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Ocean mixing in deep-sea trenches: New insights from the Challenger Deep, Mariana Trench

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ABSTRACT

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41 **Reliable very deep shipborne SBE 911plus Conductivity Temperature Depth (CTD) data**

42 **to within 60 m from the bottom and Kongsberg EM122 0.5°×1° multibeam echosounder**

43 **data are collected in the Challenger Deep, Mariana Trench. A new position and depth are**

44 **given for the deepest point in the world's ocean. The data provide insight into the**

45 **interplay between topography and internal waves in the ocean that lead to mixing of the**

46 **lowermost water masses on Earth. Below 5,000 m, the vertical density stratification is**

47 **weak, with a minimum buoyancy frequency N = 1.0±0.6 cpd, cycles per day, between 6,500**

48 **and 8,500 m. In that depth range, the average turbulence is coarsely estimated from**

49 **Thorpe-overturning scales, with limited statistics to be ten times higher than the mean**

50 **values of dissipation rate $\varepsilon_T = 3\pm2\times10^{-11}$ m^2 s^{-3} and eddy diffusivity $K_{zT} = 2\pm1.5\times10^{-4}$ m^2 s^{-1}**

51 **estimated for the depth range between 10,300 and 10,850 m, where N = 2.5±0.6 cpd.**

52 **Inertial and meridionally directed tidal inertio-gravity waves can propagate between the**

53 **differently stratified layers. These waves are suggested to be responsible for the observed**

54 **turbulence. The turbulence values are similar to those recently estimated from CTD and**

55 **moored observations in the Puerto Rico Trench. Yet, in contrast to the Puerto Rico**

56 **Trench, seafloor morphology in the Mariana Trench shows up to 500 m-high fault scarps**

57 **on the incoming tectonic plate and a very narrow trench, suggesting that seafloor**

58 **topography does not play a crucial role for mixing.**

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60 Keywords: new deepest point estimate; Challenger Deep; Mariana Trench; improved deep CTD

61 observations; turbulence parameter estimates; inertio-gravity tidal wave

62 propagation

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1. Introduction

Life exists at great ocean depths in the ocean's hadal zone to water depths of over 10,000 m in deep sea trenches like the Mariana Trench (Jamieson, 2015; Gallo et al., 2015; Nunoura et al., 2015). As the deepest life requires sufficient supply of nutrients and energy in form of chemical species, the ocean, even at these great depths, has to be in motion and cannot be stagnant. In analogy with the atmosphere, where breathing by inhaling of oxygen would be impossible without turbulent motions, life in the hadal zone requires turbulent rather than laminar flows for survival. Mainly due to the logistical problems imposed by the large hydrostatic pressure which normal oceanographic equipment does not withstand, little is known about the physical oceanography of deep trenches and nothing about the physics that govern the turbulent processes. For example, turbulence microstructure profiles do not go deeper than 6,000 m to date. As an indicator for upper trench turbulence, recent yearlong high-resolution temperature measurements from around about 6,000 m just below the 'top' of the Puerto Rico Trench suggest turbulence generation by the interaction of large-scale 20-100 days periodic boundary currents with near-inertial and tidal internal wave breaking (van Haren and Gostiaux, 2016). Turbulence estimates from these data correspond to estimates from shipborne Conductivity Temperature Depth CTD data averaged over a suitable depth range of 600 m and the overturn shapes suggest shear-convective turbulence (van Haren, 2015). These shear-convective turbulent mixing processes found in trenches are quite different in magnitude from the mainly shear-driven turbulence found in deep passages through ridges and between islands (e.g., Polzin et al. ,1996; Lukas et al., 2001; Alford et al., 2011). However, in both cases turbulence is inherently pulse- and intermittent-like with overturn sizes reaching 200 m.

The only moored and hourly sampled measurements so far, by Taira et al. (2004), near the deepest point on Earth, the bottom of the Challenger Deep--Mariana Trench, showed typical current speeds of 0.04 m s^{-1} with a dominant semidiurnal tidal periodicity. Although Taira et al. (2004) did not show internal wave band spectra they mentioned sub-peaks at diurnal and inertial frequencies. These data already suggested that waters are not stagnant. The observed

91   semidiurnal currents may be related to internal tides, whether propagating from remote source

92   Luzon Strait (Morozov, 1995) or from local source Mariana Arc (Jayne and St. Laurent, 2001).

93   Water characteristics are also barely known near the bottom of the Challenger Deep (Table

94   1). This is mainly because few oceanographic research vessels are equipped with cables that

95   are more than 11 km long. As an alternative and following discrete inverse thermometer

96   readings from R/V Vityaz in the late 1950's, a small free-falling water-sampling device

97   equipped with reversing thermometers was dropped to the bottom of the Challenger Deep in

98   1976 (Mantyla and Reid, 1978). These data are also used by Taira et al. (2005) as a reference

99   for the first deep CTD-cast attached to a custom-made titanium wire in the Challenger Deep,

100   down to 10,877 m. They used a SeaBird Electronics SBE-911 CTD at a station that was 40 km

101   east of the site where Mantyla and Reid (1978) deployed their water sampler. Manned (Gallo

102   et al., 2015) and un-manned (Nunoura et al., 2015) submarines carried CTDs (an SBE-49 in the

103   latter case), but these data have not been analyzed and published for detailed water

104   characteristics.

