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**Stratified-turbulence observations dominated
by slantwise downward convective warm-
water periods in the deep Mediterranean**

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27 **Abstract.** A three-dimensional mooring array holding nearly 3000 high-resolution temperature sensors
28 in 2500-m deep Mediterranean-Sea waters is used for a yearlong study on different sources of turbulent
29 waterflows, which are vital for life. Although temperature differences are found never larger than
30 0.01°C, daily, weekly, and seasonal variations are observed. With a delay of about a week, the deep-sea
31 stratified turbulence tracks atmospheric disturbances, which are found 35% more energetic in winter
32 than in summer. About half the time, relatively warm stratified waters are moved to near the seafloor
33 from 100's of meters higher levels. Combined internal-wave and sub-mesoscale eddy-induced motions
34 lead to slantwise downward convective warm-water periods that are half an order of magnitude more
35 turbulent than those induced via general geothermal heating from below, and about one order of
36 magnitude more turbulent than those from open-ocean processes. The analysis estimates that eddy-
37 induced stratified turbulence is likely more important for deep-sea life than rare, not observed, deep
38 dense-water formation at the abyssal-plain mooring site.

39

40 **1 Introduction**

41 Irreversible, energy-consuming turbulence is indispensable for life everywhere on earth, including in
42 the deep sea. Most ocean turbulence is generated near its boundaries, with an important input via the
43 breaking of internal waves at steeply sloping underwater topography (Eriksen, 1982; Thorpe, 1987).
44 Observational details of turbulence are still scarce from the abyssal deep sea that is commonly
45 considered to be ‘quiescent and stagnant’. One of the key aspects in the breaking of internal waves is
46 the ‘warming phase’, when, e.g., an internal wave moves downslope of topography and
47 characteristically (re-)stratifies waters near the seafloor with near-homogeneous waters due to
48 convection turbulence above (van Haren and Gostiaux, 2012). Restratification renders efficient turbulent
49 mixing during the following upslope phase. Indirectly topography-connected are, perhaps less known,
50 slantwise-moving, warm turbulent waters (van Haren and Dijkstra, 2021). This turbulence generation is
51 possibly related with near-inertial waves and (sub-)mesoscale eddies in basins like the Mediterranean
52 Sea where tides and stratified conditions are both weak. As will be demonstrated in this paper, these
53 warm waters have the potential to generate larger turbulent mixing than in the open-ocean interior.

54 Under weakly stratified conditions, the potential of deep-sea turbulence generation via downward or
55 slantwise moving waters should be compared with local general geothermal heating through the seafloor
56 (e.g., Pasquale et al., 1996), and with deep dense-water formation from the surface (e.g., Marshall and
57 Schott, 1999). Sparse shipborne microstructure profiling has provided estimates of turbulence
58 contributions from different sources across the Northwestern Mediterranean in which geothermal
59 heating is found more important than internal wave breaking (Ferron et al., 2017). However, details of
60 relevant processes are lacking and require high-resolution moored observations over prolonged periods
61 of time.

62 In a stably-stratified environment like the sun-heated ocean, downward motions of warm water
63 reaching the deep seafloor seem impossible and counter-intuitive in relation with irreversible convection
64 turbulence involving motion in three spatial dimensions ‘3D’. Natural, body-forced buoyancy-driven
65 convection (e.g., Dalziel et al., 2008; Ng et al., 2016) applies to denser waters moving down and less
66 dense waters moving up in narrow plumes, i.e. in terms of temperature variations: warmer waters
67 moving up and cooler waters moving down. In the ocean, such buoyancy-driven convection turbulence

68 occurs regularly in the upper $O(10)$ m near the surface during nighttime (e.g., Brainerd and Gregg, 1995)
69 and possibly lower 100 m above the seafloor due to geothermal heating, depending on the local
70 stratification. It can also occur as deep dense-water formation after specific preconditioning of
71 stratification reduction near the surface in localized areas like polar seas and the Mediterranean during
72 brief irregular, rare periods (Marshall and Schott, 1999). An exemption can occur when salinity
73 dominates over temperature in contributing to density variations: if downward moving warm waters are
74 sufficiently saltier than their environment, cooler and fresher waters may move up.

75 A reversible, also 3D, process occurs when internal-wave motions affect the stratified environment
76 (e.g., LeBlond and Mysak, 1978). Such motions may displace relatively warm waters downward during
77 a particular wave-phase, and cooler waters up. However, such displacements will not overturn and
78 vertically mix the different water masses.

79 Nevertheless, an apparent counter-intuitive combination of irreversible and reversible processes was
80 occasionally observed above a slope of Great Meteor Seamount where, in the weakly stratified waters
81 underneath internal waves, convection turbulence was observed (van Haren, 2015). Whilst in those
82 North-Atlantic Ocean waters the effects of salinity could not be excluded, confirmation was found in
83 similar observations from fresh-water alpine Lake Garda (van Haren and Dijkstra, 2021). As these
84 observations showed similarity with the warming phase of a nonlinear wave breaking above a sloping
85 seafloor (e.g., van Haren and Gostiaux, 2012), it was suggested that the convection underneath internal
86 waves was either generated via shear moving convection tubes slantwise, or via wave-accelerations
87 overcoming vertical density differences in internal-forcing overcoming reduced gravity, instead of
88 body-forcing overcoming gravity as in natural convection.

89 Similar to the effect of large-scale shear, planetary slantwise convection in the direction of the Earth's
90 rotational vector may be brought about by the horizontal Coriolis parameter f_h (Marshall and Schott,
91 1999). The convection can be induced via resonantly-forced standing inertial waves under homogeneous
92 conditions (McEwan, 1973). At mid-latitudes, apparently-stable stratification having buoyancy
93 frequencies of $N = f_h, 2f_h$ or $4f_h$ occur in marginal stability due to planetary slantwise convection (van
94 Haren, 2008).

95 In this paper, the investigation is further pursued of slantwise downward convective warm-water
96 periods inducing stratified-turbulence that occur frequently in the deep Western Mediterranean. For this
97 purpose, a nearly half-cubic-hectometer large 3D mooring-array is constructed holding about 3000 high-
98 resolution temperature T-sensors and deployed at a 2500-m deep seafloor. Turbulence calculations are
99 made based on Ellison (1957) and Thorpe (1977) overturning scales for a full observational year to
100 investigate potential seasonal variations therein. Governing physics processes that indirectly may affect
101 deep-sea life by inducing or transporting sufficient turbulence for nutrient and oxygen supply are:
102 atmospheric-disturbance generated near-inertial waves, boundary-flow instability generating sub-, $O(1)$
103 km, and meso-, $O(10-100)$ km, scale eddies. Motions associated with these processes dominate dispersal
104 of water masses in seas and oceans. They are all greatly affected by the rotation of the Earth. They are
105 however not considered to be part of irreversible ‘small-scale’ turbulent mixing. How they transfer
106 energy from the surface to deep-sea small-scales of turbulence dissipation is not yet fully established.

107 The materials and methods are described in Section 2, with an extension of turbulence calculation
108 methods in Appendix A2. The results are presented in Section 3, which starts with a yearlong overview
109 timeseries designating local variations in stratification conditions in comparison with the larger
110 background stratification. The analysis then highlights periods under differently stratified conditions,
111 before presenting yearlong timeseries of turbulence values, their improved statistics and short-scale
112 variations. The various mechanistics and their effects on the deep sea are discussed in Section 4.

113

114 **2 Materials and Methods**

115 Sufficiently resolved temperature time series mainly from single mooring lines in well-stratified waters
116 have proven useful for calculating turbulence values (van Haren and Gostiaux, 2012; Cimatoribus et al.,
117 2014) using methods proposed by Ellison (1957) and Thorpe (1977). The original methods need some
118 adaptation to be used in very weakly stratified waters like in the deep Mediterranean Sea. Turbulence
119 values are calculated for dissipation rate of turbulent kinetic energy, henceforth ‘turbulent dissipation
120 rate’, and, for a few instances, turbulent diffusivity.

121

122

123 2.1 Sampling network

124 In order to get some insight in the generation and development of deep-sea turbulence, 2925
125 independent, self-contained high-resolution NIOZ4 temperature ‘T’-sensors were distributed in nearly
126 half-a-million cubic meters of seawater. Details of construction and deployment of the large-ring
127 mooring can be found in van Haren et al. (2021). The large number of independent T-sensors is expected
128 to improve statistics of turbulence values, in an environment where all dynamics is captured by
129 temperature variations of less than 0.01°C. The small temperature variations put a large strain on the
130 technical capabilities of quantifying the deep-sea turbulence.

131 With two supplementary T-sensors registering tilt information above and below, 63 T-sensors were
132 taped at 2-m intervals to 45 vertical lines 125-m tall that each were tensioned to 1.3 kN by a single buoy
133 on top. The T-sensors were located between nominally $h = 1.5\text{-}125.5 \pm 0.5$ m above seafloor and recorded
134 data at a rate of once per 2 s. Three buoys, of lines 1.4, 3.5 and 5.7 (henceforth throughout the text, the
135 original naming ‘group.line’ is shortened without period; for layout see Appendix A1), held a single-
136 point Nortek AquaDopp current meter at $h = 126$ m, which recorded data once per 600 s. The lines were
137 attached at 9.5-m horizontal intervals to a steel-cable grid that was tensioned inside a 70-m diameter
138 steel-tube ring functioning as a 140-kN anchor.

139 The ensemble ‘large-ring mooring’ was deployed on the $<1^\circ$ flat and 2458-m deep seafloor of 42°
140 $49.50'N$, $006^\circ 11.78'E$ just 10 km south of the steep continental slope, 5 km from its abyssal-plain foot,
141 of the Northwestern Mediterranean Sea, in October 2020 (Fig. 1a). At the site’s (mid-)latitude $f_h = 1.08f$.
142 For calibration and reference purposes, a single shipborne Conductivity Temperature Depth ‘CTD’
143 profile was obtained to $h = 0.5$ m, about 1 km horizontally from the mooring site during the deployment
144 cruise. Meteorological data were obtained from Island Station ‘Porquerolles’, $43^\circ 0'N$, $006^\circ 12'E$.

145 With the aid of Irish Marine Institute Remotely Operated Vehicle “Holland I” all 45 vertical lines
146 with T-sensors were successfully recovered in March 2024. Of the lines, 43 were mechanically in good
147 order. Line 18 was hit by the drag parachute, which functioned as a stabiliser during the free-fall
148 deployment, whereby 10 sensors were lost. Line 65 was about 0.5-m lower than nominal because of a
149 loop near the cable grid. Fig. A1 shows the numbering of the lines, which were ordered in six groups

150 for synchronisation purposes. As with previously deployed NIOZ4 T-sensors (for details see van Haren,
151 2018), the individual clocks were synchronised to a single standard clock every 4 hours, so that all T-
152 sensors were sampled within 0.02 s. Line 36 did not register synchronization, possibly due to an electric
153 cable failure. Three T-sensors leaked and <10 were shifted in position due to a tape malfunctioning.
154 After calibration, some 20 extra T-sensors are not further considered due to electronics (noise) problems.
155 In total, 2882 out of 2925 T-sensors functioned as expected for the first 20 months after deployment,
156 with remaining bias due to electronic drift resulting in deviations from absolute accuracy. Depending
157 on the period and type of analysis considered, between 50 and 150 T-sensors showed too large bias
158 requiring additional attention during post-processing of the records from the weakly stratified deep sea.

159 Due to unknown causes all T-sensors switched off unintentionally when the file size on the memory
160 card reached 30 MB. This may have to do with a formatting or programming error. The dataset is 20
161 months long for sensors only registering temperature, and 3 months for those with tilt, though about 2.5-
162 3 years is possible with the present setup.

163

164 **2.2 Post-processing of temperature data**

165 With respect to previous NIOZ4 T-sensor version, improvements of the electronics resulted in about
166 noise levels of 0.00003°C and a doubling in battery life. As described in van Haren (2018), calibration
167 yielded a relative precision of $<0.0005^{\circ}\text{C}$, and $<0.0001^{\circ}\text{C}$ after careful post-processing. Bias due to
168 instrumental electronic drift of $<0.001^{\circ}\text{C mo}^{-1}$ after aging was primarily corrected by referencing daily
169 averaged vertical profiles, which must be stable from a perspective of turbulent overturning in a stratified
170 environment, to a smooth polynomial without instabilities. In addition, because vertical temperature
171 (density) gradients are so small in the deep Mediterranean, reference was made to periods of typically
172 one hour duration that were homogeneous with temperature variations smaller than instrumental noise
173 level (van Haren, 2022). Such periods were on days 350, 453, and 657 in the existing records. This
174 secondary correction included low-pass noise filtering ‘lpf’ of data with time. Under near-homogeneous
175 conditions, a tertiary correction involved lpf of data in the vertical. Temperature records were pressure-

176 corrected by transferring to Conservative Temperature Θ (IOC et al., 2010) using CTD's mean local
177 salinity value. Henceforth, Θ will be named 'temperature', for short.

178

179 **2.3 Turbulence values calculation for different stratification periods**

180 Given the consistent and tight temperature-density relationship (Section 3), corrected temperature data
181 allowed for calculations of turbulence values using the overturning displacement method of Thorpe
182 (1977) by reordering density instabilities. Details for the method of calculating turbulence values from
183 moored T-sensor data from well-stratified waters is given in van Haren and Gostiaux (2012). Here, the
184 method is applied under weakly stratified conditions in which buoyancy frequency $N \ll 10f$, f denotes
185 the local inertial frequency, the vertical Coriolis parameter. As the method by Thorpe (1977) is typically
186 applied to profiling data, the vertical spacing of 2 m of moored T-sensors is sufficient to resolve
187 Ozmidov (1965) and largest overturning scales, as has been verified in (van Haren and Gostiaux, 2012).
188 The transition to the energetic turbulence regime occurs for scales >1.8 m under $N = f$ using the
189 buoyancy-Reynolds number threshold given in (Gargett et al., 1984).

