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Tidal mixing processes amid small-scale, deep-ocean topography

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Abstract The nature of tide-topography interaction reflects the topographic scales experienced by water parcels during their tidal excursions. In the deep ocean these scales are typically subkilometer, yet direct observations of tidal processes on such scales are lacking. At one site, a saddle amid steep and complex Mid-Atlantic Ridge topography, observations reveal tidally pulsed, bottom-trapped fronts, overflows, and lee waves in response to a tide combined with a mean flow of similar amplitude. The tidal pulsing of the fronts and overflows was only evident locally, and their phase became unpredictable over scales of hundreds of meters. Enhanced turbulence in a 100–200 m thick bottom boundary layer had an estimated dissipation rate of $2.6 \times 10^{-2}\text{Wm}^{-2}$, exceeding the large-scale average of tidal dissipation in mid-ocean ridge environments but by less than an order of magnitude. This site was not a dissipation “hotspot,” and the processes observed could provide widely distributed mixing to the meridional overturning circulation.

1. Introduction

Mixing in the deep ocean draws buoyant water downward. This provides potential energy to the water column which, in part, sustains the meridional overturning circulation (MOC). Since tides are a key agent of deep mixing [Munk and Wunsch, 1998; Wunsch and Ferrari, 2004], the magnitude and distribution of tidal mixing profoundly influence the structure and timescales of the MOC [Hasumi and Suginohara, 1999; Jayne, 2009], and consequently influence the ability of the deep ocean to buffer the climate system [Meehl et al., 2011].

Tidal currents encountering sloping topography lose energy, both to dissipative processes that drive local mixing, and to radiated internal waves. In view of the sensitivity of the MOC to the distribution of mixing, it is important to understand the balance between local dissipation and radiation and the location and efficiency of their ultimate delivery of energy to mixing. These aspects are highly process dependent and poorly understood. A key difficulty is that tidal processes reflect topography on the scales experienced by water parcels during their tidal excursions, and, in the deep ocean, these scales are typically less than a kilometer. Abyssal topography is rich in these scales, with mid-ocean ridges in particular providing vast tracts of complex, multiscale topography. While observational confirmation of the role of tidal mixing in mid-ocean environments comes from a spring-neap modulation of dissipation [Ledwell et al., 2000], direct, process level observations in such environments are restricted to topographic features that are considerably larger than the tidal excursion [Alford et al., 2014; Klymak et al., 2008; Van Haren et al., 2005] or cases where the mean flow dominates the tide [Alford et al., 2013; St Laurent and Thurnherr, 2007; Thurnherr et al., 2005]. An additional difficulty is that ocean bathymetry is not consistently known at subkilometer scales. Satellite altimetry, while providing global bathymetric coverage, resolves only scales greater than 25 km [Smith and Sandwell, 1997] and higher resolution ship-based surveys are only patchily available. Merged best available topographies, therefore, significantly underestimate both bottom slope and excursion-scale complexity [Becker and Sandwell, 2008; Zibelman et al., 2009]. The likely significance of unresolved scales in tidal energy conversion [Melet et al., 2013] poses a challenge for global models of tidal dissipation [Nycander, 2005; Simmons et al., 2004], which also face computational limits on their resolution. While such models appear to be converging on the known (astronomically derived) energy loss from the tide, they potentially remain deficient at the process level and vary in the location of their delivery of energy to mixing [Green and Nycander, 2013].

For an isolated topographic feature, the processes governing tide-topography interaction depend on the slope of the flanks of the feature, the vertical density gradient, geographic latitude (via the Coriolis parameter $\beta$), and the horizontal scale of the topography relative to the tidal excursion [Garrett and Kunze, 2007]. The complexity of the situation is compounded if there is nontidal flow or three-dimensionality of background flows.
or topography. When topographic scales are large compared to the tidal excursion, individual water parcels experience the topography as a slope. This case is relatively well understood, although turbulent and complex behavior may occur. When the bottom slope is near-critical (i.e., close to the internal tide propagation angle), near-bed flow becomes highly sheared and upslope-propagating bores, resembling gravity currents, may be generated [Aucan et al., 2006; Gayen and Sarkar, 2011; Gemmrich and Van Haren, 2001]. Similar bores have also been observed at noncritical slope angles [Van Haren, 2005; Van Haren et al., 2005] and may result from internal waves impinging on a slope [Hosegood and van Haren, 2004].