105   In this paper, we report on new high-resolution SBE-911 CTD casts into the Challenger

106   Deep, reaching a depth of 10,851 m in 10,907 m water depth at a location about 2 km east of

107   Mantyla and Reid's position. The CTD data deliver T, Salinity S and stratification information

108   and thus information on the internal wave band below 5,000 m of the surrounding ocean floor

109   and near the maximum capability of standard oceanographic instrumentation. A second

110   objective of this experiment is to gain a first impression of the internal wave-induced turbulence

111   variation with depth down to near the bottom of the Challenger Deep. As a third objective, we

112   use the simultaneously acquired high-precision multibeam echosounder data that were

113   calibrated with the local CTD data to validate the most recent estimates of Challenger Deep's

114   greatest depth (Gardner et al., 2014).

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2. Data

Observations have been made from the German R/V Sonne above the Challenger Deep, the southernmost part of the Mariana Trench including world's deepest point (Fig. 1). We collected SeaBird SBE911plus CTD profiles using freshly calibrated T-C sensors at 11° 19.752′N, 142° 11.277′ E in 10,907±12 m water depth in November 2016. Water depth was measured using a Simrad EM122 multibeam bathymetry system with a 0.5°×1° beam angle. After an initial multibeam profile over the area using a sound velocity profile from a standard CTD cast, a 120° swath angle and a survey speed of 11 knots, the echosounder was calibrated with a sound velocity profile from the local CTD cast before a second multibeam profile was run at low speed of less than 2 knots but still with 120° swath angle. CTD data were converted into sound velocity using Delgrosso (1974) and the soundings of the first profile were recalculated using that sound velocity profile. The footprint of the R/V Sonne multibeam system is 96x192 m in 11,000 m water depth and decreases in shallower water but increases away from the nadir. Making use of the multiple overlap resulting from the low survey speed, the multibeam data were binned in a 100x100 m grid. The standard deviation of sounding depths within one cell amounts to a ±12 m uncertainty in such water depths.

We mounted the CTD horizontally at the bottom of a 1.9 m high frame, which was stripped from all other equipment and water sampling bottles. The CTD frame was lowered using a 12 km long steel cable with 18 mm diameter. Except for reduced speeds in the upper and lower 500 m, the winch speed was kept constant at a rate of 0.7 m s^{-1} for downcast and 1.0 m s^{-1} for upcast. Weather and sea conditions were good, with 4 m s^{-1} wind speeds and maximum 2 m wave height. R/V Sonne held its position to within 10 m precision during the 14 hours of CTD operation.

Although the main electronics housing and the housing of the T- and C-sensors were made of titanium and nominally rated to 10,500 m, a 'standard', nominally 7,000 dbar rated pressure p sensor was used. Nonetheless, we acquired a CTD profile over nearly the entire water column. This was achieved by first launching a normal CTD cast to 7,965 m, close to the maximum

146 depth without damaging the installed pressure sensor. After being brought back on deck, the

147 oil-filled opening leading to the pressure capillary tube was replaced by a NIOZ custom-made

148 titanium plug. Thus, the second CTD cast, deployed 6 hours after the first one, did not include

149 a pressure sensor and we had to use the length of the cable paid out as guidance for depth. This

150 required higher effort in post-processing and caution upon approaching the sea floor. With a

151 ±25 m uncertainty in water depth estimates known at the time of CTD cast and a ±5 m

152 uncertainty in cable length read-out/depth relationship, we decided to stop the CTD at 10,851

153 m while the multibeam echosounder showed 10,905 m. In retrospect, after re-analyzing the

154 multibeam data, the CTD was stopped 56±12 m from the bottom.

155   Additional post-processing of the CTD data from the second cast involved matching times

156 of the winch cable length with the CTD data. As the former was available at a rate of 1 Hz and

157 the latter at a rate of 24 Hz, the CTD data were first averaged to 1 s data. A linear relationship

158 exists between cable length and depth for the first CTD cast. This relationship was subsequently

159 used to obtain depth and pressure for the second CTD cast. Unfortunately, this relationship does

160 not account for the fact that the CTD's pressure sensor records the ship's motions by waves

161 (e.g., Trump, 1983), while cable length does not register wave motions. We attempted to correct

162 this by establishing a second relationship between the first derivatives, gradients, of depth and

163 those of the cable tension reading relative to the local tension. Although the two roughly show

164 a linear relationship, small-scale mismatches in phase and amplitude of waves are large, and

165 there was no overall improvement compared to the linear relationship between cable length

166 reading and depth. This implies that, using only the latter data as a proxy for depth, the salinity

167 data, like those of density, appear considerably noisier during the second cast with respect to

168 the first cast, while temperature data are little affected. This was remedied by applying a sharp

169 double elliptic, phase-preserving low-pass filter (Parks and Burrus, 1987) with 0.05 cps, cycles

170 per second, cut-off to all data, as proposed in van Haren (2015) to remove ship motion effects

171 by surface waves.

