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



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## ABSTRACT

Reliable very deep shipborne SBE 911plus Conductivity Temperature Depth (CTD) data to within 60 m from the bottom and Kongsberg EM122  $0.5^{\circ} \times 1^{\circ}$  multibeam echosounder data are collected in the Challenger Deep, Mariana Trench. A new position and depth are given for the deepest point in the world's ocean. The data provide insight into the interplay between topography and internal waves in the ocean that lead to mixing of the lowermost water masses on Earth. Below 5000 m, the vertical density stratification is weak, with a minimum buoyancy frequency N =  $1.0 \pm 0.6$  cpd, cycles per day, between 6500 and 8500 m. In that depth range, the average turbulence is coarsely estimated from Thorpe-overturning scales, with limited statistics to be ten times higher than the mean values of dissipation rate  $\epsilon_T = 3 \pm 2 \times 10^{-11} \text{ m}^2 \text{ s}^{-3}$  and eddy diffusivity  $K_{zT} = 2 \pm 1.5$  $\times$  10<sup>-4</sup> m<sup>2</sup> s<sup>-1</sup> estimated for the depth range between 10,300 and 10,850 m, where N = 2.5 ± 0.6 cpd. Inertial and meridionally directed tidal inertio-gravity waves can propagate between the differently stratified layers. These waves are suggested to be responsible for the observed turbulence. The turbulence values are similar to those recently estimated from CTD and moored observations in the Puerto Rico Trench. Yet, in contrast to the Puerto Rico Trench, seafloor morphology in the Mariana Trench shows up to 500 m-high fault scarps on the incoming tectonic plate and a very narrow trench, suggesting that seafloor topography does not play a crucial role for mixing.

## 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 6000 m to date. As an indicator for upper trench turbulence, recent yearlong high-resolution temperature measurements from around about 6000 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  $0.04 \text{ m s}^{-1}$  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 semidiurnal currents may be related to internal tides, whether propagating

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#### Table 1

Historic physical oceanography water characteristics observations near the deepest point on Earth in the Challenger Deep, Mariana Trench.

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

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

Water characteristics are also barely known near the bottom of the Challenger Deep (Table 1). This is mainly because few oceanographic research vessels are equipped with cables that are more than 11 km long. As an alternative and following discrete inverse thermometer readings from R/V Vityaz in the late 1950's, a small free-falling water-sampling device equipped with reversing thermometers was dropped to the bottom of the Challenger Deep in 1976 (Mantyla and Reid, 1978). These data are also used by Taira et al. (2005) as a reference for the first deep CTD-cast attached to a custom-made titanium wire in the Challenger Deep, down to 10,877 m. They used a SeaBird Electronics SBE-911 CTD at a station that was 40 km east of the site where Mantyla and Reid (1978) deployed their water sampler. Manned (Gallo et al., 2015) and un-manned (Nunoura et al., 2015) submarines carried CTDs (an SBE-49 in the latter case), but these data have not been analyzed and published for detailed water characteristics.

In this paper, we report on new high-resolution SBE-911 CTD casts into the Challenger Deep, reaching a depth of 10,851 m in 10,907 m water depth at a location about 2 km east of Mantyla and Reid's position. The CTD data deliver T, Salinity S and stratification information and thus information on the internal wave band below 5000 m of the surrounding ocean floor and near the maximum capability of standard oceanographic instrumentation. A second objective of this experiment is to gain a first impression of the internal wave-induced turbulence variation with depth down to near the bottom of the Challenger Deep. As a third objective, we use the simultaneously acquired high-precision multibeam echosounder data that were calibrated with the local CTD data to validate the most recent estimates of Challenger Deep's greatest depth (Gardner et al., 2014).