When topography contains scales that are comparable to or shorter than the tidal excursion, water parcels experience changes of gradient, permitting processes such as lee waves, hydraulically controlled flows and overflows. Although changes of gradient on the flanks of larger features are included in this case [Klymak et al., 2008; Legg and Klymak, 2008], the focus here is on features that are themselves of a size that is comparable to the excursion scale. While several studies have considered tidal interactions with excursion-scale topography from a modeling perspective [Legg and Huijts, 2006; Nikurashin and Legg, 2011], detailed observations have been lacking.

Our studies were conducted on the eastern flank of the Mid-Atlantic Ridge at 49°N (Figure 1) at a site where, in 2007, a single-day lander record had shown the passage of two cold fronts separated by a tidal period. The detailed setting, revealed by a multibeam bathymetric survey [Niedzielski et al., 2013], was a saddle, the low point (depth 2535 m) of a spur emanating from a more extensive elevated area to the west and NW (summit depth 1555 m, 2 km west of the boundary of Figure 1b), set within an extensive tract of complex, mid-ocean ridge topography. The flanks of the spur were asymmetric, with a short slope (70 m vertical, 250 m horizontal) to a small basin to the SW, and a longer slope (270 m vertical, 1000 m horizontal) to a more extensive deep
area to the NE. Time series from seafloor lander deployments, a moored thermistor chain, and “yoyo” profiles provide insights into the tidal processes of this site and their spatial variability over scales of hundreds of meters. These observations allow an estimate of local mixing energetics which will be related to the global tidal mixing budget.

2. Methods

2.1. Moored Instruments: Thermistor Chain and Lander

Data collection was during August 2009 from RRS James Cook at three sites within 500 m of the saddle (Figure 1c). At site A, 230 m to the NE of the saddle and 25 m downslope from it, a vertical array of thermistors (RBR XR420TiD + T16) was moored for 18 semidiurnal tidal cycles. The array consisted of 16 equally spaced thermistors positioned 9–109 m above the seabed and sampling with a 5 s interval. While the thermistor chain was at A, a concurrent set of observations was made on the opposite flank of the spur. A bottom lander was deployed, initially for 5–6 tidal cycles at site C1, 370 m to the SSW of the saddle and 80 m deeper than it. The lander was then redeployed for a further 14 tidal cycles at B, 15 m below the saddle and 370 m to the WNW of it on the SW flank of a slightly higher part of the spur. Lander instrumentation consisted of a Seabird 37 CTD (conductivity-temperature-depth) logger sampling at 3.4 m above the bed, and an upward looking 300 kHz Acoustic Doppler Current Profiler (ADCP) recording velocity profiles in 2 m vertical bins from 8 m to approximately 50 m above the bed. Both instruments had a 10 s sample interval. In view of the unknown lateral displacement due to horizontal currents as these moored instruments descended through the water column, their final positions on the seabed were triangulated acoustically to within an estimated 50 m. Most (10 of the 16) thermistors showed time-varying, monotonic calibration drift, which has been corrected by comparison with nearby stable thermistors using a spline temporal correction to match a vertically interpolated mean temperature.

2.2. Yoyo Temperature, Salinity, and Velocity Profiles

Concurrent with the thermistor array at A and the lander at C1, repeated vertical yoyo profiles of temperature, salinity, and velocity were collected from the ship stationed at site C2, 90 m to the NE of C1, for 24 h using a 24 Hz Seabird 911+ CTD (conductivity-temperature-depth) system. C1 and C2 were essentially collocated within the known accuracy of their locations, but not necessarily with respect to sub-100 m flow structures.