172   As SeaBird's conductivity sensor already applies calibration coefficients including factors

173 for temperature and pressure corrections, an extra step in post-processing of the second cast

174 was necessary. The acquisition computer applies calibration coefficients to the instrument's

175 sensor frequency data and it is necessary to re-apply them to the conductivity sensor frequency

176 data using the cable length proxy for pressure. Not applying these corrections leads to

177 underestimation of the salinity and density profiles as a function of pressure. This would result

178 in an apparently unstable water column under weakly stratified conditions.

179   After the additional post-processing described above, we processed the CTD data using the

180 standard procedures incorporated in the SBE-software, including corrections for cell thermal

181 mass using the parameter setting of Mensah et al. (2009) and sensor time-alignment. We

182 calculate the potential temperature θ, practical salinity S and density anomalies σ_8 referenced

183 to 8,000 dbar in order to allow comparison with the results of Taira et al. (2005). All other

184 analyses were performed using Conservative (~potential) Temperature (Θ), absolute salinity

185 SA and density anomalies σ_{11} referenced to 11,000 dbar using the gsw-software described in

186 (IOC, SCOR, IAPSO, 2010).

187   Estimates of turbulence dissipation rate $\varepsilon_T = c_1^2 d^2 N^3$ and vertical eddy diffusivity $K_{zT} =$

188 $m_1 c_1^2 d^2 N$ are made from the downcast CTD data using the method of reordering potentially

189 unstable vertical density profiles in statically stable ones, as proposed by Thorpe (1977). Here,

190 d denotes the displacements between unordered (measured) and reordered profiles. N denotes

191 the buoyancy frequency computed from the reordered profiles. Rms values are not determined

192 over individual overturns, as in Dillon (1982), but over 200 m vertical intervals that just exceed

193 the largest overturn intervals. This avoids the complex distinction of smaller overturns in larger

194 ones and allows the use of a single averaging length scale. We use standard constant values of

195 $c_1 = 0.8$ for the Ozmidov/overturn scale factor (Dillon, 1982) and $m_1 = 0.2$ for the mixing

196 efficiency (Osborn, 1980; Oakey, 1982). This is the most commonly used parameterization for

197 oceanographic data. Recently, it is challenged for young convective turbulence mostly from

198 numerical modeling (e.g., Scotti, 2015). However, the above parametrization still seems valid

199 for most deep-ocean applications provided some suitable averaging is performed over all

200 relevant overturning scales of the unique mix of shear- and convective overturning in large

201 Reynolds number flow conditions (e.g., Ferron et al., 1998; Gregg et al., 2012; Mater et al.,

202 2015). As a criterion for determining overturns from the surface wave low-pass filtered density

203 data, we only used those data of which the absolute value of difference with the local reordered

204 value exceed a threshold of 7×10^{-5} kg m^{-3}, which corresponds to applying a threshold of 1.4×10^{-3}

205 ^{3} kg m^{-3} to raw data variations (e.g., Stansfield et al., 2001; Gargett and Garner, 2008).

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3. Observations

3.1 Multibeam bathymetry data and the deepest point

209 For water depths greater than 8 km, the multibeam swath width is approximately 30 km

210 (Fig. 1; for reference: the size of the figure equals 45x37 km). The first multibeam bathymetry

211 survey was made with a standard sound velocity profile and high velocity (while sailing the

212 orange trajectory in Fig. 2, partially outside the window on the south side). After passing

213 directly over the then deepest point known (Gardner et al., 2014), indicated by 'NH' in Fig. 2,

214 a quick analysis was made of the data searching for the deepest point. At that analyzed point,

215 the deep CTD casts were made (Figs 1, 2). A second survey (blue-purple trajectory in Fig. 2)

216 was made with the sound velocity profile from the first CTD cast, thereby properly correcting

217 for local conditions. Thus, a total of three passes were made over the western portion of the

218 Challenger Deep, to within 1.5 km directly over the deepest point known. This provides a rather

219 precise estimate of the bottom topography of the middle portion of Figs 1,2, as the center of the

220 sweep is the most accurate part of a multibeam sweep.

221 Re-analysis of the multibeam data on a 100x100 m grid showed that the CTD casts were

222 made about 1.6 km west of the deepest point. Our data show that the deepest point is at

223 10,925±12 m at 11° 19.945'N, 142° 12.123'E. The horizontal position of the grid point has an

224 uncertainty of ±50 to ±100 m, depending on along-track or across-track direction. Nevertheless,

225 the position is significantly different from the one established as the deepest point by Gardner

226 et al. (2014). Their depth is significantly greater, by nearly 60 m.

227 The multibeam bathymetry data show deepening from about 2,000 m water depth on the

228 Mariana Arc to the maximum depth of 10,925 m before shoaling on the incoming plate to about

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229   5,000 m water depth for the abyssal plain not affected by subduction (Fig. 1). Steep slopes are

230   observed on the accretionary prism for the last 10 km before reaching the deformation front,

231   with water depth increasing from 8,300 m to 10,925 m. Overall the Mariana Trench is very

232   narrow: 120 km when measured at 5,000 m water depth in the study area. The data reveal

233   distinctly different morphologies of the incoming and overriding plates (Fig. 1). While the

234   overriding plate shows a chaotic morphology with 10 km-sized lobes in irregular patterns, the

235   incoming plate is generally dipping smoothly except for very high fault scarps in excess of 500

236   m. The faults strike parallel to the trench azimuth and some extend for several tens of

237   kilometers. Generally, the fault scarps dip away from the subduction zone, although two fault

238   scarps identified in the bathymetry data verge towards the subduction zone. The faults dip at

239   45 degree angles and the typical separation between major faults is between 3 and 3.5 km. The

240   deformation front separating the incoming and overriding plate generally strikes parallel to the

241   trench at N80. In the Challenger Deep a fault appears to cross-cut the deformation front from

242   the incoming plate and albeit subdued continues in a N70 direction into the accretionary prism.

