Microstructure measurements along a quasi-meridional transect in the northeastern Atlantic Ocean

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[1] This study presents vertical profiles of turbulence parameters obtained in the upper 100 m of the northeastern Atlantic Ocean along a transect from tropical permanently stratified waters to subpolar seasonally stratified waters in July–August 2009. The focus is to fully characterize the vertical mixing along this transect for further studies related to phytoplankton and nutrient distributions. Derived values of temperature eddy diffusivity $K_T$, of temperature variance dissipation rate $\chi_T$, and of turbulent kinetic energy dissipation rate $\varepsilon$ indicate a northward increase of turbulent mixing below the mixed layer. In the northern stations where the wind stress is sufficiently high, the vertical distributions of $\varepsilon$ in the mixed layer follow the wind stress scaling. The related low wind-scaling factors, and the low values of $\varepsilon$ encountered at this cruise, are in agreement with a strongly stratified upper water column at midday in summer.


1. Introduction

[2] Observations as well as results from climate models indicate that stratification patterns in the ocean may change due to global warming [Sarmiento et al., 1998; Levitus et al., 2000; Toggweiler and Russell, 2008]. Changes in stratification patterns may have major effects on the production and species composition of phytoplankton mainly due to changes in turbulent vertical fluxes of nutrients [Behrenfeld et al., 2006; Huisman et al., 2006; Omta et al., 2007]. This will subsequently impact grazing, virally induced mortality and sedimentation rates, with cascading effects on ecosystem functioning and biogeochemical fluxes [Kiorboe, 1993; Mann and Lazier, 2006]. Little is known, however, of the exact implications of stratification for these fundamental processes.

[3] The turbulence at the smallest scales, i.e., in the energy dissipation range, is a central issue when assessing the effects of the changes in stratification on phytoplankton [Fasham, 2002]. In the upper ocean, the generally stable stratification favors the largest property gradients in the vertical direction [Osborn and Cox, 1972] and hence a one-dimensional vertical approach is well-justified in the determination of the turbulent mixing. The properties of the vertical mixing due to small-scale turbulence are generally measured by so-called microstructure profilers. These are low-inertia free-falling instruments which measure the vertical temperature and/or the shear velocity distribution with a very high spatial resolution [Gregg and Cox, 1971]. Key derived turbulence quantities are the temperature eddy diffusivity $K_T$, computed from the temperature variance dissipation rate $\chi_T$, and the turbulent kinetic energy (TKE) dissipation rate $\varepsilon$; these quantities play an important role in phytoplankton dynamics [Peters and Marrasé, 2000; Huisman et al., 2006].

[4] The majority of the reported microprofiler surveys in the open ocean have been carried out in the Pacific [Osborn and Cox, 1972; Oezmidov, 1973; Gregg et al., 1985; Lombardo and Gregg, 1989; Moun, 1996; Sharples et al., 2001; Soloviev and Lukas, 2003], while there are relatively few surveys of the Atlantic Ocean [Shay and Gregg, 1986; Ruddick et al., 1997; Lozovatsky et al., 2005]. Apart from early results obtained during cruises by the Russian Academy of Sciences in the Pacific [Oezmidov, 1973], and a longitudinal transect in the North Atlantic near 52°N [Lozovatsky et al., 2005], most studies have been performed in localized regions of the open ocean or in short cruises covering no more than a few latitudinal degrees. The results indicate an upper ocean with large eddy diffusivities, highly patchy and variable, and a relatively quiescent interior deep ocean. Measured values of $K_T$ range from $10^{-2}$ m$^2$ s$^{-1}$ in the mixed layer to $10^{-5}$ m$^2$ s$^{-1}$ in the deep ocean away from boundaries. Values of $\varepsilon$ are more uniform in depth and generally range from $10^{-8}$ m$^2$ s$^{-3}$ to $10^{-6}$ m$^2$ s$^{-3}$. The vertical structures of $\varepsilon$ and $\chi_T$ in the oceanic mixed layer are strongly connected to the atmospheric forcing [Oakey and Elliott, 1982; Shay and Gregg, 1986; Soloviev et al., 1988; Lombardo and Gregg, 1989; Anis, 2006]. However, empirically determined scaling factors in the open ocean are scarce.

