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**Submarine Canyon Oxygen Anomaly Caused by Mixing and Boundary-Interior Exchange**

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**Abstract**

Closely spaced CTD stations showed elevated oxygen within Monterey Submarine Canyon. Anomalously high (2–5 μmol kg⁻¹) dissolved oxygen was found between 600–1,100 m in the O₂ minimum, co-located with a turbulence hotspot caused by convergence of upcanyon, semidiurnal internal tidal energy flux. Furthermore, the oxygen anomaly extended >10 km downcanyon at the same depth and isopycnals of a previously identified intrusion predicted from buoyancy conservation. We show that dissolved oxygen and fine suspended particles act as independent tracers to (a) validate previous microstructure observations of intense turbulence extending >400 m above the bed (mab) at the canyon hotspot, and (b) track boundary-interior exchange driven by mixing in the form of isopycnal-spreading of anomalies away from a near-boundary source. This study demonstrates the use of oxygen, commonly measured with shipboard profiling, as a tool for tracking mixing and lateral dispersal.

**Plain Language Summary**

Continental margins and complex topography at canyons play a critical role in mixing the main ocean thermocline and the vertical transport of heat and carbon. Monterey Submarine Canyon has a persistent hotspot of elevated internal wave energy dissipation between 1,200 and 700 m depth. Using previous data and new analyses we show that this mixing hotspot is also an elevated oxygen hotspot in the middle of the deep oxygen minimum zone (a basin-wide feature), and that this elevated oxygen anomaly, along with suspended particles and mixed fluid, are expelled from the near-boundary hotspot and flow offshore ~10 km within an intrusion. Results improve understanding of how continental-margin boundary layers remain stratified in the face of persistent energetic turbulence, and suggest potential use of oxygen in future studies of mixing and exchange between continental margins and ocean interior.

**1. Introduction**

Submarine canyons are important sinks and sources for internal-wave energy (Carter & Gregg, 2002; Hotchkiss & Wunsch, 1982; Petruncio et al., 1998). Collectively, canyon-influenced turbulence and deep boundary mixing are important for mixing of the main ocean thermocline (Kunze et al., 2006) with implications for nutrient, carbon and heat transport, as well as climate. Monterey Submarine Canyon, central California, is a region of energetic internal tides (Kunze et al., 2002; Petruncio et al., 1998). Several studies have identified a turbulent mixing hotspot of energy-flux convergence at which low-mode semidiurnal internal tides dissipate a significant amount of energy as they propagate upcanyon (Hall & Carter, 2011; Jachec et al., 2006; Kunze et al., 2002). Microstructure shear measurements confirm elevated turbulent dissipation, \( \varepsilon = \mathcal{O} \left(10^{-8} \text{W kg}^{-1}\right) \) within stratified fluid and extending at least 400 m above the seafloor (Kunze et al., 2012) at this hotspot. The persistence of a stratified bottom boundary layer (BBL) in the face of intense dissipation implies continuous exchange with interior waters.

In a previous study of this data set an intermediate nepheloid layer (INL) was found spreading downcanyon between 750–900 m depth from the mixing hotspot (Kunze et al., 2012). Although boundary-interior exchange is enhanced in canyons by turbidity currents, upwelling/downwelling, and currents overshooting the shelf break incision (Allen & Durrieu de Madron, 2009; Hickey et al., 1986; Puig et al., 2014). It was shown that this INL is driven by buoyancy conservation in the presence of mixing, with turbulence provided by convergence of up-canyon internal wave energy flux (Kunze et al., 2012). Although studies on the closely related phenomenon of lateral dispersal driven by gravitational collapse after stratified mixing have primarily been done in theoretical, laboratory, and numerical studies (Arneborg, 2002; McPhee-Shaw &...
Kunze, 2002; Winters, 2015), there is evidence for the importance of this process causing intrusions at continental margins (Inall, 2009; Masunaga et al., 2016; McPhee-Shaw, 2006; McPhee-Shaw et al., 2004; Moum et al., 2002). Resulting isopycnal transport of mixed fluid away from sloping boundaries has implications for the effectiveness of turbulence on buoyancy flux. It is difficult to investigate the details of intrusion behavior in the natural environment. For example, little is known about their temporal variability. How far offshore do intrusions persist? If driven by internal tides, how much do they fluctuate over tidal periods?