190 For the present deep-sea area, distinction is made between periods under environmental conditions
191 when $N \sim f$, somewhat exaggerating named stratified-water 'SW' conditions, and $N \ll f$ including
192 unstable values, named near-homogeneous 'NH' conditions. For NH, the tertiary correction is needed,
193 and, when unstable overturns exceed the 124-m vertical range of sensors, an extra correction is
194 mandatory because the Thorpe (1977) method of reordering is over-estimating displacements and
195 resulting stratification (van Haren, 2025). Such periods are difficult to trace, because of the extremely
196 small vertical temperature differences, and the selection can only be done manually as it is inadequately
197 automated.

198 For comparison with mean Thorpe (1977) method, turbulence values are also computed using
199 'Ellison'-scales (Ellison, 1957 for atmospheric data; Itsweire, 1984 for laboratory data; Moum, 1996 for
200 oceanographic microstructure profiler data). Such scales are determined from moored T-sensor time
201 series by filtering out internal wave and sub-mesoscale motions. The method is quite sensitive for the
202 precise high-pass filter 'hpf' cut-off frequency, as was noted for well-stratified Atlantic Ocean waters
203 (Cimatoribus et al., 2014 for oceanographic moored T-sensor data). Appendix A2 proposes a modified

204 version for application to moored T-sensor data under very weakly stratified conditions where filter cut-
205 off frequencies are given for SW and NH conditions in the deep Mediterranean. For both methods, mean
206 values are obtained from moored multiple-line T-sensor records after averaging over at least the largest
207 turbulence scales, over the vertical '[...]', over time '<...>', and over 45 horizontally distributed lines
208 '(...').

209

210 **3 Results**

211 **3.1 General yearlong overview and large-scale stratification**

212 The focus is on the first full year of observations to investigate potential seasonal variation in deep-sea
213 turbulence dissipation rate values. The yearlong data-overview time series in Fig. 2 demonstrates 2-30
214 day variations, in waterflow speed (Fig. 2a), temperature (Fig. 2b), horizontal waterflow difference (Fig.
215 2c), and vertical temperature difference (Fig. 2d). Such time-variability is typical for sub-meso- and
216 mesoscale motions, which are likely associated with the dynamically unstable, meandering boundary
217 current over the canyon-incised steep continental slope (Crepon et al., 1982) and which may develop
218 into eddies, with strongest flows near the surface and in winter (Alb erola et al., 1995; van Haren and
219 Millot, 2003).

220 Over the one year of observations, the deep-sea waterflow speed U seldom exceeds 0.1 m s^{-1} , with
221 little variations through the seasons. U also rather strongly varies with local inertial period, which
222 partially reflects the variable thickness of the graphical curves in Fig. 2a, and which may be understood
223 from comparing the daily-filtered time series with the original one.

224 Inertial motions do not dominate T-sensor data (Fig. 2b,d). Also in contrast with U , the temperature
225 data demonstrate a seasonal variation with relatively warmer (Fig. 2b) and more stratified (Fig. 2d)
226 waters in winter, coarsely between days 365 and 495. This seasonal variation is also found in horizontal
227 waterflow difference (Fig. 2c). The entire dynamical temperature variation over the year and up to $h =$
228 125 m from the seafloor is captured within maximum $|\Delta\Theta| < 0.01^\circ\text{C}$, and commonly amounts only a
229 few millidegrees.

230 About half the time, the temperature difference between $h = 1$ and 125 m $\Delta\Theta > 0.0002^\circ\text{C} \equiv T_{\text{thres}}$, a
231 threshold level that is about six times the standard deviation of T-sensor noise level. This temperature

232 difference provides an overall stratification resulting in $N > 0.65f$. These relatively warm SW either
 233 come from above or from the side, slantwise downward. The other half of the time $\Delta\Theta < T_{\text{thres}}$, $N < 0.65f$
 234 and stratification may be unstable, or NH conditions. Under NH, less than 0.7% of total time negative
 235 temperature differences are found exceeding the (absolute value of) threshold level and corresponding
 236 with large-scale, >125-m developed geothermal heating. As a result, at the observational site convection
 237 turbulence associated with geothermal heating is suppressed by warm waters advected into the area most
 238 of the time.

239 Although the boundary current is strongest near the surface, it manifests itself at great depths
 240 including mesoscale variations at horizontal scales $O(10-100)$ km. However, observed spatial waterflow
 241 variations over about 50-m horizontally also indicate much smaller-scale, rapidly-fluctuating differences
 242 (Fig. 2c). These variations associate, in absolute value, with warm SW conditions in approximately half
 243 the cases. No cooling, inversely stratified waters, from above are observed in the yearlong record.

244 Compared to open-ocean waters where large 100-m scale $N > 10f$, the deep Mediterranean SW are
 245 characterized by weak stratification with $N = O(f)$. Despite the relatively weak stratification, SW will
 246 prove important for turbulent mixing in the area.

247 The single shipborne CTD observations show no dominant influence of salinity over temperature
 248 governing density variations, in the lower 500 m above the seafloor (Fig. 3). Over the well-resolved
 249 stratified portion between $-2165 < z < -2055$ m, the density-temperature relationship is found to be
 250 consistent (cf. van Haren, 2025),

$$251 \quad \delta\sigma_2/\delta\Theta = -0.25 \pm 0.01 \text{ kg m}^{-3} \text{ } ^\circ\text{C}^{-1}, \quad (1)$$

252 where $\delta\sigma_2$ denotes the density anomaly referenced to a pressure level of 2×10^7 Pa. Hence, Θ can be used
 253 as tracer for density variations to quantify turbulent overturning using the Thorpe (1977) reordering
 254 method.

255 In the weakly stratified waters over a vertical range of 100 m, density stratification varies, so that N
 256 $< 1f$ or NH is found near the seafloor, and $N \approx 2f$ or SW around $z = -2050$ m (Fig. 3d). Over 25-m
 257 vertical ranges, $N \geq 2f$ can be found (Fig. 3d), and over 1-10 m ranges $N > 4f$ may be inferred from Fig.
 258 3c around $z = -2110$ and -1980 m. Such thin stratified layers are occasionally also found at greater depths

259 if the better-resolved temperature profile is investigated (Fig. 3b). While the occasional vertical
260 temperature differences of $<0.01^{\circ}\text{C}$ in Fig. 2d could result from horizontal differences or fronts, it seems
261 more likely that stratification of around $z = -2050$ m in Fig. 3 is periodically lowered by action from
262 above. Such action is expected down to about $h < 10$ m from the seafloor, a thin layer in which vertical
263 temperature differences are generally very small but not always (black graph in Fig. 2d).

264 Thus, although the moored T-sensors were located between the seafloor and $z = -2332$ m, CTD-
265 measured stratification may vary considerably with depth and time, and physical processes may lower
266 warmer water some 400 m or advecting such waters slantwise, or possibly quasi-horizontally into the
267 range of T-sensors. The precise direction of warm-water motion cannot be determined from single-
268 station profiles, but may be resolved with a properly scaled 3D mooring-array. First however, details
269 are investigated from different stratified periods using T-sensors data from a single line.

270

271 **3.2 1D-details of a stratified-water period**

272 Considering an average 5-mK-amplitude warm-water period (cf. Fig. 2b), a 1.3-day depth-time detail
273 from around day 485 is presented from single line 15 (Fig. 4). While waters seem depressed from $h >$
274 125 m, the warming occurs in variable periods of <1 hour (Fig. 4a). During the second half of the
275 warming, 0.001°C additional heat is observed near the top. Relatively warm waters reach the seafloor
276 twice within an inertial period of 0.73 day, around days 485.45 and 485.80. The warming ends with two
277 cooler-water fronts and large overturning reaching the seafloor around day 486.2.

278 The warming is depressed to within $h < 10$ m from the seafloor, with a relatively large vertical
279 temperature gradient between the cooler waters near the seafloor and the warmer waters higher up. This
280 is reflected in the increased value of 2-m-small-scale buoyancy frequency N_s in $h < 30$ m. Turbulent
281 overturns hardly occur between days 485.2 and 486.15 for $h < 5$ m, but are non-negligible in the stratified
282 waters above for $5 < h < 30$ m, and are typically 50-m large for $h > 30$ m (Fig. 4b).

283 Quantifying turbulence dissipation rate requires averaging, over all overturning scales possible, and
284 124-m vertically averaged values demonstrate variations with time over two orders of magnitude, when

285 reordered data are used (Thorpe (1977) method, black graph in Fig. 4c), or using $48 < \omega < 3000$ cpd
286 (cycles per day) band-pass filtered data (Ellison (1957) method cf. Appendix A2, cyan graph in Fig. 4c).

287 Time-depth mean values for line 15 from SW's day 485 are: turbulence dissipation rate $\langle[\varepsilon_T]\rangle =$
288 $6\pm 3\times 10^{-10} \text{ m}^2\text{s}^{-3}$ and turbulent diffusivity $\langle[K_z]\rangle = 1.5\pm 0.7\times 10^{-3} \text{ m}^2\text{s}^{-1}$ under buoyancy frequency $\langle[N]\rangle$
289 $= 2.9\pm 0.3\times 10^{-4} \text{ s}^{-1} \approx 3f$, using Thorpe (1977) method. Modified Ellison (1957)-method $\langle[\varepsilon_E]\rangle = 7\pm 3\times 10^{-$
290 $10} \text{ m}^2\text{s}^{-3}$ (Appendix A2). There is some bias to low mean turbulence values by disregarding overturns
291 below threshold, which is very low given the $<0.0001^\circ\text{C}$ precision of T-sensors after post-processing
292 (van Haren, 2018). However, this bias is well within the error of calculation as turbulence is dominated
293 by the large energy-containing overturns. These mean turbulence values are more than one order of
294 magnitude larger than open-ocean values observed in stratified waters well away from boundaries (e.g.,
295 Gregg, 1989; Polzin et al., 1997; Yasuda et al., 2021).

296

297 **3.3 1D-details of a near-homogeneous period**

298 For comparison, such turbulence values are under NH conditions between days 316.5 and 320.5 (Fig.
299 5): $\langle[\varepsilon_T]\rangle = 2\pm 1\times 10^{-10} \text{ m}^2\text{s}^{-3}$ and $\langle[K_z]\rangle = 1.1\pm 0.5\times 10^{-2} \text{ m}^2\text{s}^{-1}$ under $\langle[N]\rangle = 0.5\pm 0.2\times 10^{-4} \text{ s}^{-1} = 0.5f$.
300 These values follow partially correcting the original method by Thorpe (1977) for overturns exceeding
301 the height of instrumentation with information from manually selected NH environments that are
302 bounded by stratification above (van Haren, 2025). For periods with NH bounded by stratification
303 above, such as between days 318.6 and 318.96, their mean turbulence dissipation rate is to within 10%
304 the same as found for periods with convection turbulence due to geothermal heating: $\varepsilon_{\text{GH}} = 1.2\times 10^{-10}$
305 m^2s^{-3} . The ε_{GH} matches average geophysical heat-flux observations in the area (Pasquale et al., 1996),
306 under the condition that the mixing coefficient of $\Gamma_C = 0.5$ (van Haren, 2025), which is typical for
307 buoyancy-driven convection turbulence (Dalziel et al., 2008). For this period, $\langle[\varepsilon_E]\rangle = 1.3\pm 1\times 10^{-10} \text{ m}^2\text{s}^{-$
308 3 (Appendix A2).

309

310

311

312 3.4 Some 45-line statistics of short periods under SW and NH conditions

313 For consistency and statistics, six half-day periods are considered for computation of turbulence
314 dissipation rate values, three under SW and three under NH conditions. The computations are performed
315 for all 45 vertical lines and averages are computed over the 124-m height and half-day periods. It
316 provides a one-and-a-half order of magnitude distribution of mean turbulence dissipation rate values
317 (Fig. 6).

318 While some values are highly consistent between lines, e.g. the most energetic period on day 441,
319 others show a half-order of magnitude distribution of values like on day 459. Initially, this calculation
320 was set-up to help identify biased T-sensors and the appropriate polynomial correction. After applied
321 tertiary correction, remaining wide distributions are attributed to more general turbulence variability.

322 The statistics certainly improve turbulence dissipation rate values calculated using other
323 instrumentation and methodology, which is generally to within a factor of two at best. The six examples
324 of 45 lines provide about four times better statistics for the half-day periods (Table 1). The three NH
325 values average to $\langle[\varepsilon_T]\rangle_{\text{NH}} = 1.1 \pm 0.2 \times 10^{-10} \text{ m}^2\text{s}^{-3}$, which is well within error equivalent to ε_{GH} as
326 calculated from heat-flux measurements using mixing coefficient for convection-turbulence $\Gamma_C = 0.5$
327 (Dalziel et al., 2008), while significantly different from the value for shear-turbulence $\Gamma_S = 0.2$ typical
328 for stratified conditions (Osborn, 1980; Oakey, 1982). It thus confirms previous results (van Haren,
329 2025) and laboratory findings for convection-turbulence (e.g., Dalziel et al., 2008). The three SW values
330 average to $\langle[\varepsilon_T]\rangle_{\text{SW}} = 8 \pm 5 \times 10^{-10} \text{ m}^2\text{s}^{-3}$, noting that the standard deviations of individual mean values are
331 one order of magnitude smaller (Table 1).

332 Another consequence of the use of multiple mooring lines besides improved statistics, is some insight
333 in possible distribution of mean turbulence values. While one would expect erratic distribution over the
334 short horizontal distances <70 m, particularly NH distributions yield two-dimensional consistent images
335 such as on days 459 and 495 (Fig. 7). It provides confidence in consistency of methods used, but results
336 in a puzzling gradient in turbulence that apparently is independent of waterflow (measured at $h = 126$
337 m). For these geothermal heating- and near-inertial eddies-dominated periods the 9.5-m interval between
338 lines seems reasonably well chosen, where 100 m may have been too large.

