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Diapycnal mixing across the photic zone of the NE-Atlantic

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44 **Abstract.** Variable physical conditions such as vertical turbulent exchange, internal wave and
45 mesoscale eddy action, affect the availability of light and nutrients for phytoplankton
46 (unicellular algae) growth. It is hypothesized that changes in ocean temperature may affect
47 ocean vertical density stratification, which may hamper vertical exchange. In order to quantify
48 variations in physical conditions in the Northeast Atlantic Ocean, we sampled a latitudinal
49 transect along $17\pm 5^\circ\text{W}$ between 30 and 62°N in summer. A shipborne Conductivity-
50 Temperature-Depth CTD-instrumented package was used with a custom-made modification of
51 the pump-inlet to minimize detrimental effects of ship motions on its data. Thorpe-scale
52 analysis was used to establish turbulence values for the upper 500 m near the surface from 3 to
53 6 profiles obtained in a short CTD-yoyo, 3 to 5 h after local sunrise. From south to north,
54 temperature decreased together with stratification while turbulence values weakly increased or
55 remained constant. Vertical turbulent nutrient fluxes across the stratification ~~were~~ did not found
56 to significantly vary with latitude. This apparent lack of correspondence between turbulent
57 mixing and temperature is suggested to be due to internal waves breaking and acting as a
58 potential feed-back mechanism. Our findings suggest that nutrient availability for
59 phytoplankton in the euphotic surface waters may not be affected by the physical process of
60 global warming.
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62

63 **1 Introduction**

64 The physical environment is important for ocean life, including variations therein. For
65 example, the sun stores heat in the ocean with a stable vertical density stratification as result.
66 Generally, stratification hampers vertical turbulent exchange because of the required work
67 against (reduced) gravity before turbulence can take effect. It thus hampers a supply of nutrients
68 via a turbulent flux from deeper waters to the photic zone. However, stratification supports
69 internal waves, which (i) may move near-floating particles like phytoplankton (unicellular
70 algae) up- and down towards and away from the surface, and (ii) may induce enhanced
71 turbulence via vertical current differences (shear) resulting in internal waves breaking (Denman
72 and Gargett, 1983). Such changes in the physical environment are expected to affect the
73 availability of phytoplankton growth factors such as light and nutrients.

74 Climate models predict that global warming will reduce vertical mixing in the oceans (e.g.,
75 Sarmiento et al., 2004). Mathematical models on system stability suggest that reduced mixing
76 may generate chaotic behaviour in phytoplankton production, thereby enhancing variability in
77 carbon export into the ocean interior (Huisman et al., 2006). ~~None~~ However, none of these
78 models include potential feed-back systems like internal wave action or mesoscale eddy
79 activity. From observations in the relatively shallow North Sea it is known that the strong
80 seasonal temperature stratification is marginally stable, as it supports internal waves and shear
81 to such extent that sufficient nutrients are replenished from below to sustain the late-summer
82 bloom (van Haren et al., 1999). This challenges the current paradigm in climate models.

83 In this paper, the objective is to resolve the effect of vertical stratification and turbulent
84 mixing on nutrient supply to the photic zone of the open ocean. For this purpose, upper-500-
85 m-ocean shipborne Conductivity-Temperature-Depth CTD-observations were made in
86 association with those on dissolved inorganic nutrients during a survey along a transect in the
87 NE-Atlantic Ocean from mid- to high-latitudes in summer. Throughout the survey,
88 meteorological and sea-state conditions were favourable for adequate sampling and wind
89 speeds varied little between 5 and 10 m s⁻¹, independent of locations. All CTD-observations

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90 were made far from lateral, continental boundaries and at least 1000 m vertically away from
91 bottom topography. The NE-Atlantic is characterized by abundant (sub-)mesoscale eddies
92 especially in the upper ocean (Charria et al., 2017) and which that influence local plankton
93 communities (Hernández- Hernández et al., 2020). The area also shows continuous abundant
94 internal wave activity away from topographic sources and sinks, with the semidiurnal tide as a
95 main source from below and atmospherically induced inertial motions from above (e.g., van
96 Haren, 2005; 2007). The present research ~~is~~ complements research based on photic zone (upper
97 100 m) observations obtained along the same transect using a slowly descending turbulence
98 microstructure profiler next to CTD-sampling eight years earlier (Jurado et al., 2012). Their
99 data demonstrated a negligibly weak increase in turbulence values with significant decreases in
100 stratification going north. However, no nutrient data were presented and no turbulent nutrient
101 fluxes could be computed. In another study (Mojica et al., 2016), macro-nutrients and their
102 vertical gradients were presented for the upper 200 m and both were found to increase from
103 south to north. The present observations go deeper to 500 m, also across the non-seasonal more
104 permanent stratification. Moreover, coinciding measurements were made of the distributions
105 of macro-nutrients and dissolved iron. This allows vertical turbulent nutrient fluxes to be
106 computed. It leads to a hypothesis concerning a physical feed-back mechanism that may
107 controls changes in stratification.

108 109 **2 Materials and Methods**

110 Between 22 July and 16 August 2017, observations were made from the R/V Pelagia in the
111 Northeast Atlantic Ocean at stations along a transect from Iceland, starting around 60°N, to the
112 Canary Islands, ending around 30°N, (Fig. 1). The transect was more or less in meridional
113 direction, with stations along 17±5°W, all in the same time zone (UTC-1 h = local time LT).
114 Full water-depth Rosette bottle water sampling was performed at most stations.

