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# 1 **Focusing of baroclinic tidal energy in a canyon**

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**Abstract.**

**Key points:** Generation of internal tide in a circular canyon; Focusing of baroclinic tidal energy in the canyon centre; Generation of a baroclinic eddy due to diapycnal mixing.

Strong three dimensional focusing of internal tidal energy in the Petite Sole Canyon in the Celtic Sea is analysed using observational data and numerical modelling. In a deep layer (500-800 m) in the centre of the canyon shear variance was elevated by an order of magnitude. Corresponding large vertical oscillations of deep isotherms, and a local maximum of horizontal velocity were replicated numerically using the MITgcm. The elevated internal tidal activity in the deep part of the canyon is explained in terms of the downward propagation and focusing of multiple internal tidal beams generated at the shelf break. The near-circular shape of the canyon head and steep bottom topography throughout the canyon (steeper than the tidal beam) create favourable conditions for the lens-like focusing of tidal energy in the canyon's centre. Observations and modeling show that the energy focusing greatly intensifies local diapycnal mixing, that leads to local formation of a baroclinic eddy.

## 1. Introduction

20 Oceanic canyons are potential places for significant tidal energy conversion from the  
 21 barotropic to baroclinic modes, with major implications for water mass mixing. According  
 22 to *Hickey* [1995], nearly 20% of the Eastern Pacific shelf edge between Alaska and the  
 23 equator is dominated by steep, narrow, and abrupt canyons. Historically, the first and  
 24 most extensively studied canyon was La Jolla Canyon (California). The results by *Shepard*  
 25 [1974] and *Gordon and Marshall* [1976] showed that steep canyons can act as a trap  
 26 for tidally generated internal waves. Specifically, it was recognised that the dynamical  
 27 processes occurring in canyons strongly depend on the ratio of the maximum bottom  
 28 steepness ( $S_{topo} = \partial H/\partial l$ ) (here  $H(x, y)$  is the water depth, and  $l$  is the direction of the  
 29 seabed depth gradient vector) to the inclination of the characteristic paths of the internal  
 30 wave energy propagation

$$31 \quad S_{wave} = dz/dl = \pm[(\omega^2 - f^2)/(N^2(z) - \omega^2)]^{1/2}, \quad (1)$$

32 where  $\omega$  is the tidal frequency,  $f$  is the Coriolis parameter, and  $N(z)$  is the buoyancy  
 33 frequency. In other words, in terms of the mechanism of internal wave dynamics, the  
 34 following parameter

$$35 \quad \alpha(x, y, z) = \frac{S_{topo}}{S_{wave}} = \frac{|\partial H/\partial l|}{[(\omega^2 - f^2)/(N^2(z) - \omega^2)]^{1/2}} \quad (2)$$

36 is the principal measure of the bottom steepness that distinguishes two very different  
 37 regimes of tidal energy conversion. Schematically they are presented in Figure 1. For  
 38 specificity, the buoyancy frequency measured in a canyon of the Celtic Sea (reported in  
 39 [*Vlasenko and Stashchuk*, 2015]) is used for the analysis, see Figure 1 a.

40 In a subcritical regime, when the condition  $\alpha < 1$  is valid over the whole domain, it is  
41 mostly the lower tidal baroclinic modes that are generated. This regime is presented in  
42 Figure 1 b.

43 Over “steep” topographies, for which  $\alpha > 1$  occurs at least for some fragments of a given  
44 slope, the supercritical regime of tidal energy conversion, presented in Figure 1 c (below  
45 point C), is fulfilled substantially increasing the conversion rate [*Gerkema et al.*, 2004]. In  
46 this regime internal tidal energy is concentrated in a narrow internal tidal beam (magenta  
47 stripe in Figure 1 c) that radiates energy away from bottom fragments where  $\alpha = 1$ . The  
48 energy propagates in the tidal beam upward and downward along characteristic line (1)  
49 with group velocity  $C_g$  while the wave phase propagates across the beam with the phase  
50 speed  $C_p$ , Figure 1 c. In fact, it is not only the point C with  $\alpha = 1$  that is the area of the  
51 beam formation. A wider area A-B where the bottom inclination is close to critical is the  
52 place of the beam generation. More on different regimes of tidal energy conversion can be  
53 found in [*Vlasenko et al.*, 2005].

54 Note that intensity of the tidal beam depends not only on the relative steepness of  
55 the bottom or strength of the tidal current. It is also controlled by smoothness of the  
56 buoyancy frequency profile. In highly intermittent media with sharp vertical changes of  
57 the buoyancy frequency the downward propagated tidal energy is reflected back from the  
58 layered structures in the form of secondary tidal beams propagating to the free surface  
59 [*Grimshaw et al.*, 2010]. This process can even lead to a complete attenuation of the  
60 tidal beam [*Gerkema and van Haren*, 2012]. However, as it was shown by *Vlasenko and*  
61 *Stashchuk* [2015], the surface 1.5 km layer of the Celtic Sea (which is in the focus of the  
62 present study) is mostly unaffected by internal reflection.

63 Note that the tidal beam is a superposition of many baroclinic modes, as it was clearly  
64 shown in the analytical solution for internal tides at abrupt topographies by *St. Laurent*  
65 *et al.* [2003], and the presence of a beam is evidence of higher mode excitation. For a  
66 typical “V”-shaped canyon with  $\alpha \geq 1$ , the baroclinic tidal energy is trapped inside the  
67 canyon, being able only to propagate downward reflecting many times from canyon’s steep  
68 flanks without any opportunity of escape (see, for instance, Figure 2 in [*Balmforth and*  
69 *Peacock*, 2009]).

70 The importance of the relative bottom steepness  $\alpha$  for internal wave dynamics in canyons  
71 was acknowledged by *Petruncio et al.* [1998] in their interpretation of measurements con-  
72 ducted in another well-studied canyon, Monterey Submarine Canyon. In further analysis  
73 by *Zhao et al.* [2012], who investigated the energetic characteristics of internal waves and  
74 turbulent mixing in the canyon, it was suggested that the topographic steepness does  
75 control the energy conversion rate.

76 Some studies on the baroclinic dynamics in canyons were conducted for an idealized  
77 bottom profiles by *Baines* [1983]; *St. Laurent et al.* [2003]; *Grimshaw et al.* [1985]; *Zhang*  
78 *et al.* [2014]. A more realistic model set up, specifically, the real bottom topography, was  
79 taken in a further series of numerical experiments performed for Monterey Submarine  
80 Canyon area: *Petruncio et al.* [2002] used POM, *Jachec et al.* [2006] operated with SUN-  
81 TANS, and *Hall and Carter* [2011] used POM to investigate internal tides in the canyon  
82 area. However, the specific role of the relative steepness  $\alpha$  of the canyon topography in  
83 the distribution of the baroclinic wave energy was not investigated in any of these studies.  
84 Specifically, the structure of the baroclinic tidal field in the areas where  $\alpha > 1$ , a common  
85 occurrence, has not hitherto been discussed in detail.

86 The purpose of this paper is to interpret the three-dimensional effects of internal tidal  
87 dynamics that are seen to occur in supercritical canyons. In particular, an original obser-  
88 vational data set collected in a supercritical canyon in the Celtic Sea is analysed here in  
89 terms of the focusing of internal tidal energy radiated from the areas with critical bottom  
90 inclination.

## 2. Observations

91 The Celtic Sea is a 200 meter deep, wide shelf sea with a large number of headlands  
92 and canyons along its shelf edge, Figure 2 a. Observations analysed here were conducted  
93 on the 376-th cruise of the RRS “Discovery” (hereafter D376) in June 2012, as a part  
94 of the FASTNEt study to quantify the cross shelf transport on the NE Atlantic Ocean  
95 margin. With relevance to this paper, 14 repeat “yo-yo” CTD profiles were conducted  
96 at a station precisely in the middle of the Petite Sole Canyon presented in Figure 2 b.  
97 Vertical profiles were repeated with approximately one hour time interval to the depth of  
98 approximately 1000 m. In addition to the CTD probe, a downward looking TRDI WHM  
99 300kHz LADCP was mounted on the frame, so that each CTD profile was accompanied  
100 by a vertical profile of currents.

101 All 14 “yo-yo” temperature profiles  $T_j(z)$  ( $j = 1, 2, 3, \dots, 14$ ) are presented in Figure 3 a  
102 by blue lines. The red line shows an equilibrium temperature distribution calculated as an  
103 average temperature at each depth. Comparing all individual temperature profiles  $T_j(z)$   
104 with the average  $T(z)$  one can see that the largest deviations of every individual profile  
105 from the average take place in two zones: in the surface 150 m layer, and between 500 m  
106 and 900 m depth. Assuming that these deviations were caused by the dynamical processes  
107 developing in the canyon, the vertical displacement  $\zeta_j$  of every individual isotherm on

108 every profile from the averaged temperature profile can be calculated using the following  
 109 formula:

$$110 \zeta_j = \frac{|T_j - T|}{\partial T / \partial z}$$

111 An average profile of isotherms displacements, calculated as  $1/14 \sum_{j=1}^{14} \zeta_j(z)$ , is presented in  
 112 Figure 3 b. It shows that the absolute maximum vertical displacements of isotherms, up  
 113 to 45 m, are located in deep water, between 600 and 800 m depth. The deviation profile  
 114 calculated as a maximum displacement of the isotherm from its equilibrium state is close  
 115 to that shown in Figure 3 b.

116 Note that the maximum baroclinic horizontal velocities at the “yo-yo” station were  
 117 recorded by the LADCP in the surface layer (see also [Vlasenko *et al.*, 2014]). However,  
 118 Figure 3 b suggests that a comparable contribution of the 600-800 m depth layer to the  
 119 internal wave energy is also expected. In order to quantify the kinetic energy of dynamical  
 120 processes developing at the “yo-yo” station, an average profile of horizontal velocities for  
 121 all 14 LADCP sampling was calculated as follows

$$122 U(z) = 1/14 \sum_{j=1}^{14} \sqrt{u_j^2(z) + v_j^2(z)}.$$

123 Here  $u_j(z)$  and  $v_j(z)$  ( $j = 1, 2, \dots, 14$ ) are eastward and northward velocities.

124 The mean profile of  $U(z)$  is presented in Figure 3 c (magenta line). It shows the speed  
 125 maximum at the free surface as well as a secondary local maxima at the depth of 680 m  
 126 which coincides with the position the maximum vertical displacement and a broad region  
 127 of elevated currents between 600 m and 800 m, produced presumably by the action of  
 128 internal waves.



129 As a proxy-measure for the relative strength of diapycnal mixing, we follow the method  
130 of *Polzin et al.* [2002] in computing the buoyancy-normalised LADCP shear in overlap-  
131 ping segments (50% overlap), each of 128 meters in vertical extent. The spectra for each  
132 segment were then integrated between vertical wavelengths of 25 m to 128 m, to give 14  
133 profiles of shear variance, a property that scales with the diapycnal eddy diffusivity. The  
134 time averaged vertical profile of the shear variance (Figure 3 d) demonstrates a relative  
135 7-fold increase below 700 m depth, with a broad maximum at 780 m. The necessary seg-  
136 mentation of the data results in loss of vertical resolution in comparison with temperature  
137 or velocity profiles, but nevertheless a consistent deep maximum in fine-structure derived  
138 vertical mixing is clearly apparent.

