple flow reversals were apparent approaching the canyon head (see Fig. 13 therein). Although it has not been analyzed in great detail in previous observations, the presence of high-mode velocity structure near the head of these canyons appears to be a common feature. Next, we consider from where that structure arises and whether it influences the pattern of middepth turbulence near the canyon head reported by Albery et al. (2017) and Waterhouse et al. (2017) and this work.

b. A mechanism for enhanced turbulence: Internal tide reflection, scattering, and middepth breaking

Profiles collected at MP1 over a sustained period of 5 weeks demonstrate a relatively robust pattern: up-canyon propagating mode-1 internal tides contain most of the energy and flux and are partly standing while higher-mode waves, containing

![FIG. 8. Energy budget for the La Jolla Canyon System.](Image)

(a) Semidiurnal energy flux integrated over the canyon cross section calculated from observed values (red line) and values adjusted to account for temporal changes in energy storage (blue lines). (b) Along-canyon energy flux convergence (difference in integrated flux between lines divided by distance between center points of adjacent lines; blue line), dissipation integrated in the volume between adjacent lines (red and yellow lines), conversion (purple line). Colored stars indicate the canyon-integrated value for the corresponding quantity [i.e., $dF = (F_{SL8} - F_{SL2})$/canyon length]]. Note: for conversion, this total quantity does not include the value of the farthest offshore line (SL8). For SWIMS lines 2–3 and 5–8 quantities are plotted versus their distance away from the canyon head. For all quantities, errors due to sparse temporal sampling are computed for each location following Nash et al. (2005) and compounded. Additionally, volume integrals are calculated using either the volume between lines shoreward or the volume between lines seaward of a given line. The difference between these estimates is added to the computed error.

### Table 3. Characteristics of Monterey Canyon, Eel Canyon, and the La Jolla Canyon System including length from mouth to terminus, width at the canyon mouth, depth at the canyon mouth, bottom slope at the canyon head ($S$), average baroclinic energy flux into the canyon mouth ($F_{mouth}$), space and time averaged dissipation ($\epsilon$), and the average wavelength of the mode-1 $D_2$ internal tide offshore of the canyon mouth.

<table>
<thead>
<tr>
<th></th>
<th>Monterey Canyon</th>
<th>La Jolla Canyon System</th>
<th>Eel Canyon</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length (km)</td>
<td>97</td>
<td>24</td>
<td>35</td>
</tr>
<tr>
<td>Width (km)</td>
<td>21</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>Depth (m)</td>
<td>1600</td>
<td>650</td>
<td>1500</td>
</tr>
<tr>
<td>$S$ (head)</td>
<td>0.05</td>
<td>0.1</td>
<td>0.15</td>
</tr>
<tr>
<td>$F_{mouth}$ (kW m$^{-1}$)</td>
<td>5</td>
<td>0.2</td>
<td>7.6</td>
</tr>
<tr>
<td>$\epsilon$ (W kg$^{-1}$)</td>
<td>$4 \times 10^{-8}$</td>
<td>$3.3 \times 10^{-8}$</td>
<td>$4.3 \times 10^{-8}$</td>
</tr>
<tr>
<td>$\lambda_1$ (km)</td>
<td>82</td>
<td>27</td>
<td>70</td>
</tr>
</tbody>
</table>
less energy, are progressive with energy fluxes oriented
down-canyon (Fig. 6). Because there appears to be negli-
gible up-canyon propagation in the higher modes, the most
likely hypothesis to explain this observed pattern is that
a low-mode incident wave enters the canyon and is parti-
cally reflected and scattered into high modes that propagate back
down-canyon while dissipating. The balance of dissipation
versus flux indicates that the higher modes lose their energy
within the canyon system.

The observation of partly standing waves offshore of both
the meander and the canyon head (but progressive low-mode
waves along line SL5) suggests that reflection occurs in two
locations. The observed increase in flux between lines SL6 and
SL5 could arise from conversion from barotropic to baroclinic
motions occurring over the canyon’s sidewalls to the east of
the meander, but this is unlikely given the relatively small amount
of conversion computed (Fig. 8b). More likely, partial reflec-
tion of the D2 tide occurs off of the steep west-facing side of
the canyon wall between SL6 and SL5. In alignment with this
theory, the mode-1 ratio of \(c_{E}/E_F\) is high (4.4) on line SL6
(indicating a mostly standing wave there), but decreases to
below unity (0.8) on SL5 indicating wave progression (see
Table 2). Offshore of the meander, partially reflected waves
cancel out and reduce the observed net energy flux up-canyon.
Inshore of the meander, the up-canyon portion is not
encountering a reflected waveform, so the net up-canyon flux is
higher there. The second reflection occurs from the canyon
head where bathymetry steepens and becomes supercritical
inshore of the canyon’s bifurcation point.