339

340 3.5 Yearlong daily averaged turbulence dissipation rate values for 45 lines

341 A yearlong time series of daily-averaged turbulence values is computed for all 45 lines (Fig. 8). This
342 computation is automated, using a fixed 3rd-order polynomial for primary correction. Since the tertiary
343 correction for >125-m extending overturns is not applied manually, a criterion for excluding such
344 episodes is used. This criterion is simply based on the daily-averaged temperature difference between
345 uppermost and lowest T-sensor, per line. When $\Delta\Theta < T_{\text{thres}}$, given previously, the daily and vertical mean
346 turbulence dissipation rate is fixed to,

$$347 \quad \langle[\varepsilon_T]\rangle = \varepsilon_{\text{GH}} = 1.2 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}, \quad (3)$$

348 the mean value for geothermal heating (van Haren, 2025). This is found to occur $59 \pm 1.5\%$ of the time,
349 somewhat varying per line, and characterizes NH, besides geothermal heating. About $40 \pm 1.5\%$ of the
350 time is characterized by SW.

351 The overall, yearlong, 125-m vertical, and 45-line mean turbulence dissipation rate amounts,

$$352 \quad (\langle[\varepsilon_T]\rangle) = 2.4 \pm 0.2 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}, \quad (4)$$

353 so that the mean SW turbulence dissipation rate amounts,

$$354 \quad (\langle[\varepsilon_T]\rangle)_{\text{SW}} = 4.3 \pm 0.4 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}, \quad (5)$$

355 which is thus closely represented by the short periods of days 308 and 485 in Fig. 6, Table 1.

356 Part of the SW-turbulence is attributable to geothermal heating in a layer of typically $h = 30$ m under
357 stratified waters. This may be inferred from the vertical temperature difference in that layer that passes
358 T_{thres} during only 8% of daily periods, cf. the magenta graph in Fig. 2d. Nevertheless, the advection of
359 warmer waters suppresses geothermal heating-turbulence, possibly affecting the small-scale distribution
360 in Fig. 7. The associated 3.5-times larger turbulence dissipation rates (5) compared with geothermal-
361 heating value (3) are mainly generated by convection in conjunction with shear following internal-wave
362 breaking.

363 Considering the yearlong ‘seasonal’ variation that was suggested from the temperature (difference)
364 time series in Fig. 2, and which is represented by the logarithm of daily and vertically averaged $\langle[N]\rangle$
365 in Fig. 8c, the corresponding plot of $\langle[\varepsilon_T]\rangle$ (Fig. 8b) is more difficult to interpret, also in conjunction
366 with meteorological data (Fig. 8a). Different-line data mostly collapse on each other during winter

367 between days 365 and 495, for both $\langle[N]\rangle$ and $\langle[\varepsilon_T]\rangle$. During this period, vertical-line daily mean
368 turbulence dissipation rates most, 27 out of 41, exceed twice the mean value (4), shown by the green
369 asterisks in Fig. 8a. The 11% of time of green-asterisks occurrence average to a mean turbulence
370 dissipation rate of $7 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$.

371 The average turbulence dissipation rate for the 130-day winter period is 25% higher than the yearlong
372 mean (4). During this period, the wind work $\sim W^2$ is increased by 20% compared to its yearlong average
373 value. A rough visual correspondence is found between $|W|$ (Fig. 8a) and $\lg\langle[\varepsilon_T]\rangle$ (Fig. 8b), the former
374 leading the latter by about one week. The clearest value-collapse of turbulence and stratification data
375 from different lines is found between days 450 and 500. The early-spring period is unlikely governed
376 by deep dense-water formation due to limited meteorological forcing (Fig. 8a), but the preceding winter
377 cooling may induce enhanced sub-mesoscale activity. Although increased sub-mesoscale motions can
378 obscure near-inertial internal waves (van Haren and Millot, 2003), the transfer of energy to internal-
379 wave scales leading to breaking and turbulence is not hampered. Possibly, as near-inertial shear is
380 dominant in well-stratified waters, a shift from shear to convection turbulence may be associated with
381 the increase of sub-mesoscale activity. Such potential energy transfer will be elaborated elsewhere.

382 In contrast, 14% lower mean turbulence dissipation rate than (4) is found during the summer between
383 days 540-670. In this period, W^2 is decreased by 13% compared to its yearlong average value.

384

385 **4 Discussion and conclusions**

386 The observations show short $O(10)$ day periods of typically 0.005°C warmer waters than their
387 environment appearing from above, but also, as inferred from the 3D mooring-array (movies in van
388 Haren et al., 2026), from the sides, slantwise downward. The periods demonstrate a coarse near-inertial
389 periodicity, which is much less deterministic than a tide, and twice the inertial periodicity. Like internal
390 waves in the North-Atlantic Ocean and Lake Garda (van Haren, 2015; van Haren and Dijkstra, 2021),
391 they push stratification to within a few meters from the seafloor. The pushdown is vigorously turbulent,
392 more than one order of magnitude larger than in the open ocean away from boundaries. This relatively
393 large turbulence should not surprise as both the bulk Reynolds number $O(10^6)$ and buoyancy Reynolds

394 number $Re_b = \epsilon/(\nu N^2) = O(10^4) > 200$ are large, thereby classifying it as energetic regime (Gargett et al.,
395 1984), even in the weakly stratified deep sea. The $\nu \approx 10^{-6} \text{ m}^2 \text{ s}^{-1}$ denotes the kinematic viscosity. As
396 geothermal heating is found to be relatively weaker from below, the convection turbulence seems to be
397 driven by the, slanted, internal waves from above.

398 This slantwise convection turbulence is likely driven via (sub-)mesoscale eddies that are reinforced
399 by atmospheric disturbances (Fig. 9). As such they may be compared with deep dense-water formation
400 from sea-surface to seafloor that may occur over brief periods during a severe winter.

401

402 **4.1 Warm-water convection versus hypothetical deep-water formation**

403 While vertical motions by wintertime deep dense-water formation have been observed via moored
404 observations, also in the Mediterranean (e.g., Schott et al., 1996), and floats (Steffen and D'Asaro,
405 2002), and surface buoyancy fluxes have been estimated to be $O(10^{-7}) \text{ m}^2 \text{ s}^{-3}$ during convection events
406 (Marshall and Schott, 1999), quantification via observations of turbulence values associated with dense-
407 water formation reaching the abyssal seafloor have yet to be made (Thorpe, 2005). Unfortunately, a
408 dense-water event never reached the large-ring mooring while it was underwater. Estimates of deep-
409 convection duration are limited, albeit that some consensus exists about decadal variability or
410 occurrence of seafloor-reaching dense-water formation over a relatively short period of (less than) a
411 week (Lilly et al., 1999), maximum a month, per 8-10 years (Dickson et al., 1996; Mertens and Schott,
412 1998). It is tempting to compare coarse deep dense-water formation turbulence dissipation rate estimates
413 with those calculated from observations at the present mooring site during dominant warm-water
414 convection under SW and geothermal heating under NH.

415 Because of the lack of measurements to quantify deep dense-water formation turbulence, some
416 insight is gained from historic nocturnal convection-turbulence near the ocean surface. Microstructure
417 measurements by, e.g., Brainerd and Gregg (1995) demonstrated turbulence dissipation rate values $>$
418 $10^{-7} \text{ m}^2 \text{ s}^{-3}$ close to the surface and which decreased in the $O(10)$ m near-homogeneous layer to typically,

$$419 \quad \epsilon_{DWF} \approx 10^{-8} \text{ m}^2 \text{ s}^{-3}, \quad (6)$$

420 at a depth just above well-stratified waters below. The one order of magnitude reduced value reflects
421 the erosion of the stratification. Here, we take value (6) as a proxy for turbulence dissipation rate by an
422 event of deep dense-water formation-convection in waters just above the deep seafloor.

423 In comparison with geothermal heating's value (3), (6) is two orders of magnitude larger. Where
424 geothermal heating is quasi-permanent, deep dense-water formation rarely occurs, for example not at
425 all during the presented 20 months of observations. Two orders of magnitude difference implies
426 occurrence of (6) during one month per 8 years to match (3). This is the estimated maximum at a given
427 site.

428 In comparison with atmospheric-enforced mainly sub-mesoscale and internal-wave induced year-
429 average value (4), (6) would have to occur during 2.5 months per 8 years, or during 9 days every year.
430 This is not observed in the open Liguro-Provençal basin. It implies that, either deep dense-water
431 formation turbulence is stronger than (6) also for $z < -2000$ m, which seems unlikely, or geothermal
432 heating turbulence and especially SW turbulence are several times, SW turbulence up to one order of
433 magnitude, larger than deep dense-water formation turbulence, when averaged over a decade in time.
434 With their sources of sub-mesoscale eddies and near-inertial waves, the warm SW conditions thus seem
435 more important than deep dense-water formation for supply of fresh materials in the deep-sea area.
436 Recall that the observations are made in an area where tides, normally about half the ocean's mechanical
437 energy source, are weak.

438

439 **4.2 A comparison between turbulence from warm-water convection and geothermal heating**

440 The observed yearlong mean turbulence dissipation rate value of SW being 2.5 times that of geothermal
441 heating in the present area is the reverse of findings by Ferron et al. (2017), who find three times larger
442 geothermal heating than SW from sparse microstructure profiling across the entire Northwest
443 Mediterranean. The discrepancy may have to do with the location of the large-ring mooring, about 5 km
444 from the foot of the continental slope and most likely regularly located under an eddy or meander of the
445 well-stratified boundary current. These large-scale motions are intensified closer to the sea-surface
446 (Albérola et al., 1995; van Haren and Millot, 2003).

447 Estimating turbulence dissipation rates from the microstructure profiler plots in Ferron et al. (2017)
448 gives average values for $h = 100\text{-}600$ m (the instruments were stopped some 90 m above the seafloor)
449 of about $2.5 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$ and $7 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$, for the Western Mediterranean and specifically Ligurian
450 Sea, respectively. These values are in the same range as mean and SW values (4) and (5), respectively.
451 Both averaged microstructure profiles showed reduction in values in the lower $h = 100\text{-}200$ m to about
452 $1 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$ and $3 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$, respectively. For properly processed microstructure profiler data the
453 instrumental error is to within a factor of two for mean turbulence dissipation rate values, not considering
454 environmental variations. The shown values, from the same height above seafloor as the upper range of
455 the moored T-sensors, compare with the geothermal heating-(3) and mean-(4) values determined at the
456 large-ring mooring.

457

458 **4.3 Mechanistics behind slantwise warm-water convection in the deep sea**

459 A contribution of salt to density variations may possibly affect the turbulence values calculated from the
460 moored T-sensors under SW conditions, but density-temperature relationship across stratified layers is
461 found consistent between different years (van Haren, 2025). Also, convection turbulence under SW
462 conditions has been observed in deep alpine-lake Garda where salinity contributes little to density
463 variations (van Haren and Dijkstra, 2021).

464 The relative importance of stratified turbulence occurring in varying strength over about half the time
465 has consequences for deep-sea transport, redistribution of matter, and life. The regular replenishment is
466 partially related with atmospheric disturbances, in an indirect way. Winds do not directly affect motions
467 near the 2500-m deep seafloor. However, wind-induced near-inertial internal waves and boundary-
468 current variations affecting sub-mesoscale eddies seem to have correspondence with turbulence intensity
469 variations close to the seafloor, roughly a week after variations occur near the surface. More SW activity
470 and about 20% larger turbulence dissipation rates were found in winter when atmospheric activity was
471 correspondingly larger. Weakest stratification was found more in summer. As eddies and near-inertial
472 waves cause convection in the direction of the rotational axis to slant to the vertical under weak z -
473 direction stratification $N = O(f)$, cf. McEwan (1973); Straneo et al. (2002); Sheremet (2004); Gerkema
474 et al. (2008), the stratified turbulence may come from above and in part horizontally.

475 Mesoscale anticyclonic eddies in the central Mediterranean show observed downward vertical
476 motions along their rim (van Haren et al., 2006). Such eddies may trap and tunnel near-inertial waves
477 (Kunze, 1985), providing different internal wave behavior in the eddy's rim and core (Fig. 9).

478 Comparing periods under SW and NH conditions, deep-sea temperature variance is much larger at
479 all frequencies under SW, but a distinct spectral peak is lacking (Fig. A2). Temperature variations are
480 not governed by quasi-deterministic signals like near-inertial motions, but predominantly by sub-
481 mesoscale and turbulence motions. At most, a bulge is found near the buoyancy frequency in open-
482 ocean displacement spectra (Munk, 1980). For both the open ocean and the Mediterranean Sea, this
483 contrasts with waterflow kinetic energy spectra that show a near-inertial peak. In the deep Mediterranean
484 poorly resolved waterflow measurements such a peak is observed mainly under NH conditions, while
485 kinetic energy is higher at all other frequencies, especially at sub-inertial frequencies, under SW
486 conditions.

487 Thus, under SW all near-inertial motions are feeding, or are overwhelmed by, sub- and super-inertial
488 motions, which results in the continuous slope in temperature spectra with a plateau in the inertio-gravity
489 wave band, up to about the maximum small-scale buoyancy frequency. The wider inertio-gravity wave
490 band under NH sees more temperature variance near its bounds.

491 In between stratified turbulence periods, waters tend to become near-homogeneous whilst being
492 dominantly mixed by convection turbulence through geothermal heating. As in Rayleigh-Taylor
493 convection, plumes of geothermal heating-response in waters overlying the seafloor strongly vary with
494 time, and thus spatially (e.g., Dalziel et al. 2008; Ng et al., 2016).

