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Finescale parameterizations of energy dissipation in a region of strong internal tides and sheared flow, the Lucky-Strike segment of the Mid-Atlantic Ridge

Simon Pasquet, Pascale Bouruet-Aubertot*, Gilles Reverdin*

Sorbonne Universités- UPMC Univ. Paris 06- LOCEAN, France

Andreas Turnherr*

Lamont-Doherty Earth Observatory, Palisades, New York, USA

Lou St Laurent*

WHOI, Woods Hole, USA

Abstract

The relevance of finescale parameterizations of dissipation rate of turbulent kinetic energy is addressed using finescale and microstructure measurements collected in the Lucky Strike segment of the Mid-Atlantic Ridge (MAR). There, high amplitude internal tides and a strongly sheared mean flow sustain a high level of dissipation rate and turbulent mixing. Two sets of parameterizations are considered: the first ones (Gregg, 1989; Kunze et al, 2006) were derived to estimate dissipation rate of turbulent kinetic energy induced by internal wave breaking, while the second one aimed to estimate dissipation induced by shear instability of a strongly sheared mean flow and is a function of the Richardson number (Kunze et al, 1990; Polzin 1996).

The latter parameterization has low skill in reproducing the observed dissipation rate when shear unstable events are resolved presumably because there is no scale separation between the duration of unstable events and the inverse growth rate of unstable billows. Instead GM based parameterizations were

*Corresponding author

Email address: `pascale.bouruet-aubertot@upmc.fr` (Pascale Bouruet-Aubertot)

found to be relevant although slight biases were observed. Part of these biases result from the small value of the upper vertical wavenumber integration limit in the computation of shear variance in Kunze et al (2006) parameterization that does not take into account internal wave signal of high vertical wavenumbers. We showed that significant improvement is obtained when the upper integration limit is set using a signal to noise ratio criterion and that the spatial structure of dissipation rates is reproduced with this parameterization.

Keywords: internal tides, turbulence, mixing, parameterization, mid-atlantic ridge

1. Introduction

Measurements of oceanic microstructure in areas of strong bathymetric variations often reveal enhanced turbulence. This includes sites such as canyons (Lueck and Osborn, 1985; St Laurent et al., 2001; Carter and Gregg, 2002),
 5 seamounts (Nabatov and Ozmidov, 1988; Kunze and Toole, 1997; Lueck and Mudge, 1997) and hydraulically controlled passages between basins (Roemmich et al., 1996; Polzin et al., 1996; Ferron et al., 1998; St Laurent and Thurnherr, 2007). This intensification of turbulent mixing leads to values of eddy diffusivity on the order of 10^{-3} - 10^{-2} m^2s^{-1} that are 2-3 orders of magnitude larger
 10 than open ocean values. In most cases, both the breaking of internal waves and the instability of the low-frequency flow contribute to this intensification. Regions of strong variations of the bathymetry are indeed preferential sites for wave breaking. This breaking can be initiated by different processes, the most widely encountered in the scientific literature being wave intensification near
 15 critical reflection, the increase in nonlinearities associated with the generation of harmonics when waves reflect at the bottom (e.g. Gerkema, 2006; Tabaei et al., 2005) and the generation of waves of small vertical scales that break locally (Munk, 1981).

Secondly local mixing can be strongly enhanced by the onset of shear in-
 20 stability of the low-frequency motion passing over rough topographic structures

(e.g. Stober et al, 2008). Typically this instability occurs as a result of flow acceleration through narrow passages. Other processes such as Lee wave generation and hydraulic control can ultimately lead to mixing.

Sills are typical regions where both processes are encountered. One of the most well-known example is the strait of Gibraltar as reported for instance by Farmer and Armi (1988) and Wesson and Gregg (1994). Also fractures zones on oceanic ridges are regions of high mixing, as they consist of small passages linking deep basins (Ferron et al., 1998; MacKinnon et al., 2008; Morozov et al., 2008).

Further measurements are required to improve our knowledge of the impact at a global scale of regions of rough topography on ocean mixing. This implies a quantification of mixing but also process studies of the link between dynamical processes leading to turbulence and turbulent mixing. The most accurate quantification of mixing is obtained from microstructure measurements of shear or temperature at millimeter scale (Gregg, 1991). However as these measurements are quite rare, quantification of turbulent mixing largely relies upon finescale parameterization based on CTD (Conductivity Temperature Depth)/LADCP (Lowered Acoustic Doppler Current Profiler) data that are much more frequent (e.g. Polzin et al, 2014).

Most finescale parameterizations used in the ocean interior first aimed to estimate mixing induced by convective instability and shear instability, either from the mean flow or from the internal wave field. Parameterizations based only on the mixing generated by the breaking of internal waves, have been widely applied over the last two decades. The dissipation rate of turbulent kinetic energy is then assumed to be equal to the rate of energy transfers within a weakly non linear internal wave field. In this case energy is transferred through resonant interactions between wave triads and the rate of energy transfers scales like the square of energy. These parameterizations, tested against microstructure measurements, were analyzed in light of the dynamics and progressively refined

(see Polzin et al, 2014 for further details).

The first formulation of finescale parameterization based on shear instability resulting from wave-wave interactions and validation against microstructure
 55 measurements was performed by Gregg (1989). Using shear variance computed by differentiation over 10m, a consistency within a factor of 2 was obtained using four mid-latitude data sets. Later on, some studies revealed the inadequacy of Gregg (1989) to reproduce the observed dissipation in different dynamical situations (Wijesekera et al., 1993 [?]; Kunze et al., 2002; MacKinnon and Gregg,
 60 2005). This happens when the rate of energy transfer from large to breaking scales is not consistent with the assumed rate of wave-wave interaction. This can occur for instance when interaction with topography short-circuits the downscale cascade (Kunze et al., 2002) or when the highly variable nature of the coastal wavefield is not amenable in a general sense to statistical models
 65 of downscale energy transport (MacKinnon and Gregg, 2005). However, there are other situations where the parameterization might work in modified form, specifically in a form that accounts for a non-standard, e.g. non-GM, frequency content. Therefore, the choice of parameterization must rely on the energy distribution of the internal wave field as a function of frequency, namely on the
 70 shear to strain ratio. The idea was to take into account non-GM internal wave fields, with the introduction of a new factor involving shear to strain variance ratio (Henye, 1991). Note that this ratio is an estimate of the frequency content of the internal wave field. Polzin et al. (1995) proposed to apply a factor in order to take into account the deviation of the frequency spectrum of the
 75 observed internal wave field from the GM frequency spectrum. Involving the shear/strain ratio led to a better collapse between results and observations according to Polzin et al. (1995). Next, Gregg et al.(2003) explicitly tested the dependence of latitude on dissipation rate using a large set of data.