Two-dimensional models of internal waves encountering a
shelf with variable steepness, shape, and depth show that for an
incident mode-1 wave, energy transmitted onto the shelf in mode
1 is smaller than that of the incident wave and that proportionally
less energy is scattered into forward-transmitted high-mode
waves (Müller and Liu 2000; Kelly et al. 2013). A significant
fraction of the reflected energy is scattered to mode numbers
higher than the incident mode number (see Fig. 10 in Müller
and Liu 2000). This paradigm is also suggested in observations
made by Nash et al. (2004) on the continental slope of the Mid-
Atlantic Bight and in observations made by Hamann et al.
(2018) off the coast of La Push, Washington. There, conver-
gence of low-mode, semi-diurnal onshore energy flux is ap-
proximately balanced by a divergence of high-wavenumber
semi-diurnal offshore energy flux, and dissipation rates
observed at that location \(O(1 \times 10^{-8}) \text{ W kg}^{-1}\) suggest
that the high-mode reflected waves should dissipate on the
time scale of \(~1\text{ day}\). Stronger dissipation rates are ob-
erved at the head of the LJCS \(O(1 \times 10^{-7.5}) \text{ W kg}^{-1}\) and in
Eel Canyon \(O(1 \times 10^{-7}) \text{ W kg}^{-1}\) where we would expect
topographic focusing of internal waves to lead to addi-
tional shear and strain and subsequent breaking events. Scattered
high-mode waves would dissipate on a relatively short time
scale and would only travel a finite distance down-canyon
owing in part to their slower propagation. Consistent with this
theory, in canyons such as MC and EC where observations
have been made farther offshore from the canyon head, high-
mode waves signatures are not apparent at offshore stations

The mechanism we propose here is consistent with a simple
analytical model proposed by Waterhouse et al. (2017) in
which an incident wave enters a canyon, reflects from its steep
head, scatters, and dissipates completely before reaching the
offshore extent of the canyon. Evidence from the modal
analysis presented in this study homes in on a specific (and
potentially ubiquitous) mechanism by which internal waves
encountering a submarine canyon dissipate their energy. Many
submarine canyons steepen sharply at their nearshore termin-
us, and it is possible that focusing, reflection and scattering
that cause enhanced shear and strain near the canyon head and
lead to strong middepth turbulence is a commonality between
them. Although the theoretical model proposed by Hotchkiss
and Wunsch (1982) predicts the along-canyon location of the
enhanced turbulence observed here, the turbulence is distrib-
uted vertically throughout the water column instead of only at
critical breaking points near the bottom. Turbulence and
mixing at middepths could influence regional buoyancy fluxes
as well as the distribution of enhanced nutrients and benthic
larvae that are often found in and around submarine canyon
systems.

c. Other possible turbulence mechanisms

1) Asymmetrical breaking of locally
   generated internal tides

Our observations capture elevated depth-averaged \(\overline{\epsilon}\) over
sidewalls on lines SL5 and SL6 near the canyon’s meander.
There is asymmetry between dissipation over the northern and
southern flanks, with enhancement over the southern wall
nearly 10 times greater than enhancement over the northern
wall. The sense of the observed asymmetry agrees with the
asymmetry Zhang et al. (2014) report from hydrostatic models
of idealized shelfbreak canyons forced with alongshore-
uniform barotropic D2 tidal boundary forcing. In their study,
they find that tidal conversion occurs asymmetrically between
the left- and right-hand sides of the canyon due to multiple-
scattering effects on one side of the canyon rim. As barotropic
tidal currents rotate clockwise (counterclockwise) the phase
variation in its alignment with the canyon slope leads to con-
version occurring sequentially (looking from mouth to head)
from canyon axis toward the right-hand (left-hand) side of the
canyon over a short period. The curvature, then, of the right
hand (in this case, southern) side of the canyon has the po-
tential to create resonant internal-tide generation. Because the
generation process is generally associated with localized dis-
sipation, the strength of turbulent dissipation would be ex-
pected to be asymmetric in the manner that is observed here.