495 Geologically, even over a nearly flat sedimented seafloor underlying crustal cracks may develop
496 variable geothermal heating over distances as small as <1 km, depending on location of faults (e.g.,
497 Kunath et al., 2021). This could explain observed variation in mean turbulence dissipation rates over the
498 70-m size of the large-ring mooring, during geothermal heating. The 0.6-m high steel tubes of the large
499 ring will not affect geothermal heating up to $h = 125$ m. An inconclusive variation of mean turbulence
500 values over periods of SW conditions demonstrates larger scale variability.

501 Overall, the 45 vertical lines and nearly 3000 high-resolution T-sensors provided improved statistics
502 for daily mean turbulence dissipation rate values to within a reduced relative error of about 25%. Other

503 strengths of the mooring-array like improved spectral resolution and 3D evolution of turbulence will be
504 reported elsewhere, while short movies of 3D turbulence passages have been described in van Haren et
505 al. (2026).

506 The deep Mediterranean Sea observations are likely general for the deep ocean above a flat seafloor,
507 in the vicinity of large topography. This is because (sub-)mesoscale eddies, inertial motions and
508 geothermal heating are generally occurring. However, as tides and stratification are relatively weak in
509 the Mediterranean, compared to similar depths in the ocean, the observed turbulence processes may be
510 more difficult to detect in the ocean.

511
512 *Data availability.* Only raw data are stored from the T-sensor mooring-array. Analyses proceed via
513 extensive post-processing, including manual checks, which are adapted to the specific analysis task.
514 Because of the complex processing the raw data from the custom-made T-sensors are not made publicly
515 accessible. Current meter and CTD data are available from van Haren (2025): “Large-ring mooring
516 current meter and CTD data”, Mendeley Data, V1, <https://doi.org/10.17632/f8kfwcvtdn.1>. Atmospheric
517 data are retrieved from <https://content.meteoblue.com/en/business-solutions/weather-apis/dataset-api>.

518
519 *Competing interests.* The author has no competing interests.

520
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528

529 **Appendix A1 Layout of large-ring mooring**

530 The large-ring mooring has a diameter of nearly 70 m (Fig. A1). The eighteen 12-m long and 0.6-m
531 diameter steel tubes hold a steel-cable grid for rigidity. The cables are 9.5 m apart. At cable-intersects,
532 2.5-m diameter ‘small’ rings are mounted that each held a 125-m long mooring line with 65 T-sensors
533 below a single 1.45-kN buoy. Of eight small rings, imaginary intersects were at the steel tubes, so that
534 special off-set mounting was needed with three assist cables (van Haren et al., 2021). Upon landing at
535 the seafloor following parachute-controlled ‘free’ fall, the orientation of the ring was directed to the
536 north-northwest, pointing at 337 °N. After underwater chemical release of the buoys, the cable-grid was
537 lifted in a dome with its center $h = 2.0$ m above seafloor (van Haren, 2026 submitted).

538 The vertical mooring lines were named in six synchronisation groups, of maximum eight lines each.
539 The single synchroniser was located at the small ring of central line 51. Every half hour, the synchroniser
540 sent a clock pulse to a group. The synchronisation sequence of six pulses was repeated every four hours.
541

542 **Appendix A2 Proposed generalization of filter cut-off to compute Ellison scales**

543 Data from moored strings of high-resolution temperature sensors are potentially useful to compare two
544 different manners of calculating turbulence values. The more common method proposed by Thorpe
545 (1977) involves the reordering or sorting of unstable density overturns and the bookkeeping of their
546 vertical displacements, for vertical profiles at each time step. Turbulence values are computed following
547 averaging in the vertical, in time, or both and should include the largest of overturn scales. The method
548 requires a consistent temperature-density relationship. In near-homogeneous waters, difficulties may
549 arise in establishing such a relationship, but also in determining the size of largest overturns when they
550 outgrow the height of the string. A method is proposed to correct over-estimation under convection
551 turbulence by geothermal heating using verification via results from geophysical sampling (van Haren,
552 2025).

553 In well stratified waters, the Thorpe (1977) method has been successfully compared (e.g., Itsweire,
554 1984; Moum, 1996; Cimattorus et al., 2014), with results from the method introduced for atmospheric
555 data by Ellison (1957). In this Appendix, such a comparison is done for weakly stratified waters. A
556 modification is proposed of the Ellison (1957) method for application to data from moored instrumented
557 strings, and which is practically based on instrument performance and environmental physics conditions.

558 Ellison (1957) separated time series of potential temperature $\theta(t, z)$, which is dynamically equivalent
559 to Conservative Temperature Θ (IOC et al., 2010) in the ocean, at a fixed vertical position z in two,

560
$$\theta = \langle \theta \rangle + \theta', \tag{A1}$$

561 where $\langle \cdot \rangle$ denotes the lpf series and the prime its hpf equivalent. If multiple sensors are deployed in the
562 vertical to establish a mean vertical gradient, a scale height can be defined as,

563
$$L_E = \langle \theta'^2 \rangle^{1/2} / (d\langle \theta \rangle / dz). \tag{A2}$$

564 Itsweire (1984), using laboratory CTD-profile data, and Moum (1996), using ocean microstructure
565 profiler data, apply sorting as filter in z -direction. While their quasi-hpf data unlikely contain (linear)
566 internal waves, provided the profiles were instantaneously made and strictly vertical, they may contain
567 instrumental noise. With limited time-evolution available, it is assumed that the hpf data have worked

568 against the local stratification. The same assumption is made for Thorpe (1977) displacements. Sorting
569 works on all scales of overturns, which can be highly varying.

570 Like Thorpe-displacement ‘d’ scales, the Ellison scales of (A2) may be compared with the Ozmidov
571 (1965) scale $L_O = cL_E$ of largest possible turbulent overturns in stratified waters, so that the turbulence
572 dissipation rate reads,

$$573 \quad \varepsilon_E = c^2 L_E^2 N^3, \quad (A3)$$

574 in which the constant c needs to be established. If we take an average value of $c = 0.8$ (Dillon, 1982),
575 like commonly used for vertical root-mean-square ‘rms’ Thorpe scale $L_T = [d^2]^{1/2}$ so that $\varepsilon_T = c^2 L_T^2 N^3$,
576 one can compare average turbulence dissipation rate values between the two methods.

577 While the Thorpe (1977) method is most sensitive to proper resolution of the largest vertical overturn
578 scale, and the stratification (or buoyancy frequency) it works against, determination of Ellison scales
579 from moored-sensor time series is most sensitive for the appropriate separation between internal waves
580 and turbulent motions (Cimatoribus et al., 2014). For data from instrumented strings under well-
581 stratified Northeast-Atlantic conditions, wavelet decomposition worked using an averaging scale of
582 about $2/N$ for the lpf in (A1). Under such conditions, instrumental flaws like short-term bias of T-sensor
583 data were minimal. However, such a determination of Ellison scales is not a straightforward task under
584 weakly stratified and near-homogeneous conditions like occurring in the deep Western Mediterranean.

585 First, because such conditions imply very small variations in temperature (density), time series
586 require lpf to remove instrumental noise.

587 Second, time series require hpf to remove internal waves and (sub-)mesoscale motions. While the
588 common internal-wave band is considered between ranges f and N , for well stratified waters $N \gg f$,
589 more complex inertio-gravity wave frequency ‘ ω ’ bounds $[\omega_{\min} < f, \omega_{\max} > N]$, for large-scale mean N ,
590 have to be considered in waters where $N = O(f)$, e.g. (LeBlond and Mysak, 1978; Gerkema et al., 2008).
591 Furthermore, while average large-scale stratification hampers turbulent overturns, internal-wave
592 straining separates small thin-layer stratification, with small-vertical-scale buoyancy frequency N_s , from
593 near-homogeneous layers, with minimum buoyancy frequency N_{\min} , which may carry ditto waves
594 extending beyond the mean- N inertio-gravity wave-bounds. At the low-frequency, sub-inertial side of

595 the inertio-gravity wave-band the rare N_{\min} may combine with sub-mesoscale motions. More
 596 importantly, at the high-frequency, super-buoyancy side N_s may combine with turbulent overturns. If
 597 all relevant scales are resolved, a safe separation frequency would thus be at overall maximum $N_{\max} =$
 598 $\max(\max(N_s))$, where the maximum between brackets is determined for each profile. In practice, such
 599 a transition frequency between internal waves and turbulence is not easily determined, because it
 600 requires the small scales to resolve relevant L_O . Only in weakly stratified waters, L_O are $O(10)$ m, and
 601 resolution of 1-2 m scales should be sufficient.

602 Thus, under weakly stratified conditions, instrumental noise and short-term bias have to be corrected
 603 in t- and z-direction, respectively. A practical solution that also eliminates internal waves and sub-
 604 mesoscale motions, is application of a band-pass filter ‘bpf’ in t, an lpf and sorting in z so that per time
 605 step (A2) reads,

$$606 \quad |\theta'_{\text{bpf}}| / (d\theta_{\text{sorted}}/dz),$$

607 to which sufficient averaging is applied. Noting that stratification varies over different scales by two
 608 orders of magnitude, so that $N_s \approx 0-6f$, filter design discriminates between conditions of SW and NH.
 609 Sharp, phase-preserving double-elliptic filters (Parks and Burrus, 1987) are designed following
 610 inspection of temperature variance spectra (Fig. A2).

611 For SW, reasonable filter cut-offs are tuned for a 1.7-day period around day 308 by equating (A3) to
 612 ε_T . SW’s lpf cut-off is fixed at 3000 cpd. The reference hpf cut-off frequency $\omega_{\text{hpf,ref}}$ appeared at a small
 613 flat (0-)slope near the low end of turbulence buoyancy- $(\omega^{-7/5})$ and inertial-subrange $(\omega^{-5/3})$ slopes, where
 614 a short steep slope to internal-wave frequencies occurred. For other SW-periods, reference is not made
 615 using average large-scale N , but a better fit is found for the time average of maximum small-scale
 616 buoyancy frequencies per profile $N_m = \langle \max(N_s) \rangle$ so that,

$$617 \quad \omega_{\text{hpf}}^{\text{SW}} = (N_m / N_{m,\text{ref}})^2 \omega_{\text{hpf,ref}}. \quad (\text{A4})$$

618 For NH, the lpf cut-off is fixed at 500 cpd. For its hpf cut-off, a flat (0-) slope appeared at a frequency
 619 just higher than ω_{\max} so that, independent of measured N_m the cut-off is blocked at,

$$620 \quad \omega_{\text{hpf}}^{\text{NH}} = 3.7 \text{ cpd}. \quad (\text{A5})$$

621 This value is tuned for a period with dominant geothermal heating, for which the geophysics-determined
622 (e.g., Pasquale et al., 1996) buoyancy flux $fl/\Gamma_C = \varepsilon_{GH} = 1.2 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$. In the present (Fig. 6) and
623 previous (van Haren, 2025) data the mixing coefficient was found to amount $\Gamma_C = 0.5$, which is typical
624 for convection turbulence (Dalziel et al., 2008; Ng et al., 2016).

625 The cut-off frequency in (A5) is to within ± 0.2 cpd equivalent to $1.8(2\Omega) \approx 1.8\langle\omega_{\max}\rangle \approx 2N_m \approx$
626 $0.5\langle U\rangle/\langle L_T\rangle = 0.5\omega_o$, for the deep Western Mediterranean site. Ω denotes the Earth rotation, and U
627 the waterflow speed. The Ozmidov (1965) frequency ω_o is a natural separator between internal waves
628 and stratified turbulence. It is unknown why the filter cut-off is close to half the Ozmidov frequency.
629 Also puzzling is the lack of correspondence between (A4) and ω_o , with the factor varying between 0.3
630 and 0.9 for different periods of SW. In part this may have to do with the waterflow being measured at h
631 $= 126 \text{ m}$, below which it may be more uniform under NH- than under SW-conditions.

632 As shown in Fig. 4 for (A4), and Fig. 5 for (A5), the comparison between average turbulence
633 dissipation rate values using Ellison (1957) and Thorpe (1977) methods works well to within a relative
634 error of about 20% of the mean value. To Fig. 4, $\langle[\varepsilon_T]\rangle = 6.3 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$ and $\langle[\varepsilon_E]\rangle = 6.7 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$.
635 To Fig. 5, $\langle[\varepsilon_T]\rangle = 2.0 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$ and $\langle[\varepsilon_E]\rangle = 1.3 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$. Further tests were performed for
636 about ten 1-4 day periods of NH and SW, and all were within above relative error, provided that the data
637 post-processing was carefully done and longer periods were avoided. Especially (A5) is very sensitive
638 to small changes in filter steepness around the cut-off frequency, presumably due to its proximity to the
639 inertio-gravity wave upper bound.