115 Samples for dissolved inorganic macro-nutrients were filtered through 0.2 µm Acrodisc
116 filter and stored frozen in a HDPE pony-vial (nitrate, nitrite and phosphate) or at 4°C (silicate)

117 until analysis. Nutrients were analysed under temperature controlled conditions using a
118 QuAAtro Gas Segmented Continuous Flow Analyser. All measurements were calibrated with
119 standards diluted in low nutrient seawater in the salinity range of the stations to ensure that
120 analysis remained within the same ionic strength. Phosphate (PO_4), nitrate- plus nitrite (NO_x),
121 were measured according to Murphy and Riley (1962) and Grasshoff et al. (1983), respectively.

122 Silicate was analysed using the procedure of Strickland and Parsons (1968).

123 Absolute and relative precision were regularly determined for reasonably high
124 concentrations in an in-house standard. For phosphate, the standard deviation was 0.028 μM
125 (N = 30) for a concentration of 0.9 μM ; Hence the relative precision was 3.1%. For nitrate, the
126 values were 0.14 μM (N = 30) for a concentration of 14.0 μM , so that the relative precision was
127 1.0%. For silicate, the values were 0.09 μM (N = 15) for a concentration of 21.0 μM , so that
128 the relative precision was 0.4%. The detection limits were 0.007, 0.012 and 0.008 μM , for
129 phosphate, nitrate and silicate, respectively.

130 For dissolved iron samples, the ultraclean “Pristine” sampling system for trace metals was
131 used (Rijkenberg et al., 2015). All bottles used for storage of reagents and samples were cleaned
132 according to an intensive three step cleaning protocol described by Middag et al. (2009).
133 Dissolved iron concentrations were measured shipboard using a Flow Injection—
134 Chemiluminescence method with preconcentration on iminodiaceticacid (IDA) resin as
135 described by De Baar et al. (2008) and modified by Klunder et al. (2011). In order to validate
136 the accuracy of the system, standard reference seawater (SAFe) was measured regularly in
137 triplicate (Johnson et al. 1997).

138 At 19 out of 32 stations a yoyo consisting of 3 to 6 casts, totaling 72 casts, -of electronic
139 CTD profiles was done to monitor the temperature-salinity variability and to establish turbulent
140 mixing values from 5 to 500 m below the ocean surface. The yoyo casts were made
141 consecutively and took between 1 and 2 hours per station. They were mostly obtained in the
142 morning: at ten stations between 6 and 8 LT, at eight stations between 8 and 10 LT, and at one
143 station in the afternoon, around noon. As the observations were made in summer, the latitudinal

144 difference in sunrise was 1.5 h between the northernmost (earlier sunrise) and southernmost
145 stations. This difference is taken into account and sampling times are referenced to time after
146 local sunrise. It is assumed that the stations sampled just after sunrise more or less reflect the
147 upper ocean conditions of (late-) nighttime cooling convection so that vertical near-
148 homogeneity was at a maximum, and near-surface stratification at a minimum, while the late
149 morning and afternoon stations reflected daytime stratifying near-surface conditions due to the
150 stabilizing solar insolation.

151

152 **2.1 Instrumentation and modification**

153 A calibrated SeaBird 911plus CTD was used. The CTD data were sampled at a rate of 24
154 Hz, whilst lowering the instrumental package at a speed of 1 m s⁻¹. The data were processed
155 using the standard procedures incorporated in the SBE-software, including corrections for cell
156 thermal mass (Lueck, 1990) using the parameter setting of Mensah et al. (2009) and sensor
157 time-alignment. All other analyses were performed using Conservative Temperature (Θ),
158 absolute salinity SA and density anomalies σ_θ referenced to the surface using the Gibbs
159 SeaWater-software (IOC, SCOR, IAPSO, 2010).

160 Observations were made with the CTD upright rather than horizontal in a lead-weighted
161 frame without water samplers to minimize artificial turbulent overturning. Variable speeds of
162 the flow passing the temperature and conductivity sensors will cause artificial temperature and
163 thus apparent turbulent overturning, noticeable in near-homogeneous waters such as found near
164 the surface during nighttime convection. To eliminate variable flow speeds, a custom-made
165 assembly with pump in- and outlet tubes and tube-ends of exactly the same diameter was
166 mounted to the CTD as described in van Haren and Laan (2016). This reduces frictional
167 temperature effects of typically ± 0.5 mK due to fluctuations in pump speed of ± 0.5 m s⁻¹ when
168 standard SBE-tubing is used (Appendix A1). The effective removal of the artificial temperature
169 effects using the custom-made assembly is demonstrated in Fig. 2, in which surface wave action
170 via ship motion is visible in the CTD-pressure record, but not in its temperature variations