The Polzin et al.(1995) method has been applied to different regions and
 80 these studies have shown that the shear and strain variances are significantly underestimated when 10-100m velocity shear and isopycnal strain are omitted from the calculations because of the low signal to noise level in this scale range

(Kunze et al., 2002).

85 Another issue raised by Polzin et al.(2002) is the bias introduced by the attenuation of LADCP shear data at vertical scales smaller than 150m. To compensate for this, they proposed corrections based on comparisons with XCP data. These corrections were next introduced in the parameterization and applied extensively by Kunze et al.(2006).

90 In order to improve shear variance estimate by taking into account small vertical scales (typically smaller than 50m), different authors integrate spectral shear up to a cut off wavenumber that depends on the shear to noise ratio. For instance Naveira Garabato et al. (2004) applied this method in the Nordic Seas, a region of weak stratification and fairly high noise level, and found a significant
95 improvement in the estimates of eddy diffusivities.

Previous studies have shown that finescale parameterizations are sensitive to the shear and strain rate of underlying turbulent processes. As those are strongly related to local bathymetry, regions of small scale steep topography that
100 coincide with regions of strong mixing are a challenge. The Mid-Atlantic ridge is one of these regions, filled with narrow passages linking deeper bassins, and often associated with overflows, that can be hydraulically controlled (Thurnherr et al., 2008; Polzin et al., 1996; Hogg et al., 1982; Ferron et al., 1998). A better understanding of the mechanisms leading to mixing in these passages
105 and testing finescale parameterizations against microstructure measurements would significantly improve our understanding of the mixing generated along the MAR, which is the largest rough topographic feature in the ocean.

To answer these questions, we analyzed a data set collected near the Lucky-Strike sub-marine volcano in the middle of a narrow passage connecting two
110 north-south deep basins of the MAR, during the GRAVILUCK cruise in August 2006. This is a region of strong mixing as evidenced by microstructure measurements (St Laurent and Thurnherr, 2007). Near the bottom, dissipation rate can be locally a factor of 100 higher than open ocean values. Finescale

measurements were performed prior to every microstructure profile. We take
 115 here the opportunity to test the validity of different finescale parameterizations
 of dissipation rate in this area, discuss their application in light of the dynamics
 and eventually propose a refined parameterization.

We first introduce the data set and the dynamics is briefly described, high-
 lighting the strong semidiurnal tide and the strong shear in the deeper layer.
 120 We next introduce the different finescale parameterizations tested in this pa-
 per. Comparisons of the different parameterizations against direct turbulence
 measurements are next presented and discussed.

2. In situ-measurements from the GRAVILUCK cruise

The Graviduck cruise (Thurnherr et al, 2008) was conducted in the Lucky
 125 Strike (LS) segment of the mid-Atlantic ridge (MAR) (Fig.1). LS volcano, lo-
 cated at 32 deg 17'N 37deg 20'W, with the summit at $\sim 1500m$ depth, in the
 middle of the MAR axial valley. The volcano is surrounded by two narrow pas-
 sages or canyons: the western passage is $5km$ wide and $2060m$ deep at the sill
 and the eastern passage is $3km$ wide and $2075m$ deep at the sill. This combina-
 130 tion of volcano and canyons forms a zonal barrier in the MAR axial valley for
 the deep water flow between a southern and a northern deep basins, both with
 depths greater than $2500m$. These two basins communicate with the abyssal
 ocean through two passages, located in the East and the North West of the
 northern basin and in the South West and South East of the southern basin.
 135 A survey of the LS segment, with both CTD/LADCP profiles and microstruc-
 ture measurements, was performed in August 2006 during the GRAVILUCK
 cruise (Thurnherr et al, 2008). Fifty six CTD/LADCP profiles were collected:
 the survey was focused on the fairly narrow eastern passage (smaller than the
 internal radius of deformation even for stratification below $1600m$) with along
 140 passage sampling. A few profiles were collected along the width of the western
 passage in order to get a sense of the distribution of transport across this wider
 passage. Also a few repeated profiles were collected in the northern and southern

basins to investigate semi-diurnal tides. Among these CTD/LADCP profiles, eleven were performed just after the deep microstructure profiler deployment and eight were used in this study for direct comparison with microstructure data. The locations of these stations are displayed in Figure 1, eight of them are in the eastern canyon and three in the western canyon. Note that the position of the deep microstructure profiler, DMP, and CTD/LADCP profiles may differ by one mile since the ship was moved in order to avoid possible collision between the DMP and the rosette.

Finescale measurements were made with a CTD SBE911+ and a workhorse sentinel 300kHz LADCP downward looking. The LADCP data, acquired in 8m bins, were post-processed with the LDEO software. The accuracy of the salinity measurements, determined from the Autosal measurements and supported by a post-cruise conductivity sensor calibration, is approximately 0.001. The temperature and pressure sensors were also post-cruise calibrated, yielding accuracies of 0.001°C and 0.5 dbar respectively. CTD data were post-processed with Software Version Seasave Win32 V 5.37d, on a 1db grid.

Microstructure measurements were made using a Rockland Scientific Deep Microstructure profiler (DMP), a free falling profiler. Further details on the DMP instrument system are described at <http://www.rocklandscientific.com>. We focus on the measurements of centimeter-scale shear from which the dissipation rate of turbulent kinetic energy $\epsilon = (15/2)\nu(\partial u/\partial z)^2$ may be estimated, where ν is the molecular viscosity and isotropy is assumed. Estimates of the shear variance are done using spectral analysis in the manner described by Gregg (1999) over 1-m depth intervals of each microstructure-shear record. No spectral corrections are done to the shear spectra, as the inertial sub-range and dissipative roll-off were always well resolved. The profiler was equipped with 2 microstructure shear sensors from which the vertical shear of the two horizontal velocity components was inferred, with a sampling rate of 512Hz. Dissipation rate of turbulent kinetic energy (ϵ) was then determined from shear spectra and the vertical profile of ϵ was interpolated on a 1m grid (St Laurent and Thurnherr, 2007). Due to instrument vibrations, a noise level of $(0.5 - 1) \times 10^{-10} W.kg^{-1}$

applies to the estimates of ϵ reported for this study. For bulk Richardson number estimates, velocity and density profiles have been smoothed over 32m prior
 175 to shear and N^2 computation in order to avoid high frequency noise.