640

641 **References**

- 642 Adrián-Martínez, S., Ageron, M., Aharonian, F., Aiello, S., Albert, A., et al.: Letter of intent for
643 KM3NeT 2.0, *J. Phys. G*, 43, 084001, DOI 10.1088/0954-3899/43/8/084001, 2016.
- 644 Albérola, C. Millot, C., and Font, J.: On the seasonal and mesoscale variabilities of the Northern Current
645 during the PRIMO-0 experiment in the western Mediterranean Sea, *Oceanol. Acta*, 18, 163-192,
646 <https://archimer.ifremer.fr/doc/00096/20770/18396.pdf>, 1995.
- 647 Brainerd, K. E., and Gregg, M. C.: Surface mixed and mixing layer depths, *Deep-Sea Res. I*, 42, 1521-
648 1543, [https://doi.org/10.1016/0967-0637\(95\)00068-H](https://doi.org/10.1016/0967-0637(95)00068-H), 1995.
- 649 Cimadoribus, A. A., van Haren, H., and Gostiaux, L.: Comparison of Ellison and Thorpe scales from
650 Eulerian ocean temperature observations, *J. Geophys. Res.*, 119, 7047-7065,
651 <https://doi.org/10.1002/2014JC010132>, 2014.
- 652 Crepon, M., Wald, L., and Monget, J. M.: Low frequency waves in the Ligurian Sea during December
653 1977, *J. Geophys. Res.*, 87, 595-600, <https://doi.org/10.1029/JC087iC01p00595>, 1982.
- 654 Dalziel, S. B., Patterson, M. D., Caulfield, C. P., and Coomaraswamy, I. A.: Mixing efficiency in high-
655 aspect-ratio Rayleigh-Taylor experiments, *Phys. Fluids*, 20, 065106,
656 <https://doi.org/10.1063/1.2936311>, 2008.
- 657 Dickson, R. R., Lazier, J. R. N., Meincke, J., Rhines, P., and Swift, J.: Longterm coordinated changes
658 in the convective activity of the North Atlantic, *Progr. Oceanogr.*, 38, 241-295,
659 [https://doi.org/10.1016/S0079-6611\(97\)00002-5](https://doi.org/10.1016/S0079-6611(97)00002-5), 1996.
- 660 Dillon, T. M.: Vertical overturns: a comparison of Thorpe and Ozmidov length scales, *J. Geophys. Res.*
661 87, 9601-9613, <https://doi.org/10.1029/JC087iC12p09601>, 1982.
- 662 Ellison, T. H., Turbulent transport of heat and momentum from an infinite rough plane, *J. Fluid Mech.*,
663 2, 456-466, <https://doi.org/10.1017/S0022112057000269>, 1957.
- 664 Eriksen, C. C.: Observations of internal wave refraction of sloping bottoms, *J. Geophys. Res.*, 87, 525-
665 538, <https://doi.org/10.1029/JC087iC01p00525>, 1982.
- 666 Ferron, B., Bouruet Aubertot, P., Cuypers, Y., Schroeder, K., and Borghini, M.: How important are
667 diapycnal mixing and geothermal heating for the deep circulation of the Western

668 Mediterranean? *Geophys. Res. Lett.*, 44, 7845-7854, <https://doi.org/10.1002/2017GL074169>,
669 2017.

670 Gargett, A. E., Osborn, T. R., and Nasmyth, P. R.: Local isotropy and the decay of turbulence in a
671 stratified fluid, *J. Fluid Mech.*, 144, 231-280, <https://doi.org/10.1017/S0022112084001592>,
672 1984.

673 Gerkema, T., Zimmerman, J. T. F., Maas, L. R. M., and van Haren, H.: Geophysical and astrophysical
674 fluid dynamics beyond the traditional approximation, *Rev. Geophys.*, 46, RG2004,
675 <https://doi.org/10.1029/2006RG000220>, 2008.

676 Gregg, M. C.: Scaling turbulent dissipation in the thermocline, *J. Geophys. Res.*, 94, 9686-9698,
677 <https://doi.org/10.1029/JC094iC07p09686>, 1989.

678 IOC, SCOR, and IAPSO: The International Thermodynamic Equation of Seawater – 2010: Calculation
679 and Use of Thermodynamic Properties, Intergovernmental Oceanographic Commission,
680 Manuals and Guides No. 56, UNESCO, Paris, 196 pp,
681 <https://unesdoc.unesco.org/ark:/48223/pf0000188170>, 2010.

682 Itsweire, E. C.: Measurements of vertical overturns in a stably stratified turbulent flow, *Phys. Fluids*,
683 27, 764-766, <https://doi.org/10.1063/1.864704>, 1984.

684 Kunath, P., Chi, W.-C., Berndt, C., and Liu, C.-S.: A rapid numerical method to constrain 2D focused
685 fluid flow rates along convergent margins using dense BSR-based temperature field data, *J.*
686 *Geol. Res. Solid Earth*, 126, e2021JB021668, <https://doi.org/10.1029/2021JB021668>, 2021.

687 Kunze, E.: Near-inertial wave propagation in geostrophic shear, *J. Phys. Oceanogr.*, 15, 544-565,
688 [https://doi.org/10.1175/1520-0485\(1985\)015<0544:NIWPIG>2.0.CO;2](https://doi.org/10.1175/1520-0485(1985)015<0544:NIWPIG>2.0.CO;2), 1985.

689 LeBlond, P. H., and Mysak, L. A.: *Waves in the Ocean*, Elsevier, New York, 602 pp, 1978.

690 Lilly, J. M., Rhines, P. B., Visbeck, M., Davis, R., Lazier, J. R. N., Schott, F., and Farmer, D.: Observing
691 Deep Convection in the Labrador Sea during Winter 1994/95, *J. Phys. Oceanogr.*, 29, 2065-
692 2098, [https://doi.org/10.1175/1520-0485\(1999\)029<2065:ODCITL>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<2065:ODCITL>2.0.CO;2), 1999.

693 Marshall, J., and Schott, F.: Open-ocean convection: Observations, theory, and models, *Rev. Geophys.*
694 37, 1-64, <https://doi.org/10.1029/98RG02739>, 1999.

695 McEwan, A. D.: A laboratory demonstration of angular momentum mixing, *Geophys. Fluid Dyn.*, 5,
696 283-311, <https://doi.org/10.1080/03091927308236121>, 1973.

697 Mertens, C., and Schott, F.: Interannual variability of deep-water formation in the Northwestern
698 Mediterranean, *J. Phys. Oceanogr.*, 28, 1410-1424, [https://doi.org/10.1175/1520-0485\(1998\)028<1410:IVODWF>2.0.CO;2](https://doi.org/10.1175/1520-0485(1998)028<1410:IVODWF>2.0.CO;2), 1998.

700 Moum, J. N.: Energy-containing scales of turbulence in the ocean thermocline, *J. Geophys. Res.*, 101,
701 14,095-14,109, <https://doi.org/10.1029/96JC00507>, 1996.

702 Munk, W. H.: Internal wave spectra at the buoyant and inertial frequencies, *J. Phys. Oceanogr.* 10, 1718-
703 1728, [https://doi.org/10.1175/1520-0485\(1980\)010<1718:IWSATB>2.0.CO;2](https://doi.org/10.1175/1520-0485(1980)010<1718:IWSATB>2.0.CO;2), 1980.

704 Ng, C. S., Ooi, A., and Chung, D.: Potential energy in vertical natural convection, *Proc. 20th*
705 *Australasian Fluid Mech. Conf.*, Paper 727, 4 pp.,
706 [https://people.eng.unimelb.edu.au/imarusic/proceedings/20/727 Paper.pdf](https://people.eng.unimelb.edu.au/imarusic/proceedings/20/727%20Paper.pdf), 2016.

707 Oakey, N. S.: Determination of the rate of dissipation of turbulent energy from simultaneous temperature
708 and velocity shear microstructure measurements, *J. Phys. Oceanogr.*, 12, 256-271,
709 [https://doi.org/10.1175/1520-0485\(1982\)012<0256:DOTROD>2.0.CO;2](https://doi.org/10.1175/1520-0485(1982)012<0256:DOTROD>2.0.CO;2), 1982.

710 Osborn, T. R.: Estimates of the local rate of vertical diffusion from dissipation measurements, *J. Phys.*
711 *Oceanogr.* 10, 83-89, [https://doi.org/10.1175/1520-0485\(1980\)010<0083:EOTLRO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1980)010<0083:EOTLRO>2.0.CO;2),
712 1980.

713 Ozmidov, R. V.: About some peculiarities of the energy spectrum of oceanic turbulence, *Dokl. Akad.*
714 *Nauk SSSR*, 161, 828-831, <https://www.mathnet.ru/eng/dan/v161/i4/p828>, 1965.

715 Parks, T. W., and Burrus, C. S.: *Digital Filter Design*, John Wiley Sons, New York, 342 pp, 1987.

716 Pasquale, V., Verdoya, M., and Chiozzi, P.: Heat flux and timing of the drifting stage in the Ligurian–
717 Provençal basin (northwestern Mediterranean), *J. Geodyn.*, 21, 205-222,
718 [https://doi.org/10.1016/0264-3707\(95\)00035-6](https://doi.org/10.1016/0264-3707(95)00035-6), 1996.

719 Polzin, K. L., Toole, J. M., Ledwell, J. R., and Schmitt, R. W.: Spatial variability of turbulent mixing in
720 the abyssal ocean, *Science*, 276, 93-96, DOI: 10.1126/science.276.5309.93, 1997.

721 Schott, F., Visbeck, M., Send, U., Fischer, J., Stramma, L., and Desaubies, Y.: Observations of deep
722 convection in the Gulf of Lions, northern Mediterranean, during the winter of 1991/ 92, *J. Phys.*

723 Oceanogr., 26, 505-524, <https://doi.org/10.1175/1520->
724 0485(1996)026<0505:OODCIT>2.0.CO;2, 1996.

725 Sheremet, V. A.: Laboratory experiments with tilted convective plumes on a centrifuge: A finite angle
726 between the buoyancy force and the axis of rotation, *J. Fluid Mech.*, 506, 217-244,
727 <https://doi.org/10.1017/S0022112004008572>, 2004.

728 Steffen, E. L., and D'Asaro, E.: Deep convection in the Labrador Sea as observed by Lagrangian floats,
729 *J. Phys. Oceanogr.*, 32, 475-492, <https://doi.org/10.1175/1520->
730 0485(2002)032<0475:DCITLS>2.0.CO;2, 2002.

731 Straneo, F., Kawase, M., and Riser, S. C.: Idealized models of slantwise convection in a baroclinic flow,
732 *J. Phys. Oceanogr.*, 32, 558-572, <https://doi.org/10.1175/1520->
733 0485(2002)032<0558:IMOSCI>2.0.CO;2, 2002.

734 Thorpe, S. A.: Turbulence and mixing in a Scottish loch, *Phil. Trans. Roy. Soc. Lond. A*, 286, 125-181,
735 <https://doi.org/10.1098/rsta.1977.0112>, 1977.

736 Thorpe, S. A.: Current and temperature variability on the continental slope, *Phil. Trans. Roy. Soc. Lond.*
737 *A*, 323, 471-517, <https://doi.org/10.1098/rsta.1987.0100>, 1987.

738 Thorpe, S. A.: *The turbulent ocean*, Cambridge Univ Press, Cambridge, 439 pp, 2005.

739 van Haren, H.: Abrupt transitions between gyroscopic and internal gravity waves: the mid-latitude case,
740 *J. Fluid Mech.*, 598, 67-80, <https://doi.org/10.1017/S0022112007009524>, 2008.

741 van Haren, H.: Instability observations associated with wave breaking in the stable-stratified deep-
742 ocean, *Phys. D*, 292-293, 62-69, <https://doi.org/10.1016/j.physd.2014.11.002>, 2015.

743 van Haren, H.: Philosophy and application of high-resolution temperature sensors for stratified waters,
744 *Sensors*, 18, 3184, doi:10.3390/s18103184, <https://doi.org/10.3390/s18103184>, 2018.

745 van Haren, H.: Thermistor string corrections in data from very weakly stratified deep-ocean waters,
746 *Deep-Sea Res. I*, 189, 103870, <https://doi.org/10.1016/j.dsr.2022.103870>, 2022.

747 van Haren, H.: Corrected values of turbulence generated by general geothermal convection in deep
748 Mediterranean waters, *Ocean Dyn.*, 75, 46, <https://doi.org/10.1007/s10236-025-01693-4>, 2025.

749 van Haren, H., and Dijkstra, H. A.: Convection under internal waves in an alpine lake, *Env. Fluid Mech.*,
750 21, 305-316, <https://doi.org/10.1007/s10652-020-09774-2>, 2021.

751 van Haren, H. and Gostiaux, L.: Detailed internal wave mixing observed above a deep-ocean slope, *J.*
752 *Mar. Res.*, 70, 173-197, https://elischolar.library.yale.edu/journal_of_marine_research/337,
753 2012.

754 van Haren, H., and Millot, C.: Seasonality of internal gravity waves kinetic energy spectra in the
755 Ligurian Basin, *Oceanol. Acta*, 26, 635-644, [https://doi.org/10.1016/S0399-1784\(03\)00062-8](https://doi.org/10.1016/S0399-1784(03)00062-8),
756 2003.

757 van Haren, H., Millot, C., and I. Taupier-Letage, I.: Fast deep sinking in Mediterranean eddies, *Geophys.*
758 *Res. Lett.*, 33, L04606, doi:10.1029/2005GL025367, 2006.

759 van Haren, H., Bakker, R., Witte, Y., Laan, M., and van Heerwaarden, J.: Half a cubic hectometer
760 mooring-array 3D-T of 3000 temperature sensors in the deep sea, *J. Atmos. Ocean. Technol.*,
761 38, 1585-1597, <https://doi.org/10.1175/JTECH-D-21-0045.1>, 2021.

762 van Haren, H., Adriani, O., Albert, A., Alhebsi, A. R., Alshalloudi, S., et al.: Whipped and mixed warm
763 clouds in the deep sea, *Geophys. Res. Lett.*, 53, e2025GL119998,
764 <https://doi.org/10.1029/2025GL119998>, 2026.

765 Yasuda, I., Fujio, S., Yanagimoto, D., Lee, K.J., Sasaki, Y., Zhai, S., Tanaka, M., Itoh, S., Tanaka, T.,
766 Hasegawa, D., Goto, Y., and Sasano, D.: Estimate of turbulent energy dissipation rate using
767 free-fall and CTD-attached fast-response thermistors in weak ocean turbulence, *J. Oceanogr.*,
768 77, 17-28, <https://doi.org/10.1007/s10872-020-00574-2>, 2021.