243

244   *3.2 Comparison of CTD data*

245   All data at pressures exceeding 5,000 dbar, the approximate level of the surrounding ocean

246   floor to the southeast, from the two CTD casts are presented in Figure 3. They show good

247   agreement, with downcasts (red and blue) and upcasts (light-blue and black) better aligned

248   between themselves than down- with upcasts (red/light-blue, blue/black) from a particular CTD

249   cast. This is due to dynamic pressure by the water flow during the lowering and hoisting of the

250   CTD frame that affects the pump speed in the TC-duct (van Haren, 2015; Uchida et al., 2015).

251   This variation in water flow speed passing the T- and C-sensors artificially modifies the

252   temperature measured due to variations in skin friction. Temperature varies by about 3×10^{-4} °C

253   per m s^{-1} flow speed change. For a horizontally mounted system, temperature is directly related

254   to pressure variations with time, i.e. the lowering speed (van Haren and Laan, 2016). This is

255   confirmed here, with the downcast 2 (blue) and downcast 1 (red) profiles being consistently

256   warmer than the upcast 2 (black) and upcast 1 (magenta) profiles, respectively (Fig. 3a). The

257 observed T-difference associates with the sign-change in velocity between up- and downcast

258 and with the reduction in hoisting/lowering speeds towards the bottom of the profiles. This

259 artificial temperature variation is transferred to conductivity measurements, being inversely T-

260 dependent besides being directly p-dependent. Artificial variations are consequently visible in

261 salinity (Fig. 3b) and density anomaly data (Fig. 3c) as consistently lower values for the down-

262 compared to the upcasts. On top of these large-scale, slow variations between down- and upcast

263 data, faster variations are due to surface waves generating ship's motions, i.e. the apparent noise

264 in the data. Whilst these are small for cast 1, they are 2-5 times larger for cast 2, as we cannot

265 correct for short-term pressure variations when using the cable length read-out data as a proxy

266 for pressure. For the subsequent analyses this is not a problem, as surface wave effects will be

267 filtered out (van Haren, 2015), see Section 3.3.

268   The present data are statically stable on the large, 200 m-vertical scale (Fig. 3c; better

269 inferable from temperature data Fig. 3a). Potential temperature steadily decreases with pressure

270 (Fig. 3a). Salinity steadily increases with pressure (Fig. 3b). As a result, with both salinity and

271 temperature contributing positively, density increases towards the bottom, albeit mainly due to

272 salinity variations in the lower 2,500 dbar (Fig. 3c). In that range, stratification increases

273 towards the bottom as well, also mainly due to salinity variations. The weakest stratification

274 exists higher -up, between about 6,500 and 8,500 dbar.

275

276 *3.3 Turbulence parameter estimates*

277   In the range of 5,000 to 6,000 dbar where stratification is relatively strong, the observations

278 in Fig. 3 show vertical excursions due to internal wave motions by the differences between,

279 especially, temperature profiles. At a particular temperature, the change in pressure exceeds

280 100 dbar, a large excursion but typical for internal wave motions in the deep ocean. At greater

281 depths however, these data are not suitable for internal wave 'observations' because 'short-

282 'scale surface wave effects dominate the variations. The artificial surface wave motions also

283 need to be removed in order to estimate turbulence parameters. A low-pass filter with a 0.05

284 cps cut-off allows separating the well-defined surface wave peak due to ship motion coupled

285 down the wire centered around 0.1 cps in all primary parameters (Fig. 4). Considering the 0.7

286 m s^{-1} lowering speed of the CTD frame, 15 m and larger structures like overturns are thus

287 retained, which suffices to resolve the most energetic turbulence overturning scales estimated

288 to be 100 m in weakly stratified waters.

289 The filtered data show a smooth large-scale variation with depth, except for some saline

290 water intrusion in the upper 300 m with a maximum salinity around 125 m (Fig. 5). Except for

291 this anomaly, temperature continuously decreases (Fig. 5a), while salinity (Fig. 5b) and density

292 (Fig. 5c) continuously increase with depth, when smoothed over scales >200 m. In terms of the

293 buoyancy frequency, the stratification is strong in the upper 25-125 m of the water column, and

294 steadily decreases with depth down to halfway the water column. Below $z = -5,000$ m, $N < 2.5$

295 cycles per day (cpd), and reaches a minimum of $N = 1\pm0.6$ cpd = $2.5\pm1.5f$ between 6,500 and

296 8,500 m (Fig. 5d). The standard deviation designates the same error as found in very weakly

297 stratified waters in the deep Mediterranean (van Haren and Millot, 2006). These N-values are

298 very close to tidal frequencies, but the inertio-gravity wave band includes the tidal frequencies

299 so that their motions can represent freely propagating waves (further discussion in Section 4).