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171 record. For example, at station 32 the CTD was lowered in moderate sea state conditions with
172 surface waves of maximum 2 m crest-trough. The surface waves are recorded by pressure
173 variations as a result of ship motions (Fig. 2a). In the upper 40 m near the surface, the waters
174 were near-homogeneous, with temperature variations well within ± 0.5 mK (Fig. 2b). The ΔT -
175 variations did not vary with the surface wave periodicity of about 10 s. No correlation is found
176 between data in Fig. 2b and Fig. 2a. This effective removal of ship motion in CTD-temperature
177 data is confirmed for the entire 500 m depth-range in average spectral information (Fig. 2c-e).
178 In the power spectra, the pressure gradient $dp/dt \sim$ CTD-velocity shows a clear peak around 0.1
179 cps, short for cycles per s, which correspond to a period of 10 s. Such a peak is absent in both
180 spectra of temperature T and density anomaly referenced to the surface σ_θ . The correlation
181 between dp/dt and T is not significantly different from zero (Fig. 2d,e). With conventional
182 tubing and tube-ends, the surface wave variations would show in such ΔT -graph (van Haren
183 and Laan, 2016). Without the effects of ship motions, considerably less corrections need to be
184 applied for turbulence calculations (see below).

185

186 **2.2 Ocean turbulence calculation**

187 Turbulence is quantified using the analysis method by Thorpe (1977) on density (ρ)
188 inversions of less dense water below a layer of denser water in a vertical (z) profile. Such
189 inversions are interpreted as turbulent overturns of mechanical energy mixing. Vertical
190 turbulent kinetic energy dissipation rate (ϵ) is a measure of the amount of kinetic energy put in
191 a system for turbulent mixing. It is proportional to turbulent diapycnal flux (of density) $K_z dp/dz$.
192 In practice it is determined by calculating overturning scales with magnitude $|d|$, just like
193 turbulent eddy diffusivity (K_z). The vertical density stratification is indicated by dp/dz . The
194 turbulent overturning scales are obtained after reordering the potential density profile $\sigma_\theta(z)$,
195 which may contain inversions, into a stable monotonic profile $\sigma_\theta(z_s)$ without inversions
196 (Thorpe, 1977). After comparing raw and reordered profiles, displacements $d = \min(|z -$
197 $z_s|) \cdot \text{sgn}(z - z_s)$ are calculated that generate the stable profile. Then,

198 $\epsilon = 0.64d^2N^3$ [m²s⁻³], (1)

199 where $N = \{-g/\rho(dp/dz + g\rho/c_s^2)\}^{1/2}$ (e.g., Gill, 1982) denotes the buoyancy frequency (~square-
 200 root of stratification, squared as is clear from the equation) computed from the reordered profile.

201 Here, g is the acceleration of gravity and c_s the speed of sound reflecting pressure-
 202 compressibility effects. N is computed over a typical vertical length-scale of $\Delta z = 100$ m, which
 203 more or less represents the scale of large internal waves that are supported by the density
 204 stratification. The numerical constant of 0.64 in (1) follows from empirically relating the
 205 overturning scale magnitude with the Ozmidov scale L_O of largest possible turbulent overturn
 206 in a stratified flow: $(L_O/d)_{rms} = 0.8$ (Dillon, 1982), a mean coefficient value from many
 207 realizations. Using $K_z = \Gamma\epsilon N^{-2}$ and a mean mixing efficiency coefficient of $\Gamma = 0.2$ for the
 208 conversion of kinetic into potential energy for ocean observations that are suitably averaged
 209 over all relevant turbulent overturning scales of the mix of shear-, current differences, and
 210 convective, buoyancy driven, turbulent overturning in large Reynolds number flow conditions
 211 (e.g., Osborn, 1980; Oakey, 1982; Ferron et al., 1998; Gregg et al., 2018), we find,

212 $K_z = 0.128d^2N$ [m²s⁻¹]. (2)

213 This parametrization is also valid for the upper ocean, as has been shown extensively by Oakey
 214 (1982) and recently confirmed by Gregg et al. (2019). The inference is that the upper ocean
 215 may be weakly stratified at times, but stratification and turbulence vary considerably with time
 216 and space. Sufficient averaging collapses coefficients to the mean values given above. This is
 217 confirmed in recent numerical modeling by Portwood et al. (2019).

218 As K_z is a mechanical turbulence coefficient it is not property-dependent like a molecular
 219 diffusion coefficient that is about 100-fold different for temperature compared to salinity. K_z is
 220 thus the same for all turbulent transport calculations no matter what gradient of what property.
 221 For example, the vertical turbulent flux of dissolved iron is computed as $K_z d(DFe)/dz$.

222 According to Thorpe (1977), results from (1) and (2) are only useful after averaging over
 223 the size of a turbulent overturn instead of using single displacements. Here, root-mean-square-
 224 displacement values d_{rms} are not determined over individual overturns, as in Dillon (1982), but
 225 over 7 m vertical intervals (equivalent to about 200 raw data samples) that just exceed average

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226 L_0 . This avoids the complex distinction of smaller overturns in larger ones and allows the use
227 of a single length scale of averaging. As a criterion for determining overturns we only used
228 those data of which the absolute value of difference with the local reordered value exceeds a
229 threshold of $7 \times 10^{-5} \text{ kg m}^{-3}$, which corresponds to applying a threshold of $1.4 \times 10^{-3} \text{ kg m}^{-3}$ to raw
230 data variations (e.g., Galbraith and Kelley, 1996; Stansfield et al., 2001; Gargett and Garner,
231 2008). Vertically averaged turbulence values, short for averaged ε - and K_z -values from (1) and
232 (2), can be calculated to within an error of a factor of two to three, approximately. As will be
233 demonstrated below, this is considerably less spread in values than the natural turbulence values
234 variability over typically four orders of magnitude at a given position and depth in the ocean
235 (e.g., Gregg, 1989).