2) Critical breaking of higher harmonics

Although not examined closely in this dataset, the process of
reflection can siphon energy to the higher harmonics of the
fundamental wave (Rodenborn et al. 2011). This energy si-
phoning would contribute to energy flux convergence in the
semi-diurnal band (examined in Fig. 8), but since dissipation
balances flux convergence on the whole, the results suggest
that these higher-frequency waves do not escape the system
and dissipate their energy therein. High-frequency waves can
(and seemingly do) carry energy away from the reflection point and cause enhanced turbulence at depths and locations not predicted by considering semidiurnal flux convergence alone.

The criticality of a given slope is determined relative to the propagation angle of an incident wave (which is set by the near bottom stratification and the frequency of the wave). Waves of higher frequency have steeper propagation angles such that topography that may be reflective to the fundamental wave frequency can be critical or subcritical to the higher harmonics. These waves, then, may propagate farther up-canyon and break where they encounter steeper topography matching their steeper propagation angle. Higher harmonics with steeper propagation angles may also undergo critical breaking over the steeper canyon walls, leading to elevated dissipation higher in the water column. To verify whether this mechanism contributes to the patterns of enhanced dissipation observed here, further work including modeling would be required.

d. Contextualizing canyon turbulence

The observations presented here represent the first comprehensive examination of internal tides and turbulence in the LJC, which is only the sixth submarine canyon in which some measure of both dissipation and internal tide energy flux has been made. The others include Monterey Canyon (Wain et al. 2013), Ascension Canyon (AC; Gregg et al. 2011), Eel Canyon (Waterhouse et al. 2017), Juan de Fuca Canyon (JdeF; Alford and MacCready 2014), and Gaoping Canyon (GC; Lee et al. 2009). If we assume that the measured value of $\epsilon$ averaged over space and time ($\overline{\epsilon}$) occurs uniformly throughout the canyon’s volume and integrates thereover ($\overline{\rho \overline{\epsilon} dV}$), and that the average value of up-canyon energy flux incident at the canyon mouth ($F_o$) occurs everywhere along the cross-sectional area of the canyon mouth ($\int_A F_o dA$), we can compare these roughly estimated terms of the canyon energy budget (Fig. 9).

Despite the fact that these canyons differ in location and geometry and are exposed to a wide range of internal wave energy flux from offshore, the markers in Fig. 9 mostly fall just below the 1:1 line, which indicates that most energy flux incident at the canyon mouth is balanced by volume-integrated dissipation. It also suggests that, like we observe in the LJC, the conversion in most of these canyon systems is negligible compared to the energy flux incident at the canyon mouth. The exceptions include Eel Canyon and Juan de Fuca Canyon. In AC dissipation is observed only over the canyon axis and accounts for one quarter of the incident flux—likely due to significant dissipation occurring over the canyon’s sidewalls (Waterhouse et al. 2017). In JdeF tides are weak but cross-shore pressure gradients drive strong subtidal up-canyon flows (Denman et al. 1981; Waterhouse et al. 2009). A sill within the canyon induces a hydraulic jump, and the majority of the observed turbulence occurs there (Alford and MacCready 2014). Because of the nontidal nature of these conditions, we expect dissipation to exceed internal wave energy fluxes as shown in Fig. 9.

Although the relationship shown here is based on very simplified metrics, it suggests that the net internal wave energy that enters a canyon at the offshore extent is primarily dissipated within the canyon walls rather than depositing any significant energy into small-scale processes or flows on the adjacent shelf.

6. Summary and conclusions

Measurements of velocity, density, and turbulence made throughout the La Jolla Canyon System were made in order to characterize the dynamics that drive enhanced mixing in this submarine canyon system. The most important results are summarized here:

• The La Jolla Canyon System is relatively steep and reflective to the semidiurnal tide. Mode-1 waves exhibit a partly standing character, while higher modes tend to be progressive in character and directed offshore, suggesting reflection and scattering of incident low-mode waves.

• Semidiurnal energy flux in the system is primarily in mode 1 and up-canyon. Flux integrated over the cross-sectional area of the canyon at each line decreases from offshore ($240 \pm 15$ kW) to onshore ($5 \pm 0.3$ kW). Smaller energy fluxes are apparent in higher modes and are mostly oriented down-canyon, consistent with reflection and scattering.