300 The turbulence estimates computed from the two CTD downcasts using Thorpe

301 displacements after reordering yield comparable results (Fig. 6). Vertical averages over the

302 range between 5,000 and 7,750 m are within a factor of two: For the first CTD-cast we find ε_T

303 $= 1.4\pm1\times10^{-10}$ m^2 s^{-3} and $K_{zT} = 1\pm0.7\times10^{-3}$ m^2 s^{-1} while for the second cast $\varepsilon_T = 2.3\pm1.5\times10^{-10}$

304 m^2 s^{-3} and $K_{zT} = 1.5\pm1\times10^{-3}$ m^2 s^{-1}. Vertically, the plotted 200 m averaged energy dissipation

305 rate (Fig. 6b) and eddy diffusivity values (Fig. 6c) vary over nearly three orders of magnitude.

306 This is common for ocean turbulence, although in stronger turbulence/stratification four orders

307 of magnitude variation occur (e.g., Gregg, 1989). Averaged over ranges of 500-1,000 m, the

308 variation is a factor of about 10. Highest values and largest rms overturn displacements (Fig.

309 6d) are found in the weakest stratification, between 6,500 and 8,500 m. For the near-bottom

310 range [10,300 10,850] m averages of $\varepsilon_T = 3\pm2\times10^{-11}$ m^2 s^{-3} and $K_{zT} = 2\pm1.5\times10^{-4}$ m^2 s^{-1} are

311 estimated. These values are rather uncertain as they are determined from only 3 or 4 overturns.

312

4. Discussion

Originally (Thorpe, 1977), the method of overturn displacements is a statistical estimate of turbulence rather than an event by event comparison. The standard error of such average estimates is a factor of 2-3 over 100-200 m intervals, which is also typical for microstructure profiler estimates (e.g., Oakey, 1982; Gregg, 1989). The present two CTD profiles (down to 8,000 m) obtained 6 hours apart, i.e. half a semidiurnal tidal period, and the single profile to near the bottom provide limited data for statistics to which the instrumental error contributes only modestly. However, Oakey (1982) showed that the Thorpe overturn – Ozmidov scale relationship is spread over one order of magnitude around the above mean. This is mainly due to various stages of mixing being patchy in place and time. Recently, this relationship has been questioned for 'convective, young' turbulence, from numerical modeling (Scotti, 2015) and observations (Mater et al., 2015). Mater et al. (2015) demonstrate that after suitable averaging ocean data, either in time or space, the canonical mean c_1- and m_1-values used here are retrieved for typical ocean conditions (in fact they even find $c_1 = 1$). This is no surprise, because a mix of shear- and convection-induced, young and old turbulence exist concurrently in the ocean interior, even under weak stratification. It is the reason the method has been and still is widely used in oceanography (e.g., Thorpe, 1977; Seim and Gregg, 1994; Galbraith and Kelley, 1996; Ferron et al., 1998; Stansfield et al., 2001; Gargett and Garner, 2008; Alford et al., 2011; Gregg et al., 2012). Single CTD profiles do not distinguish between these processes. As averaging is limited for the present CTD-data, we here also refer to CTD data (van Haren, 2015) and long-term moored high-resolution temperature data from the Puerto Rico Trench (van Haren and Gostiaux, 2016).

The observed turbulence values are comparable to those calculated from CTD down to 7,000 m for the Puerto Rico Trench (van Haren, 2015) and yearlong moored high-resolution temperature sensors near 6,000 m (van Haren and Gostiaux, 2016), the latter depth being just below the level of the seafloor surrounding the Puerto Rico Trench, see the light-blue value ranges in Figs 6b,c labeled 'PRT'. The moored PRT observations in particular showed tidal,

340   inertial and sub-inertial motions and their interaction affecting the turbulence 2,000 m above

341   the bottom of that trench. When the sub-inertial motions, which have a 20-100 days periodicity

342   and are probably evidencing a meandering boundary flow, advected warm water over the

343   trench, turbulence increased by a factor of 100, $\varepsilon \sim O(10^{-11}) \rightarrow O(10^{-9})$ m^{2}s^{-3}. The larger

344   turbulence was mainly caused by up to 200 m high convection tubes and associated secondary

345   shear instabilities along its edges (van Haren and Gostiaux, 2016). Considering the high

346   buoyancy Reynolds numbers $Re_b = \varepsilon/\nu N^2 = O(10^4)$ and these secondary shear instabilities, this

347   resembles estuarine mixing which was found (Holleman et al., 2016) with a large inertial

348   subrange to have high efficiency values $m_1 > 0.2$. This contrasts with lake and modeling results

349   that suggest mixing efficiency values half an order of magnitude lower than those of Osborn

350   (1980) at such Re_b (e.g., Bouffard and Boegman, 2013). For the present data, the

351   parameterization by Bouffard and Boegman (2013) for field data yields about one order of

352   magnitude lower K_z in the range of weak stratification between 6,000 and 8,500 m, see dashed

353   lines in Fig. 6c. Without knowledge on, e.g., the extent of the inertial subrange, the particular

354   type of turbulence and hence the mixing efficiency cannot be established. Comparing Holleman

355   et al. (2016) and Bouffard and Boegman (2013)'s results, the physics processes of turbulence

356   generation appear to be different in lakes and ocean. It would be good to test the condition of

357   marginal stability of destabilizing shear just balancing the stable stratification in lakes.