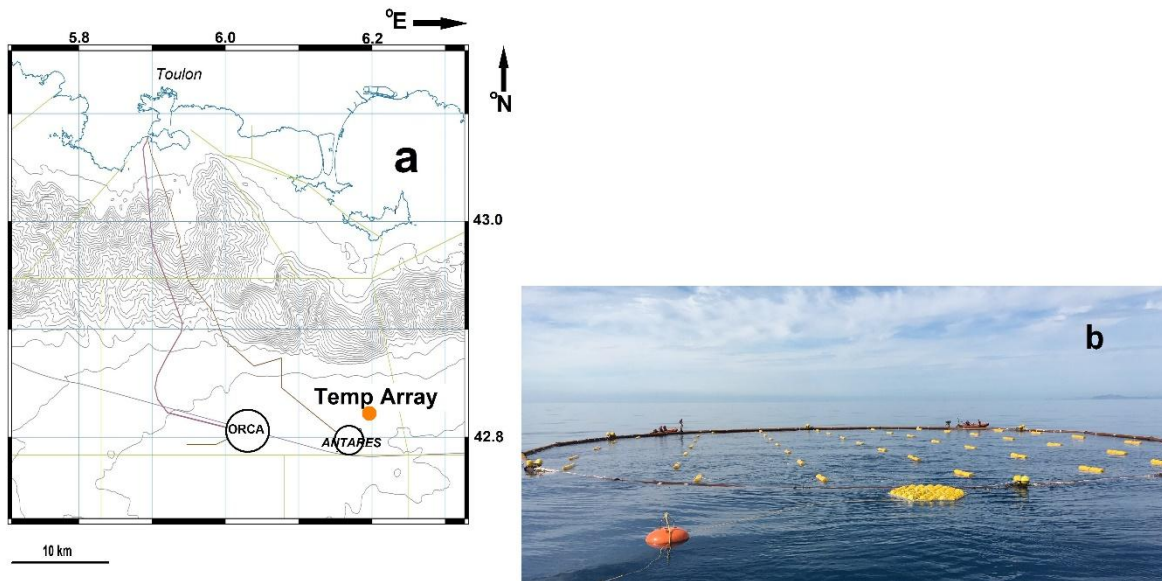
769

770 **Table 1.** General first and second moment statistics calculated for 45-line, 124-m height, and half-day
 771 period turbulence dissipation rate values of Fig. 6. Two environmental conditions are characterized:
 772 SW = stratified-water, NH = near-homogenous, and may include geothermal heating. The two
 773 conditions are separated by a criterion equivalent of $N = 0.65f$.

774

775	<i>Day</i>	$\varepsilon_T [m^2 s^{-3}]$	<i>Cond.</i>
776	308	$4.0 \pm 0.8 \times 10^{-10}$	SW
777	441	$1.5 \pm 0.1 \times 10^{-9}$	SW
778	459	$1.0 \pm 0.2 \times 10^{-10}$	NH
779	495	$1.0 \pm 0.2 \times 10^{-10}$	NH
780	485	$4.8 \pm 0.4 \times 10^{-10}$	SW
781	652	$1.4 \pm 0.1 \times 10^{-10}$	NH

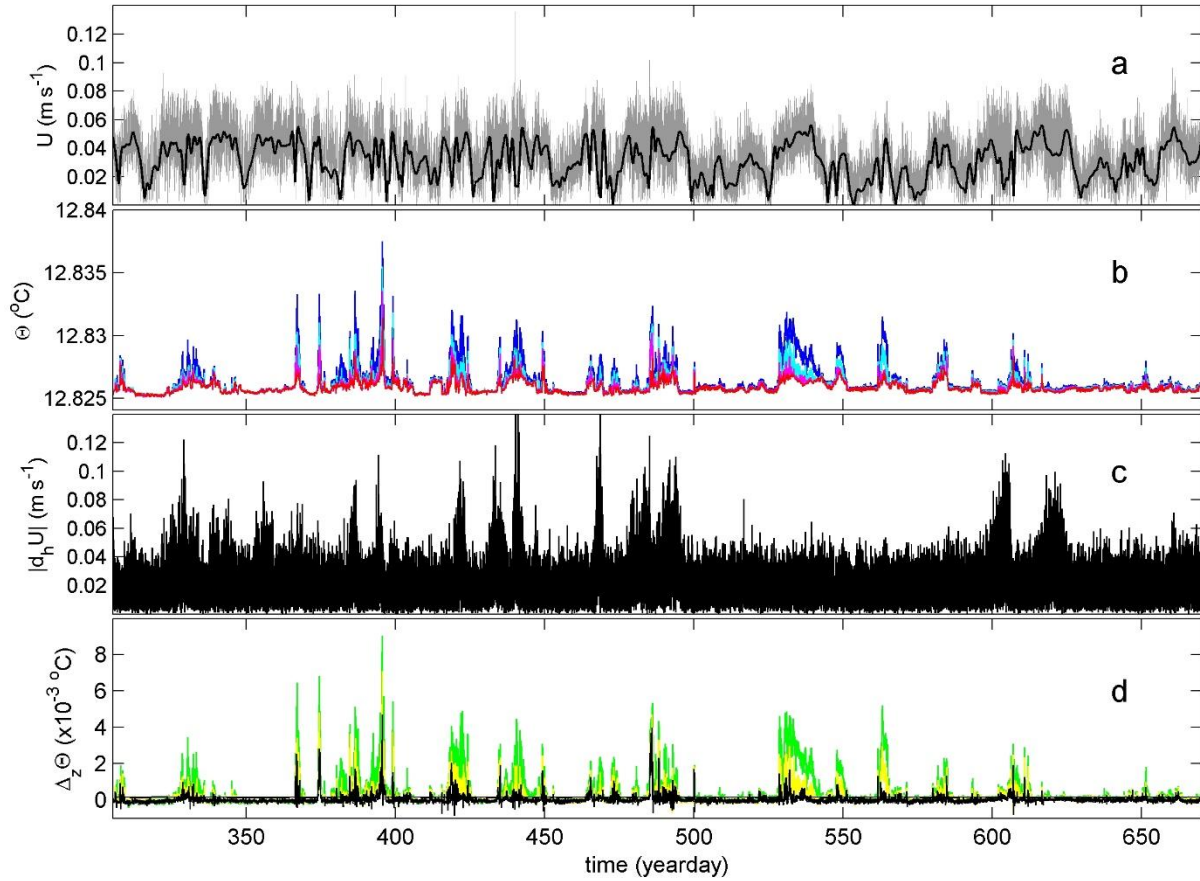
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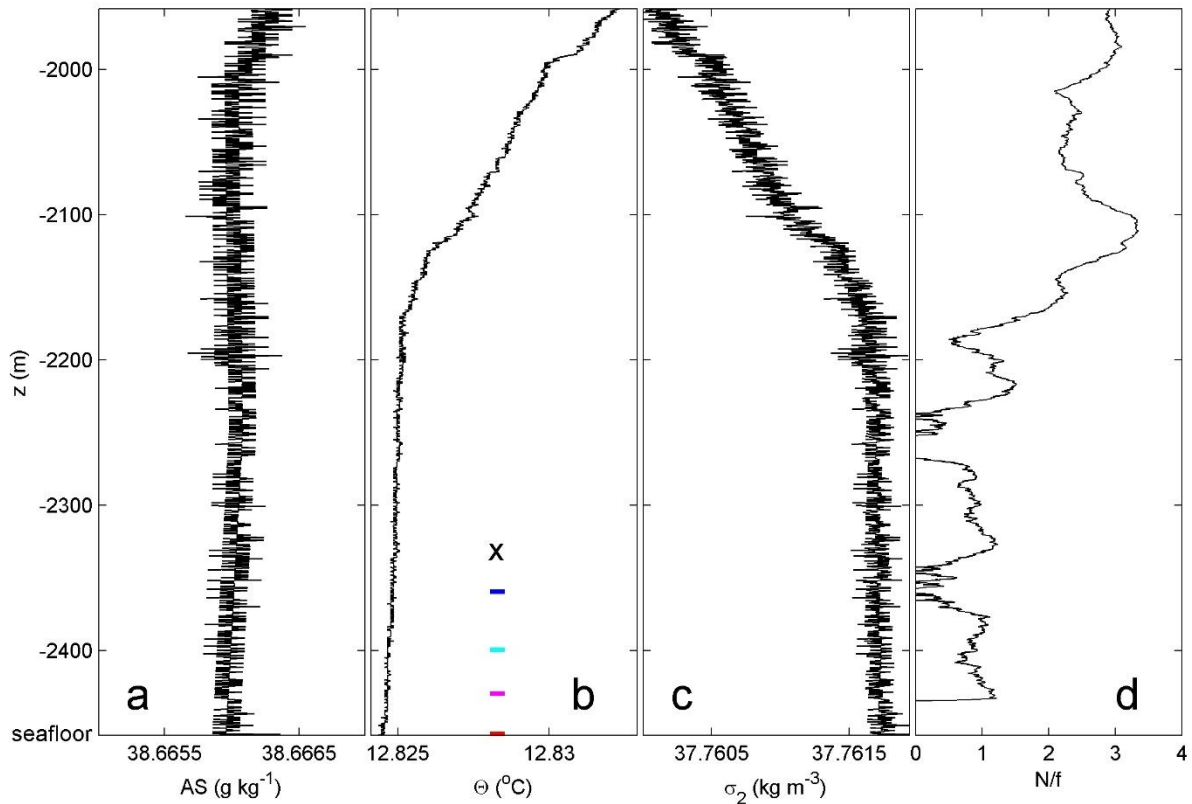
783

784 **Figure 1.** Large-ring mooring site and deployment. (a) Location named "Temp Array" (orange dot) on
 785 map off southern France. The mooring is well east of main neutrino-telescope site "ORCA" of KM3NeT
 786 (Adrián-Martinez et al., 2016) and just northeast of the former "ANTARES" neutrino-telescope site.
 787 Meteorological station is at Porquerolles Island, 20 km due North of Temp Array. Isobaths are drawn
 788 every 100 m. The grey lines indicate Toulon-harbour approach sectors. (b) At sea, during deployment
 789 finalizing the opening of air valves before sinking. The near part of the large steel-tube ring is already
 790 underwater. Almost all buoys of the 45 small-ring compacted vertical lines are visible (for layout see
 791 Appendix A1).

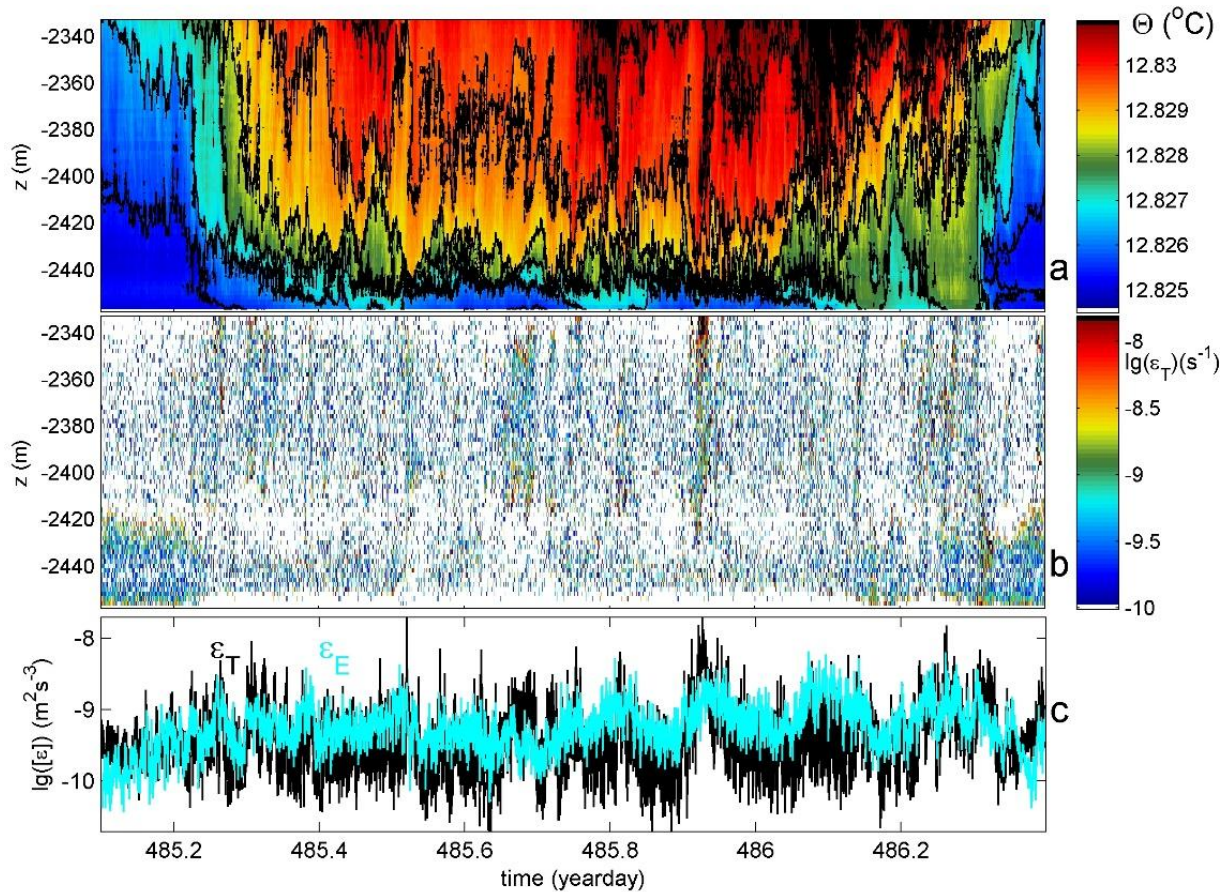
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793
 794 **Figure 2.** Time series of T-sensor and current meter data, for the first year of sampling data. Time in
 795 days of year 2020, +366 in 2021. (a) Unfiltered waterflow speed at $h = 126$ m above seafloor (grey)
 796 and daily filtered (black). (b) Conservative Temperature from vertical line 25, at $h = 1.5$ (red), 29.5
 797 (magenta), 59.5 (cyan) and 99.5 m (blue), corrected for drift and referenced to CTD-data of Fig. 3b.
 798 (c) Amplitude of horizontal waterflow differences. (d) Vertical temperature differences between the
 799 lowest T-sensor and those above from b., between 99.5 and 1.5 (green), between 59.5 and 1.5 m
 800 (yellow), between 29.5 and 1.5 m (black).

