>>>We are grateful for the comments on our manuscript from the reviewer. We feel that this new version of the paper is much stronger as the result of the comments we received on the original manuscript. We have addressed all of the comments and have detailed our response to specific comments below. Our response to each comment is bulleted and in italics below the relevant comment behind>>>

Anonymous Referee #1
Received and published: 4 September 2020
Review for manuscript # os-2020-73 "Diapycnal mixing across the photic zone of the NE-Atlantic" by van Haren et al.

Formal review:
The authors discuss dissipation rates of turbulent kinetic energy, eddy diffusivities and vertical turbulent nutrient fluxes inferred from upper-ocean hydrographic and nutrient data taken during a cruise on a transect from 60_N to 30_N along about 17_W in the North Atlantic. Inferred eddy diffusivities and vertical turbulent nutrient fluxes in the upper thermocline (<500m depth) did not vary with latitude. However, from south to north stratification in the upper thermocline weakened by a factor of 5. The authors claim that the lack of correspondence between turbulent mixing and stratification (temperature) suggest that nutrient availability for phytoplankton in the euphotic surface waters may not be affected by global warming.
While this paper is fairly well written and addresses scientifically relevant question such as an advancing quantitative understanding of the role of mixing in sustaining biological production in the near surface layers of the ocean, the current version of the manuscript has major deficiencies. In particular, I find that the results presented in the manuscript are not sufficient to support the authors' interpretations and conclusions. Furthermore, a statistical analysis of uncertainties inherent to the results needs to be added.
>>>We thank the reviewer for the appreciation. We have now attempted to substantiate support for our interpretations. Uncertainties to the results are further explained.

Major concerns
Personally, I per se agree with the statement that climate warming and associated increase of upper-ocean stratification will not necessarily lead to a decrease of turbulent mixing in the thermocline or a decrease of vertical turbulent nutrient fluxes. Certainly, there are also arguments that support an enhanced energy flux into internal waves due to increasing stratification (which has also been suggest by several previous publications, e.g. DeCarlo et al., 2015). However, to me, the data analysis presented here does not permit to draw any conclusions on this issue. This is because (1) the data is inadequately resolving average mixing quantities. Turbulent mixing in the ocean exhibits a near log-normal frequency distribution and elevated mixing events occur infrequently. However, these elevated mixing events are dominantly responsible for the vertical turbulent fluxes of solutes in the ocean. The 60+ profiles (I am guessing here as no numbers are provided in the manuscript) that may represent turbulence conditions over a period of 3 to 4 hours at the 15 to 20 individual stations are certainly inadequate to draw any conclusions on average turbulence quantities at different latitudes. The variability of turbulent mixing is also reflected by (2) the individual estimates of vertical turbulent nutrient fluxes available from the limited individual stations along the transect. Fluxes vary by three orders of magnitude (Figures 7, 8, 9). Again, an analysis of their statistical uncertainty would show the ambiguity of any trend analysis. Finally, (3) I cannot approve the approach chosen here as a whole. Comparing the strength of upper thermocline mixing at different latitudes cannot lead to any conclusions on local changes of the strength of turbulent mixing e.g. due to locally increasing stratification. The regions where measurements were taken combine very different external forcing and internal wave environments making it impossible to relate mixing strength to a single parameter.
>>>In reply to point (1) We are aware of the near-lognormal PDF of turbulence dissipation data, actually it is one of the reasons to plot our data in log-fashion. We do not agree that the number of profiles cannot say anything about average quantities, as the spread is clearly given. Of course one can

compute average values from that, also considering that every 24 Hz sampled profile is binned in 7 m vertical bins (200 data points), that are again grouped in several layers (down to 500 m, or 70 bins). We wonder if the reviewer hereby discards all observational oceanographic turbulence work? Much effort goes into such observational work. Point (2) Yes, that is precisely what we indicated in the original manuscript: two (to four) orders of magnitude variability. The statistics is thereby given: the spread around the mean, considering the instrumental and methodological error of about half an order of magnitude. Point (3) We do not agree with this statement, because all sampling is done in the upper 500 m where the local water depth was at least 1100 m, and, except for 3 stations, most stations were over (much) deeper waters >2000 m. So, sampling was well away from bottom topography, in the NE-Atlantic where semidiurnal tides, and inertial motions, dominate the internal wave field, in summertime under overall moderate-good weather conditions across the entire survey. As a result, the dominant convection (in the upper 20-30 m) and internal wave induced mixing (in the stratified layers below) are much less variable across the transect due to different forcing than due to the highly intermittent occurrence of turbulent bursts as the reviewer correctly indicates above. Those bursts are inherent to turbulence, and less so dependent on the generation process. We added text to better explain this, lines 419-421:' If shear-induced turbulence in the upper ocean is dominant it may thus be latitudinally independent (Jurado et al., 2012; deeper observations present study). There are no indications that the overall open ocean internal wave field and (sub)mesoscale activities are energetically much different across the mid-latitudes.'

As a revision strategy, I would advise the authors to remove the discussion on mixing and nutrient fluxes in a changing climate from the manuscript. Instead, the focus could be shifted to a detailed discussion of an upper-ocean nutrient budget including statistical uncertainties and a comparison to the net community production.
>>> The outcome of our paper is the suggestion that climate change might not affect fluxes as strongly as current paradigm suggests. The intention is to inspire discussion/further research. The nutrient budget and comparison to the net community production have been described by Mojica et al (2016), which we will not repeat in our paper which is more oriented to physics processes than biology. We explained this better now. Our manuscript is an extension of that work.