#### 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  $\pm$  12 m water depth in November 2016. Water depth was measured using a Simrad EM122 multibeam bathymetry system with a  $0.5^{\circ} \times 1^{\circ}$  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 96  $\times$  192 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 100  $\times$  100 m grid. The standard deviation of sounding depths within one cell amounts to  $a \pm 12 \text{ m}$ 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  $0.7 \text{ m s}^{-1}$  for downcast and  $1.0 \text{ m s}^{-1}$  for upcast. Weather and sea conditions were good, with 4 m s<sup>-1</sup> wind speeds and maximum 2 m wave height. R/V Sonne held its position to within 10 m precision during the 14 h 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 7000 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 7965 m, close to the maximum depth without damaging the installed pressure sensor. After being brought back on deck, the oil-filled opening leading to the pressure capillary tube was replaced by a NIOZ custommade titanium plug. Thus, the second CTD cast, deployed 6 h after the first one, did not include a pressure sensor and we had to use the length of the cable paid out as guidance for depth. This required higher effort in post-processing and caution upon approaching the sea floor. With  $a \pm 25$  m uncertainty in water depth estimates known at the time of CTD cast and a  $\pm$  5 m uncertainty in cable length read-out/depth relationship, we decided to stop the CTD at 10,851 m while the multibeam echosounder showed 10,905 m. In retrospect, after re-analyzing the multibeam data, the CTD was stopped 56  $\pm$  12 m from the bottom.

Additional post-processing of the CTD data from the second cast involved matching times of the winch cable length with the CTD data. As the former was available at a rate of 1 Hz and the latter at a rate of 24 Hz, the CTD data were first averaged to 1 s data. A linear relationship exists between cable length and depth for the first CTD cast. This relationship was subsequently used to obtain depth and pressure for the second CTD cast. Unfortunately, this relationship does not account for the fact that the CTD's pressure sensor records the ship's motions by waves (e.g., Trump, 1983), while cable length does not register wave motions. We attempted to correct this by establishing a second relationship between the first derivatives, gradients, of depth and those of the cable tension reading relative to the local tension. Although the two roughly show a linear relationship, small-scale mismatches in phase and amplitude of waves are large, and there was no overall improvement compared to the linear relationship between cable length reading and depth. This implies that, using only the latter data as a proxy for depth, the salinity data, like those of density, appear considerably noisier during the second cast with respect to the first cast, while temperature data are little affected. This was remedied by applying a sharp double elliptic, phase-preserving low-pass filter (Parks and Burrus, 1987) with 0.05 cps, cycles per second, cut-off to all data, as proposed in van Haren (2015) to remove ship motion effects by surface waves.

As SeaBird's conductivity sensor already applies calibration coefficients including factors for temperature and pressure corrections, an extra step in post-processing of the second cast was necessary. The acquisition computer applies calibration coefficients to the instrument's sensor frequency data and it is necessary to re-apply them to the conductivity sensor frequency data using the cable length proxy for pressure. Not applying these corrections leads to underestimation of the salinity and density profiles as a function of pressure. This would result in an apparently unstable water column under weakly stratified conditions.

After the additional post-processing described above, we processed the CTD data using the standard procedures incorporated in the SBEsoftware, including corrections for cell thermal mass using the parameter setting of Mensah et al. (2009) and sensor time-alignment. We calculate the potential temperature  $\theta$ , practical salinity S and density anomalies  $\sigma_8$  referenced to 8000 dbar in order to allow comparison with the results of Taira et al. (2005). All other analyses were performed using Conservative (~ potential) Temperature ( $\Theta$ ), absolute salinity SA and density anomalies  $\sigma_{11}$  referenced to 11,000 dbar using the gsw-software described in (IOC, SCOR, IAPSO, 2010).