Figure 2. Time series during yoyo profiling on 9 August 2009 at C2. (a) Horizontal velocity vectors (y axis is oriented to north), averaged over the bottom 500 m, from lowered ADCP. (b) Vertical temperature structure. Black dots indicate Thorpe overturning scales exceeding 25 m. The vertical structure of the tidally averaged TKE dissipation rate (ε) is shown to the right. (c) Temperature at 4 m above the bed from the bottom lander at nearby site C1. (d) Velocity at 8 m above the bed from the bottom lander at site C1. Times are relative to low water at A.
A total of 77 profiles were obtained, spanning from a depth of 2150 m to within 10 m of the bed. A pair of upward and downward looking 300 kHz Acoustic Doppler Current Profilers (ADCPs) was mounted on the CTD frame. The vessel held position using Dynamic Positioning during this period. Depth-averaged velocities have been calculated under the assumption that when bottom tracking was not available there was no horizontal motion of the instrument package. This leads to an error with an estimated RMS amplitude of 0.8 cm s\(^{-1}\), an order of magnitude smaller than the observed variability. The sea state was slight.

2.3. Dissipation Estimation From Overturns

Using calibrated 24 Hz temperature profiles from the yoyo, a conservative method was used to calculate the vertical extent of overturns of the water column [Thorpe, 1977]. Setting the noise level of the Seabird thermistor to 0.002°C and defining the Thorpe scale \((L_T)\) as the root-mean-square of the vertical displacements of the monotonically reordered temperature profiles, the Ozmidov scale \((L_O)\) was determined [Ferron et al., 1998] according to \(L_O = 0.95L_T\) and the turbulent kinetic energy dissipation rate \(\varepsilon = \langle N^2 \rangle L_O^2\). The angle brackets here denote an average over turbulent (overturning) patches and \(N\) is the buoyancy frequency of the reordered water column.

3. Observed Tidal Dynamics

In the central North Atlantic, the tide is dominated by the lunar semidiurnal \((M_2)\) constituent. The predicted depth-mean \(M_2\) flow at the study site (TPXO7.2 tidal model [Egbert and Erofeeva, 2002]) is largely east-west with current ellipses having a major axis of 3.6 cm s\(^{-1}\) and a tidal excursion of 506 m (Figure 1). The TPXO model does not resolve topography smaller than its quarter-degree resolution, however, and local steering of tidal flows is expected. The observed velocity, averaged over the bottom 500 m of the water column during the yoyo at C2, showed a tidally pulsed flow peaking at 7–8 cm s\(^{-1}\) to the SW (Figure 2a). This flow largely stalled on each tidal cycle but did not reverse to the NE, a pattern consistent with tidal currents comparable in amplitude to the TPXO predictions superimposed on a background flow to the SW of similar magnitude.

![Figure 3. Thermistor chain time series of temperature in the bottom 109 m of the water column at site A, 7–16 August 2009. Time is in hours relative to local low water for the M2 tidal constituent (TPXO7.2 model). Inset numbers give the sequential tidal cycle within the data set. Only those tidal cycles where sharp bottom fronts were observed are shown, with cycle 15 included as representative of the cycles on which a sharp front was not observed.](image-url)
Two distinct phenomena were observed in response to this tidal forcing: the pulsing of dense water across the saddle during the strongest flow to the SW, and the development and release of internal lee waves. Bottom-trapped pulses of relatively cool (dense) water (Figure 3) at the thermometer array site, A, traveled to the SW across the saddle (the direction is inferred from the near-bed current associated with these features in the 2007 observations). Of these cool pulses, 10 of 18 were led by a sharp front at which near-bed temperatures dropped by at least 0.05°C in 10 min, nearly half of the total observed temperature range at this site and equivalent to a density increase of approximately 0.01 kg m$^{-3}$ (temperature and density were related monotonically). On no occasion was there a similarly sharp rise in temperature. The cool pulses were clearly tidal, with sharp fronts clustering in phase between 5.1 and 7.3 h after local low water (Figure 4) as the N-S flow at C2 stalled prior to building to the SW (Figure 2a). The fronts were traveling upslope, approaching the crest of the spur. No such fronts were observed on the opposite phase of the tide. The vertical and temporal structure of the fronts varied considerably (Figure 3), with their thickness ranging from O(10 m) to more than 100 m. Typical characteristics were a sloping leading edge and an elevated head with overturning on its trailing side followed by a depression and gradual thickening of the cool layer. In these respects they strongly resembled the upslope-propagating bores described by Van Haren [2005] and modeled by Gayen and Sarkar [2011]. There was no clear spring-neap cycle to the occurrence, timing, or structure of these fronts (spring tides occurred early in the record and neap tides toward the end). The role of slope criticality in their formation is unknown; slopes in the region are largely supercritical (steeper than the internal tide propagation angle; Figure 1c) with transitions through criticality on the flanks of summits and troughs, including near A. The tidal phasing at A suggests generation downslope of this site on the NE flank of the spur. The tidal asymmetry of this process (cool pulses pass over the saddle to the SW, but not to the NE on the opposite phase of the tide) is likely imposed by the background flow; however, topographic asymmetry may also play a role.