Here we demonstrate boundary-interior exchange using dissolved oxygen as a tracer, in addition to suspended particles, from a high-spatial resolution CTD survey, and examine temporal behavior of the intrusion. Oxygen and transmissometer beam attenuation coefficient are used as independent tracers to (a) validate previous microstructure inferences of intense turbulent mixing extending ~400 m above the seafloor at the hotspot, (b) track boundary-interior exchange in the form of isopycnal spreading of these signals downcanyon from the hotspot source, (c) suggest that intrusion transport is balanced by lateral diffusivity.

2. Methods

CTD profiles were collected 18–30 August, 2008, in water depths 700–900 m along the thalweg (Figure 1). These data were previously used to study internal tide energetics, stratified turbulent mixing and boundary-interior exchange (Kunze et al., 2012). Observations and modeling demonstrate that the stratified boundary layer between 1,200 and 700 m water depth is a zone of intense internal tide dissipation, with depth-integrated semidiurnal energy-flux of 3–4 kW m⁻¹ (Hall & Carter, 2011; Jachec et al., 2006; Kunze et al., 2002). Here, the connection between this mixing hotspot and a layer of anomalous dissolved oxygen (DO) and transmissometer beam attenuation coefficient (BAC) is examined for evidence of lateral dispersal. Stations span the roughly 600–900 m range of the eastern North Pacific oxygen minimum zone (OMZ) (e.g., Chan et al., 2008), a global-scale feature formed by a balance between respiration and sinking carbon with ventilation and mixing of higher-oxygen waters from below. In the ~750–900 m range it is difficult to distinguish mid-water-column features (the previously observed INL) from features associated with bottom processes, so we focus on stations 22 and downcanyon; 930–1,790 m bottom depth. Most profiles reached to within 10–20 mab, but offshore of station 42 maximum measured depth was 1,400 m; thus stations 49–61 did not measure bottom boundary or bottom nepheloid layers. Stations 32, 42, and 45 were occupied for 10–12 h with between 9 and 12 sequential profiles, covering a semidiurnal tidal cycle.

DO was measured with a Seabird SBE43 oxygen sensor and BAC with a Wetlabs 25 cm pathlength transmissometer. 1 L water samples from within and outside of the INL were filtered through pre-dried, pre-weighted 0.4 mm pore-size, 47 mm-diameter filters, then dried and weighed for suspended particulate mass concentration. BAC is linearly related to particle concentration (Morris, 2011), as found by previous transmissometer response studies for fine grain size and low concentrations (1–5 mg l⁻¹) in deep canyons and continental slopes (Baker & Lavelle, 1984; Gardner, 1989; McPhee-Shaw et al., 2004).

Because internal-wave heaving is up to 300 m, DO, and BAC are examined in potential density σθ space. DO anomalies are defined relative to the two deepest stations, 58 and 61, just outside of where the canyon incises the shelf and the average profile represents background waters from the outer continental shelf and upper continental slope; a positive anomaly represents elevated oxygen above the background O₂ minimum. The background profile is a least-squares fit of DO at stations 58 and 61 to a cubic polynomial of σθ, yielding coefficients 1.88 × 10⁻¹⁰, b: −1.51 × 10⁻⁸, c: 4.06 × 10⁻⁷ d: −3.63 × 10⁻⁶, R² = 0.99).