139 Intensification of vertical and horizontal motions in the layer between 500 and 800 m  
140 depth, classified above as a consequence of internal waves action is confirmed by Fig-  
141 ure 4, where the temperature and eastward horizontal velocity profiles recorded at all  
142 14 CTD stations are shown as time series. Both panels reveal semidiurnal periodicity  
143 of the recorded signals, with strong baroclinic contributions (note the intensification of  
144 horizontal currents and vertical oscillation in the surface 200 m, which is consistent with  
145 Figure 3). The most interesting feature of Figure 4 a is clear evidence of intensive semid-  
146 iurnal periodicity below 500 m depth (white dashed line in panel a) with relatively quite  
147 background between 500 and 200 m.

148 Horizontal currents also reveal semidiurnal periodicity, most clearly seen in the surface  
149 400 m layer. At first glance this periodicity can be attributed to barotropic tidal motions.  
150 However, weakening of the tidal signal below 400 m, and its vertical intermittency in the  
151 deep layers suggests that there should also be a strong contribution of a baroclinic signal.

152 Moreover, analysing the relative location of minima and maxima of the velocity pattern  
153 in space and time one can conclude that Figure 4 b can be treated as evidence of evolution  
154 of a baroclinic tidal beam with its phase moving upward (as it is presented in Figure 1 c).  
155 Co-phase segments of the beam are marked in Figure 4 b by white dashed lines. It is  
156 interesting to note that similar features have been found also in the model output which  
157 was applied in this paper to replicate the tidal dynamics in the canyon area. The results  
158 of modelling are discussed in the next section.

159 In order to bring more clarity to the interpretation of these relatively sparse observations  
160 and to understand the reasons for apparent internal wave energy and vertical mixing  
161 intensification in the deep part of the canyon, a series of numerical experiments was  
162 conducted.

### 3. Model results

163 The fully non-linear non-hydrostatic MITgcm was used to model internal tides in the  
164 canyon and the surrounding area (see Figure 2 b). In the main part of the model domain  
165 the horizontal and vertical resolutions were 100 m and 10 m, respectively. In order to avoid  
166 any spurious boundary reflections, an exponentially increasing horizontal grid step near  
167 the lateral boundaries was used that guaranteed an accurate numerical solution within  
168 the internal model domain without any signals reflected from the boundaries during at  
169 least 10 tidal cycles.

170 The tidal forcing was set in the model by a tidal potential added to the right hand side  
171 of the momentum balance equations. Its intensity was chosen using TPXO8.0 [*Egbert and*  
172 *Erofeeva, 2002*] in such a way as to reproduce tidal velocities recorded by moored ADCP  
173 current meters deployed during D376 that have shown the predominance of semi-diurnal

174 M2 tidal harmonic [Vlasenko et al., 2014]. Spatial distribution of tidal ellipses is shown in  
 175 Figure 2 d. A vertical stratification was introduced into the model after setting the tidal  
 176 forcing for a homogeneous fluid. The temperature and salinity profiles were taken from  
 177 the direct CTD measurements at the yo-yo station.

178 The Richardson number dependent parametrization for vertical viscosity  $\nu$  and diffu-  
 179 sivity  $\kappa$  introduced in [Pacanowski and Philander, 1981] was used:

$$180 \quad \nu = \frac{\nu_0}{(1 + \beta \text{Ri})^n} + \nu_b, \quad \kappa = \frac{\nu}{(1 + \beta \text{Ri})} + \kappa_b.$$

181 Here Ri is the Richardson number,  $\text{Ri} = N^2(z)/(u_z^2 + v_z^2)$ ,  $u$  and  $v$  are the components of  
 182 horizontal velocity;  $N(z)$  is the buoyancy frequency  $N^2(z) = -g/\rho(\partial\rho/\partial z)$  in which  $g$  is  
 183 the acceleration due to gravity and  $\rho$  is the water density;  $\nu_b=10^{-5} \text{ m}^2 \text{ s}^{-1}$  and  $\kappa_b=10^{-5} \text{ m}^2$   
 184  $\text{ s}^{-1}$  are the background viscosity, and diffusivity, respectively;  $\nu_0=1.5 \cdot 10^{-2} \text{ m}^2 \text{ s}^{-1}$ ,  $\beta=5$   
 185 and  $n=1$  are the adjustable parameters. Such a parametrisation increases  $\nu$  and  $\kappa$  in  
 186 areas where the Richardson number is small. The horizontal viscosity and diffusivity were  
 187 set to a constant value of  $0.5 \text{ m}^2 \text{ s}^{-1}$ . More details on the model initialization and input  
 188 parameters can be found in Vlasenko et al. [2014].

189 The principal question to be addressed by the modelling efforts is to identify the cause  
 190 of the highly energetic internal wave activity in the centre of the canyon below 500 m  
 191 depth. The modelling evidence of this intensification in deep water is seen in Figure 5  
 192 where the amplitude of the model-predicted horizontal velocities

$$193 \quad U_{\max}(x, y, z) = \sqrt{u_{\max}^2(x, y, z) + v_{\max}^2(x, y, z)}.$$

194 is presented. Here  $u_{\max}(x, y, z)$  and  $v_{\max}(x, y, z)$  are amplitudes of the eastward and  
 195 northward velocities found over one tidal cycle at the position  $(x, y, z)$ .

196 Nine horizontal slices of the velocity  $U_{\max}$  at depths of between 300 m and 700 m pre-  
 197 sented in Figure 5 reveal quite a curious tendency. In the surface layers, i.e. shallower  
 198 than 500 m, the wave energy is mostly concentrated at the periphery of the canyon corre-  
 199 sponding to the shelf break area. However, below this depth the regions with high energy  
 200 concentration are mostly located within the centre of the canyon, not around its edge.  
 201 This finding is consistent with the observational profiles shown in Figures 3 b and c that  
 202 reveal the energy maximum at approximately 700 m depth recorded at the CTD station  
 203 in the middle of the canyon.

204 Such a focusing of internal wave energy in the canyon's centre can be explained in terms  
 205 of a superposition of several tidal beams generated at the shelf edge on the periphery of  
 206 the canyon and radiating downward toward the centre of the canyon, as it is shown in  
 207 Figure 1 c. Indeed, analysis of the bottom steepness (2) has shown that  $\alpha \ll 1$  in the  
 208 surrounding shelf area, but  $\alpha > 1$  in the central part of the canyon, see Figure 6 a. The  
 209 red zones here, situated along the shelf break, separate the areas of subcritical shelf from  
 210 the supercritical abyssal part of the canyon.

211 It is important to note that the positions of the potential generation sites of tidal beams  
 212 (as sectors A-C shown in Figure 1 c) correlate very well with the locations of the internal  
 213 body force (IBF) introduced by *Baines* [1982] for quantification of the tidal energy  
 214 conversion. In a two-dimensional  $(x, z)$  case the IBF reads:

$$215 \quad F = \rho Q z \frac{N(z)^2}{\omega} \left( \frac{1}{H(x)} \right)_x .$$

216 Here  $Q$  is water discharge produced by tides and  $\rho$  is the reference density. Being inte-  
 217 grated from the surface to the bottom this formula gives an overall efficiency of the internal  
 218 tide generation for the whole water column,  $\Phi = \int_{-H}^0 F dz$ . In a three-dimensional case

219  $(x, y, z)$  the vertically integrated IBF in every point is calculated using the discharge and  
220 the derivative  $\partial(1/H)/\partial l$ , where  $l$  in the direction of the depth gradient, as shown in for-  
221 mulae (2). The spatial distribution of the vertically integrated IBF  $\Phi$  shown in Figure 6 b  
222 was calculated for the stratification presented in Figure 1 a and the tidal forcing depicted  
223 in Figures 2.

224 Figure 6 a demonstrates that the main part of the canyon topography is supercritical  
225 for semi-diurnal internal tidal waves with the maximum IBF concentrated around the  
226 canyon rim. As a result, according to theory [Vlasenko *et al.*, 2005], the internal tide in  
227 the canyon should take a form of tidal beams generated at the shelf break around the  
228 canyon periphery which radiate downward toward the centre of the canyon, as is shown  
229 in the scheme depicted in Figure 7. Bearing in mind that the canyon head has a near-  
230 circular shape, it is expected that it can function like an optical lens focusing wave energy  
231 into its centre. Evidence for that interpretation is presented in Figure 8 a for the vertical  
232 cross-section depicted in Figure 2 b by a white line. Three tidal beams can be identified  
233 in Figure 8 a (the characteristic lines, equation (1), are shown here by thin white lines).  
234 The tidal beam  $a-b-c$  is generated at the shelf break point, at  $b$ , and propagates downward  
235 along characteristic line  $b-c$ . The internal wave beam pattern, resembling a St. Andrews  
236 cross, is generated at the saddle point  $e$ . The tidal energy propagates from this point along  
237 four characteristic lines,  $e-g$ ,  $e-d$ ,  $e-h$ , and  $e-f$ . It is interesting and relevant to note that  
238 the two tidal beams generated at the opposite sides of the canyon, viz.  $b-c$  and  $e-f$  meet  
239 in the centre of the canyon in the layer between 600 and 800 m depths which is consistent  
240 with the position of the deep-water maxima in vertical displacement, horizontal velocity  
241 and shear variance seen in Figures 3 d.

242 Observational evidence of the tidal beam can also be found in Figures 4 b where the  
243 northward horizontal velocity recorded at the yo-yo station is presented as a time series.  
244 It correlates well with Figures 9 d where similar signal, predicted by the MITgcm at  
245 the position of the yo-yo station, is presented. Consistency of both patterns is obvious,  
246 although the observational one is less contrasted and more noisy. Both panels reveal  
247 upward propagation of the wave phase (shown by dashed white lines in Figures 4 b and  
248 9 b), which can be treated in terms of the tidal beam evolution schematically presented  
249 in Figures 1 c.

250 Model predicted vertical displacements of the isotherms below 400 m depth are a bit  
251 larger than those recorded in-situ (compare Figures 4 a and 9 a), although both panels  
252 demonstrate evidence of semidiurnal periodicity (find white dashed lines) and relatively  
253 "calm" internal wave activity in the intermediate layers (between 150 and 350 m).

254 As seen from Figure 6, similar conditions of the energy focusing discussed above, i.e.  
255 position of sub-, and supercritical areas, are valid also for many other cross-sections pass-  
256 ing through the centre of the canyon. In other words, the tidal energy is converted at  
257 many particular areas around the canyon periphery, and is then radiated towards its cen-  
258 tre and accumulated there at the depths of between 500 and 800 m, as is observed in the  
259 CTD/LADCP analysis.

260 Such focusing of wave energy in the canyon's centre should increase the associated  
261 level of local water mixing there, since energy cannot accumulate indefinitely. Elevated  
262 shear variance is testament to greater elevated mixing at depth (Figure 3 d). Model  
263 results demonstrate a similar increase of vertical diapycnal mixing. Figure 10 shows  
264 that the coefficient of vertical diffusivity  $\kappa$  in the centre of the canyon calculated using

265 the Richardson number dependent parametrization [*Pacanowski and Philander, 1981*]  
266 increases from the background level  $\kappa_b=10^{-5} \text{ m}^2$  above 500 m depth up to  $10^{-2} \text{ m}^2$  in the  
267 layer from 600 m to 800 m depth where the tidal beam meet in the centre of the canyon.  
268 This dramatical increase suggests strong diapycnal mixing in the place of tidal beam  
269 focusing. As a result of intensified local diapycnal mixing, a quasi stationary density  
270 (temperature) gradient is formed across the canyon, as shown in Figure 8 b (the same  
271 cross-section as in Figure 8 a). Convergence and divergence of isotherms at the depths of  
272 between 500 m to 900 m is clearly seen here from two pairs of isotherms colored in white  
273 and magenta. Initially (before the model run) the distance between both (parallel) isolines  
274 was equal to 100 m: the upper and lower “white” isotherms were initially at depths of  
275 650 m and 750 m, respectively; the upper and lower “magenta” isotherms were initially  
276 at depths of 800 m and 900 m. After ten cycles of tidal action the distance between the  
277 two groups of isotherms was modified: at some positions they converged, and at others  
278 they diverged. Figure 11 shows the difference between isotherms initially centred at  
279 700 m (Figure 11 a) and at 850 m (Figure 11 b). It is interesting that the convergence and  
280 divergence of isolines is opposite for the two depth pairs, Figure 11 a, b.