3. The Lucky Strike segment: overview and dynamics

The analysis of GRAVILUCK data provided an overview of the dynamics, both of the low frequency, sub-inertial, flow of deep waters and of super-inertial
 180 frequency motions. The deep flow has been characterized from all LADCP profiles collected during the cruise and from one-year mooring data in the eastern passage (Pasquet et al, 2011). In summer 2006, the deep flow is directed northward and extends up to 1400m as reported in St Laurent and Thurnherr (2007). Currents are on the order of 10cm.s^{-1} at the sill, and can reach values close to 20cm.s^{-1} downstream but more importantly the current is strongly
 185 sheared (Fig.2) and possibly subject to shear instabilities. CTD profiles at the site showed that the stratification is uniform between 100m and 800m within a factor of two, and decreases with depth below 900m towards the bottom (Fig.3), roughly coinciding with the summits of the rift valley walls. Thurnherr et al. (2008) showed that the valley region was significantly less stratified than the
 190 external ridge flanks at the same depth. The along-stream density structure is characteristic of an hydraulically controlled flow, a typical situation for submarine Valleys. This assumption has been confirmed by St Laurent and Thurnherr (2007), who found critical Froude numbers near the sill.

The one-year mooring data indicate that, as expected, the deep water flow is
 195 most of the time northward while a reversed southward flow is often observed at intermediate depth (Pasquet et al, 2011). The depth of reversal varies with time, typically within $[100, 600]\text{m}$ height above bottom at the sill.

At higher frequencies, the dynamics are dominated by semi-diurnal tides. They
 200 were best characterized during the cruise near the sill by bottom mounted ADCP data. The barotropic semi-diurnal tidal flow is clearly present on the along channel velocity component while the across channel velocity component reveals a

more baroclinic structure (not shown).

Further information on internal tides was collected during the cruise from
 205 two fixed point stations, in the northern and in the southern basins. Profiles
 performed at 6h time interval provided a basic overview of the variability over
 half a semi-diurnal period. Figure 4 shows an example of this variability between
 two successive profiles (separated by half the semi-diurnal period). The vertical
 dependency of these profiles is evocative of high order vertical modes. Internal
 210 waves were also identified at higher frequencies from repeated CTD profiles over
 typically 6 hours near the surface, thus providing complementary information
 on the internal wave field over the area. This is illustrated on Figure 5, showing
 time depth plots of shear inferred from ship mounted ADCP on which isopycnals
 from repeated CTD measurements are superimposed. One can see simultaneous
 215 oscillations of narrow layers of current shear with the isopycnals. The vertical
 scale of these shear layers is typically within $[30m, 100m]$ which is of the same
 order of magnitude as the vertical scale of velocity variations inferred from the
 LADCP profiles (Fig.4).

220 Measurements of dissipation rate of turbulent kinetic energy (ϵ) by the DMP
 showed a significant increase of dissipation rate above the bottom, above and
 downstream the sill (St. Laurent and Thurnherr, 2007). Interestingly, these
 deep regions of enhanced dissipation with values greater than $10^{-7} \text{ W.kg}^{-1}$
 match areas where critical values of Ri are numerous (see profiles 9, 11, 12 and
 225 8 in Figure 6). This is in contrast with typical ϵ values in the interior that are on
 the order of $3 \times 10^{-10} \text{ W.kg}^{-1}$ except for local increases in dissipation with values
 up to $10^{-8} \text{ W.kg}^{-1}$ that are sometimes correlated with low Richardson number
 values (e.g. profiles 9 and 10 and 12 in Figure 6). Average profiles of ϵ displayed
 in Figure 7 underline the bottom intensification of turbulence downstream of the
 230 sill. Accordingly, values of eddy diffusivities were as high as $\sim 10^{-2} \text{ m}^2.\text{s}^{-1}$ over
 the first 200m above bottom (Fig.7). Downstream, diffusivities are between
 $[10^{-3}, 10^{-2}] \text{ m}^2.\text{s}^{-1}$, with a decrease of two orders of magnitude in the water
 column above it. There diffusivities are smaller though significant with an

average value of $\sim 10^{-4}m^2.s^{-1}$. The latter could reflect mixing induced by
 235 internal tides. The role of shear instability as a mechanism for mixing has been
 inferred from the Richardson number. Statistics of Ri number showed that
 39% of Ri number are below the critical value within the first 500m above the
 bottom while the percentage decreases down to 6% elsewhere. This could reflect
 the role of shear instability of the mean flow and internal waves as one of the
 240 major mechanism for mixing within a few hundred meters above the bottom.

4. Finescale Parameterizations

As the dynamics combine both strong internal tides and a mean shear flow,
 two kinds of parameterizations were tested here. We tested Garrett and Munk
 (GM) based parameterizations, in which shear and strain variances are com-
 245 pared to GM values (Garrett and Munk, 1975). We test the application of
 these parameterizations in situations of high energy levels and small-scale in-
 ternal waves typical of the LS area. Secondly, a parameterization based on
 the Richardson number, which aimed to estimate dissipation induced by shear
 instability, is also tested.