3. Results and Interpretation

DO, BAC, and DO anomaly profiles Figure 2 (a–c) suggest grouping the stations into the following categories: hotspot, transition, and outer-canyon sites. The hotspot stations (22–38), span an 8 km reach, and characterized by homogenous elevated DO between 2 and 5 μmol kg⁻¹ relative to background, with the anomaly extending from the seafloor to roughly σθ ~ 27.05, (640 ± 100 m depth). Elevated BAC extended to similar heights, with a more intense signal within 150–200 mab, likely due to bottom nepheloid layers or interleaving layers from nearby slope bottom bathymetry (Armi, 1978; McPhee-Shaw, 2006). Outer-cany-
yon sites are stations 58 and 61. Stations 42–53 occupy a transition zone between elevated DO in the hotspot and the background oxygen minimum offshore. At these sites, DO is only distinguishable from background between isopycnals ~[27.20–27.35], within a layer where DO anomaly and BAC were elevated and exhibited an offshore gradient between the hotspot and outer-canyon stations. This intrusion thinned offshore (Figures 2c and 2d) and spread into stratified water with average buoyancy frequency \( <N> = 4.5 \times 10^{-3} \text{ rad s}^{-1} \). As emphasized by Kunze et al. (2012), near-bottom waters were stratified as well, with average stratification over 200 mab of \( <N> = 3.5 \times 10^{-3} \text{ rad s}^{-1} \). Well-mixed layers were rare: Bottom mixed layers were observed extending to the deepest measurement only twice: cast 8 of station 32 (50 mab), and cast 2 of station 27 (40 mab).

Profile time series of BAC, DO, and DO anomaly at three stations show semidiurnal heaving of 50–100 m in the bottom ~400 m (Figure 3). The same DO and BAC intrusion apparent in Figure 2 stands out as a continuous feature at about 800–900 m depth. The layer showed strong isopycnal fidelity at each location. In other words, neither constituent migrated off its original density surface during the 12 h but instead was...
advected up and down by internal-tide excursions. This demonstrates that these scalars were passive on internal-wave timescales and that fine particles did not settle over a tidal timescale. The intrusion was \( \sim 400–500 \) m a b and \( \sim 250–400 \) m a b at stations 45 and 42, respectively. This height above the seafloor within stratified interior waters demonstrates decoupling of the layer from local sediment erosion. Station 32 was shallower, within the hotspot, and during the first five hours the layer was near the seafloor. However even then it was within stratified waters, that is, not part of a well-mixed bottom layer. At transition stations 42 and 45, upward displacement of deep waters was apparent not just in the BAC and DO anomaly, but even in DO > 1,000 m (\( \sim \) hours 4–8, top panel). Because of the shape of the background OMZ, DO has a strong negative vertical gradient at depths >900 m (Figure 2a). This suggests simple upcanyon advection of dense, high-DO water by the low-mode internal tide, and a likely source for anomalies in profiles (Figure 2d).

Along-canyon variability of the anomaly can be examined via tracer-tracer plots (Figures 4 and 5). A potential temperature (\( \theta \)) – salinity (S) diagram color-coded by DO anomaly reveals a clear distinction between

**Figure 2.** Profiles with density \( \sigma_\theta \) (left axis of panel (a), of (a) dissolved oxygen (DO), (b) beam attenuation coefficient (BAC), (c) BAC for transition stations, and (d) DO anomaly at transition stations. For sites with multiple casts, the first cast was used to plot only one cast per station at each site. Average depth corresponding to each isopycnal grid is labeled on right axis of panel (d). Red lines mark rough intrusion boundaries (\( \sigma_\theta = 27.2–27.35 \)).
profiles with elevated anomaly (≥2.5–5 μmol kg⁻¹) and those without, deeper than σθ ∼ 27.1 (680 m) and above 27.3 (Figure 4). Elevated anomalies fall on a relatively straight line in T-S space. In contrast, waters in this density range with little or no anomaly bow toward lower salinity compared to those with elevated DO. The shallower, high-salinity California Countercurrent (Collins et al., 2000) cannot explain why outer-canyon sites have lower salinity than those upcanyon. However the straight line can be explained by diapycnal mixing over the observed depth range. This is consistent with, and confirms our interpretation of, the oxygen anomaly (yellow to light green, Figure 4) being caused by vertical mixing at the hotspot by high turbulent diffusivity ($K_ρ ≃ 16 \times 10^{-4}$ m² s⁻¹, Kunze et al., 2012) of higher oxygen from above and below the O₂ minimum. The convergence between the straight line of the highest DO anomaly (≈4–5 μmol kg⁻¹) and other water masses occurs at σθ ∼ 27.0, an average depth range of 600 m. This implies that mixing at the hotspot, the only sites to experience such high DO anomalies, extends from the bottom up to approximately