281 The formation by diapycnal mixing of quasi-stationary horizontal pressure gradients  
282 suggests the existence of geostrophically balanced baroclinic eddies. A quasi-stationary  
283 eddy is clearly seen in Figure 12 in the velocity vector fields at depths of 450 m, 600 m and  
284 700 m. It was obtained by subtraction of the barotropic tidal flow (at each point it was  
285 found as an average value of the velocity through the whole water column from surface  
286 to the bottom) and by averaging of all fields over several tidal periods in order to exclude  
287 periodic tidal motions and to reduce noise. The eddy around the topography bank is

288 seen at a depth of 450 m where the tidal beam is still located close to the canyon flank  
 289 and the shelf edge. However, vortical motions are absent in the centre of the canyon at  
 290 this depth (450 m). According to the findings presented above, the tidal energy is mostly  
 291 concentrated beneath 450 m and in the centre of the canyon. Here a baroclinic eddy with  
 292 anti-cyclonic rotation at 600 m depth and cyclonic rotation at 700 m depth is clearly seen.

293 Note that the correct choice of the model parameters for the turbulent closure that  
 294 produces an adequate eddy pattern is not always possible without validation of the model  
 295 output against observational data. In our case we used observations by *Inall et al.* [2000]  
 296 conducted over the Malin Sea continental slope adjacent to the considered here area.  
 297 Direct in-situ measurements of the background turbulence allowed them to quantify the  
 298 vertically integrated coefficient of turbulent diffusion in the range  $(5-12) \cdot 10^{-4}$  ( $\text{m}^2\text{s}^{-1}$ ).  
 299 Integration of the vertical profile of diffusivity shown in Figures 10 produced by the model  
 300 returns the value of  $7.5 \cdot 10^{-4}$  ( $\text{m}^2\text{s}^{-1}$ ) (shown by a dashed line) which is well inside of the  
 301 in-situ observed range.

302 To find observational evidence of the predicted baroclinic eddy generated in the middle  
 303 of the canyon, the following analysis of the LADCP data was performed. First, the  
 304 barotropic velocities  $U_j^{\text{bar}}$  and  $V_j^{\text{bar}}$  were found by averaging of the instant velocity  
 305 profiles:

$$306 \quad U_j^{\text{bar}} = \frac{1}{H_j} \int_0^{H_j} u_j(z) dz;$$

$$307 \quad V_j^{\text{bar}} = \frac{1}{H_j} \int_0^{H_j} v_j(z) dz.$$



308  $j = 1, 2, 3, \dots, 14$ . Second, the barotropic tidal signal was removed from the sampling  
 309 data using the following procedure:

$$310 \quad u_j^{\text{int}} = u_j - U_j^{\text{bar}};$$

$$311 \quad v_j^{\text{int}} = v_j - V_j^{\text{bar}},$$

312 It is clear that the remaining baroclinic signals  $u_j^{\text{int}}$  and  $v_j^{\text{int}}$  contain both mean currents  
 313 and internal waves. They are also not free from a random signal that always is present  
 314 in any observational data set. In order to reduce the noise average vertical profiles of  
 315 both components were calculated as follows:  $u^{\text{int}}(z) = 1/14 \sum_{j=1}^{14} u_j^{\text{int}}(z)$  and  $v^{\text{int}}(z) =$   
 316  $1/14 \sum_{j=1}^{14} v_j^{\text{int}}(z)$ . In addition, vertical averaging of  $u^{\text{int}}(z)$  and  $v^{\text{int}}(z)$  in 100 m thick  
 317 layers with centres at depths 600 m and 750 m was conducted. The resulting vectors  
 318 along with the model predicted vectors are presented in Figures 13 a and 13 b. Looking  
 319 at the averaged velocity vectors one can discern a good consistency. It seems that the  
 320 direction of the black arrows in Figures 13 a and 13 b is consistent with that predicted by  
 321 the model, i.e. the anti-cyclonic and cyclonic rotation in depths 600 m and 750 m layers,  
 322 respectively. This can be considered as further evidence that tidal energy focusing in the  
 323 centre of the canyon is responsible for the formation of baroclinic eddies there.

#### 4. Two-dimensional versus three-dimensional focusing

324 The importance of three-dimensional effects in focusing of baroclinic tidal energy in the  
 325 centre of a circular canyon (as shown in Figure 7) can be demonstrated and quantified  
 326 more accurately by comparing three-dimensional (3D) and two-dimensional (2D) simpli-  
 327 fied cases. In doing so a series of numerical experiments was conducted for two idealized  
 328 canyons, shown in Figures 14. The shape of both canyons in the  $x$ -direction was the same,

329 i.e.  $H(x) = H_0 + H_m \cos^2(\pi x/2l)$ , although in a 2D case the canyon was indefinitely long  
 330 in the  $y$ -direction to exclude influence of the  $y$ -derivatives, but its 3D counterpart has  
 331 circular form, see Figures 14 b. The parameters of the bottom topography were taken  
 332 realistic, i.e.  $H_0 = 200$  m,  $H_m = 800$  m,  $l = 7.5$  km. The buoyancy frequency in this series  
 333 of experiments was taken according to profile shown in Figures 1 a, but the intensity of  
 334 the tidal forcing was taken ten times weaker than in Section 3 in order to reduce the  
 335 influence of non-linear terms and to show the effect of tidal energy focusing more clearly.

336 Two systems of tidal beams are generated at either sides of both canyons. In Figure  
 337 14 they are presented as a series of instant horizontal velocity fields both for 2D (upper  
 338 panels) and 3D (bottom panels) canyons. Tidal energy is radiated upward and downward  
 339 from the shelf edges propagating in narrow bands along characteristic lines (1). Downward  
 340 propagating tidal beams meet in the centre of the canyons at the depth of about 550m  
 341 and amplify each other creating a spots of high tidal activity. Propagating further down-  
 342 ward the tidal beams experience multiple reflections from the canyon flanks and produce  
 343 amplification of tidal energy in the canyon centres. As a result of narrowing cross-section  
 344 of canyons down to the deep the intensity of tidal motion increases. This effect is clearly  
 345 seen in all six panels presented in Figure 15.

346 The comparison analysis of the top and bottom panels shows that the structure of the  
 347 wave fields in two- and three-dimensional canyons is a bit different. Note also quite a  
 348 different level of intensity of baroclinic motions in both canyons. They look weaker near  
 349 the surface in the 3D case, but stronger in its deep part.

350 In order to quantify the difference more accurately the amplitude of the baroclinic tidal  
 351 energy  $E = 0.5\rho(u^2 + v^2 + w^2 + N^2\xi^2)$  was calculated (here  $u$ ,  $v$ , and  $w$  are the components

of the velocity vector,  $N$  is the buoyancy frequency,  $\xi$  is isopycnal displacement, and  $\rho$  is  
the water density). The amplitude values of energy density are presented in Figure 16  
both for 2D and 3D canyons. Four control points with maximum of local tidal energy have  
been chosen for the comparative analysis: the point **a** near the shelf break where tidal  
beams are generated, points **b** and **c** in the deep part where the beams intersect, and point  
**d** with the highest level of concentration of baroclinic tidal energy (near the bottom). The  
appropriate values of the energy density were 24, 31, 32 and 43  $Jm^{-3}$  for the 2D canyon,  
and 16, 36, 61 and 50  $Jm^{-3}$  for the 3D one. From this series of experiments it is clear that  
a 2D canyon is more efficient in terms of the tidal energy conversion demonstrating 1.5  
larger value of baroclinic tidal energy at the shelf break. At the same time the intensity of  
the baroclinic tidal motions in the deep is much higher in the three-dimensional case. In  
fact, the 3D canyon works as an optical lens focusing the tidal energy from all directions  
into a focal centre. Even though the tidal beams generated at the shelf break of the 3D  
canyon are 1.5 times weaker than in a 2D case, the 3D focusing provides two times larger  
density of baroclinic tidal energy in the deep part of the canyon than its 2D counterpart.

The increase of baroclinic tidal energy in the deep part of the 3D canyon compared to  
that in surface layers results in larger level of diapycnal mixing and formation of stronger  
geostrophically adjusted eddies near the bottom. This fact is clearly seen in Figure 17  
where vectors of the velocity fields averaged over one tidal cycle at depths 800 m, and  
900 m are presented. These two sections coincide with the positions of the maximum  
of tidal energy concentration depicted in Figure 16 b by letters **c** and **d**. Both patterns  
demonstrate evidence of an anticyclonic eddy with the velocities at the level of  $1\text{ cm s}^{-1}$ .  
Note, however, that the currents near the bottom at the depth of 900 m on average are

375 two time stronger than that at 800 m depth. This is consistent with Figure 16 b which  
376 demonstrates higher energy concentration near the bottom.

## 5. Discussion and conclusions

377 Submarine canyons are common bathymetric features at many of the world's shelf edge  
378 regions. They can trap internal wave energy holding it towards the head of canyons in  
379 a converging wave-guide that can lead to a high level of turbulent mixing there. This  
380 mechanism was discussed by *Baines* [1983]; *Gardner* [1989]; *Gordon and Marshall* [1976];  
381 *Hotchkiss and Wunsch* [1982]. Note, however, that the aforementioned papers appeal  
382 predominantly to a two-dimensional concept of this mechanism. In reality one should  
383 operate with three-dimensional characteristic surfaces of a 3D wave equation rather than  
384 with the characteristic lines of its 2D counterpart. As a result of the three-dimensionality,  
385 the tidal beams emanating from flanks of a concave topography can focus in its centre  
386 producing a spot with high levels of internal wave energy and mixing. This paper deals  
387 with the three-dimensional aspects of this focusing mechanism.

388 The possibility of intensification of baroclinic tidal energy due to wave interference  
389 was recently reported in a number of theoretical papers. *Carter* [2010] analysing model  
390 output for baroclinic tide in the Monterrey Bay region testified interference of internal  
391 waves generated at different sectors of the bay. It was found that the model predicted up  
392 to 5 times increase of the baroclinic tidal flux close to the Monterrey canyon axis located  
393 in the centre of the bay. It was hypothesized there that the effect was created thanks  
394 to topographic focusing, although this fact was not clarified, specifically in terms of the  
395 beam-like structure of baroclinic tides or supercriticality of the bottom topography.

396 Similar effects of the interference of baroclinic tidal energy radiated from scattered  
397 multiple sources were also reported by *Rainville et al.* [2010] for the Hawaiian Ridge area.  
398 As distinct from the present study, mostly horizontal interference of the baroclinic tidal  
399 waves was studied, however the effect of baroclinic tidal energy superposition was clearly  
400 demonstrated.