4.1. GM based parameterizations

4.1.1. Gregg 1989 parameterization

The first finescale parameterization was proposed by Gregg (1989) (1), who
 used the HWF model (Henyey et al, 1986) to estimate ϵ . It is defined as:

$$\epsilon_{\text{Gregg89}} = 7 \times 10^{-10} \langle N^2/N_0^2 \rangle \langle S_{10}^4/S_{GM}^4 \rangle W.kg^{-1} \quad (1)$$

255 Brackets denote an average over a given depth interval whose length is 320m
 for consistency for the comparison with the Kunze et al (2006) parameteriza-
 tion, N is the buoyancy frequency, N_0 is the GM buoyancy frequency set to
 $5.2 \times 10^{-3}s^{-2}$, S_{10} is the current shear estimated by differentiation over 10m
 and S_{GM} is the GM current shear equivalent to $1.91 \times 10^{-5}(\frac{N}{N_0})^2$. This parame-
 260 terization expressed in terms of shear normalized with GM assumes a constant

shear to strain ratio, R_ω , equal to that of GM. In fact this ratio varies with the frequency content of the internal wave field. To take this effect into account, Polzin et al (1995) introduced an additional factor to G89, $h(R_\omega)$ (see as well Kunze et al, 2006; Gregg et al, 2003). The parameterization then reads (e.g.,
 265 Cuypers et al, 2012):

$$\epsilon = h(R_\omega)\epsilon_{Gregg89} \quad (2)$$

where R_ω is the square of the shear to strain ratio. R_ω is computed following the procedure described in Kunze et al (2006) with shear spectra computed over $320m$ segments while strain spectra are computed over $256m$ segments, it is given by:

$$R_\omega = \left(\frac{\int_{I_1} S[V_z/\bar{N}] dk_z}{\int_{I_2} S[\xi_z] dk_z} \right)_{GM} \quad (3)$$

270 with $I_1 = [\frac{2\pi}{320}, \frac{2\pi}{150}] rad.m^{-1}$ and $I_2 = [\frac{2\pi}{150}, \frac{2\pi}{10}] rad.m^{-1}$, $S[V_z/\bar{N}]$ and $S[\xi_z]$ are the buoyancy normalized shear and strain spectra, respectively. Each factor is normalized by its GM counterpart, as indicated with 'GM'. The upper integration limit for shear is chosen in order to avoid instrument noise at smaller wavelengths while the lower integration limit for strain is chosen to avoid contamination by background stratification (see appendix for details). $h(R_\omega)$ is
 275 given by:

$$h(R_\omega) = \frac{3(R_\omega + 1)}{2\sqrt{2}R_\omega\sqrt{R_\omega - 1}} \quad (4)$$

This formulation, eq.(2), is more appropriate than the original one as shown for instance by Cuypers et al (2012). In Cuypers et al (2012) measurements were performed in the Mediterranean Sea within anticyclonic eddies where the
 280 internal wave field strongly departs from the GM internal wave field. They validated this formulation against microstructure measurements performed within the upper 100m including the seasonal pycnocline. Note that the factor $h(R_\omega)$ is smaller than 1 in the bottom boundary layer where the shear to strain ratio is higher than the GM value, which could lead to an underestimate of ϵ . Hence
 285 we tested an hybrid formulation:

$$\epsilon_{G89} = \max(h(R_\omega), 1)\epsilon_{Gregg89} \quad (5)$$

We tested another method computing vertical shear using a spectral method in order to avoid side effects of finite differencing (which are compensated by a 2.11 factor as expressed in equation (24) in Gregg, 1989) and applying a low-pass filter with a $1/32m^{-1}$ cutoff. To do so, we compute the Fast Fourier Transform (Fast Fourier Transform) of the whole velocity profile, we then set to zero in Fourier space all length scales smaller than 32m. The shear is obtained in Fourier space by multiplying by k_z . The final shear profile is obtained by taking the inverse FFT. The 32m GM shear variance is computed instead of the 10m GM shear variance for consistency (see eq.(1)).

4.1.2. Kunze et al (2006) parameterization

We tested a modified version of the parameterization first introduced by Gregg et al. (2003) and modified by Kunze et al. (2006) in which the shear variance is computed in spectral space. Following Kunze et al (2006) we first compute diapycnal diffusivity, K_{K06} which is given by:

$$K_{K06} = K_0 \frac{\langle V_z^2 \rangle^2}{\langle V_z^2 \rangle_{GM}^2} j(f/N) h(R_\omega) \quad (6)$$

V_z^2 is the shear variance which is obtained by integrating the shear spectrum on the vertical wavenumber interval $[\frac{2\pi}{320} rad.m^{-1}; \frac{2\pi}{150} rad.m^{-1}]$ and the following terms are defined as follows:

$$\begin{aligned} K_0 &= 5 \times 10^{-6} m^2.s^{-1} \\ j(f/N) &= \frac{f \cosh^{-1}(N/f)}{f_{30} \cosh^{-1}(N_0/f_{30})} \end{aligned} \quad (7) \quad (8)$$

f is the Coriolis parameter, f_{30} is the Coriolis value at 30N, (8) takes into account the variation with latitude (Gregg et al., 2003). Note that we have taken into account in our calculation the corrections of LADCP shear proposed by Polzin et al. (2002) and Thurnherr et al. (2012) (see appendix for further details).

Dissipation rate is then inferred as:

$$\epsilon_{K06} = 5K_0 K_{K06} \bar{N}^2 \quad (9)$$

with $\bar{N}^2$ is the buoyancy frequency square averaged over the 320m length segment.

315 *4.1.3. Criterion for the definition of an upper integration limit for the computation of shear variance in Kunze et al (2006)*

The vertical wavenumber cutoff of $\frac{2\pi}{150} \text{ rad.m}^{-1}$ defined by Kunze et al (2006) was chosen to be sufficiently small to avoid noise contamination. On the other hand shear variance might be underestimated if shear increases for high vertical wavenumbers beyond the upper integration limit. Thus the criterion should 320 define the largest upper integration limit while keeping a reasonable signal to noise ratio. We discuss here the procedure to find out such a criterion. For that we will consider averaged shear spectra which are computed over two depth intervals, from 0 to 300m above the bottom and from 300m above the bottom 325 to the surface (Fig.8). Averaged shear spectra from 0 to 300m above the bottom are clearly more energetic than averaged shear spectra further up and they are well above GM spectra. In both cases their shape is similar: spectra are blue for length scale larger than 100m, and are white between 100m and $\sim 30\text{m}$. The strong increase of the shear beyond 30m is due to the increase in noise. It is 330 thus clear that shear variance computed with an upper integration limit of $\frac{2\pi}{150}$ will be underestimated but that the upper integration limit should not exceed $\sim 30\text{m}$ to avoid noise contamination.