Figure 3. Twelve-hour time-series of beam attenuation coefficient (top panels), dissolved oxygen (DO) anomaly (middle panels), and DO (bottom panels) at stations 45, 42, and 32 with columns from down-to upcanyon (see map, Figure 1). Red curves denote isopycnals σθ = 27.20–27.35, bounding the intrusion. Bottom depths vary by ±60 m due to ship drift, and average 1,370 m for Station 45, 1,210 m for Station 42, 1,080 m for Station 23.
600 m depth, which is ~300–400 m above the seafloor depending on station depth. This tracer evidence independently supports previous microstructure measurements of elevated turbulence in the stratified water column extending >300 mab (Kunze et al., 2012). Within the intrusion density range (arrows), interme-

Figure 4. T-S diagram along with potential density contours for stations 16 through 61, spanning roughly the range of the oxygen minimum zone and features discussed in text. Data are colored by DO anomaly magnitude (colorbar on right). Arrows mark upper and lower bounds of the intrusion within the transition sites.

Figure 5. Scatterplot of beam attenuation coefficient (BAC) versus dissolved oxygen (DO) anomaly within the intrusion layer (ρθ = 27.20–27.35), colored by height above the seafloor (colorbar to right). There is a relatively tight relationship between BAC and DO anomaly for much of the range spanning the transition sites. DO anomaly is more scattered in the upper right (ellipse) associated with data within 250 mab in hotspot stations where isopycnals impinge the bottom in a turbulent stratified bottom boundary layer.
Anomalous DO and BAC signals at the hotspot intrusion source are formed by vertical mixing ($K_\rho \sim 10^{-3} \text{m}^2 \text{s}^{-1}$) in a stratified turbulent bottom boundary layer. Turbulence is generated by convergence of upcanyon internal tide energy-flux. Assuming steady-state buoyancy conservation to maintain BBL stratification requires exchange with the stratified interior with BBL outflow of 10–50 m day$^{-1}$; $U \sim O \left( 10^{-3} \text{ m s}^{-1} \right)$ at $\sigma_z = 27.17–27.31 \left( \zeta \sim 740–900 \text{ m} \right)$ sandwiched between inflows (Kunze et al., 2012). This offshore transport and tracer gradients described here can be used to determine the dominant processes governing intrusion dispersal in stratified interior water where turbulence is weaker ($K_\rho \sim 10^{-4} \text{ m}^2 \text{s}^{-1}$) than near the boundary. We consider steady-state 2-D (along-canyon and depth) tracer conservation

$$0 = U \frac{\partial C}{\partial x} + K_s \frac{\partial^2 C}{\partial x^2} + K_\rho \frac{\partial^2 C}{\partial z^2}.$$  \hfill (1)

consistent with the intrusion being continuous and contiguous over at least two weeks. $C$ is the concentration of DO, BAC or some other water-mass property, $x$ and $z$ the along-canyon and vertical coordinates, $U$ the along-canyon velocity, $K_s$ the along-canyon isopycnal diffusivity, $K_\rho$ the turbulent diapycnal diffusivity. Cross-canyon transport is assumed negligible because canyon width < Rossby radius, and internal sinks and sources are ignored. Terms (i), (ii) and (iii) scale as $U/L$, $K_s/L^2$, and $K_\rho/HL^2$ where $L \sim 10 \text{ km}$ between the hotspot and background stations and $H \sim 100–150 \text{ m}$ (Figure 2). Term (iii) is an order of magnitude smaller than advective term (i). This yields a dominant horizontal balance between terms (i) and (ii), resulting in effective isopycnal diffusivity $K_s$ that scales as $UL \sim O \left( 1 \text{ m}^2 \text{s}^{-1} \right)$. This result is similar to estimates from deliberate tracer-release experiments for coastal scales 1–10 km (Inall et al., 2013; Ledwell et al., 1998; Okubo, 1971). This boundary-interior exchange mechanism might cause global off-margin dispersal of BBL fluid in the range of 6–12 Sv, based on a conjectural estimate of volume transport ($jDLU$) using the
following parameters: $U$ described above, continental slope height $D$, global continental-slope length $L$ ($3 \times 10^5$ m), and $f$ the fraction of continental slope involved in mixing and exchange, to which we assign a conservative 5%–10%.