Following the passage of each front, cool water persisted at A for several hours during the strongest flow to the SW at C2. It is assumed that each cool pulse continued for the 230 m to the summit and spilled down the slope to the SW as a gravity current with an estimated frontal speed [Shin et al., 2004] of 0.07 m s$^{-1}$ ($c = \sqrt{g' H}$ where $g' = g \Delta \rho / \rho$ is the reduced gravity, cool layer thickness $H = 50$ m, frontal density step $\Delta \rho = 0.01$ kg m$^{-3}$, ambient density $\rho = 1040$ kg m$^{-3}$, and $g$ is the acceleration due to gravity). This speed is comparable to the peak depth-mean current. An estimate of the resultant energy flux $E$ across the spur can be obtained from the sum of kinetic and potential energy fluxes based on water of depth $H$ with density anomaly $\Delta \rho$ traveling at speed $c$. The kinetic energy density of this flow is $\rho c^2/2$ and its potential energy density
Lee wave formation occurs when upstream flow \( U \), high vertical modes are the first to be stalled and there is a tendency for upstream flow to be blocked beneath the level of the topography [Klymak et al., 2010]. As the flow speed increases, the lowest mode stalls at a topographic Froude number \( Fr = U/hN \approx 1 \) (where \( h \) is the height of the topography). Lee waves then steepen to the point of overturning [Baines, 1995] and form a downstream hydraulic jump. In the present case, \( Fr = 1 \) is realized at near-peak flow \( U = 0.07 \text{ m s}^{-1} \) for an ambient buoyancy frequency \( N = 10^{-3} \text{ s}^{-1} \) and \( h = 70 \text{ m} \), representing the SW flank of the spur. In an oscillating tidal flow, an additional requirement for the formation of a hydraulic jump is that the slope is steep (supercritical) [Legg and Klymak, 2008], as in the present case. Observations to date of tidal lee waves and hydraulic jumps in the deep ocean have been associated with significantly taller topography than is considered here [Alford et al., 2014; Klymak et al., 2008; Pinkel et al., 2012], combined with stronger flows and smaller Froude number. The present observations provide confirmation of lee wave formation due to small-scale topography, a situation which has previously been restricted to numerical studies [Legg and Huijts, 2006].

Overturbs larger than 25 m, corresponding here to a dissipation rate \( \epsilon > 2.0 \times 10^{-7} \text{ W kg}^{-1} \), reveal an approximately semidiurnal periodicity to mixing centered around the 3.22°C isotherm (~250 m above the bed, W1 and W2 in Figure 2b), and independently within the 100 m to 200 m thick bottom layer (B1 to B3, Figure 2b). B1 and B2 persist for several hours and are associated with peak flows and lee wave formation and release. B3 occurs earlier in tidal phase, is more strongly bottom trapped, and is associated with a patch of cool water in the bottom tens of meters with a sharp leading front (Figure 2c). Passage of the front is followed by near-bed flow to the south (Figure 2d). We cautiously identify this as a downslope-propagating gravity
current. W1 and W2 occur higher in the water column, with near-opposite tidal phasing to B1 and B2. Although three-dimensional complexity of the site is such that the source of these features is uncertain, they resemble patches of vertical straining and mixing that can result from lee wave formation and release [Klymak et al., 2008].