• An increase in flux between lines on- and offshore of the canyon’s meander suggest partial reflection and partial transmission of the incident low-mode wave from the steep topography at the meander.
The highest values of depth-averaged and depth-integrated dissipation occur within 3 km of the canyon head (lines SL2–3) and are associated with strain and up-canyon velocity.

Offshore, turbulence is enhanced primarily near the bottom and over the sidewalls; onshore, turbulence occurs at mid-depths and is associated with periods of elevated shear and strain.

Although flux convergence between adjacent profiles is not always balanced by the dissipation observed locally, flux convergence over the canyon as a whole is balanced by dissipation over the volume of the canyon within error bars—a finding that is consistent with limited observations available from other canyon systems.

From this set of tidally resolving, cross-canyon measurements, the observed patterns in flux and its partitioning between the modes suggests that mode-1 waves incident on the canyon reflect and scatter backward at a bend along the canyon axis and at the head of the canyon where the thalweg becomes steep. High-mode reflected waves induce additional shear and strain, leading to enhanced turbulence near at middepth near the canyon head. A review of data presented in previous canyon studies suggests that enhanced middepth mixing may be prevalent in both shelf-incising and slope canyons.

The distribution of dissipation both laterally and in the vertical has important implications for regional water mass transformation and biological interactions near canyons worldwide. For slope canyons that connect to deep isopycnals, understanding the physics and correctly parameterizing the distribution of mixing driven by internal waves in canyons could greatly improve our ability to represent and accurately predict future changes in ocean circulation and heat storage (Melet et al. 2013).

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Data availability statement. Shipboard data for this experiment can be found at https://doi.org/10.7284/906929 and https://doi.org/10.7284/907217. Data from the Del Mar mooring can be accessed at http://mooring.ucsd.edu/dev/delmar1/. Wind data from the Scripps Pier can be accessed at http://cdip.ucsd.edu/themes/?d2=p705s:073:st:1. Please contact mhamann@ucsd.edu for access to datasets.

APPENDIX

Thorpe Scale Validation

In February 2017, a short experiment (PLUMEX) was conducted in the vicinity of the LJCS. During this experiment, profiles were made over the axis of the LJC over the course of 10.5h using the Modular Microstructure Profiler (MMP). MMP is a loosely tethered body that measures temperature, conductivity, pressure, oxygen, and microshear and temperature as it free falls vertically at 0.6 m s$^{-1}$ such that its shear probes are able to resolve turbulent velocity fluctuations over the full spectrum of shear variance (Oakey 1982; Wesson and Gregg 1994). By integrating under these well-resolved spectra, an estimate for the value of $\epsilon/\nu_{15}$ is obtained directly. Values obtained from this direct method can then be compared to those obtained by other methods. During LaJIT2, no usable microstructure measurements were recovered from SWIMS, and we therefore estimate dissipation $\epsilon_{OT}$ from regions of statically unstable fluid (overturns) following the Thorpe scale method described by Thorpe (1977), Dillon (1982), and Gargett and Garner (2008).

Because turbulence tends to coincide with overturns, values of $\epsilon_{OT}$ tend to agree well with values of $\epsilon$ from shear variance in a time-averaged sense (Ferron et al. 1998; Alford et al. 2006).

To justify using $\epsilon_{OT}$ in this particular region where stratification is relatively strong, we compare $\epsilon_{OT}$ to the concurrent estimates of $\epsilon$ made from MMP. This comparison is laid out...
Fig. A2. Scatterplot of $\epsilon$ vs $e_{\mathrm{OT}}$ for all overturns detected below 30-m depth. Red dots indicate binned and averaged values, and error bars show the standard deviation of values within each bin.

in example profiles of $e_{\mathrm{OT}}$ and $\epsilon$ in Fig. A1 and point-by-point in Fig. A2. In a bin-averaged sense, $e_{\mathrm{OT}}$ is biased slightly high (particularly for lower values of $\epsilon$). However, in a time-averaged sense there is good agreement in the vertical profile of $\epsilon$ and $e_{\mathrm{OT}}$, and the average value computed for the two methods agree well within a factor of 2 ($\langle e_{\mathrm{OT}} \rangle = 1.05 \times 10^{-8}$, $\langle \epsilon \rangle = 1.10 \times 10^{-8}$).

REFERENCES