401 The most recent model study conducted by *Zhang et al.* [2014] for an idealized and su-  
402 percritical for semi-diurnal M2 tide canyon confirmed an asymmetry of internal tide near  
403 the canyon, which presumably is a consequence of the along-shore effects of propagating  
404 internal tidal wave. In fact, they focus mostly on resonant effects of internal tide gener-  
405 ation in the canyon and its beam-like onshore propagation alongside explanation of the  
406 reasons for the asymmetry of tidal fields. With relevance to the present study, *Zhang et al.*  
407 [2014] reported the beam-like structure of baroclinic tides near the supercritical canyon,  
408 with a difference in deepward and shoreward structures, although their interference in the  
409 canyon area was not demonstrated.

410 Model output always allows us to study the process of wave focusing in detail, however  
411 in reality it is quite difficult to observe this effect in-situ. For the Celtic Sea we have  
412 found not only theoretical but also observational evidence of the baroclinic tidal energy  
413 focusing in the middle of the canyon. The measurements were conducted during D376 in  
414 the centre of the Petite Sole Canyon situated at the shelf edge of the Celtic sea (see Figures  
415 2). CTD and LADCP data collected at a station in the middle of the canyon revealed  
416 large vertical oscillations of isotherms (up to 45 m, Figures 2b) and local maximum of  
417 horizontal currents (up to  $0.12 \text{ ms}^{-1}$ , Figures 3c) in the layer between 500 and 800 m.

418 The possibility of such an inherently three dimensional focusing mechanism of baroclinic  
419 tidal energy was confirmed in a series of numerical experiments conducted using the  
420 MITgcm forced by M2 tidal harmonic. The high internal tidal activity in the deep part  
421 of the canyon (Figures 5) is treated here in terms of downward propagation and focusing  
422 of internal tidal beam generated at the shelf break. The specific circular shape of the  
423 canyon, coupled with the steep bottom topography below the shelf break (steeper than  
424 tidal beam) in all parts of the canyon create favourable conditions for the tidal energy  
425 focusing in the canyon's centre, see Figure 7. Both observations and MITgcm simulations  
426 have also shown that the tidal energy focusing intensifies local diapycnal mixing (Figures  
427 3d and 10), that can lead to formation of a baroclinic eddy below 450m depth in the  
428 central part of the canyon, see Figures 12. Evidence consistent with the presence of the  
429 cyclonic and anti-cyclonic rotation in the canyon centre was found also in in-situ data.

430 The importance of the results is that the effect of the focussing of baroclinic tidal energy  
431 in 3D configuration is quite a typical situation in many areas. This means that the results  
432 on the baroclinic tidal energy focusing reported here can have much wider application  
433 rather than just circular-shape canyons.

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436 <http://www.sams.ac.uk/fastnet>.

## References

437 Baines, P.G. (1982,) On internal tide generation models, *Deep Sea Res. I*, 29, 307–339.

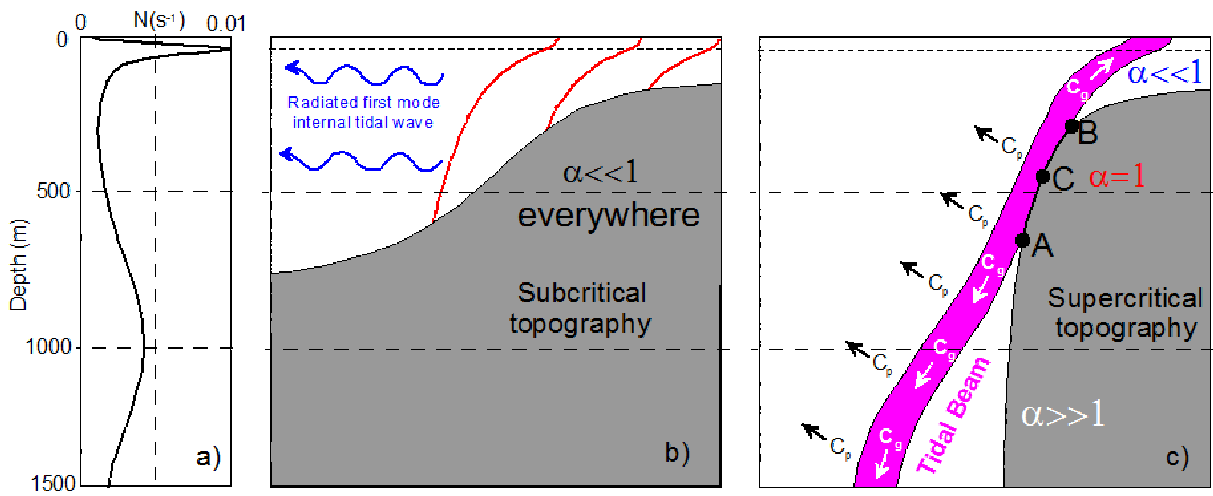
- 438 Baines, P.G. (1983), Tidal motion in submarine canyons a laboratory experiment, *J. Phys.*  
439 *Oceanogr.*, *13*, 310–328.
- 440 Balmforth, N.J. and T. Peacock (1983), Tidal conversion by supercritical topography, *J.*  
441 *Phys. Oceanogr.*, *13*, 310–328.
- 442 Carter, G.S. (2010), Barotropic and baroclinic  $M_2$  tides in Monterey Bay region, *J. Phys.*  
443 *Oceanogr.*, *39*, 1965–1974.
- 444 Egbert, G.D., and S.Y. Erofeeva (2002) Efficient inverse modeling of barotropic ocean  
445 tides, *J. Atmos. Oceanic Technol.*, *19*(2), 183–204.
- 446 Gardner, W.D. (1989), Periodic resuspension in Baltimore Canyon by focusing of internal  
447 waves, *J. Geophys. Res.*, *94*(C12), 18,185–18,194.
- 448 Gerkema, T., F.A. Lam, and L.R. Maas (2004), Internal tides in the Bay of Biscay:  
449 conversion rates and seasonal effects *Deep-Sea Res. II*, *51*, 2995–3008.
- 450 Gerkema, T., and H. van Haren (2012), Absence of internal tidal beams due to non-  
451 uniform stratification. *J. of Sea Res.*, *74*, 2–7.
- 452 Gordon, R.L., and N.F. Marshall (1976), Submarine canyons: internal wave traps? *Geoph.*  
453 *Res. Let.*, *3*(10), 622–624.
- 454 Grimshaw, R.H.J., P.G. Baines, and R.C. Bell (1985), The reflection and diffraction of  
455 internal waves from the junction of a slit and a half-space, with application to submarine  
456 canyons. *Dyn. Atmos. and Oceans*, *9*, 85–120.
- 457 Grimshaw R., E. Pelinovsky, and T. Talipova (2010), Non-reflecting internal wave beam  
458 propagation in the deep ocean *J. of Phys Oceanogr.*, *40*, 802-813.
- 459 Hall, R.A., and G.S. Carter (2011), Internal tides in Monterrey Submarine Canyon, *J.*  
460 *Phys. Oceanogr.*, *41*, 186–204.

- 461 Hickey, B.M. (1995), Coastal submarine canyons, *Topographic Effects in the Ocean: 'Aha*  
462 *Huliko'a Hawaiian Winter Workshop*, Honolulu, HI, University of Hawaii at Manoa,  
463 95–110.
- 464 Hotchkiss, F.S., and C. Wunsch (1982), Internal waves in Hudson Canyon with possible  
465 geological implications, *Deep-Sea Res.*, *29(4A)*, 415–442.
- 466 Inall, M.E., T.P. Rippeth, and T.J.Sherwin (2000), Impact of nonlinear waves on the  
467 dissipation of internal tidal energy at a shelf break, *J. Geophys. Res.*, *115(4)*, 8687–  
468 8705, doi:10.1002/2013JC009708.
- 469 Jachec, S.M., O.B. Fringer, M.G. Gerritsen, and R.L. Street (2006), Numerical simulation  
470 of internal tides and the resulting energetics within Monterey Bay and the surrounding  
471 area, *Geoph. Res. Let.*, *33*, doi:10.1029/2006GL026314.
- 472 Kunze, E., L.K. Rosenfeld, G.S. Carter, and M.C.Gregg (2002), Internal waves in Mon-  
473 terrey Submarine Canyon, *J. Phys. Oceanogr.*, *32*, 1890–1913.
- 474 Pacanowski, R.C., and S.G.H. Philander (1981) Parameterisation of vertical mixing in  
475 numerical models of Tropical Oceans, *J. Phys. Oceanogr.*, *11*, 1443-1451.
- 476 Petrucio, E.T., L.K. Rosenfeld, and J.D. Paduan (1998), Observations of the internal  
477 tide in Monterey Canyon, *J. Phys. Oceanogr.*, *28*, 1873–1903.
- 478 Petrucio, E.T., J.D. Paduan and L.K. Rosenfeld (2002), Numerical simulation of the  
479 internal tide in a submarine canyon, *Ocean Modell.*, *4*, 221–248.
- 480 Polzin, K., E. Kunze, J. Hummon, and E. Firing (2002) The finescale response of lowered  
481 ADCP velocity profiles, *J. Atmos. and Oceanic Technol.*, *19*, 205–224.
- 482 Rainville, L., T.M.S. Jonston, G.S. Carter, M.A. Merifield, R.Pincel, P.F. Worcester, and  
483 B.D. Dushaw (2010), Interference pattern and propagation of the  $M_2$  internal tide south

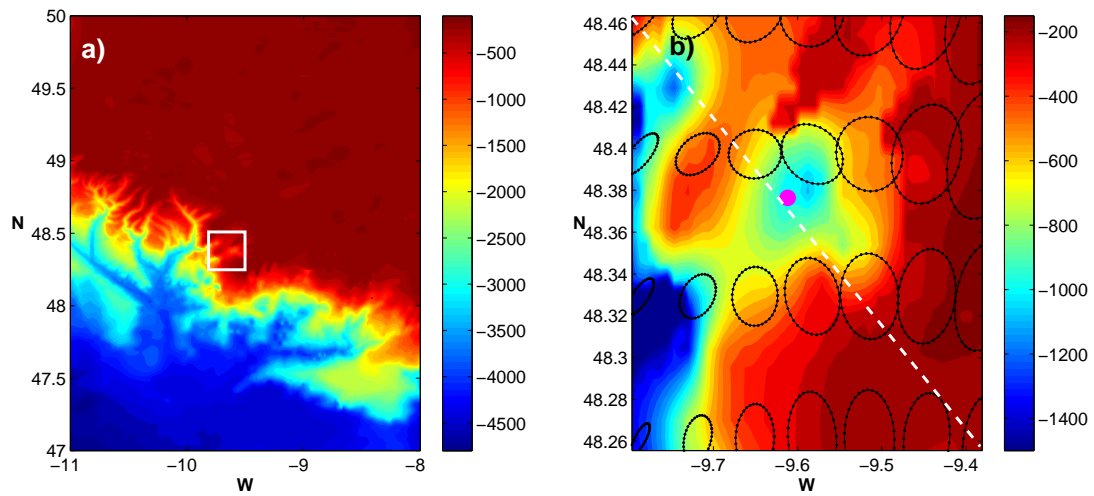


- 484 of the Hawaiian Ridge, *J. Phys. Oceanogr.*, *40*, 311–325.
- 485 Shepard, F. P. (1974), “Internal waves” advancing along submarine canyons, *Science*, *19*,
- 486 195–198.
- 487 St. Laurent, L., S. Srtringer, C. Garrett, and D. Perrault-Joncas (2003), The generation
- 488 of internal tides at abrupt topography, *Deep-Sea Res. I*, *50*, 987–1003.
- 489 Thurnher, A.M., L.C. St.Laurent, K.G. Speer, J.M. Toole and J.R. Ledwell (2005), Mixing
- 490 associated with sills in a canyon on the Midocean Ridge flank, *J. Phys. Oceanogr.*, *35*,
- 491 1370–1381.
- 492 Vlasenko V., and N. Stashchuk (2015) Internal tides near the Celtic Sea shelf
- 493 break: A new look at a well known problem *Deep-Sea Res.*, *103*, 24-36,
- 494 <http://dx.doi.org:10.1016/j.dsr.2015.05.003>.
- 495 Vlasenko V., N. Stashchuk, and K. Hutter (2005), Baroclinic tides: theoretical modeling
- 496 and observational evidence. Cambridge University Press. 365 pp.
- 497 Vlasenko V., N. Stashchuk, M.Inall, and J.Hopkins (2014), Tidal energy conversion in a
- 498 global hot spot: on the 3D dynamics of baroclinic tides at the Celtic Sea shelf break,
- 499 *J. Geophys. Res.*, *119*(6), 3249–3265, doi:10.1002/2013JC009708.
- 500 Wain, D.J., M.C. Gregg, M.H. Alford, R.-C. Lien, R.A. Hall, and G.S. Carter (2013),
- 501 Propagation and dissipation of the internal tide in upper Monterrey Canyon, *J. Geophys.*
- 502 *Res.*, *18*, 4855–4877, doi:10.1002/jgrc.20368.
- 503 Zhang,W.Z., T.F. Duda, and I.A. Udovychenkov (2014) Modeling and analysis of internal-
- 504 tide generation and beamlike onshore propagation in the vicinity of shelfbreak canyons,
- 505 *J. Phys. Oceanogr.*, *44*, 834–849.