[5] In order to investigate how changes in the vertical stratification affect phytoplankton communities, the variation of the vertical turbulent mixing along a transect with
variable background stratification is fundamentally important, and it should be measured with state-of-the-art methods. The STRATIPHYT-I cruise (Las Palmas, Reykjavik, 15 July to 9 August 2009) has been designed for this purpose and it has provided daytime measurements of vertical turbulent mixing, phytoplankton, and nutrients along a gradient from strong permanent stratification in the tropics to weak seasonal stratification in the subpolar region. This change in background vertical density gradients along the section provides an ideal opportunity for the comparative study of different stratification regimes.

The primary objective of this study is to provide a detailed description of the vertical distribution of small-scale turbulent mixing measured in the STRATIPHYT-I cruise. In sections 2 and 3 we present the cruise data and the methodology chosen to derive the turbulence quantities $K_T$, $\chi_T$, and $\epsilon$. Results follow in section 4, together with the analysis of the relation between atmospheric forcing and the turbulence quantities. Further details are included in Appendix A and in the auxiliary material. A summary of the results together with conclusions is provided in section 5.

2. Cruise Data

The work presented here is based on temperature and conductivity microstructure profiles measured during the STRATIPHYT-I cruise on board of the R/V Pelagia along the transect from Las Palmas de Gran Canaria (27°53’N, 15°22’W) to Reykjavik (64°6’N, 21°50’W), via Cork (51°47’N, 8°16’W). During one month (15 July to 9 August 2009), the research vessel stopped at 32 CTD stations, achieving approximately two stations per day, each station spaced apart 50 nautical miles. At 15 CTD stations, high-resolution profiles of temperature and conductivity in the upper 100 m of the ocean were measured with a microstructure profiler (section 2.1). Standard meteorological parameters were also continuously recorded during the cruise by an onboard meteorological station (section 2.2).

2.1. Microstructure Profiler Data

2.1.1. Experimental Details

Temperature and conductivity were measured by a commercial microstructure profiler (Self Contained Autonomous Microprofiler, SCAMP), a battery-powered free-fall profiler designed to collect data up to 100 m depth, at a sampling rate of 100 Hz, and at a vertical fall velocity of 10 cm s$^{-1}$ [Stevens et al., 1999]. SCAMP’s slow falling speed allowed to collect data at a high vertical resolution which is needed to characterize the small-scale turbulent motions. The 0.76 m long profiler (diameter 0.07 m, mass 6 kg) included two fast response thermistors, a more accurate and a stable one, a conductivity pair, and a pressure sensor. The accuracy of the fast and precision thermistors were 0.05°C and 0.02°C, respectively. The accuracy characteristics of the conductivity pair were 0.45 S/m and 0.02 S/m, and for the pressure sensor, 0.5% of the full scale range. The lowest in situ noise level of the temperature measurements was equivalent to approximately 4 $10^{-9}$ °C s$^{-2}$ Hz$^{-1}$. SCAMPs have been used in a number of other field studies, most of them in coastal regions and lakes (see http://www.pme.com/HTML Docs/Scamp_Home.html for a complete record).

During our cruise, the SCAMP was deployed approximately from 11 h to 15 h, with each vertical cast taking about 20 min. In order to obtain statistically robust and unbiased estimates of mean values of turbulent parameters, a large number of casts is desired. Severe weather conditions, and the impossibility to use SCAMP while the CTD was operating, were constraints to deploy the instrument, which led to a total of 148 profiles over 15 CTD stations. The geographical locations of the stations where the SCAMP was used are shown in Figure 1; details on the station properties and the number of profiles in each station are presented in Table 1. Stations with less than 3 profiles were discarded from the analysis of the station-averaged results. The largest number of profiles were obtained at the stations where the ship stayed two days (stations 17 and 30).