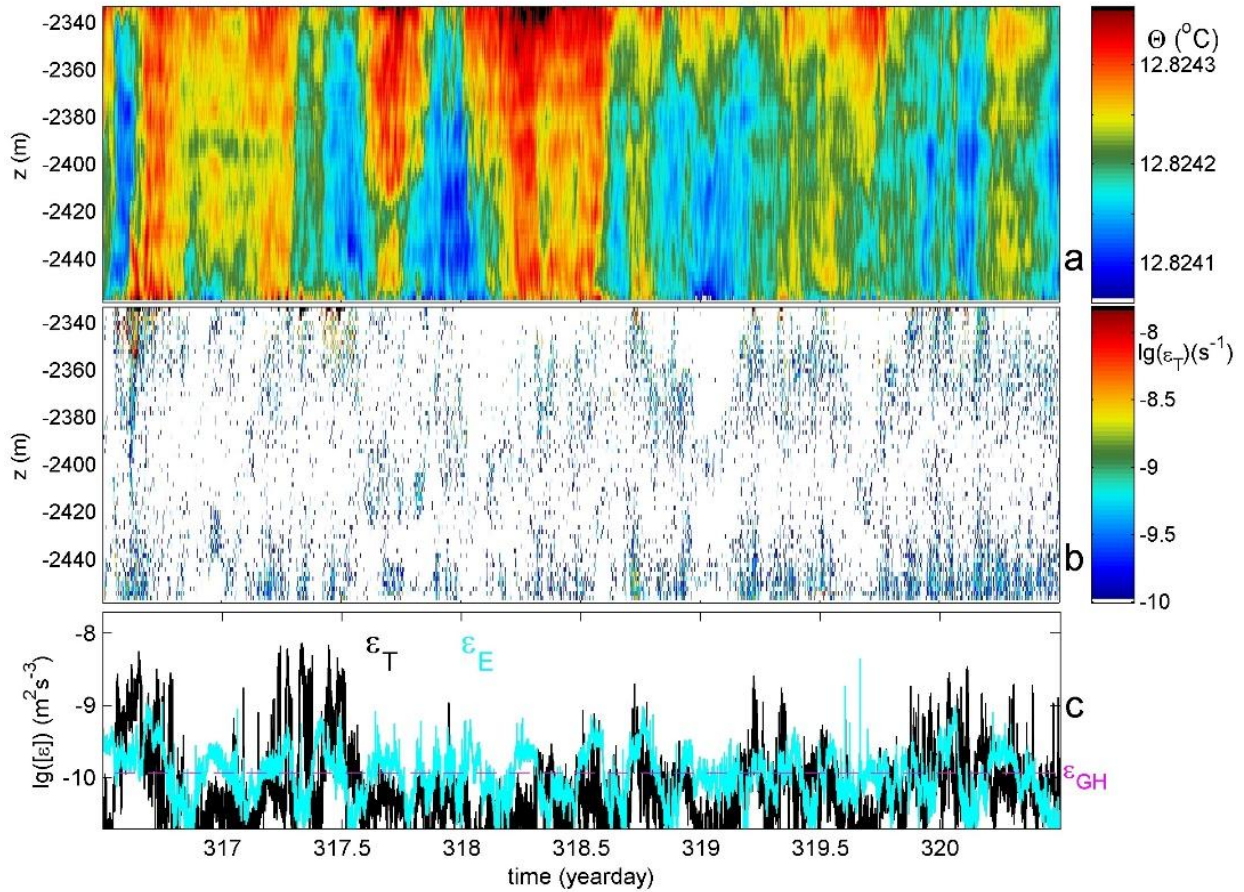


801
 802 **Figure 3.** Lower 500 m of shipborne CTD profile obtained to within 0.5 m from the seafloor (at $z = -2458$
 803 m). (a) Absolute Salinity, with x-axis range similar to that of b. in terms of contribution to density
 804 variations. (b) Conservative Temperature. The colored ticks indicate the vertical positions of the four T-
 805 sensors of which data are displayed in Fig. 2b. The 'x' indicates the position of the current meter. (c)
 806 Density anomaly referenced to $2 \times 10^7 \text{ N m}^{-2}$. (d) Ratio of 25-m scale buoyancy frequency 'N' over local
 807 inertial frequency 'f'.
 808



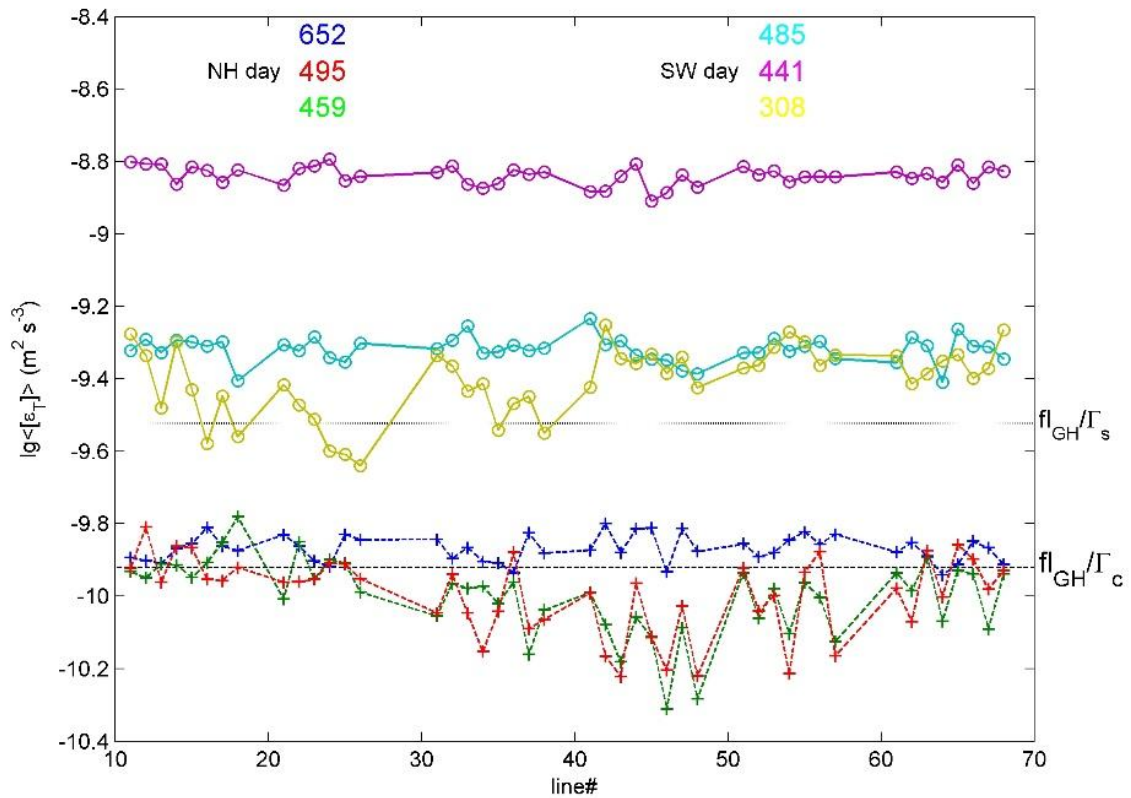
809
 810 **Figure 4.** A 1.3-day period of relatively strong stratification with maximum small-scale buoyancy
 811 frequency $N_{\max} = 6f$, for data from vertical line 15. (a) Time-depth plot of Conservative Temperature
 812 with black contours every 0.001°C . The horizontal axis is at the seafloor. (b) Logarithm of non-averaged
 813 turbulence dissipation rate from data in a. using Thorpe (1977) method. (c) Time series of logarithm of
 814 data from b. averaged over 124-m vertical extent of T-sensors (black), compared with calculations using
 815 Ellison (1957) method (cyan) with high-pass filter ‘hpf’ cut-off from Fig. A2a.

816

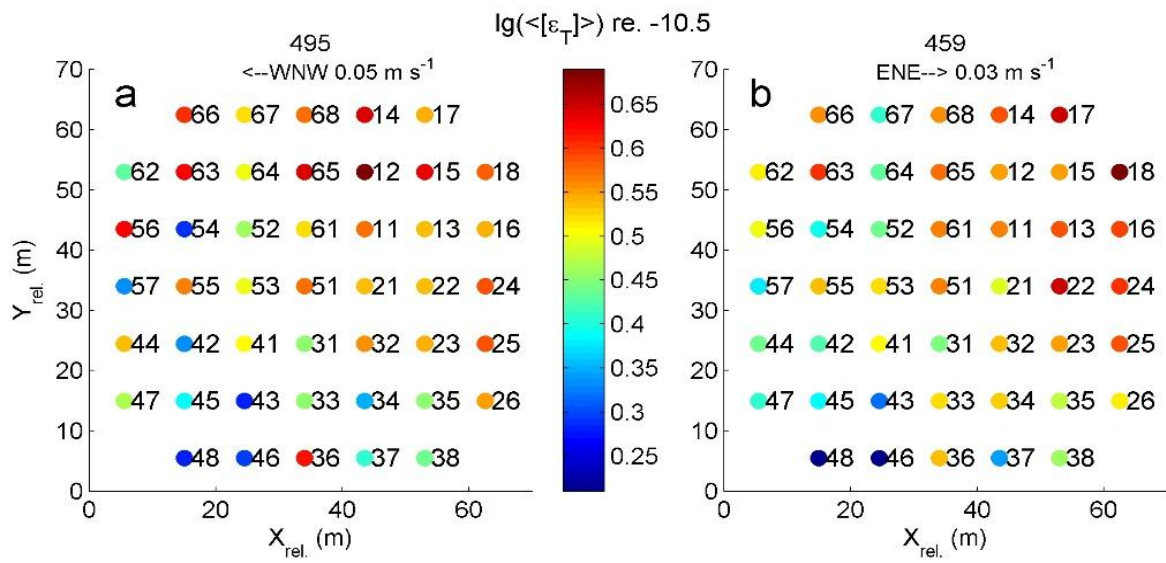


817
 818 **Figure 5.** As Fig. 4, but for four days of near-homogeneous conditions with mean $N \approx 0.5f$, very weak
 819 stratification alternated with convectively unstable periods. For c., the hpf cut-off for determining ε_E is
 820 shown in Fig. A2b and the magenta-dashed line indicates the average turbulence dissipation rate
 821 attributed to geothermal heating (see text).