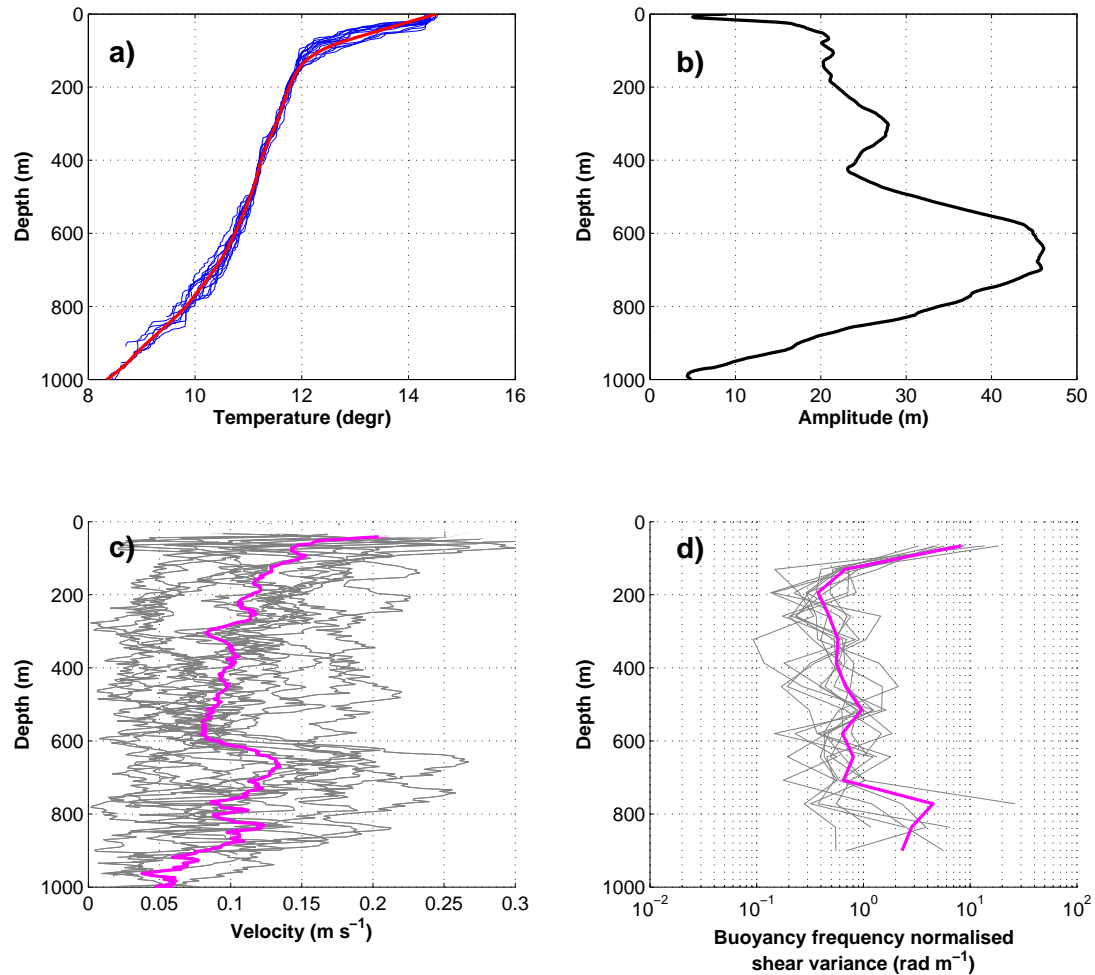
506 Zhao,Z., M.H. Alford, R.C. Lien, and M. C. Gregg (2012) Internal tides and mixing in  
507 submarine canyon with time-varying stratification, *J. Phys. Oceanogr.*, *42*, 2121–2142.



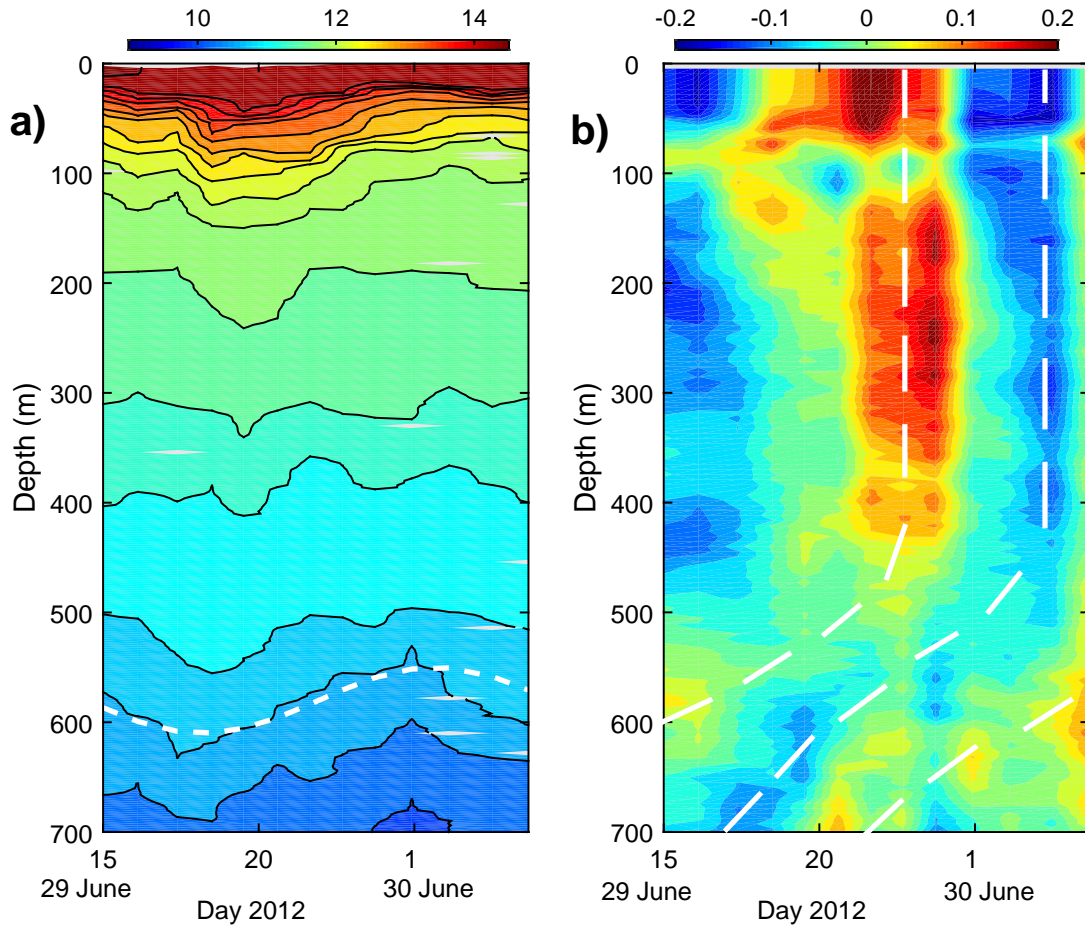
**Figure 1.** a) Buoyancy frequency measured in the Celtic Sea. b)-c) Schemes of the generation regime over subcritical (b) and supercritical (c) topographies. Red lines in panel (b) show characteristics (1). Magenta area in panel (c) depicts a tidal beam.



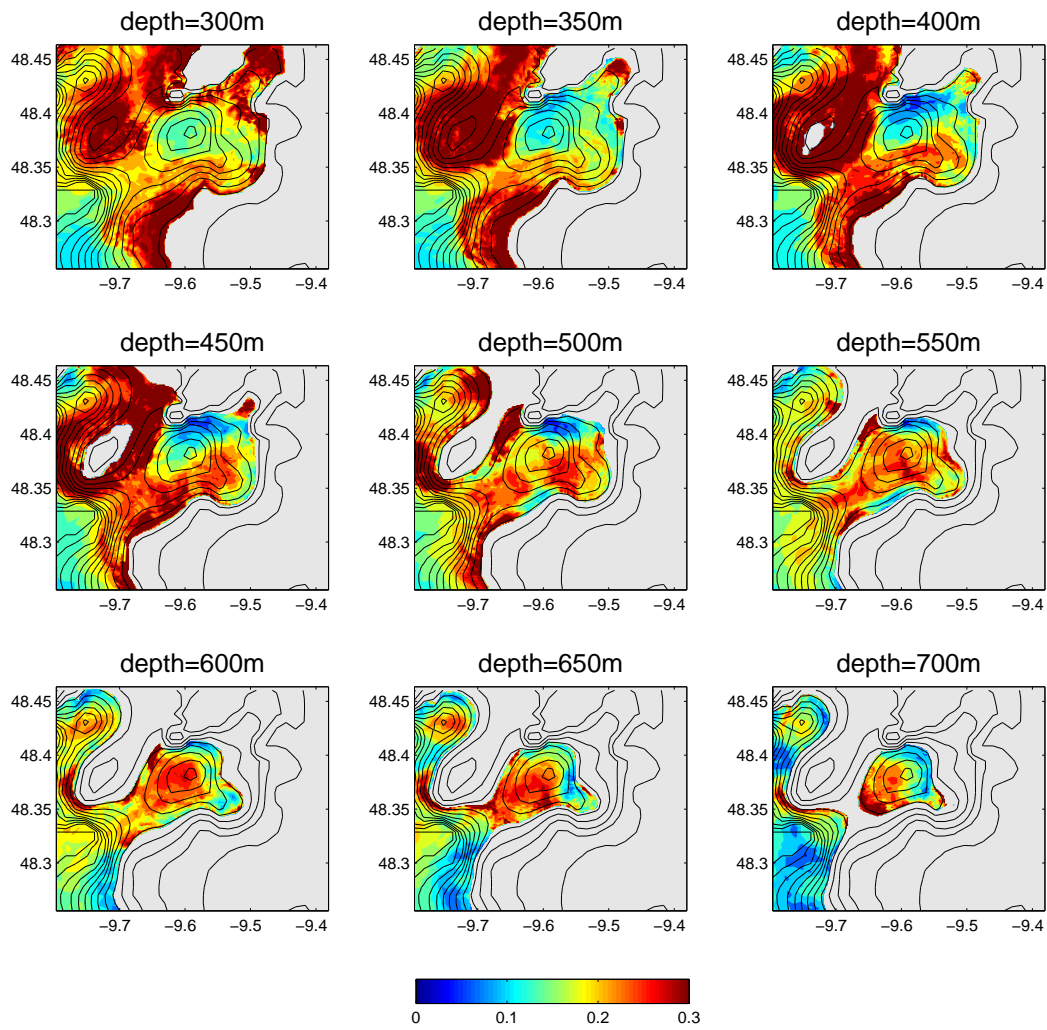
**Figure 2.** a) Bathymetry of the Celtic Sea. White rectangle shows the position of the canyon. b) Zoom of the Petite Sole Canyon area. The position of the yo-yo CTD station is depicted by a magenta dot. Tidal ellipses showing the intensity of the model forcing are presented by black contours.