**Fig. 1.** Map of the western part of the Challenger Deep, Mariana Trench. Depths greater than 7000 m are coloured (cf. Fig. 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, < 2 kn, 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. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Estimates of turbulence dissipation rate  $\varepsilon_T = c_1^2 d^2 N^3$  and vertical eddy diffusivity  $K_{zT} = m_1 c_1^2 d^2 N$  are made from the downcast CTD data using the method of reordering potentially unstable vertical density profiles in statically stable ones, as proposed by Thorpe (1977). Here, d denotes the displacements between unordered (measured) and reordered profiles. N denotes the buoyancy frequency computed from the reordered profiles. Rms values are not determined over individual overturns, as in Dillon (1982), but over 200 m vertical intervals that just exceed the largest overturn intervals. This avoids the complex distinction of smaller overturns in larger ones and allows the use of a single averaging length scale. We use standard constant values of  $c_1 =$ 0.8 for the Ozmidov/overturn scale factor (Dillon, 1982) and  $m_1 = 0.2$ for the mixing efficiency (Osborn, 1980; Oakey, 1982). This is the most commonly used parameterization for oceanographic data. Recently, it is challenged for young convective turbulence mostly from numerical modeling (e.g., Scotti, 2015). However, the above parametrization still seems valid for most deep-ocean applications provided some suitable averaging is performed over all relevant overturning scales of the unique mix of shear- and convective overturning in large Reynolds number flow conditions (e.g., Ferron et al., 1998; Gregg et al., 2012; Mater et al., 2015). As a criterion for determining overturns from the surface wave low-pass filtered density data, we only used those data of which the absolute value of difference with the local reordered value exceed a threshold of  $7\times10^{-5}$  kg m $^{-3}$ , which corresponds to applying a threshold of 1.4  $\times~10^{-3}$  kg m $^{-3}$  to raw data variations (e.g., Stansfield et al., 2001; Gargett and Garner, 2008).

#### 3. Observations

#### 3.1. Multibeam bathymetry data and the deepest point

For water depths greater than 8 km, the multibeam swath width is approximately 30 km (Fig. 1; for reference: the size of the figure equals  $45 \times 37$  km). The first multibeam bathymetry survey was made with a standard sound velocity profile and high velocity (while sailing the orange trajectory in Fig. 2, partially outside the window on the south side). After passing directly over the then deepest point known (Gardner et al., 2014), indicated by 'NH' in Fig. 2, a quick analysis was made of the data searching for the deepest point. At that analyzed point, the deep CTD casts were made (Figs. 1 and 2). A second survey (blue-purple trajectory in Fig. 2) was made with the sound velocity profile from the first CTD cast, thereby properly correcting for local conditions. Thus, a total of three passes were made over the western portion of the Challenger Deep, to within 1.5 km directly over the deepest point known. This provides a rather precise estimate of the bottom topography of the middle portion of Figs. 1 and 2, as the center of the sweep is the most accurate part of a multibeam sweep.

Re-analysis of the multibeam data on a  $100 \times 100$  m grid showed that the CTD casts were made about 1.6 km west of the deepest point. Our data show that the deepest point is at  $10,925 \pm 12$  m at  $11^{\circ}$  19.945'N, 142° 12.123'E. The horizontal position of the grid point has an uncertainty of  $\pm$  50 to  $\pm$  100 m, depending on along-track or across-track direction. Nevertheless, the position is significantly different from the one established as the deepest point by Gardner et al. (2014). Their



**Fig. 2.** Zoom of Fig. 1 with non-white colours indicating depths below 10,900 m. Each square indicates the average depth within a  $100 \times 100$  m 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. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

depth is significantly greater, by nearly 60 m.

The multibeam bathymetry data show deepening from about 2000 m water depth on the Mariana Arc to the maximum depth of 10,925 m before shoaling on the incoming plate to about 5000 m water depth for the abyssal plain not affected by subduction (Fig. 1). Steep slopes are observed on the accretionary prism for the last 10 km before reaching the deformation front, with water depth increasing from 8300 m to 10,925 m. Overall the Mariana Trench is very narrow: 120 km when measured at 5000 m water depth in the study area. The data reveal distinctly different morphologies of the incoming and overriding plates (Fig. 1). While the overriding plate shows a chaotic morphology with 10 km-sized lobes in irregular patterns, the incoming plate is generally dipping smoothly except for very high fault scarps in excess of 500 m. The faults strike parallel to the trench azimuth and some extend for several tens of kilometers. Generally, the fault scarps dip away from the subduction zone, although two fault scarps identified in the bathymetry data verge towards the subduction zone. The faults dip at 45° angles and the typical separation between major faults is between 3 and 3.5 km. The deformation front separating the incoming and overriding plate generally strikes parallel to the trench at N80. In the Challenger Deep a fault appears to cross-cut the deformation front from the incoming plate and albeit subdued continues in a N70 direction into the accretionary prism.