236

237 **3 Results**

238 **3.1 Physical parameters**

239 An early morning vertical profile of density anomaly in the upper 500 m at a northern
240 station (Fig. 3a) is characterized by a near-homogeneous layer of about 15 to 40 m, which is
241 above a layer of relatively strong stratification and a smooth moderate stratification deeper
242 below. In the near-homogeneous upper layer, in this example $z > -30$ m, relatively large
243 turbulent overturn displacements can be found of $d = \pm 20$ m (Fig. 3b): so called large density
244 inversions. For $-200 < z < -30$ m, large turbulent overturns are few and far between. Turbulence
245 dissipation rate (Fig. 3c) and eddy diffusivity (Fig. 3d) are characterized by relatively small
246 displacement sizes of less than 5 m. For $z < -200$ m, displacement values weakly increase with
247 depth, together with stratification ($\sim N^2$; Fig. 3e). Between $-30 < z < 0$ m, turbulence dissipation
248 rate values between $< 10^{-11}$ and $> 10^{-8} \text{ m}^2 \text{ s}^{-3}$ are similar to those found by others, using
249 microstructure profilers (e.g., Oakey, 1982; Gregg, 1989), lowered acoustic Doppler current
250 profiler or CTD-Thorpe scale analysis (e.g., Ferron et al., 1998; Walter et al., 2005; Kunze et
251 al., 2006). Here, eddy diffusivities are found between $< 10^{-5}$ and $3 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$ and these values
252 compare with previous near-surface results (Denman and Gargett, 1983). The relatively small

253 $|d| < 5$ m displacements (Fig. 3b) are genuine turbulent overturns, and they resemble ‘Rankine
254 vortices’, a common model of cyclones (van Haren and Gostiaux, 2014), as may be best visible
255 in this example in the large turbulent overturn near the surface. The occasional erratic
256 appearance in individual profiles, sometimes still visible in the ten-profile means, reflects
257 smaller overturns in larger ones.

258 A mid-morning profile at a southern station shows different characteristics (Fig. 4),
259 although 500 m vertically averaged turbulence values are similar to within 10% of those of the
260 northern station. This 10% variation is well within the error bounds of about a factor of two. At
261 this southern station, the near-surface layer is stably stratifying (Fig. 4a) and displays few
262 overturning displacements (Fig. 4b), while the interior demonstrates rarer but occasional
263 intense turbulent overturning (at $z = -160$ m in Fig. 4), presumably due to internal wave
264 breaking. At greater depths, stratification ($\sim N^2$; Fig. 4e) weakly decreases, together with ε (Fig.
265 4c) and K_z (Fig. 4d).

266 Latitudinal overviews are given in Fig. 5 for: Average values over the upper $z > -15$ m,
267 which covers the diurnal mainly convective turbulent mixing range from the surface, average
268 values between $-100 < z < -25$ m, which covers the seasonal strong stratification, and average
269 values between $-500 < z < -100$ m, which covers the more permanent moderate stratification.
270 Noting that all panels have a vertical axis representing a logarithmic scale, variations over
271 nearly four orders of magnitude in turbulence dissipation rate (Fig. 5a) and eddy diffusivity
272 (Fig. 5b) are observed between ~~casts at the same station~~~~individual average values~~. This
273 variation in magnitude is typically found in near-surface open-ocean turbulence microstructure
274 profiles (e.g., Oakey, 1982). Still, considerable variability over about two orders of magnitude
275 is observed between the ~~3 to 6 cast~~ averages ~~from the at a different~~~~particular~~ stations. This
276 variation in station- and vertical averages far exceeds the instrumental error bounds of a factor
277 of two (0.3 on a log-scale), and thus reveals local variability. The turbulence processes occur
278 ‘intermittently’.

279 The observed variability over two orders of magnitude between yoyo-casts at a single
280 station may be due to active convective overturning during early morning in the near-

281 homogeneous upper layer, or due to internal wave breaking and sub-mesoscale variability
282 deeper down. Despite the large variability at stations, trends are visible between stations in the
283 upper 100 m over the 32° latitudinal range going poleward: Buoyancy frequency (~ square root
284 of stratification) steadily decreases significantly (p-value < 0.05) given the spread of values at
285 given stations, while turbulence values vary insignificantly with latitude as they remain the
286 same or weakly increase by about half an order of magnitude (about a factor of 3). At a given
287 depth range, turbulence dissipation rate roughly follow a log-normal distribution with standard
288 deviations well exceeding half an order of magnitude. The comparison of latitudinal variations
289 with the (log-normal) distribution are declared insignificant with p > 0.05 when the mean values
290 are found within 2 standard deviations (see Appendix A2). This is not only performed for
291 turbulence dissipation rate, but also for other quantities. The trends suggest only marginally
292 larger ~~(convective)~~ turbulence going poleward, which is possibly due to larger cooling from
293 above and larger internal wave breaking deeper down going poleward. It is noted that the results
294 are somewhat biased by the sampling scheme, which changed from 3 to 4 h after sunrise
295 sampling at high latitudes to 4 to 5 h after sunrise sampling at lower latitudes, see the sampling
296 hours after local sunrise in (Fig. 5d). Its effect is difficult to quantify, but should not show up
297 in turbulence values from deeper down ($-500 < z < -100$ m).