358   The range of turbulence values and vertical scales matches that of the present CTD

359   observations deeper into the Mariana Trench. The turbulence level is about three orders of

360   magnitude larger than molecular diffusion. Speculating, it may be sufficient to refresh its deep

361   waters. Future mooring measurements should refine the present deep trench estimates, in order

362   to establish the relevant time scales of possibly tidal and monthly periodicity.

363   Moored high-resolution measurements may also shed light on the turbulence generation of

364   the deep Mariana Trench waters. Here only suggestions can be given as to the possible source

365   of turbulence generation. The observed minimum stability was N = 1 cpd = 2.5f. The factor of

366   2.5 is identical to the one found in marginally stable waters of the Western Mediterranean

367   showing sharp transitions to fully homogeneous layers above and below (van Haren and Millot,

368   2006). Under such weakly stratified conditions, the semidiurnal internal tidal motions, with

369   dominant lunar component M_2 at 1.93 cpd, cannot propagate freely in zonal direction. (Diurnal

370   tidal currents, also freely propagating internal waves because $f \approx 0.4$ cpd, seem to be relatively

371   weak, Taira et al., 2004). This is because $M_2 > N$, so that semidiurnal internal tidal motions are

372   outside the commonly used internal wave band $[f, N]$ under the traditional approximation. Here,

373    $f_v = 2\Omega\sin\varphi \equiv f$ denotes the vertical Coriolis parameter of the earth rotational vector $\mathbf{\Omega}$ at latitude

374   φ, or the inertial frequency. Under the traditional approximation for generally moderate-strong

375   stratification, the effects of the horizontal Coriolis parameter $f_h = 2\Omega\cos\varphi$ are neglected. Its

376   effects are zero for zonal propagation, because $f_s = f_h\sin\alpha = 0$ for angle α with respect to the

377   zonal direction. However, for non-zonal wave propagation in weakly stratified waters, the

378   inertio-gravity wave band $[\sigma_{min}, \sigma_{max}] = 1/\sqrt{2}\cdot[(A-B)^{1/2}, (A+B)^{1/2}]$ where $A = N^2+f^2+f_s^2$ and $B =$

379    $(A^2-(2fN)^2)^{1/2}$, extends below f and above N (e.g., LeBlond and Mysak, 1978; version from

380   Gerkema et al., 2008). In meridional, north-south direction $\alpha = \pi/2$ the inertio-gravity wave

381   band thus has a maximum frequency range and internal tides at M_2 can always propagate freely

382   because $\sigma_{max} \geq 2\Omega$, while $[\sigma_{min}, \sigma_{max}] = [0.4f, 2.5N]$ for $N = 1$ cpd (Fig. 5d). Of all internal wave

383   motions only waves at f can propagate through the entire water column in all directions, even

384   where $N = 0$ and when they propagate in zonal direction $\alpha = 0$. Waves at other frequencies

385   depend on N, φ and α for free propagation or trapping.

386   Under marginally stable conditions, the stratification is supporting the maximum

387   destabilizing shear before breakdown. In the Mediterranean, such shear is induced by 150 m

388   large inertial waves as tides are very weak (Perkins, 1972). In the Challenger Deep, internal

389   tidal waves are more energetic, inferring from observed tidal amplitudes of 0.04 m s^{-1} (Taira et

390   al., 2004), than in the deep Mediterranean. Here, following numerical modelling, internal tides

391   are either remotely generated in Luzon Strait (Morozov, 1995) or locally at the Mariana Arc

392   (Jayne and St. Laurent, 2001). They likely contribute to shear, possibly after local nonlinear

393   interaction with waves of different frequencies, if propagating more meridionally than zonally

394   and possibly of equal amplitude considering the equivalent levels of (in)stability and

395   turbulence. As the Challenger Deep's long axis is nearly parallel to east-west, the main internal

396   tide propagation path is expected between the two walls north and south of the Deep. As

397   mentioned above, the behaviour of such large internal waves in the weakly stratified deep-sea

398   well away from surface and bottom boundaries is understood (LeBlond and Mysak, 1978),

399   whether the restoring force is dominantly reduced gravity for internal gravity waves for $N > 0$,

400   or momentum for gyroscopic waves for $N = 0$.