## 3.2. Comparison of CTD data

All data at pressures exceeding 5000 dbar, the approximate level of the surrounding ocean floor to the southeast, from the two CTD casts are presented in Fig. 3. They show good agreement, with downcasts (red and blue) and upcasts (light-blue and black) better aligned between themselves than down- with upcasts (red/light-blue, blue/black) from a particular CTD cast. This is due to dynamic pressure by the water flow during the lowering and hoisting of the CTD frame that affects the pump speed in the TC-duct (van Haren, 2015; Uchida et al., 2015). This variation in water flow speed passing the T- and C-sensors artificially modifies the temperature measured due to variations in skin friction. Temperature varies by about  $3 \times 10^{-4}$  °C per m s<sup>-1</sup> flow speed change. For a horizontally mounted system, temperature is directly related to

pressure variations with time, i.e. the lowering speed (van Haren and Laan, 2016). This is confirmed here, with the downcast 2 (blue) and downcast 1 (red) profiles being consistently warmer than the upcast 2 (black) and upcast 1 (magenta) profiles, respectively (Fig. 3a). The observed T-difference associates with the sign-change in velocity between up- and downcast and with the reduction in hoisting/lowering speeds towards the bottom of the profiles. This artificial temperature variation is transferred to conductivity measurements, being inversely T-dependent besides being directly p-dependent. Artificial variations are consequently visible in salinity (Fig. 3b) and density anomaly data (Fig. 3c) as consistently lower values for the down- compared to the upcasts. On top of these large-scale, slow variations between down- and upcast data, faster variations are due to surface waves generating ship's motions, i.e. the apparent noise in the data. Whilst these are small for cast 1, they are 2-5 times larger for cast 2, as we cannot correct for short-term pressure variations when using the cable length read-out data as a proxy for pressure. For the subsequent analyses this is not a problem, as surface wave effects will be filtered out (van Haren, 2015), see Section 3.3.

The present data are statically stable on the large, 200 m-vertical scale (Fig. 3c; better inferable from temperature data Fig. 3a). Potential temperature steadily decreases with pressure (Fig. 3a). Salinity steadily increases with pressure (Fig. 3b). As a result, with both salinity and temperature contributing positively, density increases towards the bottom, albeit mainly due to salinity variations in the lower 2500 dbar (Fig. 3c). In that range, stratification increases towards the bottom as well, also mainly due to salinity variations. The weakest stratification exists higher -up, between about 6500 and 8500 dbar.

## 3.3. Turbulence parameter estimates

In the range of 5000 to 6000 dbar where stratification is relatively strong, the observations in Fig. 3 show vertical excursions due to internal wave motions by the differences between, especially, temperature profiles. At a particular temperature, the change in pressure exceeds 100 dbar, a large excursion but typical for internal wave motions in the deep ocean. At greater depths however, these data are not suitable for internal wave 'observations' because 'short-'scale surface wave effects dominate the variations. The artificial surface wave motions also need to be removed in order to estimate turbulence parameters. A lowpass filter with a 0.05 cps cut-off allows separating the well-defined surface wave peak due to ship motion coupled down the wire centered around 0.1 cps in all primary parameters (Fig. 4). Considering the  $0.7\ m\ s^{-1}$  lowering speed of the CTD frame, 15 m and larger structures like overturns are thus retained, which suffices to resolve the most energetic turbulence overturning scales estimated to be 100 m in weakly stratified waters.