298 Between $-500 < z < -100$ m, no clear significant trend with latitude is visible in the
299 turbulence values (Fig. 5a,b), although $[K_z]$ weakly increases with increasing latitude at all
300 levels between $-500 < z < 0$ m, while buoyancy frequency stratification significantly decreases
301 (Fig. 5c). The ~~deeper~~ data from well-stratified waters deeper down thus show the same
302 latitudinal trend as unambiguously confirm the observations from the near-surface layers. Our
303 turbulence values from CTD-data also confirm previous results by Jurado et al. (2012) who
304 made microstructure profiler observations from the upper $z > -100$ m along the same transect.
305 Their results showed turbulence values remain unchanged over 30° latitude or increase by at
306 most one order of magnitude, depending on depth level. Their ‘mixed’ layer ($z > -25$ m)
307 turbulence values are similar to our $z > -15$ m values and half to one order of magnitude larger
308 than the present deeper observations. The slight discrepancy in values averaged over $z > -25$ m

309 may point at either i) a low bias due to a too strict criterion of accepting density variations for
310 reordering applied here, or ii) a high bias of the ~10-m largest overturns having similar velocity
311 scales (of about 0.05 m s^{-1}) as their 0.1 m s^{-1} slowly descending SCAMP microstructure profiler.
312 At greater depths, $-500 < z < -100 \text{ m}$, it is seen in the present observations that the spread in
313 turbulence values over four orders of magnitude at a particular station is also large. This spread
314 in values suggests that dominant turbulence processes show similar intermittency in weakly (at
315 high-latitudes $N \approx 10^{-2.5} \text{ s}^{-1}$) and moderately (at mid-latitudes $N \approx 10^{-2.2} \text{ s}^{-1}$) stratified waters,
316 respectively, for given resolution of the instrumentation.

317 Mean values of N are larger by half an order of magnitude in the seasonal pycnocline than
318 those near the surface and in the more permanent stratification below (Fig. 5). Such local
319 vertical variations in N ~~have~~are the same range of variation as observed horizontally across
320 latitudes $[30, 62]^\circ$ per depth level.

321

322 3.2 Nutrient distributions and fluxes

323 Vertical profiles of macro-nutrients generally resemble those of density anomaly in the
324 upper $z > -500 \text{ m}$ (Fig. 6). In the south, low macro-nutrient values are generally distributed over
325 a somewhat larger near-surface mixed layer. The mixed layer depth, defined as the depth at
326 which the temperature difference with respect to the surface was 0.5°C (Jurado et al., 2012),
327 varies between about 20 and 30 m on the southern end of the transect and weakly becomes
328 shallower with latitude (Fig. 7a). This weak trend may be expected from the summertime wind
329 conditions that also barely vary with latitude (Fig. 7b,c). In contrast, the euphotic zone, defined
330 as the depth of the 0.1% irradiance penetration level (Mojica et al., 2015), demonstrates a clear
331 latitudinal trend decreasing from about 150 to 50 m (Fig. 7a). For $z < -100 \text{ m}$ below the seasonal
332 stratification, vertical gradients of macro-nutrients are large (Fig. 6b-d). Macro-nutrient values
333 become more or less independent of latitude at depths below $z < -500 \text{ m}$. Dissolved iron profiles
334 differ from macro-nutrient profiles, notably in the upper layer near the surface (Fig. 6a). At
335 some southern stations, dissolved iron and to a lesser extent also phosphate, have relatively

336 high concentrations closest to the surface. These near-surface concentration increases suggest
337 atmospheric sources, most likely Saharan dust deposition (e.g., Rijkenberg et al., 2012).

338 As a function of latitude in the near-surface ‘mixed’ layer (Fig. 87), the vertical turbulent
339 fluxes of dissolved iron and phosphate (representing the macro-nutrients, for graphical reasons,
340 see the similarity in profiles in Fig.6b-d) ~~is-are~~ found constant or insignificantly ($p > 0.05$)
341 increasing (Fig. 87d). Here, the mean eddy diffusivity values for the near-surface layer as
342 presented in Fig. 5 are used for computing the fluxes. It is noted that in this layer turbulent
343 overturning (Figs 3b, 4b) is larger and nutrients are mainly depleted (Fig. 6), except when
344 replenished from atmospheric sources. Hereby, lateral diffusion is not considered important.
345 More interestingly, the vertical turbulent fluxes of nutrients across the seasonal pycnocline (Fig.
346 98) are found ambiguous or statistically independently varying with latitude (Fig. 98d).
347 Likewise, the vertical turbulent fluxes of dissolved iron and phosphate are marginally constant
348 with latitude across the more permanent stratification (Fig. 109). Nitrate fluxes show the same
349 latitudinal trend, with values around $10^{-6} \text{ mmol m}^{-2} \text{ s}^{-1}$. Such values are of the same order of
350 magnitude as reported for the interior of the Saint Laurence seaway (Cyr et al., 2015). Overall,
351 the vertical turbulent nutrient fluxes across the seasonal and more permanent stratification
352 resemble those of the physical vertical turbulent mass flux, which is equivalent to the
353 distribution of turbulence dissipation rate and which is latitude-invariant (Fig. 5a).