Vertical mixing and dissipation associated with the observed locally intensified tidal phenomena were large. Peak TKE (Turbulent Kinetic Energy) dissipation rates, derived from the largest individual overturns, were $\epsilon = 2.1 \times 10^{-6} \text{ W kg}^{-1}$ within the bottom features B1 to B3 and $\epsilon = 2.5 \times 10^{-6} \text{ W kg}^{-1}$ within the water column features W1 and W2. Averaging over two tidal cycles gives a depth-integrated dissipation rate of $2.6 \times 10^{-2} \text{ W m}^{-2}$.

While the spatial complexities of sites such as this cannot be unraveled from a small number of sampling locations, the observations provide insight into the phenomena to be expected in such environments and their associated energetics. In particular, they emphasize the fact that the lateral scales of variability reflect the subkilometer scales of the tidal excursion, the internal Rossby radius, and similar topographic scales.

4. Upscaling

Astronomical observations place the dissipation of tidal energy in the ocean at around 3.5 TW [Munk and Wunsch, 1998]. Satellite altimeter estimates suggest that, of this, around 1 TW is lost by the tides in the deep ocean as a result of topographic interactions [Egbert and Ray, 2001], contributing nearly half of the 2.1 TW required to support the MOC [Munk and Wunsch, 1998; Wunsch and Ferrari, 2004].

Of the 1 TW of tidal energy that is lost in the deep ocean, estimates of the locally dissipated component vary widely [Munk and Wunsch, 1998; St Laurent and Garrett, 2002]. Assuming that 0.7 TW of the 1 TW is dissipated locally [Munk and Wunsch, 1998] and that this dissipation is evenly distributed over the 23% of the Earth’s surface that is covered by mid-ocean ridges [Heezen, 1969] yields a mean spatial density of local tidal dissipation of $6 \times 10^{-5} \text{ W m}^{-2}$. This is around a quarter of the dissipation that we have estimated from the bottom 500 m of the yoyo and half the dissipation estimated from gravity current energetics downslope of the saddle. Dissipation at this site is therefore elevated above the expected large-scale spatial mean, but not by orders of magnitude. While dissipation “hot spots” have been identified in several deep-ocean locations [Alford et al., 2013; MacKinnon et al., 2008; Pinkel et al., 2012], such sites are, by definition, atypical. In contrast, less dissipative phenomena, such as those described here, could be widely present in complex bathymetric environments, setting the fluid dynamical character of more typical mid-ocean ridge sites.

5. Conclusions

Currents tend to be weak in the deep ocean. In this weakly stratified environment, however, even relatively low-energy flows may be highly turbulent. Tidal processes are expected to reflect the subkilometer scales of the tidal excursion, of the internal Rossby radius, and similar scales in the seafloor topography. Attempts to observe tidal physics on these scales, however, are confounded by inherent complexity, variability and three dimensionality as well as by the practical problems of observing the deep ocean. These include the difficulty of accurately positioning instrumentation to sub-100 m accuracy from a vessel several kilometers above. Nevertheless, we have described tidally pulsed fronts, overflows, and lee waves from an essentially randomly selected site. The tidal phase locking of these phenomena, while clear near to their generation site (a saddle), was rapidly lost over just a few hundred meters. These were very local phenomena which did not always clearly reveal their tidal origin. The ubiquity of such topography suggests that such phenomena may be widely present in deep-ocean environments but also site specific in their character, reflecting local topography and the relative local importance of tidal and background low-frequency flow. While global general circulation models are far from resolving topography and physics on the tidal excursion scale, it is suggested that parameterizations of tidal dissipation and mixing would better reflect process-level physics if they reflected regional differences in the small-scale character and complexity of topography.
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