The filtered data show a smooth large-scale variation with depth, except for some saline water intrusion in the upper 300 m with a maximum salinity around 125 m (Fig. 5). Except for this anomaly, temperature continuously decreases (Fig. 5a), while salinity (Fig. 5b) and density (Fig. 5c) continuously increase with depth, when smoothed over scales > 200 m. In terms of the buoyancy frequency, the stratification is strong in the upper 25-125 m of the water column, and steadily decreases with depth down to halfway the water column. Below z = -5000 m, N < 2.5 cycles per day (cpd), and reaches a minimum of N = 1  $\pm$  0.6 cpd = 2.5  $\pm$  1.5 f between 6500 and 8500 m (Fig. 5d). The standard deviation designates the same error as found in very weakly stratified waters in the deep Mediterranean (van Haren and Millot, 2006). These N-values are very close to tidal frequencies, but the inertio-gravity wave band includes the tidal frequencies so that their motions can represent freely propagating waves (further discussion in Section 4).

The turbulence estimates computed from the two CTD downcasts using Thorpe displacements after reordering yield comparable results (Fig. 6). Vertical averages over the range between 5000 and 7750 m are



Fig. 4. Nearly unsmoothed,  $\sim$  3 degrees of freedom, spectra of 1-s averaged CTD data from 6,800–7960 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.

σ (cps)

within a factor of two: For the first CTD-cast we find  $\epsilon_T = 1.4 \pm 1 \times 10^{-10} \, m^2 \, s^{-3}$  and  $K_{zT} = 1 \pm 0.7 \times 10^{-3} \, m^2 \, s^{-1}$  while for the second cast  $\epsilon_T = 2.3 \pm 1.5 \times 10^{-10} \, m^2 \, s^{-3}$  and  $K_{zT} = 1.5 \pm 1 \times 10^{-3} \, m^2 \, s^{-1}$ . Vertically, the plotted 200 m averaged energy dissipation rate (Fig. 6b) and eddy diffusivity values (Fig. 6c) vary over nearly three orders of magnitude. This is common for ocean turbulence, although in stronger turbulence/stratification four orders of magnitude variation occur (e.g., Gregg, 1989). Averaged over ranges of 500–1000 m, the variation is a factor of about 10. Highest values and largest rms overturn displacements (Fig. 6d) are found in the weakest stratification, between 6500 and 8500 m. For the near-bottom range [10,300–10,850] m averages of  $\epsilon_T = 3 \pm 2 \times 10^{-11} \, m^2 \, s^{-3}$  and  $K_{zT} = 2 \pm 1.5 \times 10^{-4} \, m^2 \, s^{-1}$  are estimated. These values are rather uncertain as they are determined from only 3 or 4 overturns.

### 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 **Fig. 3.** Lower 6000 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 8000 dbar. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

CTD profiles (down to 8000 m) obtained 6 h 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 c1- and m1-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 7000 m for the Puerto Rico Trench (van Haren, 2015) and yearlong moored high-resolution temperature sensors near 6000 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 Fig. 6b,c labeled 'PRT'. The moored PRT observations in particular showed tidal, inertial and sub-inertial motions and their interaction affecting the turbulence 2000 m above the bottom of that trench. When the sub-inertial motions, which have a 20-100 days periodicity and are probably evidencing a meandering boundary flow, advected warm water over the trench, turbulence increased by a factor of 100,  $\varepsilon \sim O(10^{-11}) \rightarrow O(10^{-9}) \text{ m}^2 \text{s}^{-3}$ . The larger turbulence was mainly caused by up to 200 m high convection tubes and associated secondary shear instabilities along its edges (van Haren and Gostiaux, 2016). Considering the high buoyancy Reynolds numbers  $Re_b = \epsilon/\nu N^2 = O(10^4)$  and these secondary shear instabilities, this resembles estuarine mixing which was found (Holleman et al., 2016) with a large inertial subrange to have high efficiency values  $m_1 > 0.2$ . This contrasts with lake and modeling results that suggest mixing efficiency values half an order of magnitude lower than those of Osborn