354

355 4 Discussion

356 Practically, the upright positioning CTD while using an adaptation consisting of a
357 ~~sophisticated~~ custom-made equal-surface inlet worked well to minimize ship-motion effects on
358 variable flow-imposed temperature variations. This improved calculated turbulence values
359 from CTD-observations in general and in near-homogeneous layers in particular. The indirect
360 comparison with previous microstructure profiler observations along the same transect (Jurado
361 et al., 2012) confirms the same trends, although occasionally turbulence values were lower (to
362 one order of magnitude in the present study). This difference in values may be due to the time

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363 lapse of 8 years between the observations, but more likely it is due to inaccuracies in one or
364 both methods. It is noted that any ocean turbulence observations cannot be made better than to
365 within a factor of two (Oakey, pers. comm.). In that respect, the standard CTD with the here
366 presented adaptation is a cheaper solution than additional microstructure profiler observations.
367 Although the general understanding, mainly amongst modellers, is that the Thorpe length
368 method overestimates diffusivity (e.g., Scotti, 2015; Mater and Venayagamoorthy, 2015), this
369 view is not shared amongst ocean observers (e.g., Gregg et al., 2018). In the large parameter
370 space of the high Reynolds number environment of the ocean, turbulence properties vary
371 constantly, with an interminglement of convection and shear-induced turbulence at various
372 levels. Given sufficient averaging, and adequate mean value parametrization, the Thorpe length
373 method is not observed to overestimate diffusivity. This property of adequate and sufficient
374 averaging yields similar mean parameter values in recent modelling results estimating a mixing
375 coefficient near the classical bound of 0.2 in stationary flows for a wide range of conditions
376 (Portwood et al., 2019). It is noted that diffusivity always requires knowledge of stratification
377 to obtain a turbulent flux, and it is better to consider turbulence dissipation rate for
378 intercomparison purposes. -Nevertheless, ~~future research may it would be good to~~ perform a
379 more extensive comparison between Thorpe scale analysis data and deeper microstructure
380 profiler data ~~but that is not the scope of this work.~~

381 While our turbulence values are roughly similar to those of others transecting the NE-
382 Atlantic over the entire water depth (Walter et al., 2005; Kunze et al., 2006), the focus in the
383 present paper is on the upper 500 m because of its importance for upper-ocean marine biology.
384 Our study demonstrates a significant decrease of stratification with increasing latitude and
385 decreasing temperature that, however, does not lead to significant variation in turbulence values
386 and vertical turbulent fluxes. These findings can suggest that global warming may not
387 necessarily lead to a change in vertical turbulent exchange. We hypothesize that internal waves
388 may drive the feed-back mechanism, participating in the subtle balance between destabilizing
389 shear and stable (re)stratification as outlined below.

390 Molecular diffusivity of heat is about $10^{-7} \text{ m}^2 \text{ s}^{-1}$ in seawater, and nearly always smaller than
391 turbulent diffusivity in the ocean. The average values of K_z during our study were typically 100
392 to 1000 times larger than molecular diffusivity, which implies turbulent diapycnal mixing
393 drives vertical fluxes despite the relatively slow turbulence compared to surface wave breaking.
394 Depending on the gradient of a substance like nutrients or matter, the relatively slow turbulence
395 may not necessarily provide weak fluxes $K_z d(\text{substance})/dz$ into the photic zone. In the central
396 North Sea, a relatively low mean value of $K_z = 2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ comparable to values over the
397 seasonal pycnocline here, was found sufficient to supply nutrients across the strong summer
398 pycnocline to sustain the entire late-summer phytoplankton bloom in near-surface waters and
399 to warm up the near-bottom waters by some 3°C over the period of seasonal stratification (van
400 Haren et al., 1999). There, the turbulent exchange was driven by a combination of tidal currents
401 modified by the stratification, shear by inertial motions driven by the Coriolis force (inertial
402 shear) and internal wave breaking. Such drivers are [also](#) known to occur in the open ocean,
403 although to unknown extent.

404 The here observed [\(lack of\)](#) latitudinal trends of ϵ , K_z and N [yieldare](#) more or less the same
405 [information](#) as the vertical trends in these parameters at all stations. [In the vertical f](#)For $z < -$
406 200 m, turbulence values of ϵ and K_z weakly vary with stratification. This is perhaps unexpected
407 and contrary to the common belief of stratification hampering vertical turbulent exchange of
408 matter including nutrients. It is less surprising when considering that increasing stratification is
409 able to support larger shear. Known sources of destabilizing shear include near-inertial internal
410 waves of which the vertical length-scale is relatively small compared to other internal waves,
411 including internal tides (LeBlond and Mysak, 1978).