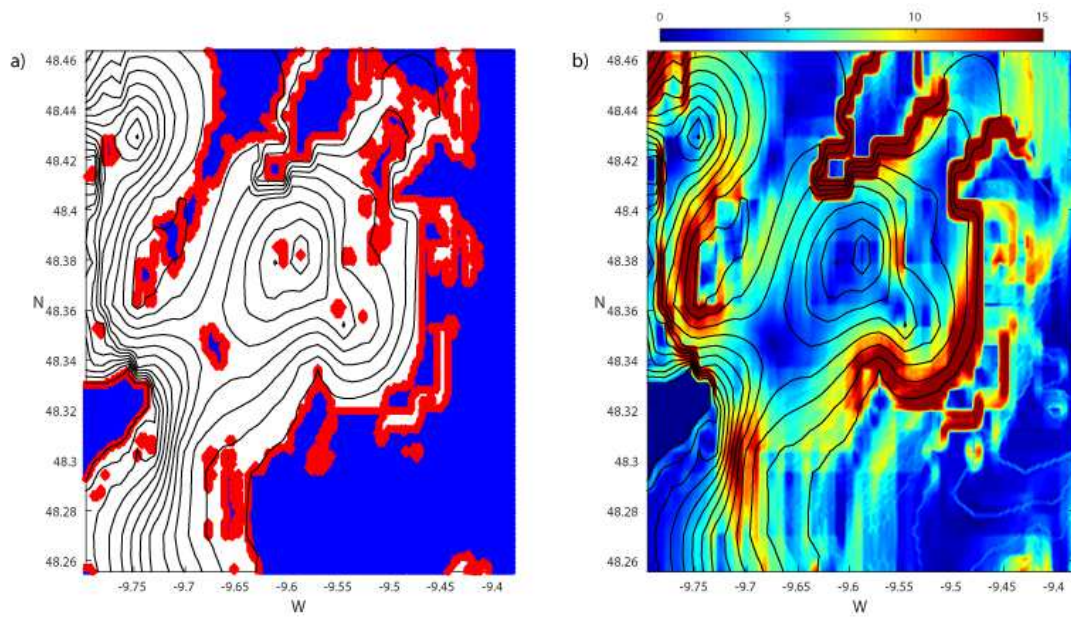
**Figure 3.** a) Temperature profiles recorded at 14 “yo-yo” CTD stations (blue lines) along with the average temperature profile (red line). b) Depth dependent amplitude of isotherms deviation. c) Horizontal velocity profiles recorded by LADCP (grey lines) and their mean profile (magenta line). d) Distribution of shear variance. All profiles are shown in grey and the magenta line depicts their average.



**Figure 4.** a) Temperature ( $^{\circ}\text{C}$ ) and b) eastward horizontal velocity ( $\text{ms}^{-1}$ ) recorded at the "yo-yo" CTD station and presented as time series. Dashed lines in panels (b) show upward propagation of the tidal beam phase, as it is schematically shown in Figure 1 c.

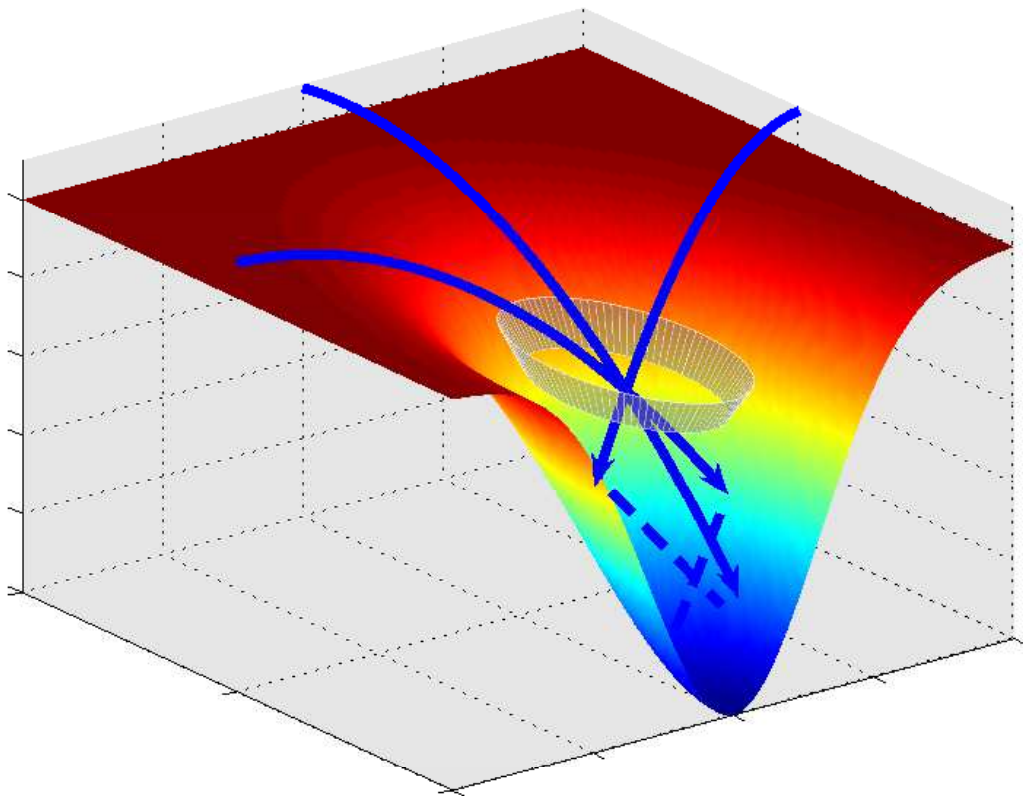


**Figure 5.** Horizontal distribution of the amplitudes of horizontal velocity at different depths.

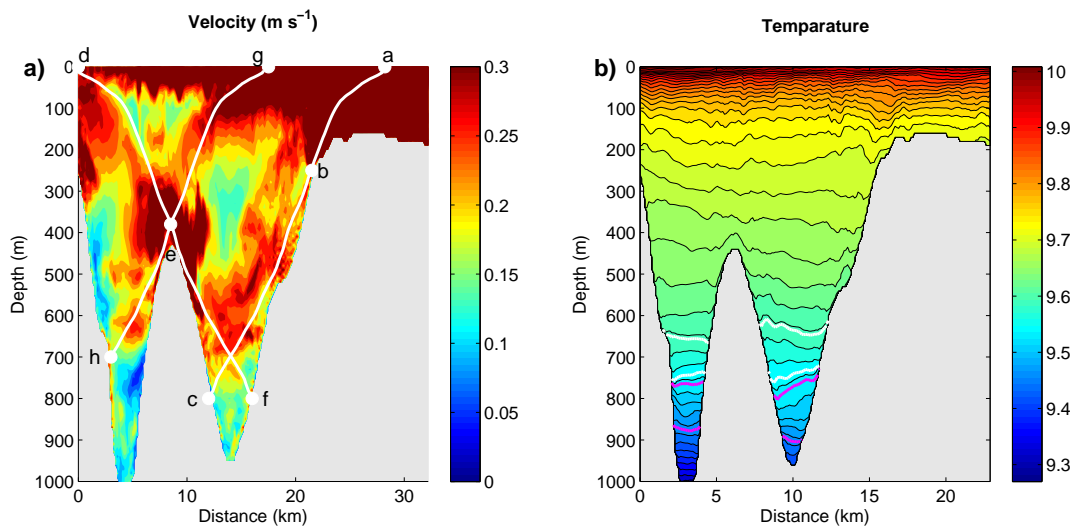


**Figure 6.** a) Spatial distribution of parameter  $\alpha$ : blue areas  $\alpha < 1$ , red stripes  $\alpha = 1$ , and clear areas with  $\alpha > 1$ . b) Positions of vertically integrated internal body force  $\Phi$  ( $\text{m}^2\text{s}^{-2}$ ).