Some specific comments
Line 294 – 299, discussion of nutrient fluxes in the mixed layer and Fig. 7. I find the discussion of macro-nutrient fluxes in the mixed layer erroneous. First of all, vertical gradients of macro-nutrients are mostly insignificant. Macro-nutrient concentrations determined by a QuAAtro autoanalyser usually have accuracies of 0.1mM if CRM standards were used (please add details of uncertainties inherent to the nutrient concentrations to the methods section). To me, the differences between macro-nutrient concentrations measured at 10m and 25m depth are mostly smaller than measurement uncertainties.
>>>The reviewer is right, we should have given the precision and detection limits. Without that info, the interpretation is not well substantiated. However, the accuracy is much better than assumed by the reviewer, as it is e.g. 0.028 µM for phosphate. We added the information now in the Methods section of the revised manuscript.

*Absolute and relative precision for reasonably high concentrations in an in-house standard that is often measured.*

	S.D. (µM)	N	concentration	rel SD
PO4	0.028	30	0.9	3.1%
NO3	0.143	30	14.0	1.0 %
Si	0.088	15	20.99	0.42%

The method detection limit was calculated during the cruise using the standard deviation of ten samples containing 2% of the highest standard used for the calibration curve and multiplied with the student's value for n=10, thus being 2.82. (M.D.L = Std Dev of 10 samples x 2.82)

	µM
PO4	*0.007*
NH4	*0.010*
NO3+NO2	*0.012*
NO2	*0.003*
Si	*0.008*

Line 300 – 303, discussion of nutrient fluxes below the mixed layer. As stated in the above, individual estimates of nutrient fluxes vary by three orders of magnitude and a statement about how the nutrient fluxes vary with latitude (i.e. with stratification) is inadequate. What may be interesting to the reader is the magnitude of average regional fluxes that could be compared to previous estimates (see e.g. Cyr et al., 2015). Presented results should also include nitrate/nitrite fluxes as the relative vertical turbulent fluxes of reactive nitrogen species and phosphorous could be of interest to a broader scientific community.
>>>We are happy to compare with works from others, noting that Cyr et al. presented work at 2 stations in an estuary, which may be difficult to compare with the open ocean. We choose to graphical display macronutrient phosphorous representing other nutrients that show similar latitudinal trends. Attached is a version of Fig. 9 demonstrating the little extra information if we conclude NOx. We have to rescale panel c. We have now given global figures for nitrate fluxes. As mentioned, in this paper we are mainly interested in latitudinal and stratification trends and trends for phosphate fluxes precisely represent those for nitrate fluxes (blue and green lines in panel d in the figure below).
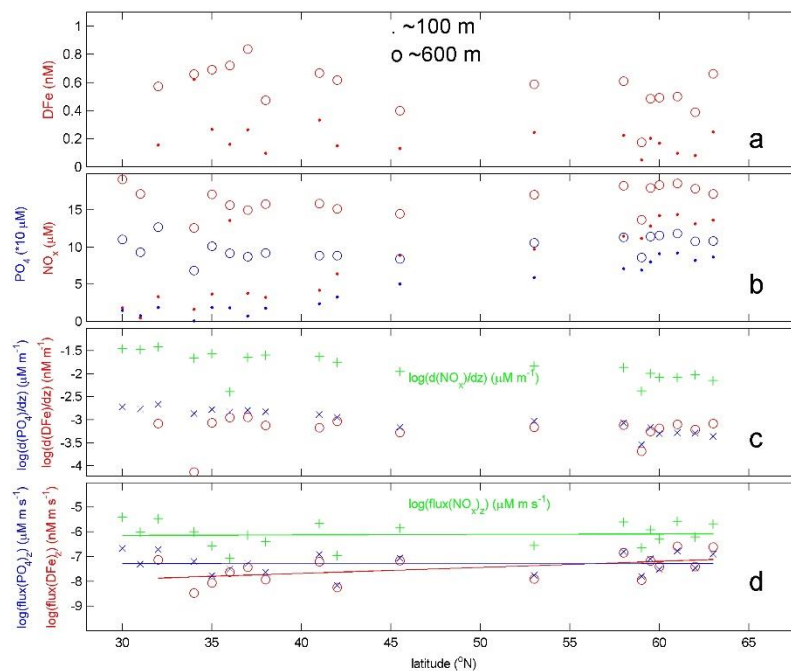
Literature
Cyr, F., D. Bourgault, P. S. Galbraith, and M. Gosselin (2015), Turbulent nitrate fluxes inthe Lower St. Lawrence Estuary, Canada, J. Geophys. Res. Oceans, 120, 2308–2330,doi:10.1002/2014JC010272.

DeCarlo, T. M., K. B. Karnauskas, K. A. Davis,and G.T.F. Wong (2015), Climate modulates internal wave activity in the Northern South China Sea, Geophys. Res.Lett., 42, 831–838, doi:10.1002/2014GL062522.

>>>We are grateful for the comments on our manuscript from the reviewer. We feel that this new version of the paper is much stronger as the result of the comments we received on the original manuscript. We have addressed all of the comments and have detailed our response to specific comments below. Our response to each comment is bulleted and in italics below the relevant comment behind>>>

Anonymous Referee #2
Received and published: 6 September 2020

The manuscript titled "Diapycnal mixing across the photic zone of the NE-Atlantic" by Haren et al. quantified the upper ocean nutrient flux using a custom modified CTD and nutrient measurements at discrete depths from a latitudinal transect along 17_5_W between 30 and 62_N in summer. The authors observed no increase in vertical mixing or diapycnal nutrient flux from south to north, where the temperature increased. Further, they opined that nutrient supply by diapycnal flux to the euphotic zone might not be affected by the physical process of global warming. It is a well-written manuscript and presents an interesting take on the ocean biophysical coupling in the global warming scenario. However, I feel that the authors jumped into a conclusion without providing enough evidence to support their say. Hence I recommend major revision.