BAC and DO anomaly provide two distinct signatures of mixing at the hotspot and subsequent along isopycnal dispersal. But we should consider alternative possibilities. For example, might the outer-canyon sites themselves have anomalously low oxygen instead of representing background conditions, undermining the interpretation that the hotspot is an anomaly? This is not plausible. Apparent oxygen utilization increase along the 26.6 isopycnal in the northern Pacific (Mecking et al., 2008) affects waters shallower than those examined here, as does variability associated with the slope undercurrent (Collins et al., 2000). Could there be a causal relationship between the two variables not explained by co-dispersal from the mixing hotspot? There are some scenarios in which increased oxygen might be associated with particles; in the euphotic zone, for example, photosynthesizing phytoplankton and algae introduce oxygen to the water column briefly before it is remineralized. However, the intrusion layer is well below the euphotic zone. Fluorescence above instrument noise level of ~0.03 mg m$^{-3}$ was measured only above 100 m.

What would happen if DO were not semi-conservative with respect to BAC, but instead underwent Lagrangian $\frac{dO}{dt} \propto \Delta$BAC (i.e., internal source or sink caused by particles) following along-isopycnal dispersion of benthic particles within the INL? At depths below the OMZ, $dO/dt$ is dominated by abyssal benthic respiration with consumption rates 10–50 μmol m$^{-3}$ yr$^{-1}$ (Emerson & Hedges, 2008). How much anomaly attenuation offshore of the hotspot might be attributed to respiration by benthic particles instead of lateral diffusion? Using a timescale of a year (based on length and times scales above) for offshore transport, a rough estimate of oxygen depletion is 0.05–0.1 μmol kg$^{-1}$. This is two orders of magnitude weaker than the anomaly observed. We conclude that DO is independent of BAC and that the two constituents are primarily affected by advection and diffusion once the water mass leaves the boundary hotspot.

We are unaware of other studies using CTD oxygen profiles to identify intrusions and examine characteristics of boundary-interior exchange. Perhaps this could become a more widely used technique. However, it requires mixing to coincide with a strong property anomaly, which is not always the case. Whether processes found in Monterey Canyon are typical of other canyons and slopes is not known. 20 km-resolution NE Pacific CALCOFI surveys found Monterey Canyon 737 dbar site site to be characterized for many decades by 50% higher oxygen than elsewhere at that depth (Collins et al., 2010). Our study confirms this anomaly persists at higher spatial resolution, and establishes a link to internal wave energy flux convergence. Deep oxygen has been decreasing globally for decades (Bograd et al., 2008; Breitburg et al., 2018) and 2–5 μmol kg$^{-1}$ is the same order of magnitude as climate-change-related oxygen decrease in some eastern Pacific waters (Czeschel et al., 2012). This emphasizes the importance of siting sustained ocean monitoring stations (e.g., Garcon et al., 2019) away from anomalies that may exhibit shorter space and time scales. Some higher trophic level organisms respond to variations in abyssal anoxia (for example Humboldt Squid modify abyssal dive patterns in response to OMZ depth (Gilly et al., 2012)); it may be worth investigating whether this oxygen anomaly hotspot might invoke behavioral responses in certain pelagic organisms.

Data Availability Statement
Data are publicly available at University of Washington Research Works Archive: http://hdl.handle.net/1773/46929.

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