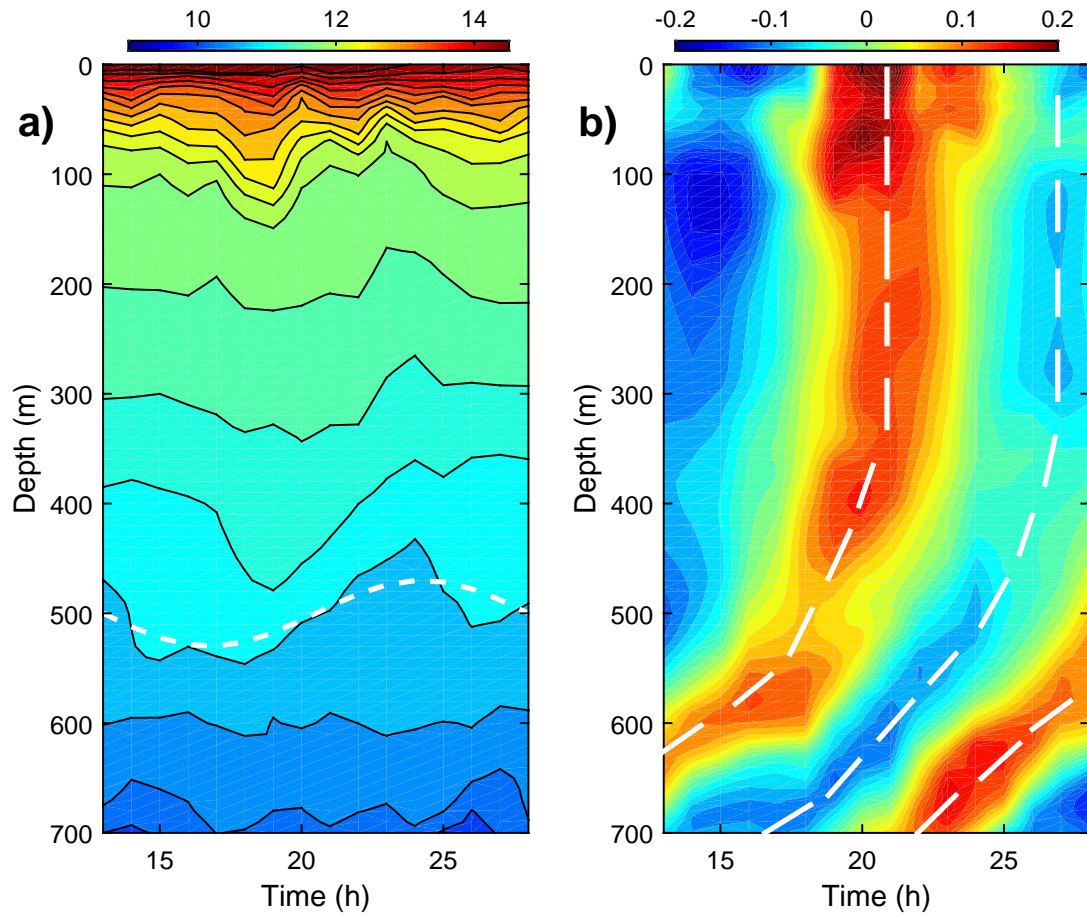




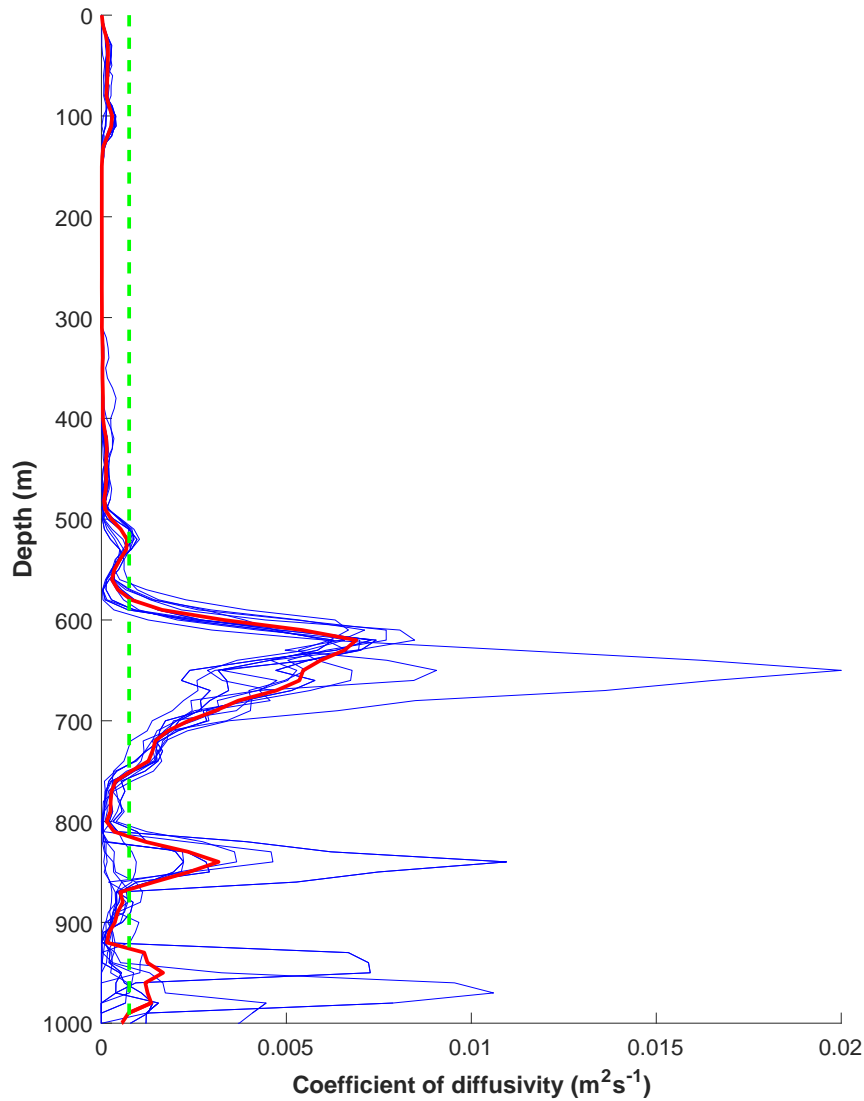
**Figure 7.** Scheme of interference of tidal beams in the centre of an idealized canyon with formation of a baroclinic eddy.



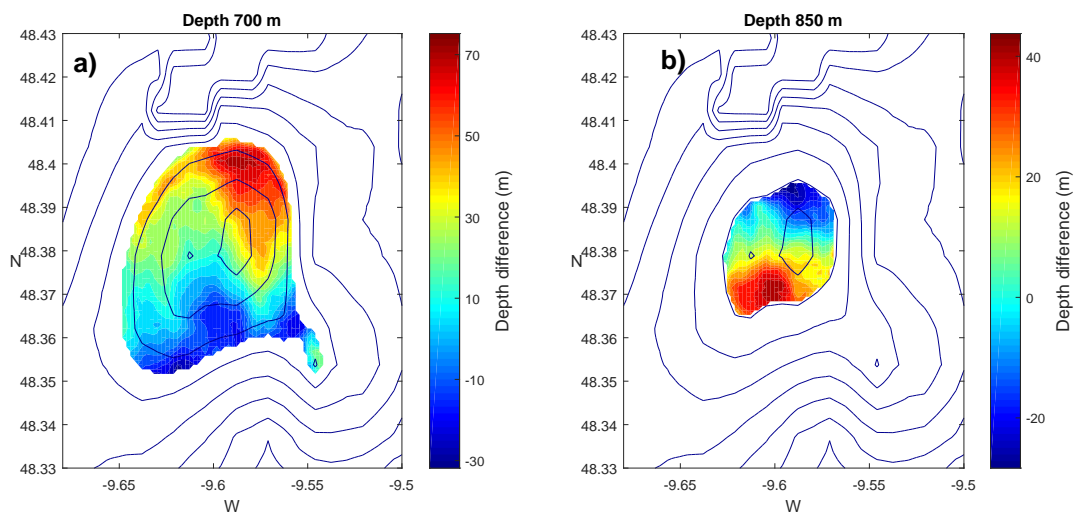
**Figure 8.** a) The largest values of horizontal velocity calculated for the cross-section shown in Figure 2 a. White contours depict characteristic lines (1). b) Average temperature distribution with upper white contour representing isotherm  $10.55^{\circ}\text{C}$  and the lower one  $10.1^{\circ}\text{C}$ . Similar values for two magenta isolines are  $10.0^{\circ}\text{C}$  and  $9.3^{\circ}\text{C}$ .



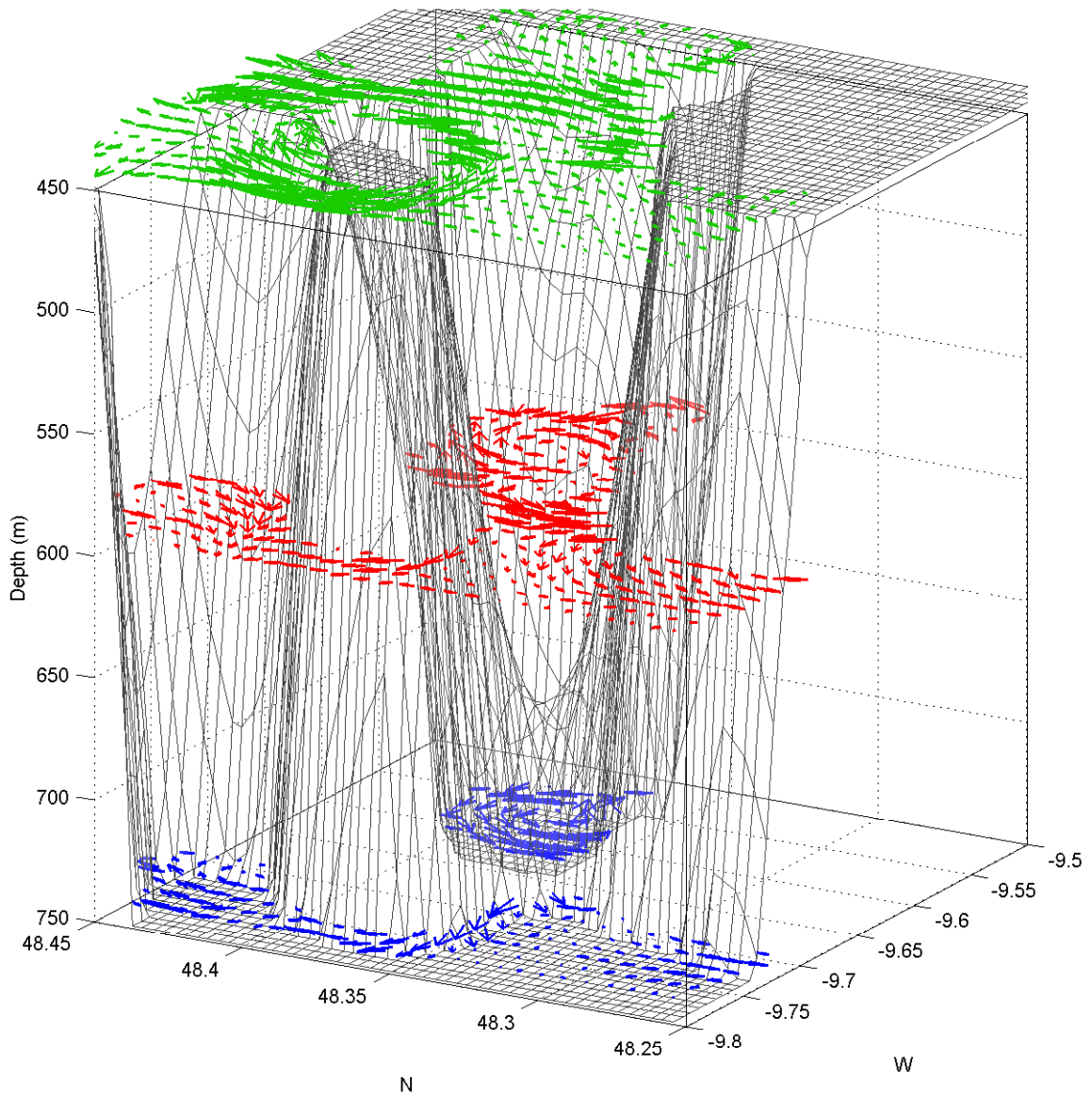
**Figure 9.** Model predicted a) temperature ( $^{\circ}\text{C}$ ) and b) eastward horizontal velocity ( $\text{ms}^{-1}$ ) for the position of the "yo-yo" CTD station presented as time series. Dashed lines in the velocity panels (b) show upward propagation of the tidal beam phase, as it is schematically shown in Figure 1 c.