>>>Thank you for the appreciation. To be noted, temperature decreased, not increased, from south to north and we like to add that stratification, the medium to support internal gravity waves, also decreased.

Major Comments

L63-96 The introduction needs a more general introduction to the oceanography of the region. Especially knowledge of bathymetry, background internal wave field, eddy kinetic energy, and wind conditions during summer.

>>>We have no objection to add information on the North-Atlantic Ocean in general. However, the observations were made in the upper 500 m, and water at all stations were >1000 m deep with only 3 <2000 m. Local bottom topography did not influence the internal wave field directly. We added this consideration now in the revised manuscript. We also added that the survey was done in summer time,with in general moderate to good weather conditions, no big storms. We have no information on the eddy kinetic energy at the time, other than the generally excepted view from literature, which we also added to the revised manuscript.

L123 In the Thorpe length calculation section, please mention the lowering speed of the CTD. A slow lowering can resolve overturns efficiently. In the mixed layer, the Thorpe method will consider it as a large overturn. How you will justify the validity of diffusivity within the mixed layer, where N2 is weak. A brief discussion on lowering speed of CTD and justification for the diffusivity within the mixed layer will give clarity to the reader.

>>> We agree with the reviewer that we should have added the lowering speed, it was 1 m $^{-1}$s. Yes, slow lowering resolves overturns better, but in doing so it is lowered obliquely through the overturns in case of non-zero background flow, which is nearly always present. A completely and thoroughly mixed layer hardly ever exists, but the stratification is often weak in the upper 20-30 m while is varies in height and time. For the validity of choice of parameters we refer to the extensive work by Oakey (1982) who demonstrated upper ocean parametrization to vary over at least one order of magnitude but, given enough data points, with a particular average value as used here. This is confirmed in more recent works (Gregg et al 2018) for ocean observations and Portwood et al (PRL2019) for modelling work. We added this information to the manuscript at P14: 'Although the general understanding, mainly amongst modellers, is that the Thorpe length method overestimates diffusivity (e.g., Scotti, 2015; Mater and Venayagamoorthy, 2015), this view is not shared amongst ocean observers (e.g., Gregg et al., 2018). In the large parameter space of the high Reynolds number environment of the ocean, turbulence properties vary constantly, with an interminglement of convection and shear-induced turbulence at various levels. Given sufficient

averaging, and adequate mean value parametrization, the Thorpe length method is not observed to overestimate diffusivity. This property of adequate and sufficient averaging yields similar mean parameter values in recent modelling results estimating a mixing coefficient near the classical bound of 0.2 in stationary flows for a wide range of conditions (Portwood et al., 2019). It is noted that diffusivity always requires knowledge of stratification to obtain a turbulent flux, and it is better to consider turbulence dissipation rate for intercomparison purposes. Nevertheless, future research may perform a more extensive comparison between Thorpe scale analysis data and deeper microstructure profiler data.'

L256-258 Substantiate the surface cooling and internal wave breaking using data.
>>>This was indeed not clear. We meant that the main process in the upper layer is convective surface cooling, and internal wave breaking in the more stratified layers below. We changed the text l.289-291: The trends suggest only marginally larger turbulence going poleward, which is possibly due to larger cooling from above and larger internal wave breaking deeper down.

L264-265 I could not understand this sentence.
>>>Perhaps 'confirm' was misused here; we meant to say that the deeper layers show the same latitudinal trend in turbulence and stratification values as the upper layer. We rephrased the sentence to' The data from well-stratified waters deeper down thus show the same latitudinal trend as the observations from the near-surface layers.' (line 298-300).