401   There are several remarks to be made about a comparison between the present CTD data

402   and those collected by Taira et al. (2005). First, the present data are statically stable on the

403   large, 100 m-vertical scale (Fig. 3c). This is not the case in Fig. 6 of Taira et al. (2005), where

404   a statically unstable layer of nearly 2,000 dbar extent can be observed between 7,500 and 9,500

405   dbar. This is very unlikely to happen even in the deep ocean. Considering the S-bend shape of

406   the profile in Taira et al. (2005), the apparent instability is likely due to a slight under-correction

407   of pressure in the conductivity measurement. This also explains similar behavior in their salinity

408   data. Second, Taira et al. (2005) also observe potential temperature to steadily increase with

409   pressure above 7,000 dbar, which confirms observations by Mantyla and Reid (1978). While it

410   may be true that salinity may compensate temperature to obtain neutral stratification, we do not

411   observe it in our data. In addition Mantyla and Reid (1978) did not observe a change between

412   their 7,000 and 11,000 dbar salinity data points. Mantyla and Reid's presentation of a nearly

413   4,000 m large unstable water column is therefore not very likely. Third, absolute values are

414   slightly different. In order to match values with those of Mantyla and Reid (1978), Taira et al.

415   (2005) subtracted 0.004°C and 0.011 psu from their potential temperature and salinity data,

416   respectively. It is noted that the difference may have been caused by the 40 km spacing between

417   the two sites, or it may be related to a variation in time between the respective measurements.

418   The near-bottom data of Mantyla and Reid (1978) were T = 2.462 °C and S = 34.699 psu at

419   10,890 m, providing $\theta = 1.0076$ °C using the gsw-software (IOC, SCOR, IAPSO, 2010). Forty

420   years later at 40 m higher, we observe $\theta = 1.0101$ °C and S = 34.705 psu. The 0.0025°C and

421   0.006 psu higher values compared to those of Mantyla and Reid (1978) are about half of the

422   differences subtracted by Taira et al. (2005).

15

423   The observed increase in density with depth below 8,500 m is dominantly governed by

424   increasing salinity. This confirms previous observations of Taira et al. (2005), even though their

425   data were probably slightly under-correcting conductivity for pressure effects. The shape of the

426   profiles suggests a relatively recent inflow of dense, salty and cool, water, possibly Circumpolar

427   Deep Water (e.g., Wu et al., 2015, using WOCE-data) that has not been homogenized yet. Such

428   profile shape of increasing density towards the bottom has been observed in other deep waters,

429   e.g. in the Maui Deep (Lukas et al., 2001) and in the Mediterranean soon after deep dense water

430   formation events (Schroeder et al., 2009). It remains to be quantified what length of time

431   'recent' actually represents, for which more observations distributed over time are needed.

432   As for the deepest point determination, our data are similar to measurements by Nakanishi

433   and Hashimoto (2011), who used a traditional but conservative approach governed by the

434   resolution of their echosounder that is several times less accurate (2°x2° beam opening angle)

435   than the one used on R/V Sonne. Gardner et al. (2014), on the other hand, used a statistical

436   approach in order to determine both the location and the depth of the deepest point, which

437   would be roughly 60 m deeper than our and other previous measurements (Nakanishi and

438   Hashimoto, 2011 and references therein). This discrepancy must be related to the use of the

439   correct sound velocity profile, which is crucial for accurate depth determinations in very deep

440   water. Gardner et al. (2014) used Deep Blue XBT's that only measure sound velocity for the

441   upper 760 m, which is clearly insufficient for extrapolation of the sound velocity profile down

442   to 12,000 m (as required by Simrad multibeam systems). In contrast, our CTD measurement

443   extended down to 8,000 m whereas Nakanishi and Hashimoto (2011) had CTD data down to

444   the bottom but less focused depth-soundings.

445   The incoming oceanic plate shows a number of typical bend faults (Ranero et al., 2003) that

446   in case of the Mariana Trench are parallel to the trench (Nakanishi and Hashimoto, 2011). Some

447   of the bend faults show up to 500 m-high fault scarps. It seems likely that the particular great

448   water depth in the Challenger Deep compared to other deep-sea trenches and other segments of

449   the Mariana Trench is augmented by normal faulting which adds to the already great water

450   depth resulting from the very steeply dipping subducting slab (Fryer et al., 2003).

451   Comparing the morphology, we find that the Mariana Trench and the Puerto Rico Trench

452   are quite different (Fig. 7). The Puerto Rico Trench is almost 100 km wider and less deep than

453   the Mariana Trench. It is also much smoother as it lacks the up to 500 m-high fault scarps,

454   which separate the seafloor depressions of the Mariana Trench into different compartments.

455   From this compartmentalization it could be speculated that internal waves, having wavelengths

456   O(100-1,000) m, may more easily reach the bottom waters in the Puerto Rico Trench than the

457   Mariana Trench, and that mixing should be stronger. However, the present limited turbulence

458   data show hardly any difference between the two trenches (Fig. 6). This suggests that seafloor

459   morphology on the 500 m-scale of our observations plays only a minor role for boundary layer

460   effects such as breaking of internal waves for turbulence in deep sea trenches. This may be a

461   challenging subject for future study.