Hence we define in the following a criterion for the wavenumber cut-off based on signal to noise ratio, following Naveira Garabato et al. (2004). To compute noise 335 spectra we took a rms noise of individual measurements of 3.2cm.s^{-1} (Polzin et al., 2002) with a correction factor of $\sqrt{3}$ applied by Kunze et al. (2006) and a spectral correction (From Thurnherr, 2012 and appendix) , leading to the following expression for the vertical wavenumber noise spectrum:

$$S_{noise} = \frac{3 \times (3.2\text{cm.s}^{-1})^2 \times k_z^2}{120 \times \bar{N}^2} \frac{1}{\text{sinc}^8(\frac{k_z \times 8}{2\pi})} \quad (10)$$

340 In their study, Naveiro-Garabato et al. (2004) used a signal to noise ratio (SN) of 3 as the criterion to define the upper integration limit. The same criterion

applied on our data led to a majority of k_c beyond $\frac{2\pi}{32} rad.m^{-1}$. This is an issue as this scale is close to the bin size ($8m$), leading to a strong dependence on the spectral corrections. We therefore decided to use instead $SN = 5$, which led to an upper integration limit usually on the order of $\frac{2\pi}{32} rad.m^{-1}$ (see Fig.8). In the cases where the upper integration limit defined by $SN = 5$ exceeds $\frac{2\pi}{32} rad.m^{-1}$, we took this value to avoid any influence of spectral corrections on our estimates. Note that this wavenumber domain should encompass the typical scale of vertical shear as inferred from Figures 4 and 5.

4.2. Richardson number based parameterization

We also investigated whether a parameterization based upon Richardson number would be adequate provided that shear instability is one of the main mechanisms leading to turbulence in the LS segment. As opposed to the last parameterizations involving a computation in spectral space, this parameterization leads to a local estimate of dissipation with a 32m vertical resolution, to be compared with the length of the segments used in K06 equal to 320m.

Kunze, Williams and Briscoe (1990). We implemented the parameterization described in Kunze, Williams and Briscoe (1990) (KWB), and then tested by Polzin (1996). This parameterization aims at quantifying the rate of dissipation of turbulent kinetic energy produced by shear instability. The KWB parameterization is defined as follows:

$$\epsilon_{KWB} = \frac{fr\Delta z^2}{96} \langle N^3 \left(\frac{1}{Ri} - \frac{1}{Ri_c} \right) \left(\frac{1}{\sqrt{Ri}} - \frac{1}{\sqrt{Ri_c}} \right) \rangle \quad (11)$$

where fr is the fraction of the water column which is unstable ($Ri < Ri_c$), Δz is the interval over which S^2 and N^2 are computed ($\Delta z = 8$ here), $S^2 = (U_z^2 + V_z^2)$ and the angle brackets represent an average of all unstable events over a segment. Ri_c is defined here to be the Richardson number criterion below which a dynamical event can be unstable. As in Polzin (1996), we tested the following values for Ri_c : 0.25, 0.33, 0.5 and 1. We adopt in the following $Ri_c = 1$ which resolves more unstable events. Dissipation is estimated over

370 320m segment as for other parameterizations. The only difference is that only
 segments where more than 25% of unstable events are encountered are retained
 for the comparison. This restricts the comparison to the lower part of the water
 column, where most unstable events occur.

5. Test of the parameterizations

375 An overall comparison of the different parameterizations is first displayed
 with a scatter plot of the different ϵ (ϵ_{K06} , ϵ_{K06-SN} , ϵ_{G89} and ϵ_{KWB}) as a
 function of ϵ_{DMP} (Fig.9). Notice that both ϵ_{G89} and ϵ_{KWB} which are com-
 puted on a 10m and 8m vertical grid have been averaged over 320m segments
 used for the calculation of ϵ_{K06} . We first compare the different parameteriza-
 380 tions with scatter plots in Figure 9 and with quantitative information in Table 1.

GM based parameterizations reproduces rather well the dissipation as op-
 posed to Ri based parameterization which significantly underestimates the dis-
 sipation rate through the whole water column (Fig.9). The small number of
 385 points (Fig.9.d) results from the fact that most unstable events resolved by our
 measurements are in the lower layer near the bottom where the typical scale of
 the instabilities is larger as a result of stronger shear and weaker stratification
 compared to the interior. ϵ_{G89} and ϵ_{K06} perform about the same (Fig.9.a and
 c, Table 1) though slight differences are observed: ϵ_{G89} tends to underestimate
 390 ϵ_{DMP} in the interior whereas ϵ_{K06} tends to underestimate ϵ_{DMP} in the lower
 part of the water column. The plot with ϵ_{K06-SN} has less scatter compared
 to that of ϵ_{K06} with predictions significantly improved in the lower part of the
 water column. This points out the contribution of small scale shear variance,
 not taken into account by ϵ_{K06} , that contributes to an adequate estimate of
 395 ϵ_{DMP} above the bottom.

The performance of the different parameterizations can be characterized as
 well with histograms of the ratio between ϵ parameterized and ϵ_{DMP} , shown
 in Figure 10. Basic statistics have been performed with mean value, standard

deviation and skewness in order to quantify the distributions. The histogram
 400 for ϵ_{KWB} emphasises its underestimation of ϵ_{DMP} , with a mean value of $\frac{\epsilon}{\epsilon_{DMP}}$
 equal to $10^{-0.96}$ (0.11). The other parameterizations have a mean value of
 $\frac{\epsilon}{\epsilon_{DMP}}$ close to 1, within $[10^{-0.13}, 10^{0.15}]$ ([0.74; 1.41]). The standard deviation
 lies within a factor of $[10^{0.39}, 10^{0.49}]$, corresponding to a ratio of [2.45; 3.09], the
 smallest standard deviation being obtained with ϵ_{K06-SN} .