L284-286 The nutrient flux depends on the eddy diffusivity and the nutrient concentration gradient, which changes dramatically with depth. The nutrient fluxes thus may vary with two-or-three orders difference. In the manuscript, nutrient flux is calculated using a low-resolution profile of nutrients. Does this discrete measurement introduce bias to the flux calculation? What is the typical depth of the euphotic zone in the study region?
>>>We would have liked a denser sampling of nutrients, but that was impossible in the cruise plan. On the other hand, the large gradients in nutrients are indeed in the vertical, and variations in the horizontal plane are less strong. We note that, due to overturn sizes over which we must average, turbulence is gridded in equally large vertical distances. The typical depth of the 0.1% irradiance penetration is about 50 to 100 m, see the figure panel 'a' below in which we compare this depth with the 'mixed layer depth', defined as the depth at which the temperature difference with respect to the surface was 0.5°C (as in Jurado et al 2012). We have added this information on p.12: 'The mixed layer depth, defined as the depth at which the temperature difference with respect to the surface was 0.5□C (Jurado et al., 2012), varies between about 20 and 30 m on the southern end of the transect and weakly becomes shallower with latitude (Fig. 7a). This weak trend may be expected from the summertime wind conditions that also barely vary with latitude (Fig. 7b,c). In contrast, the euphotic zone, defined as the depth of the 0.1% irradiance penetration level (Mojica et al., 2015), demonstrates a clear latitudinal trend decreasing from about 150 to 50 m (Fig. 7a).'
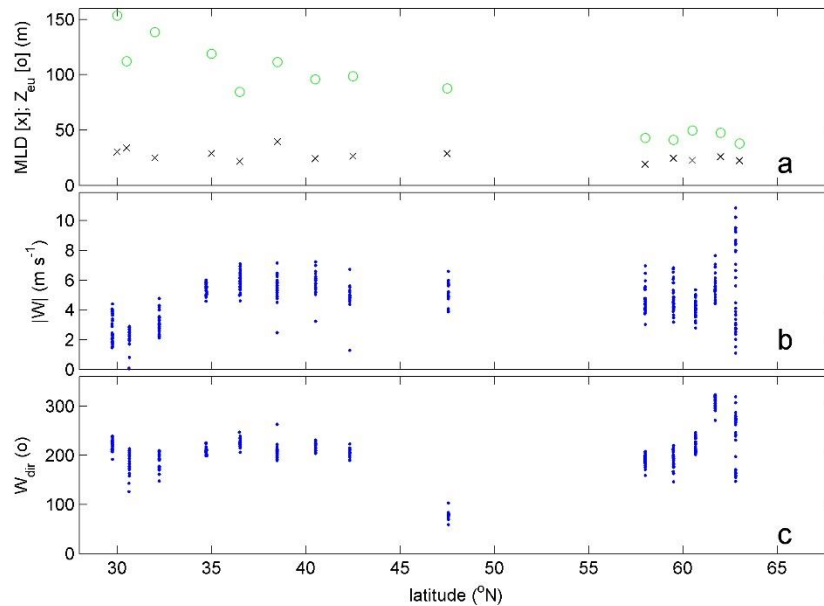
(New) Figure 7. Latitudinal transect of near-surface layers and wind conditions measured at stations during the observational survey. (a) Mixed layer depth (x) and euphotic zone (o). (b) Wind speed. (c) Wind direction.

L318-320 General understanding is that the Thorpe length method overestimates the diffusivity (Mater and Venayagamoorthy 2015; Alberto Scotti 2015).

>>>That is indeed a general understanding amongst modellers, but not amongst ocean observers (e.g. Gregg et al 2018). In the high Reynolds number environment of the ocean turbulence properties vary constantly, an interminglement of convection and shear-induced turbulence at various levels. Given sufficient averaging, and adequate mean value parametrization, the Thorpe length method does not overestimate diffusivity, see also recent modelling results by Portwood et al (PRL2019). It is noted that diffusivity always requires knowledge of stratification to obtain a turbulent flux, and it is better to consider turbulence dissipation rate for intercomparison. We clarified this in the revised manuscript (lines 364-377): 'Although the general understanding, mainly amongst modellers, is that the Thorpe length method overestimates diffusivity (e.g., Scotti, 2015; Mater and Venayagamoorthy, 2015), this view is not shared amongst ocean observers (e.g., Gregg et al., 2018). In the large parameter space of the high Reynolds number environment of the ocean, turbulence properties vary constantly, with an interminglement of convection and shear-induced turbulence at various levels. Given sufficient averaging, and adequate mean value parametrization, the Thorpe length method is not observed to overestimate diffusivity. This property of adequate and sufficient averaging yields similar mean parameter values in recent modelling results estimating a mixing coefficient near the classical bound of 0.2 in stationary flows for a wide range of conditions (Portwood et al., 2019). It is noted that diffusivity always requires knowledge of stratification to obtain a turbulent flux, and it is better to consider turbulence dissipation rate for intercomparison purposes. Nevertheless, future research may perform a more extensive comparison between Thorpe scale analysis data and deeper microstructure profiler data'.

L328-329 Here you can add a detailed discussion on how internal waves can be a feedback mechanism to counteract the suppression of mixing by increased stratification.

>>> Although originally it was merely meant as an introductory sentence to the paragraphs below, we see reviewer's point and we added it in the revised manuscript. (line 384-386 and pages that follow): 'We hypothesize that internal waves may drive the feed-back mechanism, participating in the subtle balance between destabilizing shear and stable (re)stratification as outlined below.'.

L344-364 Authors need to provide data evidence to prove that Internal wave energy/eddy kinetic energy is more in Northern stations, and thus, the relatively increased stratification (compared to south) could not suppress the diapycnal flux of nutrients to the euphotic zone from deeper depths. This will give the readers a better understanding of the lack of correspondence between temperature /stratification and diapycnal flux with latitude.

>>>There seems to be a misunderstanding here: Stratification is less in the north, compared to the south. We have emphasized this in the text. We would have loved to include direct observational information on internal wave and eddy kinetic energy but we do not have such data available in the present study. Instead, we refer to previous work in which we had such data. Using that information, we now better tried to explain as the reviewer suggests. In the discussion we support our suggested hypothesis with the (previous) observation that the state of ocean is one of marginal stability, in which stratification is a subtle balance between internal wave shear and -breaking.

One could employ the GM spectrum calculated using gridded Historical data sets (ARGO) to give an idea on the background Internal wave energy. However, I won't insist on doing this analysis.

>>>We think it is better that modelers take up this task, they will perform much better than we can on this.

A discussion on the meteorological conditions during the observation period is also warranted. What if the southern stations were characterized with anomalously calm weather that mixing was inactive and became comparable to the northern stations.

>>>This is a good idea, we added this information. For the information of the reviewer: meteorological conditions were moderate to good throughout the cruise, see for Wind Speed the panel b (and c for Wind direction) in the figure given above. This is the new Fig. 7 now.