412 The dominance of inertial shear over shear by internal tidal motions (internal tide shear),
413 together with larger energy in the internal tidal waves, has been observed in the open-ocean,
414 e.g. in the Irminger Sea around 60°N (van Haren, 2007). The frequent atmospheric disturbances
415 in that area generate inertial motions and dominant inertial shear. Internal tides have larger
416 amplitudes but due to much larger length scales they generate weaker shear, than inertial
417 motions. Small-scale internal waves near the buoyancy frequency are abundant and may break

418 sparsely in the ocean interior outside regions of topographic influence. However, larger
419 destabilizing shear requires larger stable stratification to attain a subtle balance of ‘constant’
420 marginal stability (van Haren et al., 1999). Not only storms, but other geostrophic adjustments,
421 such as frontal collapse, may generate inertial wave shear also at low latitudes (Alford and
422 Gregg, 2001), so that overall latitudinal dependence may be negligible. If ~~dominant~~, shear-
423 induced turbulence in the upper ocean is dominant it may thus be latitudinally independent
424 (Jurado et al., 2012; deeper observations present study). There are no indications that the overall
425 open ocean internal wave field and (sub)mesoscale activities are energetically much different
426 across the mid-latitudes.

427 Summarizing, our data imply that the vertical nutrient fluxes did not vary with latitude and
428 stratification and thus from a physical environment perspective, nutrient availability and
429 corresponding summer phytoplankton productivity and growth are not expected to change
430 under future environmental changes like global warming. We invite future observations and
431 numerical modelling to further investigate this suggestion and associated feed-back
432 mechanisms such as internal wave breaking.

433

434 *Competing interests.* The authors declare that they have no conflict of interest.

435

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437 contributions to the sea-operations. J. van Heerwaarden and R. Bakker made the CTD-
438 modification.

439

440 APPENDIX A1

441 **Modification of CTD pump-tubing to minimize RAM-effects**

442 The unique pump system on SeaBird Electronics (SBE) CTDs, foremost on their high-
443 precision full ocean depth shipborne and cable-lowered SBE911, minimizes the effects of flow
444 variations (and inversions) past its T-C sensors (SeaBird, 2012). This reduction in flow
445 variation is important, because the T-sensor has a slower response than the C-sensor. As data
446 from the latter are highly temperature dependent, besides being pressure dependent, the precise
447 matching of all three sensors is crucial for establishing proper salinity and density
448 measurements, especially across rapid changes in any of the parameters. As flow past the T-
449 sensor causes higher measurement values due to friction at the sensor tip, flow-fluctuations are
450 to be avoided as they create artificial T-variations of about 1 mK s m^{-1} (Larson and Pedersen,
451 1996).

452 However, while the pump itself is one thing, its tubing needs careful mounting as well, with
453 in- and outlet at the same depth level (Sea-Bird, 2012). This is to prevent ram pressure $P = \rho U^2$,
454 for density ρ and flow speed U . Unfortunately, the SBE-manual shows tubing of different
455 diameter, for in- and outlet. Different diameter tubing leads to velocity fluctuations of $\pm 0.5 \text{ m}$
456 s^{-1} past the T-sensor, as was concluded from a simple experiment by van Haren and Laan
457 (2016). The flow speed variations induce temperature variations of $\pm 0.5 \text{ mK}$ and are mainly
458 detectable in weakly stratified waters such as in the deep ocean, but also near the surface as
459 observed in the present data. Using tubes of the same diameter opening remedied most of the
460 effect, but only if the surface of the tube-opening is perpendicular to the main CT-motion as in
461 a vertically mounted CTD. If it is parallel to the main motion as in a horizontally mounted CTD,
462 the effect was found to be adverse. The make-shift onboard experiment in van Haren and Laan
463 (2016) has now been cast into a better design (Fig. A1), of which the first results are presented
464 in this paper.

465 APPENDIX A2

466 **PDFs of vertically averaged dissipation rate in comparison with latitudinal trends**

467 Ocean turbulence dissipation rate generally tends to a nearly log-normal distribution (e.g.,
468 Pearson and Fox-Kemper, 2018), so that the probability density function (PDF) of the logarithm
469 of ϵ -values is normally distributed and can be described by the first two moments, the mean
470 and its standard deviation. It is seen in Fig. A2a that the overall distribution of all present data
471 indeed approaches lognormality, despite the relatively large length-scale used in the
472 computations (cf., Yamazaki and Lueck, 1990). When the data are split in the three depth levels
473 as in Fig. 5a, it is seen that ϵ in the upper $z > -15$ m layer is not log-normally distributed due to
474 a few outlying high values confirming an ocean state dominated by a few turbulence bursts
475 (Moum and Rippeth, 2009), whereas ϵ in the deeper more stratified layers is nearly log-
476 normally distributed.

477 When we compare the mean and standard deviations of the distributions with the extreme
478 values of the latitudinal trends as computed for Fig. 5a it is seen that for none of the three depth
479 levels the extreme values are found outside one standard deviation from the mean value. In fact,
480 for deeper stratified waters the extreme values of the trends are found very close to the mean
481 value. It is concluded that the mean dissipation rate does not show a significant trend with
482 latitude, at all depth levels. The same exercise yields extreme buoyancy frequency values lying
483 outside one standard deviation from the mean values for well-stratified waters, from which we
484 conclude that stratification significantly decreases with latitude. This is inferable from Fig. 5c
485 by investigating the spread of mean values around the trend line.