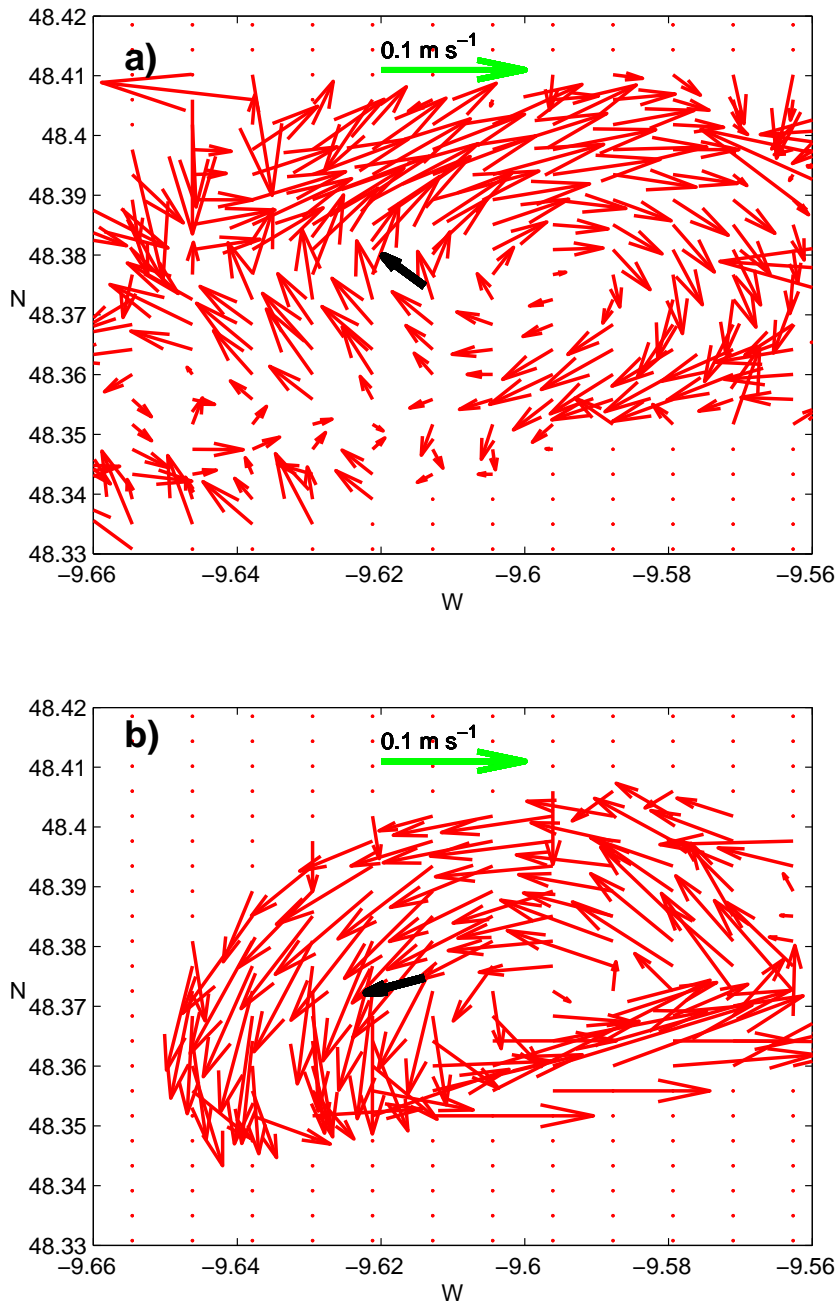
**Figure 10.** The coefficient of vertical diffusivity  $\kappa$  at the position of the yo-yo station calculated using the *Pacanowski and Philander* [1981] parametrization. Blue lines show instant profiles for the moments of yo-yo sampling, and the red line depicts an average profiles. Dashed line depicts the depth-averaged value,  $7.5 \cdot 10^{-4}$  ( $\text{m}^2\text{s}^{-1}$ ).



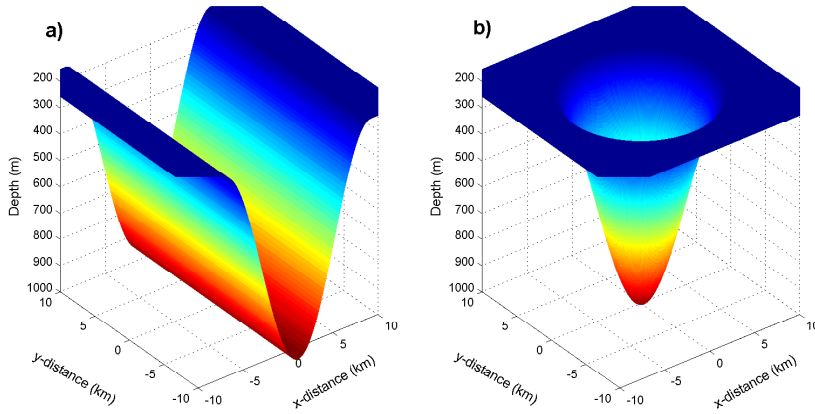
**Figure 11.** a) The distance between isotherms shown in Figure 8 b in white. b) The distance between isotherms shown in Figure 8 b in magenta.



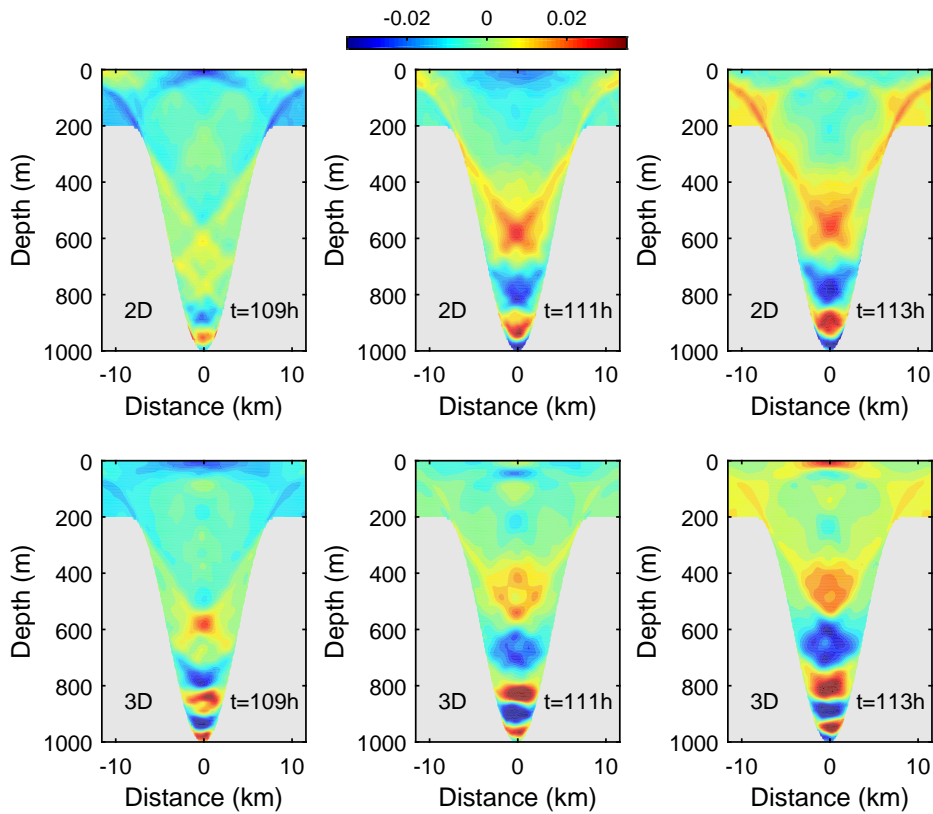
**Figure 12.** Velocity vectors of the model predicted baroclinic eddy generated inside the canyon. The currents are shown at depths 450 m (green arrows), 600 m (red arrows) and 700 m (blue arrows).



**Figure 13.** Model predicted anti-cyclonic and cyclonic eddies at the depths of 600 m (a) and 750 m (b). Average baroclinic currents recorded by LADCP at the same depths are shown by black arrows.

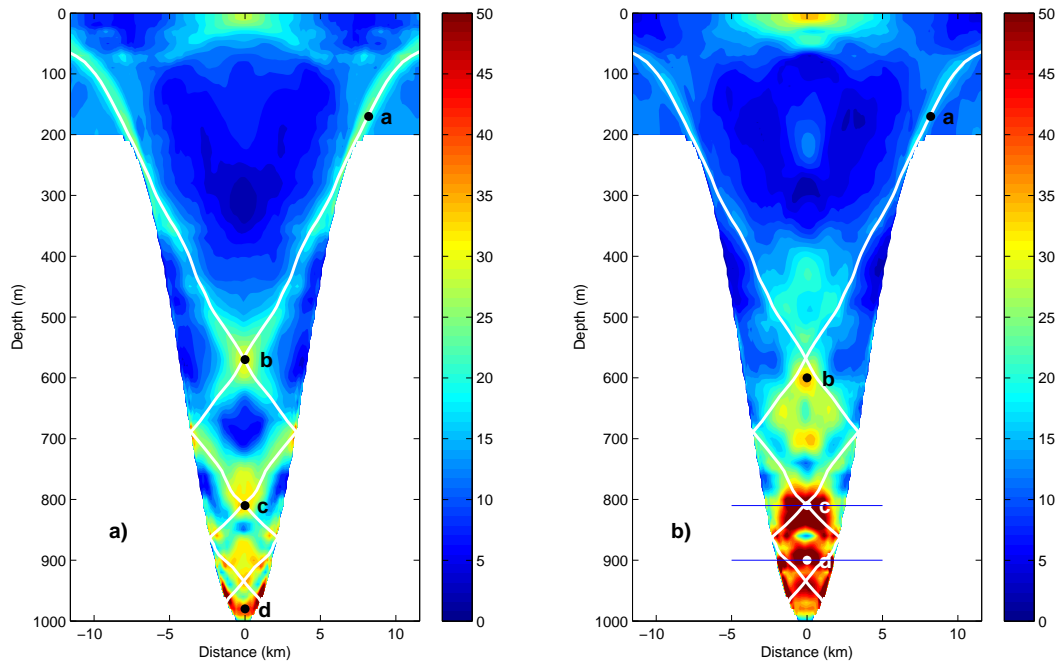


**Figure 14.** Bottom topography of idealised a) two-dimensional and b) three-dimensional canyons.

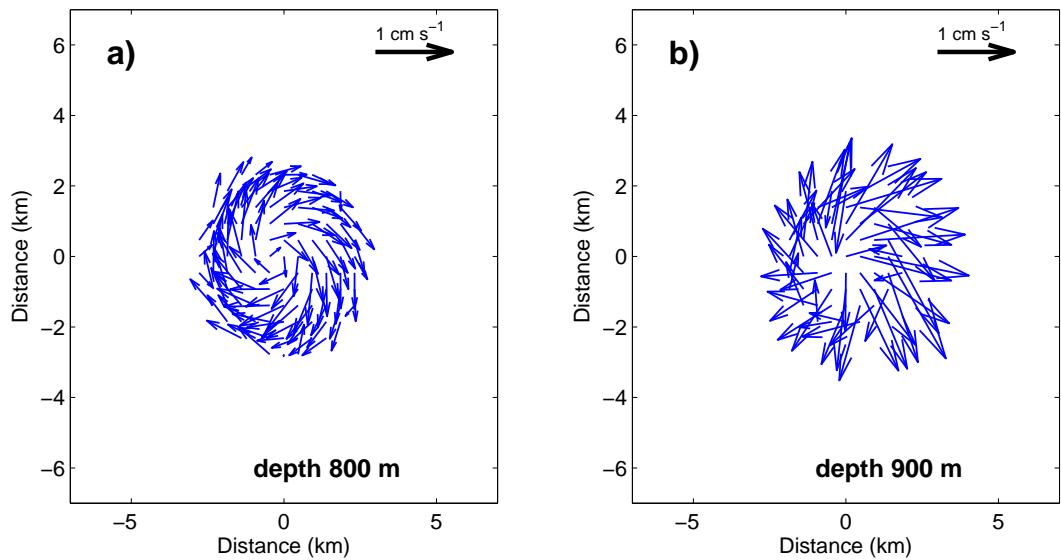


**Figure 15.** Instant horizontal velocities in the two-dimensional (top panels) and three-dimensional (bottom panels) canyons. The time after beginning of the numerical experiment is shown in the bottom right corner of each panel.





**Figure 16.** Amplitude of the baroclinic tidal energy ( $\text{Jm}^{-3}$ ) in a) two-dimensional and b) three-dimensional canyons. Dotted white contours depict characteristic lines (1).



**Figure 17.** Model predicted anti-cyclonic eddies in an ideal circular canyon.