References

Scotti, A. (2015). Biases in Thorpe-scale estimates of turbulence dissipation. Part II: energetics arguments and turbulence simulations. Journal of Physical Oceanography, 45(10), 2522-2543.

Mater, B. D., Venayagamoorthy, S. K., St. Laurent, L., & Moum, J. N. (2015). Biases in Thorpe-scale estimates of turbulence dissipation. Part I: Assessments from largescale overturns in oceanographic data. Journal of Physical Oceanography, 45(10), 2497-2521.

>>>We are grateful for the comments on our manuscript from the reviewer. We feel that this new version of the paper is much stronger as the result of the comments we received on the original manuscript. We have addressed all of the comments and have detailed our response to specific comments below. Our response to each comment is bulleted and in italics below the relevant comment behind>>>

Anonymous Referee #3
Received and published: 20 September 2020

This is an observational study of diapycnal mixing and the corresponding nutrient flux in the upper ocean across a quasi-latitudinal transect in the Northeast Atlantic Ocean. The data cover a rather long distance from 30deg.N to 62deg.N. The measurements were mainly temperature and conductivity profiles (from which the density or potential density profiles were obtained) with a carefully modified CTD system, and the turbulent kinetic energy dissipation rate and the diapycnal diffusivity were estimated based on the overturning (Thorpe) scale analysis. In general, the methodology of the analysis is reasonable, and useful information on turbulent mixing characteristics along the transect is obtained. However, I cannot recommend this manuscript for publication in the present form due to the major concerns as detailed in the following.
>>>Thank you for the appreciation of the outline and methodology.

First of all, I find the major point that the authors try to make (i.e., "nutrient availability for phytoplankton in the euphotic surface waters may not be affected by the physical process of global warming") is not convincing at all. For me, the point is not even relevant to what the data have shown. Obviously, the exact response of the upper ocean to global warming could be rather complicated, and I do agree with the authors that the global warming may not necessarily lead to a change in vertical turbulent exchange, but the results presented in the manuscript are by no means evidence for this. One may expect to see clear trend of upper ocean mixing (and corresponding material fluxes) at a certain location under continuing warming, but in such a large region covering more than 30 degrees, the underlying dynamics controlling diapycnal mixing could be very different from place to place, thus the spatial difference in mixing seen along the transect cannot be simply taken as a result of the difference in stratification (or "warming" by solar radiation).
>>>There is general consensus that upper ocean convection and interior ocean internal wave breaking are the dominant turbulence generating mechanisms in the upper 50 – 100 m and in the deeper 100(50)-500 m stratified waters, respectively. These mechanisms are universal and not particularly location dependent, but do depend on variations in stratification. Turbulence is such an intermittent process that variations over four orders magnitude occur, at the same location (e.g. Gregg, JGR1989). Such variability is found in the present observations too, e.g. as indicated in old l.242 and 243. This is much larger variability than particular variation in turbulence generation processes in the ocean interior, away from boundaries. In other words, the sources may not greatly vary their energy content along a transect compared with turbulence intermittency. The trend in spatial difference in mixing along a transect can be taken as a result of difference in stratification. Along these lines we have added text (l 233-235) 'As will be demonstrated below, this is considerably less spread in values than the natural turbulence values variability over typically four orders of magnitude at a given position and depth in the ocean (e.g., Gregg, 1989).'

More technically, although I do appreciate the authors' efforts in estimating turbulence and mixing characteristics from carefully conducted CTD measurements (via overturning scale analysis), I cannot be convinced by the subtle mixing (and flux) variability revealed by their estimates. As well acknowledged by the authors, even with the microstructure measurements one cannot expect to get an estimate with insignificant uncertainty. I agree that the overturning (Thorpe) scale analysis could be very useful in getting a rough estimate of mixing intensity when more direct measurements are not available, but using it to reveal subtle spatial (or temporal) variability could be misleading. For this purpose, direct

microstructure measurements are certainly much more reliable. On the other hand, ocean turbulence is certainly a stochastic process with both significant dynamical variability (which could be taken as deterministic linked to certain dynamical processes generating turbulence) and intermittency. As such, for the purpose of evaluating spatial variability of turbulent mixing, one should look at turbulence statistics. How many data points are used to get the reported averages? How does the PDF in each corresponding depth range look like? What are the confidence intervals of the reported averages? Are the noted differences/variabilities really significant?