462

463   **5. Conclusions**

464   The following can be concluded from the present shipborne CTD and multibeam-

465   bathymetry data in the weakly stratified waters below 5,000 m in the Challenger Deep, south

466   Mariana Trench. A high-precision CTD was successfully lowered to within 60 m from the

467   bottom, near the deepest point on Earth. The profile demonstrated that the weakest stratification

468   represented by N = 2.5f is found in the layer between 6,500 and 8,500 m and not near the

469   bottom. Purely homogeneous large-scale layers are not observed. After low-pass filtering to

470   remove surface wave effects, turbulence parameters are estimated. Turbulence estimates show

471   variations typical for open-ocean conditions, well away from surface and bottom boundary

472   layers. The limited comparison of turbulence data of the Mariana Trench with those of the

473   Puerto Rico Trench does not show significant differences possibly suggesting that trench

474   morphology does not play a crucial role for turbulence, even though turbulence is likely

475   generated by internal wave motions. Although stratification is relatively weak, it can support

476   near-inertial and tidal internal waves all the way to the bottom, provided internal tides propagate

477   well away from zonal direction. The observed turbulence is speculated to be sufficient to

478 support life in the Challenger Deep. A new position and depth for the deepest point in the

479 world's oceans have been defined.

480

481 **Acknowledgements**

486

487   **Table 1.** Historic physical oceanography water characteristics observations near the deepest

488   point on Earth in the Challenger Deep, Mariana Trench.

489

Observations	year	reference
Free-fall sampler (for S) w. reversing T	1976	Mantyla and Reid (1978)
Moored current observations	1987	Taira et al. (2004)
First SeaBird-911 CTD	1992	Taira et al. (2005)
SeaBird-911 CTD w. corr. p & turb. estimates	2016	present data

495

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Figure 1. Map of the western part of the Challenger Deep, Mariana Trench. Depths greater than 7,000 m are coloured (cf. Figure 7 for location). Grey indicates no or bad data. The red triangle indicates the CTD location and the blue and black lines represent the ship's track during rapid, 11 kn, and slow, < 2kn, multibeam bathymetry survey lines, respectively. Fault interpretations are entirely based on bathymetry interpretation assuming that the faults verge towards the steep side of the breaks of slope.

Figure 2. Zoom of Fig. 1 with non-white colours indicating depths below 10,900 m. Each square indicates the average depth within a 100x100m grid cell and all sounding above 10,900 invalidated. 'NH' indicates the depth and location of the deepest point as determined by Gardner et al. (2014). 'So' indicates the new depth and location of the deepest point, as determined from the R/V Sonne. The arrows indicate the ship's track in- (upper) and out- (lower) of the area.

Figure 3. Lower 6,000 dbar detail of the 2016 CTD profiles from the Challenger Deep for comparison with Figs 5 and 6 in Taira et al. (2005). Parameters are plotted as a function of negative pressure in decibars, as in Taira et al., (2005). In red and light-blue, the down- 'Dc1' and up-profiles 'Uc1' of cast 1, respectively. In blue and black, down 'Dc2 'and up 'Uc2' for cast 2, respectively. (a) Potential temperature. (b) Practical salinity, with the horizontal axis range matching that of (a) in terms of contributions to density variations. (c) Potential density anomaly, referenced to 8,000 dbar.

Figure 4. Nearly unsmoothed, ~3 degrees of freedom, spectra of 1-s averaged CTD data from 6,800-7,960 m range. The magenta spectrum is the <0.05 cps low-pass filtered version of the blue pressure-spectrum. Conductivity is from cast 2, Temperature from cast 1.

Figure 5. Full depth overview of 1-s averaged downcast CTD data. Cast 1 in red, cast 2 in blue. (a) Conservative Temperature. (b) Absolute Salinity. (c) Density anomaly referenced to

630   11,000 dbar. (d) Buoyancy frequency smoothed over 100 dbar vertical intervals. The thin

631   black dashed curves indicate the minimum (barely visible to the left of the f-line) and

632   maximum inertio-gravity wave bounds for meridional wave propagation (see text). The

633   inertial frequency f, the horizontal Coriolis parameter f_h (green) and the semidiurnal lunar

634   tidal frequency M_2 (dashed magenta) are indicated. The light-blue line indicates Puerto

635   Rico Trench 'PRT' mean N over the indicated vertical range (van Haren, 2015).

636

637   **Figure 6**. Lower 6,000 m of turbulence characteristics computed from <0.05 cps low-pass

638   filtered downcast data applying a threshold of 7×10^{-5} kg m^{-3}. In red CTD cast 1, in blue

639   cast 2. (a) Unordered, 'raw' profile of density anomaly referenced to 11,000 dbar. (b)

640   Logarithm of dissipation rate computed from the profiles in (a), averaged over 200 m

641   intervals. The light-blue lines indicate PRT-values: the vertical line a mean from CTD data

642   (van Haren, 2015), the horizontal dashed line the spread of 4-day, 200-m averages from

643   high-resolution moored temperature time series (van Haren and Gostiaux, 2016). (c) As

644   (b), but for eddy diffusivity. The dashed profiles indicate values using the parameterization

645   proposed for lake data by Bouffard and Boegman (2013). (d) Rms overturn displacements

646   following reordering of the profiles in (a).

647

648   **Figure 7**. Comparison of the bathymetry of the Puerto Rico Trench (top) and the Mariana

649   Trench (bottom) based on the 9.1 ETOPO-1 version of satellite altimetry-derived by Smith

650   and Sandwell (1997). The insets show the bathymetry along the white transects on the

651   maps. Stars indicate CTD positions. Black box in the lower panel indicates the location of

652   Figure 1.

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654

655

Deformation front

Puerto Rico Trench

van Haren and Gostiaux, 2016

Mariana Trench

this study