2.1.2. Data Processing

The waves and the tension of the cable at the low-end point of the profile yield highly noisy microstructure data that need a strong refinement prior to usage. Selected starting and ending segments in the casts, which could not be recovered by any de-noising procedure, were removed from the analysis by trimming the profiles of temperature, conductivity and pressure according to the depth gradient and to the vertical gradient of the falling speed of the instrument. Sharpening and smoothing, and second-order Butterworth Brick-Wall filter, were then performed using a recursive
filter technique as described by Fozdar et al. [1985]. Salinity was derived from the conductivity and temperature, and the density was computed using the United Nations Educational, Scientific and Cultural Organization (UNESCO) [1981] equation of state for seawater. The depth was derived from the pressure sensor by knowing the density of the surrounding water. The molecular kinematic viscosity, $\nu$ (m$^2$ s$^{-1}$), which is weakly dependent on salinity and pressure but strongly dependent on temperature, was computed from the polynomial approximation reported by the ISW-SST (1999) tool box.

[11] Values of the vertical gradient of the temperature fluctuations, $\partial T'/\partial z$ (the $'$ indicating a trimmed-sharpened-smoothed-filtered quantity, $T$ the temperature, and $z$ the vertical coordinate), were derived from the time-based derivative of the temperature fluctuations, $\partial T'/\partial t$, divided by the falling speed of the microstructure profiler, $v_{instr}$. $\partial T'/\partial t$ was obtained from the de-trended (with respect to the temporal mean) time-based derivatives of the temperature. $v_{instr}$ was obtained from the linear least squares differentiation of the scan-based depth signal (fitting windows of 50 scans each around each scan), multiplied by the sample rate of the instrument (100 Hz).

[12] The falling velocity of the SCAMP, $v_{instr}$, must satisfy the applicability of Taylor’s [1938] hypothesis of “frozen turbulence” to convert time records into vertical profiles. This assumption, common in turbulence microstructure studies, was verified by comparing the estimated turbulence velocities with the probe velocity. The turbulence velocity scales as $u \sim (\epsilon L)^{1/3}$, where $L$ is the root mean square of the length scale of the overturning eddies. Using the values of the centered length scale $L_c$ [Imberger and Boashash, 1986], computed from the Thorpe [1977] displacements, $L$ was found to have a maximum value of $\sim 10$ m (Figure S1 in the auxiliary material), and the turbulent energy dissipation rate $\epsilon$ had a maximum value of $10^{-3}$ m$^2$ s$^{-3}$ (section 4.1). This results in a turbulent velocity scale $u < 10^{-2}$ m s$^{-1}$ which is considerably lower than the probe falling velocity, in the order of 0.1 m s$^{-1}$. Therefore, Taylor’s [1938] hypothesis was used for all data segments measured by the microprofiler.

[13] In order to obtain estimates of the turbulent quantities, a further depth-binning process was performed. Depth-binning was conducted in segments of 1 m (with 700 scans in each bin approximately). The 1-m scale was chosen as a suitable trade-off between the need of high vertical resolution and the need of statistical robustness. More details of the binning procedure are presented in Appendix A.

2.1.3. Temperature, Salinity, and Buoyancy Frequency Profiles

[14] Depth-binned and station-averaged temperature, salinity and density profiles (Figure 2) show the expected variation from a warmer and saltier water column in the south to a colder and fresher water column in the north. Stations south of 40°N (similar to station 5 in Figure 2) show a decrease of salinity and temperature with depth. Taken on its own, the decrease of salinity would produce an unstable density stratified water column. However, the effect of the temperature decrease is stronger than the effect of the salinity decrease, yielding a stably stratified water column.

[15] In the stations south of 45°N, density profiles present a series of steps as a function of depth below the pycnocline, commonly found in the permanently stratified water columns. The uncertainty in the station-averaged profiles is thus larger in the low-latitude stations as reflected in the 95% confidence limits in Figure 2. The scatter presented in the pycnocline may be caused by the internal waves that the stratified pycnocline supports.