>>>*We thank the reviewer for the appreciation of our efforts in estimating turbulence from carefully conducting CTD measurements. We like to point out that our near-surface data do show the same trends as upper 100 m microstructure profiler observations obtained a few years earlier along the same transect (Jurado et al 2012). In general, microstructure measurements are indeed more reliable for turbulence measurements, but in this case do not provide significantly different results. We note that turbulence is not by any means a deterministic process, even though its dominant generator, e.g. tide, may be deterministic. We have now additional CTD sampling numbers. The CTD sampled at 24 Hz, so 7 m ensemble averaging vertical intervals contain about 200 data points and we collected 3 to 6 CTD-casts per station. Below we add PDF of the entire dissipation rate data, averaged per vertical interval as indicated. As for the errors: they are about a factor of 3 for mean dissipation rate.*
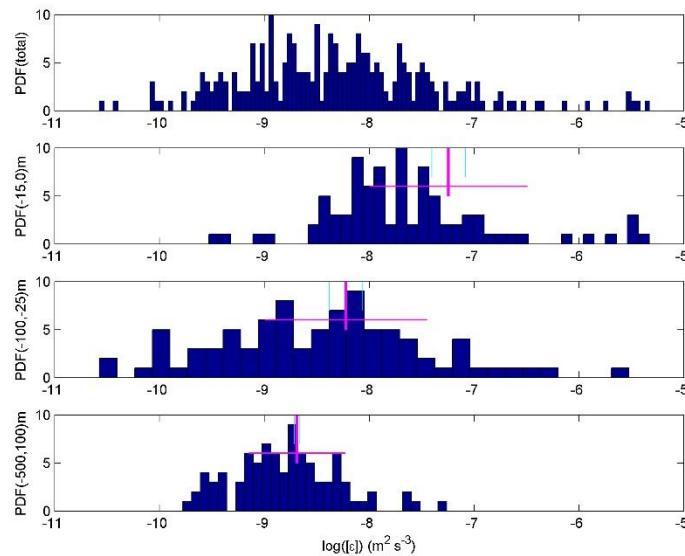
(New)Fig. A2. Probability Density Functions of logarithm of vertically averaged dissipation rate in comparison with latitudinal trend extreme values. (a) Distribution as a function of latitude for all data. (b) As a, but for the upper 15 m averages only. The mean value is given by the vertical purple line, with the horizontal line indicating +/- 1 standard deviation. The vertical light-blue lines indicate the best-fit value of the trend for 30° and 63°N. (c) As b, but for averages between -100 < z < -25 m. (d) As c, but for averages between -500 < z < -100 m.

To conclude, I agree that the reported analysis gives useful information about mixing characteristics along the sampled transect, but without clear information of the underlying mechanisms and robust constraint on the reliability and significance of the reported mixing variability, one cannot be led to the points that the authors try to make. In particular, the results presented in the manuscript do not seem to lend any support to the authors' argument on the global warming impact on upper ocean mixing and nutrient flux trend. The authors may choose to simply emphasize their mixing estimates from the overturning scale analysis, with clear indication of the underlying uncertainties.

>>>*We adapted part of the discussion to better relate to comments by the reviewers, including toning down the relation with climate change, and we hope that as such we now made our point more convincing..*

Diapycnal mixing across the photic zone of the NE-Atlantic

by Hans van Haren[*], Corina P.D. Brussaard, Loes J. A. Gerringa, Mathijs H. van Manen, Rob Middag, Ruud Groenewegen

Royal Netherlands Institute for Sea Research (NIOZ) and Utrecht University, P.O. Box 59, 1790 AB Den Burg, the Netherlands.
*e-mail: hans.van.haren@nioz.nl

Abstract. Variable physical conditions such as vertical turbulent exchange, internal wave and mesoscale eddy action, affect the availability of light and nutrients for phytoplankton (unicellular algae) growth. It is hypothesized that changes in ocean temperature may affect ocean vertical density stratification, which may hamper vertical exchange. In order to quantify variations in physical conditions in the Northeast Atlantic Ocean, we sampled a latitudinal transect along $17\pm5°$W between 30 and 62°N in summer. A shipborne Conductivity-Temperature-Depth CTD-instrumented package was used with a custom-made modification of the pump-inlet to minimize detrimental effects of ship motions on its data. Thorpe-scale analysis was used to establish turbulence values for the upper 500 m near the surface from 3 to 6 profiles obtained in a short CTD-yoyo, 3 to 5 h after local sunrise. From south to north, temperature decreased together with stratification while turbulence values weakly increased or remained constant. Vertical turbulent nutrient fluxes across the stratification were ~~did~~ not found to significantly vary with latitude. This apparent lack of correspondence between turbulent mixing and temperature is suggested to be due to internal waves breaking and acting as a potential feed-back mechanism. Our findings suggest that nutrient availability for phytoplankton in the euphotic surface waters may not be affected by the physical process of global warming.

1 Introduction

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

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

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

3

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

2 Materials and Methods

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

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

6

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 ~ 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) $K_z d\rho/dz$.

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 $d\rho/dz$. 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 $$\varepsilon = 0.64 d^2 N^3 \qquad [m^2 s^{-3}], \tag{1}$$

199 where $N = \{-g/\rho(d\rho/dz + g\rho/c_s^2)\}^{1/2}$ (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 $K_z = \Gamma \varepsilon N^{-2}$ 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 $$K_z = 0.128 d^2 N \qquad [m^2 s^{-1}]. \tag{2}$$

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

8

226 L_O. 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 7×10^{-5} kg m^{-3}, which corresponds to applying a threshold of 1.4×10^{-3} 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

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 = ±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 (~N^2; Fig. 3e). Between -30 < z < 0 m, turbulence dissipation

248 rate values between <10^{-11} and >10^{-8} m^2 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 3×10^{-3} m^2 s^{-1} and these values

252 compare with previous near-surface results (Denman and Gargett, 1983). The relatively small

9

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~~ station~~s~~. 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 ($z > \sim -25$ 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

11

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, $-500 < z < -100$ 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 $N \approx 10^{-2.5}$ s^{-1}) and moderately (at mid-latitudes $N \approx 10^{-2.2}$ 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]° 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 $z > -500$ 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 $z < -100$ 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 $z < -500$ 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} mmol m^{-2} 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

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

Formatted: Superscript
Formatted: Superscript
Formatted: Superscript

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 mayit 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.