405 In order to compare the parameterizations in a more direct way, we consider
 as well the percentage of points that estimate ϵ_{DMP} within a factor of 2, 5 and
 10 (Table ??). All GM-based parameterization provide a reasonable prediction
 of ϵ_{DMP} with more than 80% of the values estimated within a factor of 5. More
 410 precisely ϵ_{K06-SN} provides the best estimates with 46% values within a factor
 of 2 and 91% within a factor of 5. We note that Gregg (1989) and Kunze et
 al.(2006) have about the same performance with 44% and 42% of points within
 a factor of 2 and 88% and 87% when a factor 5 is considered. The use of a
 spectral method and a wavenumber cutoff in ϵ_{G89} parameterization does not
 415 lead to significant improvement. In fact the averaged ϵ is underestimated when
 finite differences are used and overestimated with pseudo-spectral computation
 with about the same shift, in absolute value, compared to 1. The dispersion
 is smaller when finite differences are used with 44% of the points predicting ϵ
 within a factor of 2, compared to 35% with the spectral method, but is compa-
 420 rable when a factor of 5 is considered (see Table 1).

Further analysis of the relevance of the parameterizations needs to be done in
 light of the dynamics. We chose here to display ϵ in the space of stratification
 and shear (Fig.11). This representation identifies in which dynamical regime
 425 each parameterization is adequate and where it needs to be improved. The
 distribution of ϵ in (N^2, S^2) space is displayed in Figure 11 for microstructure
 measurements and parameterizations. A straight line, corresponding to the
 critical Richardson number, $Ri = 0.25$, is displayed as an indicator of shear
 instability.

430 The general pattern of dissipation in (N^2, S^2) space (Fig.11.a) shows different regions of high values. Starting from low stratification, the first region of high dissipation rate is obtained for the lowest stratification where $\frac{S}{N}$ is close to 1. This region corresponds mostly to the deepest layers, weakly stratified and strongly sheared. A second region of high dissipation rate is encountered for
435 intermediate stratification and high shear. It is likely in these regions that shear contributes to the onset of wavebreaking for instance for semi-diurnal internal tides for which the shear to strain ratio is fairly large. Eventually a third region is obtained in the pycnocline, in the most stratified water, where the shear does not vary significantly. There the Richardson number is far above the critical
440 value which suggests that wavebreaking occurs at smaller vertical scales than those resolved by our measurements. ϵ_{G89} (Fig.11.b) exhibits an overall well-reproduced pattern. ϵ_{G89} better reproduces ϵ_{DMP} in regions of intermediate and strong stratification, which corresponds to the ocean interior. In regions of weak shear and intermediate stratification, ϵ_{DMP} is largely underestimated
445 while in regions of high shear and low stratification it is largely overestimated. ϵ_{K06} underestimates ϵ_{DMP} in most regions (Fig.11.a and b) while the general pattern of ϵ_{K06-SN} best reproduces that of ϵ_{DMP} (Fig.11.a and d). In fact the most striking improvement with ϵ_{K06-SN} is revealed by the distribution of ϵ in N^2/S^2 space (Fig.11.c). The highest values obtained for low N^2 high S^2 and high
450 N^2 high S^2 are now close to the observed ones, though the same biases remain: an overestimation in the $Ri < 0.25$ region and a slight underestimation in the strong N^2 /weak S^2 region. More importantly the regions of strong dissipation occurring in the ocean interior, intermediate N^2 and S^2 values, that were not present with ϵ_{K06} are now reproduced.

455 *Contribution of shear and strain variance to turbulent dissipation.* To assess the contribution of both shear and strain variances on the computed dissipation equation 9 is used. Shear variance is computed with an upper bound defined using the signal to noise ratio or up to a wavelength of 32m, whichever comes first. Strain variance is computed as explained in the appendix. Three cases are

presented:

- a shear and strain based parameterization (using R_ω), ϵ_{K06-SN}
- a shear-only parameterization, with $h(R_\omega) = 1$ in ϵ_{K06-SN}
- a strain-only parameterization, in which ϵ is defined as follows:

$$\epsilon_0 \frac{\langle \zeta_z^2 \rangle^2}{\langle \zeta_z^2 \rangle_{GM}^2} j(f/N) \quad (12)$$

Analysis of scatter plots (Fig. 12) reveals that both shear-only based parameterizations and shear/strain based parameterization reproduce better the observed dissipation than the strain based parameterization, which exhibits more scatter. It is also worth noting that the most scattered points are encountered near the bottom in the strain based parameterization. Adding the strain variance to the shear only based parameterization slightly improved statistical aspects: rms and mean values of $\log_{10}(\frac{\epsilon}{\epsilon_{DMP}})$ decrease from 0.42 to 0.4 and 0.19 to 0.12 respectively (Fig.13). Strain based parameterization on the other hand has rms value of 0.69, more than with shear based parameterizations, but mean value is better (0.12). From that we conclude that shear based variance is the most important parameter in that kind of environment.

Maps of depth-integrated ϵ values. We provide an overview of the spatial structure of dissipation in Figure 14. Dissipation rate is averaged over two layers: the interior layer (100m-1000m) (Fig.14.a) and the deeper layer (typically 0-600 above bottom) (Fig.14.b). The values of dissipation inferred from the DMP and those predicted by K06 parameterization are displayed for comparison. Most of the averaged values vary within a factor of 10 at most. In the eastern canyon where most measurements have been performed along the axis of the canyon we notice a slight intensification of dissipation near the sill, in the narrowest part of the canyon. This structure is fairly well reproduced by the parameterization except for one profile. Note that a more detailed comparison would require further measurements performed exactly at the same location owing to the high

variability of topography and flow in the area. Furthermore in the western passage there is an E/W contrast with higher dissipation on the eastern part. This contrast is consistent with the flow that is most intense on the eastern part due to the influence of rotation. Interestingly this contrast is reproduced by the parameterization but dissipation values are slightly under-estimated as previously mentioned. In the ocean interior the range of dissipation value is reduced as the effect of bottom topography is weaker. ϵ_{K06-SN} predicts dissipation within a good approximation (see for instance in the western passage).

6. Conclusion

The LS segment of the MAR is a region that combines different dynamical processes favoring the onset of turbulence with a mean flow strongly sheared and an energetic internal wave field dominated by semi-diurnal tides. As a consequence of these dynamics combined with the complex bathymetry, it is a region of intense mixing thus providing an ideal site for the validation of finescale parameterizations under conditions of high energy levels. We tested the validity of several finescale parameterizations of dissipation rate of turbulent kinetic energy by comparisons with microstructure measurements. We considered two sets of parameterizations: the first set was developed with the aim of estimating dissipation induced by internal wave breaking while the second one is a model developed for turbulence induced by shear instability and is based upon the Richardson number.