[16] The upper nearly uniform part of the profiles in Figure 2 provides an indication for the depth of the mixed layer (MLD), ranging from 20 to 45 m. As mixed layer depth, we took the depth at which the temperature difference with respect to the surface was 0.5°C [Levitus et al., 2000]. This criterion, as shown in the work of Brainerd and Gregg [1995], allows to have an estimate of the depth through which surface waters have been mixed in the recent past (where recent past often means within the preceding daily cycle).
The strength of the background stratification in the water column was quantified with the buoyancy (Brunt-Väisälä) frequency $N$ (rad s$^{-1}$)

$$N = \sqrt{-\frac{g}{\rho_0} \frac{\partial \bar{\rho}}{\partial z}}$$

where $g$ (m s$^{-2}$) is the gravitational acceleration, $\bar{\rho}$ (kg m$^{-3}$) is the trimmed-smoothed-sharpened-filtered and also depth binned (hereafter referred to as TSSFB) water density, $\rho_0$ is a reference density, and $z$ indicates the vertical coordinate (positive upward) in m.

Figure 2. Profiles of (a) temperature $T$, (b) salinity $S$, and (c) sigma-t $\sigma_t$ (density $\rho - 1000$) at 5 selected stations. The 95% confidence limits around the station-average are also shown. The legend shows the station number and the mixed layer depth (MLD). The overbar indicates a trimmed-smoothed-sharpened-filtered and binned quantity, and the angle brackets indicate a station-average.

Figure 3 presents the buoyancy frequency along the transect, with the bins with static instabilities ($N^2 < 0$) marked in grey. The static instabilities are an important source of turbulence generation in the ocean, as they involve the release of potential energy through the descent of a plume of denser water and the ascent of a plume of lighter water. The distribution of the buoyancy frequency clearly shows the evolution from the permanently stratified waters to the seasonally stratified waters, as seen in the decrease of $N$ just below the MLD in the high-latitude stations. The relative standard deviations due to bin-averaging, $RSD_{bin}$, of the values depicted in Figure 3 range between 40% and 380%, with a mean of 53%. Higher values of $RSD_{bin}$ were

Figure 3. Depth-binned profiles of the squared buoyancy frequency $N^2$, derived from the SCAMP measurements along the transect. Dotted vertical lines delimit the sampling stations and the number of the corresponding station is shown at top. The mixed layer depth (MLD) for each cast is also plotted as the thick black curve.
located close to the surface and in the thermocline (see Appendix A for details about the computation of RSDbin).

2.2. Meteorological Data

[19] The onboard meteorological station measured air temperature, humidity, wind speed and direction, downward short-wave radiation flux, and air pressure. Data was sampled every minute and stored as averages over the time of each SCAMP cast. Cloud cover and wave height were estimated from observations made every 3 hours; the rainfall rate was estimated 2 mm h\(^{-1}\) during the raining days (stations 17a, 22, and 27). Together with the sea surface temperatures and the sea surface salinity derived from the extrapolation to the surface from those measured by the microstructure profiler, the meteorological data was used to compute the net surface heat flux \(Q_0\), the net surface buoyancy flux \(B_0\), and the surface wind stress \(\tau_0\). These quantities allow to assess the relation between the atmospheric forcing and the upper ocean turbulence, as will be presented in section 4.2. The calculations of \(Q_0\) and \(\tau_0\) were performed with the Matlab Air-Sea toolbox (version 2.0, http://sea-mat.whoi.edu), which employs a simplified version of the Fairall et al. [1996] TOGA/COARE code. The calculation of the net surface buoyancy flux \((B_0 = B_t + B_s)\) was obtained from the thermal buoyancy flux \((B_t = g a Q_0 / (\rho_s C_{pw}))\), with \(a\) the thermal coefficient of expansion of seawater, \(\rho_s\) the water density at atmospheric pressure, \(C_{pw}\) the specific heat capacity of seawater) and from the haline buoyancy flux \((B_s = g b (E/C_0 P) S_0)\), with \(b\) the coefficient of haline contraction of seawater, \(S_0\) the surface salinity, and \(E/C_0 P\) the difference between the evaporation rate and the precipitation rate).