-5000

-6000

-7000

-9000

-10000

-11000 73.825

а

73.83

 $\sigma_{11} \, (\text{kg m}^{-3})$ 

E -8000



Fig. 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 11,000 dbar. (d) Buoyancy frequency smoothed over 100 dbar vertical intervals. The thin black dashed curves indicate the minimum (barely visible to the left of the f-line) and maximum inertio-gravity wave bounds for meridional wave propagation (see text). The inertial frequency f, the horizontal Coriolis parameter fh (green) and the semidiurnal lunar tidal frequency M<sub>2</sub> (dashed magenta) are indicated. The light-blue line indicates Puerto Rico Trench 'PRT' mean N over the indicated vertical range (van Haren, 2015). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

(1980) at such Re<sub>b</sub> (e.g., Bouffard and Boegman, 2013). For the present data, the parameterization by Bouffard and Boegman (2013) for field data yields about one order of magnitude lower  $K_z$  in the range of weak stratification between 6000 and 8500 m, see dashed lines in Fig. 6c. Without knowledge on, e.g., the extent of the inertial subrange, the particular type of turbulence and hence the mixing efficiency cannot be established. Comparing Holleman et al. (2016) and Bouffard and Boegman (2013)'s results, the physics processes of turbulence generation appear to be different in lakes and ocean. It would be good to test the condition of marginal stability of destabilizing shear just balancing the stable stratification in lakes.

The range of turbulence values and vertical scales matches that of the present CTD observations deeper into the Mariana Trench. The

Dc1

Dc2

73.835

-11

-10

 $\log(\epsilon_{\tau})$  (m<sup>2</sup> s<sup>-3</sup>)

b

-9

-4

-3

log(K<sub>2</sub>) (m<sup>2</sup> s<sup>-1</sup>)

turbulence level is about three orders of magnitude larger than molecular diffusion. Speculating, it may be sufficient to refresh its deep waters. Future mooring measurements should refine the present deep trench estimates, in order to establish the relevant time scales of possibly tidal and monthly periodicity.

Moored high-resolution measurements may also shed light on the turbulence generation of the deep Mariana Trench waters. Here only suggestions can be given as to the possible source of turbulence generation. The observed minimum stability was N = 1 cpd = 2.5 f. The factor of 2.5 is identical to the one found in marginally stable waters of the Western Mediterranean showing sharp transitions to fully homogeneous layers above and below (van Haren and Millot, 2006). Under such weakly stratified conditions, the semidiurnal internal tidal



С

-2

0

20

L<sub>T</sub> (m)

d

60

40

motions, with dominant lunar component M2 at 1.93 cpd, cannot propagate freely in zonal direction. (Diurnal tidal currents, also freely propagating internal waves because  $f \approx 0.4$  cpd, seem to be relatively weak, Taira et al., 2004). This is because  $M_2 > N$ , so that semidiurnal internal tidal motions are outside the commonly used internal wave band [f, N] under the traditional approximation. Here,  $f_v = 2\Omega \sin \phi = f$ denotes the vertical Coriolis parameter of the earth rotational vector  $\mathbf{\Omega}$ at latitude  $\varphi$ , or the inertial frequency. Under the traditional approximation for generally moderate-strong stratification, the effects of the horizontal Coriolis parameter  $f_h = 2\Omega cos\phi$  are neglected. Its effects are zero for zonal propagation, because  $f_s = f_b \sin \alpha = 0$  for angle  $\alpha$  with respect to the zonal direction. However, for non-zonal wave propagation in weakly stratified waters, the inertio-gravity wave band  $[\sigma_{min}]$  $\sigma_{\text{max}}$ ] = 1/ $\sqrt{2}$ ·[(A-B)<sup>1/2</sup>, (A + B)<sup>1/2</sup>] where A = N<sup>2</sup> + f<sup>2</sup> + f<sup>2</sup><sub>s</sub> and B =  $(A^2-(2fN)^2)^{1/2}$ , extends below f and above N (e.g., LeBlond and Mysak, 1978; version from Gerkema et al., 2008). In meridional, north-south direction  $\alpha = \pi/2$  the inertio-gravity wave band thus has a maximum frequency range and internal tides at M<sub>2</sub> can always propagate freely because  $\sigma_{max} \ge 2 \Omega$ , while  $[\sigma_{min}, \sigma_{max}] = [0.4 \text{ f}, 2.5 \text{ N}]$  for N = 1 cpd (Fig. 5d). Of all internal wave motions only waves at f can propagate through the entire water column in all directions, even where N = 0and when they propagate in zonal direction  $\alpha = 0$ . Waves at other frequencies depend on N,  $\phi$  and  $\alpha$  for free propagation or trapping.