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604

605 **Figure 1.** Bathymetry map of the Northeast Atlantic Ocean based on the 9.1 ETOPO-1 version
606 of satellite altimetry-derived data by Smith and Sandwell (1997). The numbered circles
607 indicate the CTD stations. Depth contours are at 2500 and 5000 m.

608
609 **Figure 2.** Test of effective removal of ship motions in CTD-data after pump in- and outlet
610 modification. Nearly raw 24 Hz sampled downcast data obtained from northern station 32
611 (cast 9). Short example time series for the 20-m depth range [10, 30] m. (a) Detrended
612 pressure (blue) and its (negative signed) first time derivative $-dp/dt$, 2-dbar-smoothed
613 (purple). (b) Detrended temperature. (c) Moderately smoothed (~ 30 degrees of freedom;
614 dof) spectra of data from the 5 to 500 m depth range. (d) Moderately smoothed (40 dof)
615 coherence between dp/dt and T from c., with dashed line indicating the 95% significance
616 level. (e) Corresponding phase difference.

617
618 **Figure 3.** Upper 500 m of turbulence characteristics computed from downcast density anomaly
619 data applying a threshold of $7 \times 10^{-5} \text{ kg m}^{-3}$. Northern station 29, cast 2. (a) Unordered, ‘raw’
620 profile of density anomaly referenced to the surface. (b) Overturn displacements following
621 reordering of the profiles in a. Slopes $\frac{1}{2}$ (solid lines) and 1 (dashed lines) are indicated. (c)
622 Logarithm of dissipation rate computed from the profiles in a., averaged over 7 m intervals.
623 (d) As c., but for eddy diffusivity. (e) Logarithm of buoyancy frequency computed after
624 reordering the profiles of a.

625
626 **Figure 4.** As Fig. 3, but for a southern station. Upper 500 m of turbulence characteristics
627 computed from downcast density anomaly data applying a threshold of $7 \times 10^{-5} \text{ kg m}^{-3}$.
628 Southern station 3, cast 4. (a) Unordered, ‘raw’ profile of density anomaly referenced to
629 the surface. (b) Overturn displacements following reordering of the profiles in a. Slopes $\frac{1}{2}$
630 (solid lines) and 1 (dashed lines) are indicated. (c) Logarithm of dissipation rate computed
631 from the profiles in a., averaged over 7 m intervals. (d) As c., but for eddy diffusivity. (e)
632 Logarithm of buoyancy frequency computed after reordering the profiles of a.

633

634 **Figure 5.** Summer 2017 latitudinal transect along $17\pm 5^\circ\text{W}$ of turbulence values for upper 15 m
635 averages (green) and averages between $-100 < z < -25$ m (blue, seasonal pycnocline) and -
636 $500 < z < -100$ m (black, more permanent pycnocline) from short yoyos of 3 to 6 CTD-
637 casts. Values are given per cast (o) and station average (heavy circle with x; the size
638 corresponds with \pm the standard error for turbulence parameters). (a) Logarithm of
639 dissipation rate. (b) Logarithm of diffusivity. (c) Logarithm of buoyancy frequency (the
640 small symbols have the size of \pm the standard error). (d) Hour of sampling after sunrise.

641

642 **Figure 6.** Upper 500 m profiles for stations at three latitudes. (a) Density anomaly referenced
643 to the surface, including profiles from Fig. 3a and 4a. (b) Nitrate plus nitrite. (c) Phosphate.
644 (d) Silicate. (e) Dissolved iron.

645

646 **Figure 7.** Latitudinal transect of near-surface layers and wind conditions measured at stations
647 during the observational survey. (a) Mixed layer depth (x) and euphotic zone (o). (b) Wind
648 speed. (c) Wind direction.

649

650 **Figure 87.** Latitudinal transect of near-surface nutrient concentrations. (a) Dissolved iron. (b)
651 Nitrate plus nitrite (red) and phosphate (blue, scale times 10). (c) Logarithm of vertical
652 gradients of values dissolved iron in a. and phosphate in b. (d). Vertical turbulent fluxes of
653 concentrations in c. using average surface K_z from Fig. 5c.

654

655 **Figure 98.** As Fig. 87, but for $-100 < z < -25$ m.

656

657 **Figure 109.** As Fig. 87, but for -600 (few nutrients sampled at 500) $< z < -100$ m.

658

659 **Fig. A1.** SBE911 CTD-pump in- and outlet modification following the findings in van Haren
660 and Laan (2016). (a) The T- and C-sensors clamped together with a structure holding in-

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661 and outlet pump-tubing of exactly the same diameter, separated at 0.3 m distance in the
662 horizontal plane. (b) The modification of a. mounted in the CTD-frame.

663

664 **Fig. A2.** Probability Density Functions of logarithm of vertically averaged dissipation rate in
665 comparison with latitudinal trend extreme values. (a) Distribution as a function of latitude
666 for all data. (b) As a, but for the upper 15 m averages only. The mean value is given by the
667 vertical purple line, with the horizontal line indicating +/- 1 standard deviation. The vertical
668 light-blue lines indicate the best-fit value of the trend for 30° and 63°N. (c) As b, but for
669 averages between $-100 < z < -25$ m. (d) As c, but for averages between $-500 < z < -100$ m.

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