The latter parameterization had low skill in reproducing well enough the observed dissipation rates. In fact, the method provides relevant estimates provided that there is a scale separation between the duration of the unstable events and that of the inverse of the billows growth rate as mentioned by Polzin (1996). A failure of this condition might explain why KWB parameterization is not relevant with our dataset.

Instead the former set of parameterizations, derived from Kunze et al. (2006)

and Gregg (1989), provided adequate estimates of ϵ_{DMP} in our dynamical context. The general pattern in (N^2, S^2) space is fairly well reproduced though some biases are observed: ϵ_{G89} tends to slightly overestimate ϵ_{DMP} in low stratification areas underestimate ϵ_{DMP} in low shear regions while K06 tends to slightly underestimate ϵ_{DMP} in most regions. The fact that K06 underestimates ϵ points out a generic problem, that of the estimate of shear variance. We find that with an upper integration limit of $\frac{2\pi}{150} rad.m^{-1}$ (as in Kunze et al, 2006) the contribution of high vertical wavenumbers which are significant here, is not taken into account. It is thus important to extend the wavenumber interval of integration by defining a cut-off wavenumber at which the shear to noise ratio is equal to a prescribed threshold. We find here that applying a criterion of signal over noise of 5 was essential to reproduce dissipation in regions of intermediate values of N^2 where internal waves and high vertical modes internal tides come into play. In comparison, Tian et al. (2009) and Naveira Garabato et al.(2004) used a threshold value of 3 for the computation of shear variance leading in their case to a cut-off vertical wavelength ranging from 50m to 90m. In our case, we obtained an upper bound most of the time on the order of 30m when we tried a shear to noise ratio set to 3, indicating that our data were less noisy or more energetic.

This study allowed us to provide a guide line for improving finescale parameterizations in different dynamical environments. The first main point is that the integration interval for K06 parameterization must include the relevant vertical scales, namely here those of the high-order vertical modes that include the significant semi-diurnal internal tide. The second main point is that the factor taking into account deviation from the GM frequency spectrum is required to reproduce the impact of higher frequency waves observed in the upper layer. Also the overestimate of dissipation in regions of weak stratification where the Richardson number is close to critical points is observed for all parameterizations. This points out a generic problem in these situations for which wave wave interactions do not fulfill the theoretical framework of GM parameterizations.

This might result from the reduced internal waves frequency range of internal waves combined with a high energy level. Apart from the weak stratification another factor is related to the canyon bathymetry which might lead to multiple internal wave reflections with enhanced interactions. In fact this was found in several studies when high energy level at finescale and/or weak stratification/canyon finescale parameterization overestimate epsilon (e.g. Bouruet-Aubertot et al, 2010 for the former and Carter and Gregg, 2002 for the latter). Others studies showed the inadequacy of finescale parameterizations to estimate dissipation rate in canyons (e.g. Kunze et al, 2002) where other processes than wave wave interactions such as critical reflections, scattering by small scale topography, internal hydraulic jumps dominate. Instead these parameterizations work fairly well in the Lucky Strike area as inferred from our observations. This suggests that internal wave cascade processes are dominant in our observations. Another interesting result is that the slight modification introduced in the G89 parameterization, with a shear to strain ratio taken into account, widens its range of application, thus providing a 10m estimate of dissipation that does not depend on spectral corrections. This method has been applied to a two mooring data set in the eastern passage at the sill and downstream, including ADCP and CTD measurements, to provide a temporal evolution of dissipation and mixing over a year (Pasquet et al, 2011; Pasquet et al, 2015).

Appendix: Implementation of Kunze et al. (2006) parameterization

We present in the following how we apply the procedure described by Kunze et al. (2006). We define 320-m segments for the shear and 256-m segments for the strain following Kunze et al. (2006). We then remove a linear trend on the velocity for each segment prior to the shear calculation and a quadratic trend on N^2 for each segment. The strain is estimated as follows:

$$\xi_z \approx \frac{N^2 - \overline{N^2}}{\overline{N^2}} \quad (13)$$

We apply a 10% Tukey window on the segments, prior to performing FFT.

We get the following spectral densities for the normalized shear and for the strain:

$$\begin{aligned} S[V_z/\bar{N}] &= \frac{1}{\bar{N}^2} \times |fft(V)|^2 k_z^2 \\ S[\xi_z] &= |fft(\xi_z)|^2 \end{aligned}$$

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As underlined by Polzin et al. (2002), spectral corrections are required due to the smoothing induced by post processing of data and LADCP motions. As the data have been processed using Velocity Inversion method (Visbeck, 2002), we followed Thurnherr et al. (2012) to determine the spectral corrections. We therefore applied the the following expression considering the fact the size of the bins are 8m:

$$\text{Shear}_{\text{Corr}}(k_z) = \text{sinc}^8\left(\frac{k_z \times 8}{2\pi}\right) \quad (14)$$

taking into account depth binning, range averaging and superensemble preaveraging. According to Thurnherr (2012), instrument tilting doesn't need to be corrected, as the maximum spectral damping associated with instrument tilting in the vertical wavenumber range resolved by the LADCP remains below 10% in the case of bin size smaller than 10m. Note also that when spectral corrections are applied, we use downcast LADCP profiles only, as suggested by Thurnherr (2012).

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For the strain, the spectral correction that takes into account the smoothing due to the interpolation operations reduces to:

$$\text{Strain}_{\text{Corr}} = \text{sinc}^2\left(\frac{k_z \times 2}{2\pi}\right) \quad (15)$$

We calculate the variances of shear and strain by integrating spectra over the interval $I_{sh} = [\frac{2\pi}{320} \text{rad.m}^{-1} \frac{2\pi}{150} \text{rad.m}^{-1}]$ for the former and over the interval $I_{st} = [\frac{2\pi}{150} \text{rad.m}^{-1} \frac{2\pi}{10} \text{rad.m}^{-1}]$ for the latter.