Under marginally stable conditions, the stratification is supporting the maximum destabilizing shear before breakdown. In the Mediterranean, such shear is induced by 150 m large inertial waves as tides are very weak (Perkins, 1972). In the Challenger Deep, internal tidal waves are more energetic, inferring from observed tidal amplitudes of 0.04 m s<sup>-1</sup> (Taira et al., 2004), than in the deep Mediterranean. Here, following numerical modeling, internal tides are either remotely generated in Luzon Strait (Morozov, 1995) or locally at the Mariana Arc (Jayne and St. Laurent, 2001). They likely contribute to shear, possibly after local nonlinear interaction with waves of different frequencies, if propagating more meridionally than zonally and possibly of equal amplitude considering the equivalent levels of (in)stability and turbulence. As the Challenger Deep's long axis is nearly parallel to eastwest, the main internal tide propagation path is expected between the two walls north and south of the Deep. As mentioned above, the behavior of such large internal waves in the weakly stratified deep-sea well away from surface and bottom boundaries is understood (LeBlond and Mysak, 1978), whether the restoring force is dominantly reduced gravity for internal gravity waves for N > 0, or momentum for gyroscopic waves for N = 0.

There are several remarks to be made about a comparison between the present CTD data and those collected by Taira et al. (2005). First, the present data are statically stable on the large, 100 m-vertical scale (Fig. 3c). This is not the case in Fig. 6 of Taira et al. (2005), where a statically unstable layer of nearly 2000 dbar extent can be observed between 7500 and 9500 dbar. This is very unlikely to happen even in the deep ocean. Considering the S-bend shape of the profile in Taira et al. (2005), the apparent instability is likely due to a slight undercorrection of pressure in the conductivity measurement. This also explains similar behavior in their salinity data. Second, Taira et al. (2005) also observe potential temperature to steadily increase with pressure above 7000 dbar, which confirms observations by Mantyla and Reid (1978). While it may be true that salinity may compensate temperature to obtain neutral stratification, we do not observe it in our data. In addition Mantyla and Reid (1978) did not observe a change between their 7000 and 11,000 dbar salinity data points. Mantyla and Reid's presentation of a nearly 4000 m large unstable water column is therefore not very likely. Third, absolute values are slightly different. In order to match values with those of Mantyla and Reid (1978), Taira et al. (2005) subtracted 0.004 °C and 0.011 psu from their potential temperature and salinity data, respectively. It is noted that the difference may have been caused by the 40 km spacing between the two sites, or it may be related to a variation in time between the respective measurements. The near-bottom data of Mantyla and Reid (1978) were T = 2.462 °C and S = 34.699 psu at 10,890 m, providing  $\theta$  = 1.0076 °C using the gsw-software (IOC, SCOR, IAPSO, 2010). Forty years later at 40 m higher, we observe  $\theta$  = 1.0101 °C and S = 34.705 psu. The 0.0025 °C and 0.006 psu higher values compared to those of Mantyla and Reid (1978) are about half of the differences subtracted by Taira et al. (2005).

The observed increase in density with depth below 8500 m is dominantly governed by increasing salinity. This confirms previous observations of Taira et al. (2005), even though their data were probably slightly under-correcting conductivity for pressure effects. The shape of the profiles suggests a relatively recent inflow of dense, salty and cool, water, possibly Circumpolar Deep Water (e.g., Wu et al., 2015, using WOCE-data) that has not been homogenized yet. Such profile shape of increasing density towards the bottom has been observed in other deep waters, e.g. in the Maui Deep (Lukas et al., 2001) and in the Mediterranean soon after deep dense water formation events (Schroeder et al., 2009). It remains to be quantified what length of time 'recent' actually represents, for which more observations distributed over time are needed.