14

390   Molecular diffusivity of heat is about 10^{-7} m^2 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_zd(substance)/dz into the photic zone. In the central

396   North Sea, a relatively low mean value of $K_z = 2\times10^{-5}$ m^2 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 fFor 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 thatthe 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

439

Modification of CTD pump-tubing to minimize RAM-effects

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

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

APPENDIX A2

PDFs of vertically averaged dissipation rate in comparison with latitudinal trends

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

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

18

487 **References**

488 Alford, M. H. and Gregg, M. C.: Near-inertial mixing: Modulation of shear, strain and

489   microstructure at low latitude, J. Geophys. Res., 106, 16,947-16,968, 2001.

490 Charria, G., Theetten, S., Vandermeirsch, F., Yelekçi, Ö and Audiffren, N.: Interannual

491   evolution of (sub)mesoscale dynamics in the Bay of Biscay, Ocean Sci., 13, 777-797, 2017.

492 Cyr, F., Bourgault, D., Galbraith, P. S. and Gosselin, M.: Turbulent nitrate fluxes inthe Lower

493   St. Lawrence Estuary, Canada, J. Geophys. Res., 120, 2308-2330,

494   doi:10.1002/2014JC010272, 2015.

495 De Baar, H. J. W. et al.: Titan: A new facility for ultraclean sampling of trace elements and

496   isotopes in the deep oceans in the international Geotraces program, Mar. Chem., 111, 4-21,

497   2008.

498 Denman, K. L. and Gargett, A. E.: Time and space scales of vertical mixing and advection of

499   phytoplankton in the upper ocean, Limnol. Oceanogr., 28, 801-815, 1983.

500 Dillon, T. M.: Vertical overturns: A comparison of Thorpe and Ozmidov length scales, J.

501   Geophys. Res., 87, 9601-9613, 1982.

502 Ferron, B., Mercier, H., Speer, K., Gargett, A. and Polzin, K.: Mixing in the Romanche Fracture

503   Zone, J. Phys. Oceanogr., 28, 1929-1945, 1998.

504 Galbraith, P. S. and Kelley, D. E.: Identifying overturns in CTD profiles, J. Atmos. Oc.

505   Technol., 13, 688-702, 1996.

506 Gargett, A. and Garner, T.: Determining Thorpe scales from ship-lowered CTD density

507   profiles, J. Atmos. Oc. Technol., 25, 1657-1670, 2008.

508 Gill, A. E.: Atmosphere-Ocean Dynamics, Academic Press, Orlando, Fl, USA, 662 pp, 1982.

509 Grasshoff, K., Kremling, K. and Ehrhardt, M.: Methods of seawater analysis, Verlag

510   Chemie GmbH, Weinheim, 419 pp, 1983.

511 Gregg, M. C.: Scaling turbulent dissipation in the thermocline, J. Geophys. Res., 94, 9686-

512   9698, 1989.

Gregg, M. C., D'Asaro, E. A., Riley, J. J. and Kunze, E.: Mixing efficiency in the ocean, Ann. Rev. Mar. Sci., 10, 443-473, 2018.

Hernández-Hernández, N. et al.: Drivers of plankton distribution across mesoscale eddies at submesoscale range, Front. Mar. Sci., 7, 667, doi:10.3389/fmars.2020.00667, 2020.

Huisman, J., Pham Thi, N. N., Karl, D. M. and Sommeijer, B.: Reduced mixing generates oscillations and chaos in the oceanic deep chlorophyll maximum, Nature, 439, 322-325, 2006.

Jurado, E., van der Woerd, H. J. and Dijkstra, H. A.: Microstructure measurements along a quasi-meridional transect in the northeastern Atlantic Ocean, J. Geophys. Res., 117, C04016, doi:10.1029/2011JC07137, 2012.

IOC, SCOR, IAPSO: The international thermodynamic equation of seawater – 2010: Calculation and use of thermodynamic properties, Intergovernmental Oceanographic Commission, Manuals and Guides No. 56, UNESCO, Paris, France, 196 pp, 2010.

Johnson, K. S., Gordon, R. M. and Coale, K. H.: What controls dissolved iron concentrations in the world ocean? Mar. Chem., 57, 137-161, 1997.

Klunder, M. B., Laan, P., Middag, R., De Baar, H. J. W. and van Ooijen, J. C.: Dissolved iron in the Southern Ocean (Atlantic sector), Deep-Sea Res. II, 58, 2678-2694, 2011.

Kunze, E., Firing, E., Hummon, J. M., Chereskin, T. K. and Thurnherr, A. M.: Global abyssal mixing inferred from lowered ADCP shear and CTD strain profiles, J. Phys. Oceanogr. 36, 1553-1576, 2006.

Larson, N., Pedersen, A. M.: Temperature measurements in flowing water: viscous heating of sensor tips, Proc. 1st IGHEM Meeting, Montreal, PQ, Canada. [Available online at http://www.seabird.com/technical_references/viscous.htm], 1996.

LeBlond, P. H. and Mysak, L. A.: Waves in the Ocean, Elsevier, Amsterdam NL, 602 pp, 1978.

Lueck, R. G.: Thermal inertia of conductivity cells: Theory, J. Atmos. Oc. Technol., 7, 741-755, 1990.

Mater, B. D., Venayagamoorthy, S. K., St. Laurent, L. and Moum, J. N.: Biases in Thorpe-scale estimates of turbulence dissipation. Part I: Assessments from largescale overturns in oceanographic data, J. Phys. Oceanogr., 45, 2497-2521, 2015.

Mensah, V., Le Menn, M. and Morel, Y.: Thermal mass correction for the evaluation of salinity, J. Atmos. Oc. Tech., 26, 665-672, 2009.