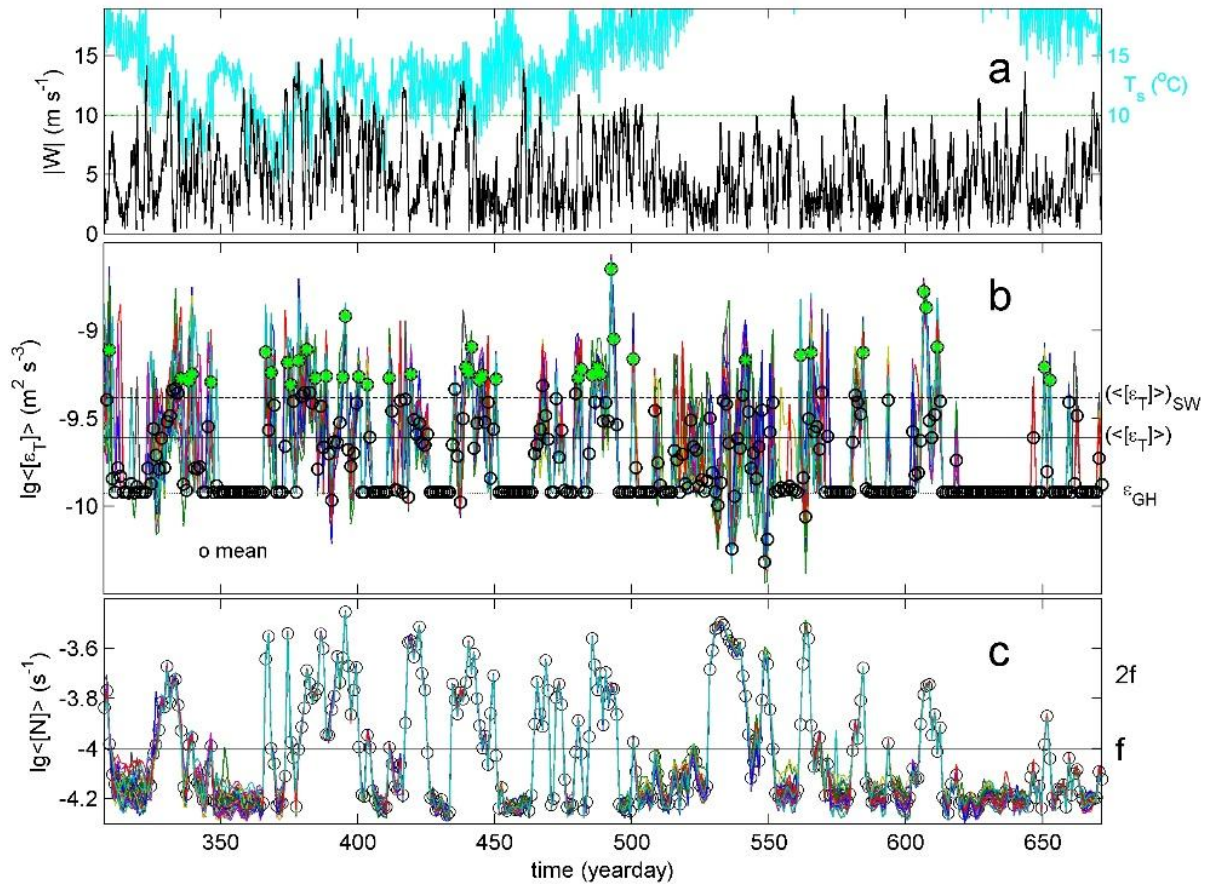
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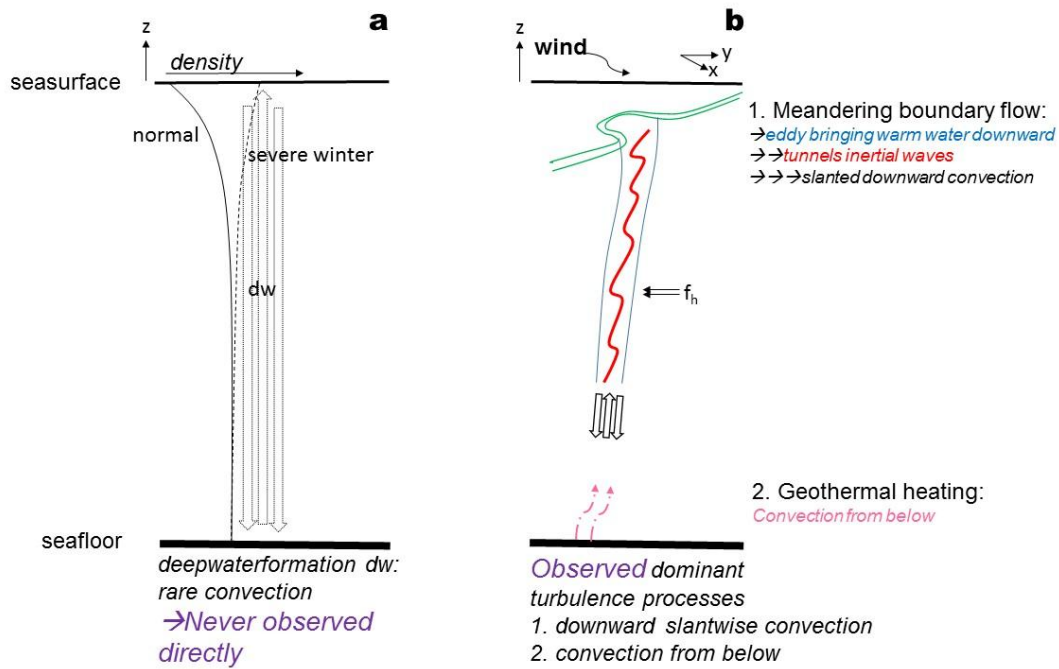
823
 824 **Figure 6.** Limited statistics of Thorpe (1977) method (logarithm of) turbulence dissipation rates averaged
 825 over 124 m vertically given for six 12-h periods indicated by day-number, as a function of all 45 lines
 826 that are indicated by their number 'line#'. Two thresholds are given as a function of general average
 827 buoyancy flux ' fl_{GH} ' from geothermal heating, divided by mixing coefficient for convection-turbulence
 828 $\Gamma_c = 0.5$ (Dalziel et al., 2008) and for shear-turbulence $\Gamma_s = 0.2$ (Osborn, 1980; Oakey, 1982). Solid
 829 lines (o) indicate Stratified-Water 'SW' conditions, dashed lines (+) indicate Near-Homogeneous 'NH'
 830 conditions.
 831



832
 833 **Figure 7.** Plan view of 45 lines indicating logarithm relative to a value of -10.5 of time- and vertical-mean
 834 turbulence dissipation rates under NH conditions on days 495 (a) and 459 (b) of Fig. 6. On top, half-day
 835 mean waterflows are indicated.
 836



837
838 **Figure 8.** Yearlong time series of 45-line, daily, and vertically averaged turbulence dissipation rate and
839 stratification values compared with meteorological data. (a) Wind speed (black) and surface temperature
840 (cyan; scale to the right) measured at the station of Porquerolles Island, 20 km north of the mooring-
841 array. The horizontal line is an arbitrary reference line below which near-surface ocean convection may
842 occur under sufficient pre-conditioning. (b) Logarithm of daily and 124-m vertically averaged Thorpe
843 (1977) method turbulence dissipation rate for all 45 lines (colour), including their mean values (black,
844 circles) of which those exceeding twice the overall mean value (green asterisks). A threshold of
845 0.0002°C is applied for $\Theta(h=125\text{ m})-\Theta(h=1\text{ m})$, below which values are forced to mean $\varepsilon_{\text{GH}} = 1.2 \times 10^{-10}$
846 $\text{m}^2\text{ s}^{-3}$, see text. The solid horizontal line indicates the overall mean value (4), the dashed line the mean
847 (5) for periods under SW conditions. (c) Logarithm of corresponding mean buoyancy frequencies from
848 reordered temperature profiles. The horizontal line indicates the local planetary inertial frequency.
849



850

851 **Figure 9.** Sketch of potentially relevant flows and turbulence processes in the large-ring mooring area.

852 (a) Rarely occurring, possibly once every 8 years during a severe winter but never directly observed

853 (Thorpe, 2005), convection-turbulence due to deep dense-water formation when the vertical density

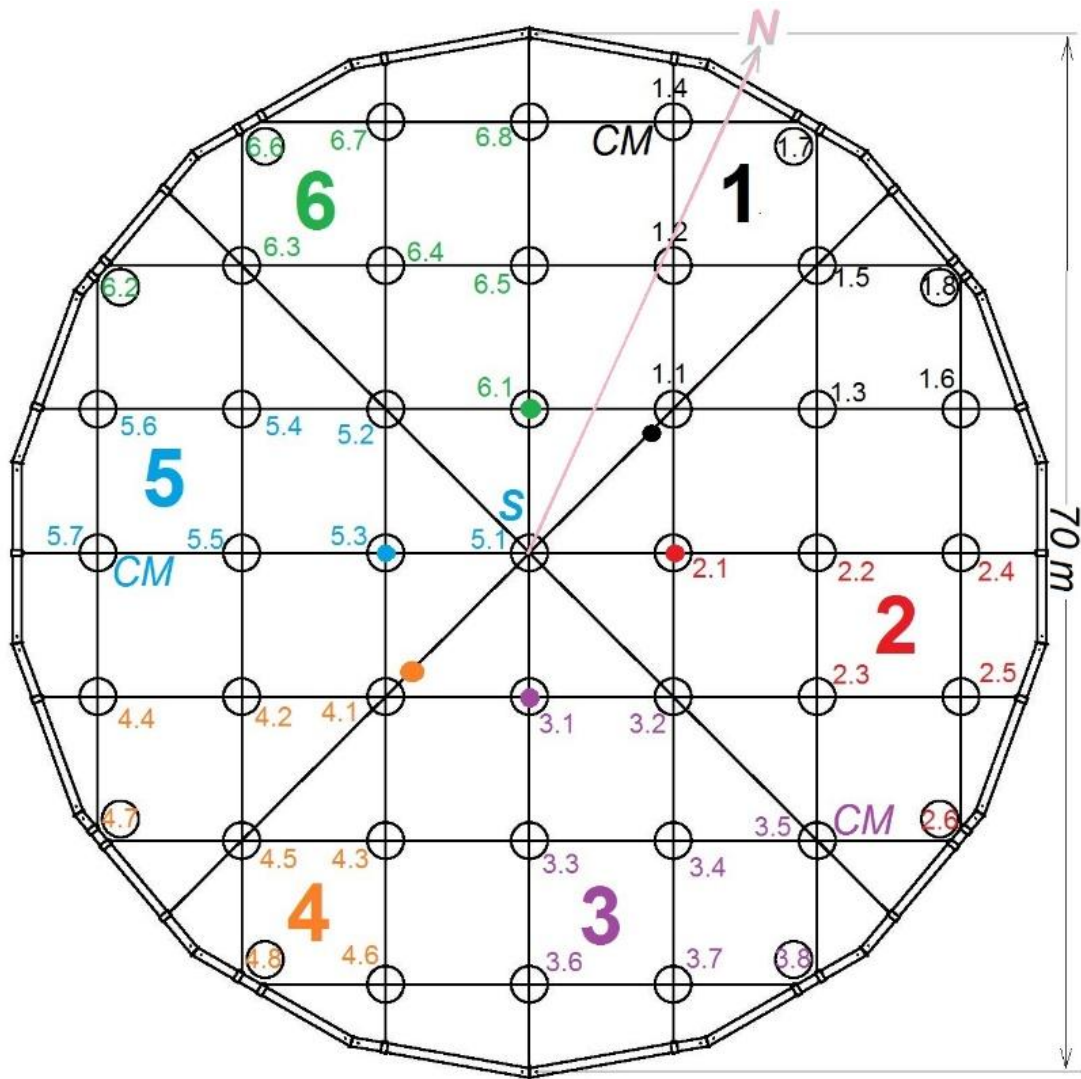
854 profile changes from stable (solid graph) to unstable (dashed graph) over the entire 2500-m depth. (b)

855 Wind-induced boundary flow meandering leading to alternating observations of deep-sea turbulence,

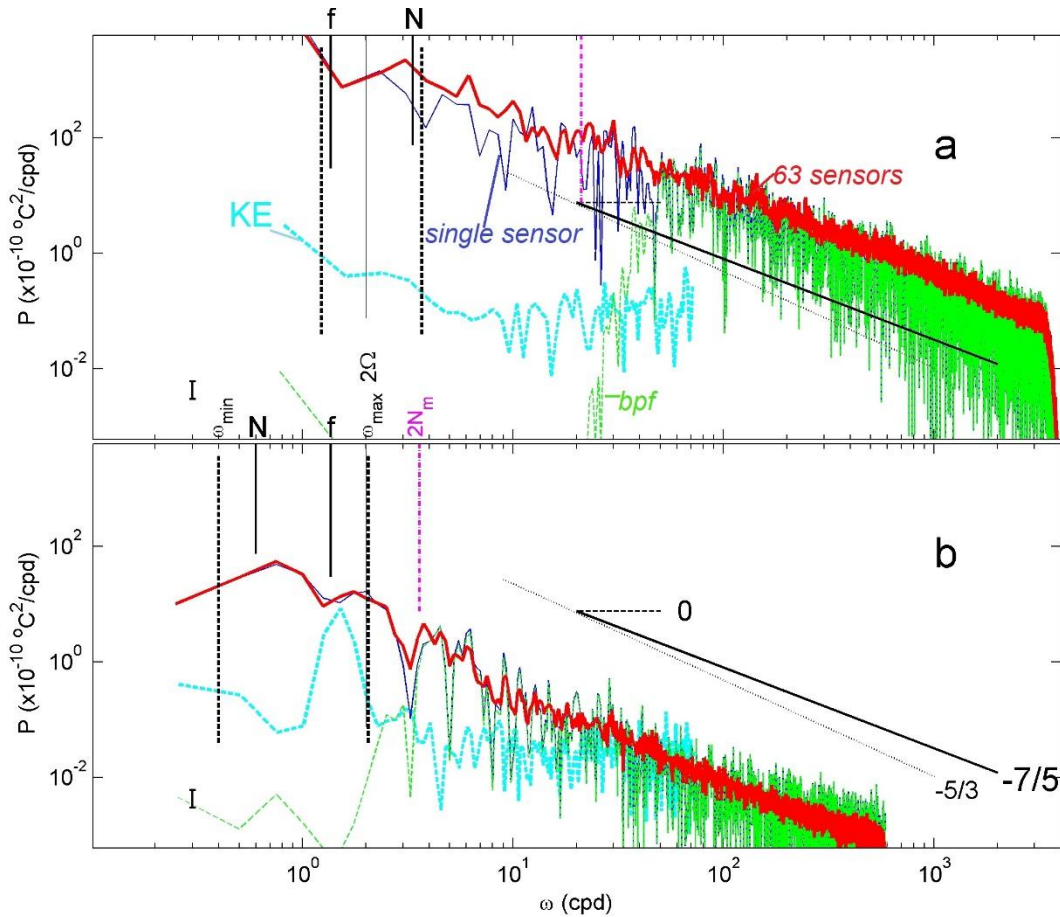
856 from above, via dominant slantwise downward convection, and from below, via geothermal heating.

857 The horizontal Coriolis parameter is indicated by f_h .

858



859
 860 **Figure A1.** Orientation and layout of the large-ring mooring viewed from above, with steel-cable grid and
 861 small rings numbered in six synchronisation groups with colour dots indicating group nodes and
 862 synchroniser 'S' at line 51. Here and elsewhere in the text, lines are indicated without period for short.
 863 Lines 14, 35 and 57 held a waterflow current meter 'CM' at the buoy.
 864



865
 866 **Figure A2.** Spectra demonstrating filter cut-off frequencies for Ellison (1957) method under SW and NH
 867 conditions. Data from line 15. Weakly smoothed (6 degrees of freedom, dof) low-pass filtered ‘lpf’
 868 spectra from a single T-sensor at mid-height (thin solid) with bpf version (thin dashed) is compared with
 869 moderately smoothed (100-dof) lpf spectrum over all 63 T-sensors (thick solid). For comparison, the
 870 corresponding weakly smoothed (10-dof) kinetic energy ‘KE’ spectra are given (thick dashed; arbitrary
 871 vertical scale), averaged over the three current meters. For reference, several frequencies and
 872 turbulence-range spectral slopes are given, see text. (a) SW period of Fig. 4, with filter cut-off following
 873 a scaling of time-mean maximum 2-m small-scale buoyancy frequency N_m and lpf cut-off at 3000 cpd
 874 (cycles per day). (b) NH period of Fig. 5, with hpf cut-off fixed near $2N_m$ and lpf cut-off at 500 cpd.
 875