As for the deepest point determination, our data are similar to measurements by Nakanishi and Hashimoto (2011), who used a traditional but conservative approach governed by the resolution of their echosounder that is several times less accurate ( $2^{\circ} \times 2^{\circ}$  beam opening angle) than the one used on R/V Sonne. Gardner et al. (2014), on the other hand, used a statistical approach in order to determine both the location and the depth of the deepest point, which would be roughly 60 m deeper than our and other previous measurements (Nakanishi and Hashimoto, 2011 and references therein). This discrepancy must be related to the use of the correct sound velocity profile, which is crucial for accurate depth determinations in very deep water. Gardner et al. (2014) used Deep Blue XBT's that only measure sound velocity for the upper 760 m, which is clearly insufficient for extrapolation of the sound velocity profile down to 12,000 m (as required by Simrad multibeam systems). In contrast, our CTD measurement extended down to 8000 m whereas Nakanishi and Hashimoto (2011) had CTD data down to the bottom but less focused depth-soundings.

The incoming oceanic plate shows a number of typical bend faults (Ranero et al., 2003) that in case of the Mariana Trench are parallel to the trench (Nakanishi and Hashimoto, 2011). Some of the bend faults show up to 500 m-high fault scarps. It seems likely that the particular great water depth in the Challenger Deep compared to other deep-sea trenches and other segments of the Mariana Trench is augmented by normal faulting which adds to the already great water depth resulting from the very steeply dipping subducting slab (Fryer et al., 2003).

Comparing the morphology, we find that the Mariana Trench and the Puerto Rico Trench are quite different (Fig. 7). The Puerto Rico Trench is almost 100 km wider and less deep than the Mariana Trench. It is also much smoother as it lacks the up to 500 m-high fault scarps, which separate the seafloor depressions of the Mariana Trench into different compartments. From this compartmentalization it could be speculated that internal waves, having wavelengths O(100–1000) m, may more easily reach the bottom waters in the Puerto Rico Trench than the Mariana Trench, and that mixing should be stronger. However, the present limited turbulence data show hardly any difference between the two trenches (Fig. 6). This suggests that seafloor morphology on the 500 m-scale of our observations plays only a minor role for boundary layer effects such as breaking of internal waves for turbulence in deep sea trenches. This may be a challenging subject for future study.

#### 5. Conclusions

The following can be concluded from the present shipborne CTD and multibeam-bathymetry data in the weakly stratified waters below 5000 m in the Challenger Deep, south Mariana Trench. A high-precision CTD was successfully lowered to within 60 m from the bottom, near the



Fig. 7. Comparison of the bathymetry of the Puerto Rico Trench (top) and the Mariana Trench (bottom) based on the 9.1 ETOPO-1 version of satellite altimetry-derived by Smith and Sandwell (1997). The insets show the bathymetry along the white transects on the maps. Stars indicate CTD positions. Black box in the lower panel indicates the location of Fig. 1.

deepest point on Earth. The profile demonstrated that the weakest stratification represented by N = 2.5 f is found in the layer between 6500 and 8500 m and not near the bottom. Purely homogeneous largescale layers are not observed. After low-pass filtering to remove surface wave effects, turbulence parameters are estimated. Turbulence estimates show variations typical for open-ocean conditions, well away from surface and bottom boundary layers. The limited comparison of turbulence data of the Mariana Trench with those of the Puerto Rico Trench does not show significant differences possibly suggesting that trench morphology does not play a crucial role for turbulence, even though turbulence is likely generated by internal wave motions. Although stratification is relatively weak, it can support near-inertial and tidal internal waves all the way to the bottom, provided internal tides propagate well away from zonal direction. The observed turbulence is speculated to be sufficient to support life in the Challenger Deep. A new position and depth for the deepest point in the world's oceans have been defined.

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