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We next compute the variance of these quantities associated with the Garrett and Munk model:

$$\langle V_z^2 \rangle_{GM} = \frac{3\pi E_0 b j_*}{2} \int_{I_{sh}} \frac{k_z^2}{(k_z + k_{z*})^2} dk_z \quad (16)$$

$$\langle \xi_z^2 \rangle_{GM} = \frac{\pi E_0 b j_*}{2} \int_{I_{st}} \frac{k_z^2}{(k_z + k_{z*})^2} dk_z \quad (17)$$

with

$$k_{z*} = \frac{\pi j_* N}{b N_0}$$

with the following constants of the model:

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- $E_0 = 6.3 \cdot 10^{-5}$ (GM dimensionless spectral level)
- $b = 1300$ m (Height scale of the thermocline)
- $J_* = 3$ (Order of vertical mode)
- $N_0 = 5.2 \cdot 10^{-3}$ rad.s⁻¹ (Reference buoyancy frequency)
- f being the inertial frequency
- f_{30} being the inertial frequency at 30N.

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We then compute the ratio of the normalized shear variance with the strain variance, called R_ω . The value of this ratio is determined for linear internal waves by their frequency, the inertial frequency and the buoyancy frequency.

620 Its general expression is:

$$R_\omega = \frac{\langle V_z^2 \rangle}{N^2 \langle \xi_z^2 \rangle} \quad (18)$$

Since normalized shear and strain variances are computed over a different wavenumber interval, I_{sh} and I_{st} , each variance is normalized by the GM variance. Note that R_ω can not be theoretically less than 1 for an internal wave ($R_\omega = 1$ for a wave with $\omega = N$). However, if this case is encountered in the calculations, ϵ is

625

set to NaN.

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7. Figures

Figure 1: Bathymetry of the area of study. The submarine volcano in the center is surrounded by 2 passages linking 2 deep basins to the north and south. DMP profiles (blue) and CTD/LADCP profiles (red) were made along the eastern passage and across the western passage.

Figure 2: Mean currents along the eastern passage. Red arrows are 50m averaged processed LADCP data. Yellow arrows are two-week averaged velocities recorded by a bottom mounted ADCP.

Figure 3: Mean N^2 profiles in s^{-2} from 11 CTD casts (blue curve), individual points from all 11 profiles are displayed as well (black points).

Figure 4: a) Baroclinic eastward velocity in the northern basin at t_0 (in blue) and $t_0 + 6h$ (in red) as a function of depth. b) same as in a) but in the southern basin.

Figure 5: Shear of the U component (a) and the V component (b) in s^{-1} from ship mounted ADCP, over the top 500m during 9 hours. Isopycnals from repeated CTD casts are superimposed in white lines.

Figure 6: Profiles of dissipation along the eastern passage. Profiles of dissipation (blue) are plotted on a logarithmic axis. The colorbar denotes $\log_{10}(1/Ri)$. The downstream direction of the deep layer flow is toward the right.

Figure 7: Profiles of dissipation (blue) and eddy diffusivity (red) averaged along the eastern passage on 50m segments. Profiles are plotted on a logarithmic axis and are functions of height above bottom (HAB). Dissipation rate from the DMP is displayed in blue ($W.kg^{-1}$) and diapycnal diffusivity is displayed in red ($m^2.s^{-1}$).

Figure 8: Vertical wavenumber spectra of normalized shear corrected with (14) (bold, solid lines), vertical wavenumber spectra of GM shear (thin, solid line), vertical wavenumber spectra of strain (dashed lines) and modelled noise spectra of the shear calculated with 10 (dotted lines). The vertical dashed lines denote where the ratio between averaged shear spectra and modelled noise spectra, SN , is equal to 5. Vertical wavenumbers of $\frac{2\pi}{150} \text{ rad.m}^{-1}$ and $\frac{2\pi}{32} \text{ rad.m}^{-1}$ are shown with thin vertical dashed lines. Shear spectra averaged over the first 300m above the bottom are in grey, and black refers to an average over shallower depths.

Figure 9: Scatter plots of parameterizations as a function of ϵ_{DMP} from ϵ_{K06} (a), ϵ_{K06-SN} (b), ϵ_{G89} (c) and ϵ_{KWB} (d) compared to the dissipation measured by the DMP. The two dotted lines delimit the area where the ratio between ϵ from the parameterizations and observation is less than a factor of 2. The colorbar denotes the height above bottom.

Figure 10: Histogram of the logarithm of the ratio between ϵ parameterized and ϵ observed for the 4 following parameterizations: ϵ_{K06} (a), ϵ_{K06-SN} (b), ϵ_{G89} (c) and ϵ_{KWB} (d). The x-axis is the logarithm of the ratio. Mean value, μ , standard deviation, σ , and skewness, γ are indicated in each case. The red curve refers to the histogram of Gregg89 computed using the FFT method described in section 4.

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 Figure 11: Distribution in the (N^2, S^2) space of ϵ_{DMP} (a), ϵ_{G89} (b), ϵ_{K06} (c) and ϵ_{K06-SN} (d). The averages were performed over 320m segments. The black line represents the threshold value of Richardson number ($Ri = 0.25$).

Figure 12: Scatter plots of parameterizations as a function of ϵ_{DMP} for: shear only based ϵ_{K06-SN} (a), strain only based ϵ_{K06-SN} (b) and shear+strain equal to ϵ_{K06-SN} (c). The two dotted lines delimit the area where the ratio between ϵ from the parameterizations and ϵ_{DMP} is less than 2. The colorbar denotes the height above bottom.

Figure 13: Histogram of the logarithm of the ratio between ϵ parameterized and ϵ_{DMP} observed for the 3 following parameterizations: shear only based ϵ_{K06-SN} (a), strain only based ϵ_{K06-SN} (b) and shear+strain equal to ϵ_{K06-SN} (c). The x-axis is the logarithm of the ratio and the y-axis is the percentage of points with this ratio. Mean value, μ , standard deviation, σ , and skewness, γ are indicated in each case.

Figure 14: Dissipation parameterized using K06 with an upper bound signal/noise=5, ϵ_{K06-SN} (circles) and dissipation measured by the DMP (squares with black edge) averaged from 100m below the surface to 1000m depth (a) and from the bottom to 600m above the bottom (b).