Middag, R., de Baar, H. J. W., Laan, P. and Bakker, K.: Dissolved aluminium and the silicon cycle in the Arctic Ocean, Marine Chemistry, 115, 176-195, 2009.

Mojica, K. D. A. et al.: Phytoplankton community structure in relation to vertical stratification along a north-south gradient in the Northeast Atlantic Ocean, Limnol. Oceanogr., 60, 1498-1521, 2015.

Mojica, K. D. A., Huisman, J., Wilhelm, S. W. and Brussaard, C. P. D.: Latitudinal variation in virus-induced mortality of phytoplankton across the North Atlantic Ocean, ISME J., 10, 500-513, 2016.

Moum, J. N. and Rippeth, T. P.: Do observations adequately resolve the natural variability of oceanic turbulence?, J. Mar. Sys., 77, 409-417, 2009.

Murphy, J. and Riley, J. P.: A modified single solution method for the determination of phosphate in natural waters, Anal. Chim. Acta, 27, 31-36, 1962.

Oakey, N. S.: Determination of the rate of dissipation of turbulent energy from simultaneous temperature and velocity shear microstructure measurements, J. Phys. Oceanogr., 12, 256-271, 1982.

Osborn, T. R.: Estimates of the local rate of vertical diffusion from dissipation measurements, J. Phys. Oceanogr., 10, 83-89, 1980.

Pearson, B. and Fox-Kemper, B.: Log-normal turbulence dissipation in global ocean models, Phys. Rev. Lett., 120, 094501, 2018.

Portwood, G. D., de Bruyn Kops, S. M. and Caulfield, C. P.: Asymptotic dynamics of high dynamic range stratified turbulence, Phys. Rev. Lett., 122, 194504, 2019.

21

566 Rijkenberg, M. J. A. et al.: Fluxes and distribution of dissolved iron in the eastern (sub-) tropical

567 North Atlantic Ocean, Glob. Biogeochem. Cycl., 26, GB3004,

568 doi:10.1029/2011GB004264, 2012.

569 Rijkenberg, M. J. A. et al.: "PRISTINE", a new high volume sampler for ultraclean sampling

570 of trace metals and isotopes, Mar. Chem., 177, 501-509, 2015.

571 Sarmiento, J. L. et al.: Response of ocean ecosystems to climate warming, Glob. Biogeochem.

572 Cycl., 18, doi:10.1029/2003GB002134, 2004.

573

574 Scotti, A.: Biases in Thorpe-scale estimates of turbulence dissipation. Part II: energetics

575 arguments and turbulence simulations, J. Phys. Oceanogr., 45, 2522-2543, 2015.

576 Sea-Bird: Fundamentals of the TC duct and pump-controlled flow used on Sea-Bird CTDs,

577 Proc. Sea-Bird Electronics Appl. note 38, SBE, Bellevue, WA, USA, 5 pp, 2012.

578 Smith, W. H. F. and Sandwell, D. T. : Global seafloor topography from satellite altimetry and

579 ship depth soundings, Science 277, 1957-1962, 1997.

580 Stansfield, K., Garrett, C., Dewey, R.: The probability distribution of the Thorpe displacement

581 within overturns in Juan de Fuca Strait, J. Phys. Oceanogr., 31, 3421-3434, 2001.

582 Strickland, J. D. H. and Parsons, T. R.: A practical handbook of seawater analysis, First

583 edition, Fisheries Research Board of Canada, Bulletin, 167, 293 pp, 1968.

584 Thorpe, S. A.: Turbulence and mixing in a Scottish loch, Phil. Trans. Roy. Soc. Lond. A, 286,

585 125-181, 1977.

586 van Haren, H.: Tidal and near-inertial peak variations around the diurnal critical latitude,

587 Geophys. Res. Lett, 32, L23611, doi:10.1029/2005GL024160, 2005.

588 van Haren, H.: Inertial and tidal shear variability above Reykjanes Ridge, Deep-Sea Res. I, 54,

589 856-870, 2007.

590 van Haren, H. and Gostiaux, L.: Characterizing turbulent overturns in CTD-data, Dyn. Atmos.

591 Oc., 66, 58-76, 2014.

592   van Haren, H. and Laan, M.: An in-situ experiment identifying flow effects on temperature

593   measurements using a pumped CTD in weakly stratified waters, Deep-Sea Res. I, 111, 11-

594   15, 2016.

595   van Haren, H., Maas, L., Zimmerman, J. T. F., Ridderinkhof, H. and Malschaert, H.: Strong

596   inertial currents and marginal internal wave stability in the central North Sea, Geophys.

597   Res. Lett., 26, 2993-2996, 1999.

598   Walter, M., Mertens, C. and Rhein, M.: Mixing estimates from a large-scale hydrographic

599   survey in the North Atlantic, Geophys. Res. Let., 32, L13605, doi:10.1029/2005GL022471,

600   2005.

601   Yamazaki, H. and Lueck, R.: Why oceanic dissipation rates are not lognormal, J. Phys.

602   Oceanogr., 20, 1907-1918, 1990.

603

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 7×10^{-5} 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 ½ (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 7×10^{-5} 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 ½

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

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

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

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

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

Figure 98. As Fig. 8$\overline{7}$, but for -100 < z < -25 m.

Figure 109. As Fig. 8$\overline{7}$, but for -600 (few nutrients sampled at 500) < z < -100 m.

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

25

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.
670
671

26

672