[20] During the cruise, the net surface heat flux was directed from the atmosphere to the ocean at most of the SCAMP casts (Figure 4). The related net surface buoyancy fluxes ranged from \(-5.8 \times 10^{-7}\) to \(-1.2 \times 10^{-8}\) W kg\(^{-1}\) (Figure 4b) and were always stabilizing the upper ocean. Note that in this paper we define the heat fluxes (and net

Figure 4. Meteorological data averaged over the time of each SCAMP profile: (a) net surface heat flux \(Q_0\) and related components (latent heat flux \(Q_{la}\), sensible heat flux \(Q_{se}\), net long-wave heat flux \(Q_{lw}\), short-wave radiation flux \(Q_{sw}\)); (b) net surface buoyancy flux \(B_0\) and related components (thermal surface buoyancy flux \(B_t\), haline surface buoyancy flux \(B_s\)); (c) surface wind stress \(\tau_0\); and (d) Monin-Obukhov length LMO, and mixed layer depth MLD. \(Q_0\) is defined positive upward, and hence for \(B_0 < 0\) the buoyancy flux is stabilizing. Each marker represents a profile, and each dotted vertical line indicates the last profile of each sampling station. The number of the corresponding station is shown at top.
surface buoyancy fluxes \( B_0 \) as negative when directed downward (from the atmosphere to the ocean) and thus \( B_0 < 0 \) indicates a buoyancy flux stabilizing the ocean. The surface wind stress ranged from 0.0 to 0.3 N m\(^{-2}\) (Figure 4c) and was derived from wind speeds that ranged from 1.0 to 12.2 m s\(^{-1}\) at 10 m height (Figure S2 in the auxiliary material). The meteorological quantities indicate a progression from more fairly constant station-averaged wind stress and more negative surface buoyancy fluxes in the lower latitude stations, toward more extreme station-averaged wind stress and buoyancy fluxes only slightly stabilizing the upper ocean.

[21] An additional meteorological-related quantity, the Monin-Obukhov length (LMO, m), was computed for each cast of the cruise as \( LMO = -u_0^2/(\kappa B_0) \), where \( u_0 = (\tau_0/\rho_0)^{1/2} \) (m s\(^{-1}\)) is the surface friction velocity and \( \kappa = 0.4 \) is the von Kármán constant. Along the transect, we find \( LMO > 0 \) (note that \( B_0 < 0 \)), indicating that the mechanically (wind) generated turbulence is suppressed by the stable stratification. Larger values of LMO, indicating a stronger wind generated turbulence or weaker stratification, tended to occur in the high-latitude stations. LMOs were generally lower than the respective MLD, indicating that the wind-induced mixing at the time of sampling could not explain the turbulence over the total depth of the mixed layer (Figure 4d).

3. Determination of Turbulent Quantities

[22] Values of the temperature eddy diffusivity, \( K_T \) (m\(^2\) s\(^{-1}\)), were determined from the temperature microstructure measurements using the Osborn and Cox [1972] model. Assumptions in the Osborn and Cox [1972] model include a stationary balance of the variance of temperature fluctuations, homogeneous and isotropic turbulence, dominance of the vertical heat fluxes over horizontal ones, and the omission of the divergence of the transport terms by the mean flow. \( K_T \) was computed from the temperature variance dissipation rate, \( \chi_T \) (°C s\(^{-1}\)), according to

\[
\chi_T = 6D_T \left( \frac{\partial T}{\partial z} \right)^2; \quad K_T = \frac{\chi_T}{2} \left( \frac{\partial T}{\partial z} \right)^{-2}
\]

where \( D_T \) is the molecular diffusivity of heat (\( \approx 1.4 \times 10^{-7} \) m\(^2\) s\(^{-1}\)). The determination of the vertical gradient of the TSSFB temperature, \( \partial T / \partial z \), and the determination of the TSSFB temperature fluctuations gradient, \( \partial T / \partial z \), and the related uncertainty, are described in Appendix A. It should be noted that both \( K_T \) and \( \chi_T \), as derived from (2), are 1-m depth-binned quantities.

[23] The TKE dissipation rate, \( \epsilon \) (m\(^2\) s\(^{-3}\)), is usually calculated from the vertical shear velocity fluctuations of the flow. Because the profiler used in this study had no sensors of velocity microstructure, \( \epsilon \) was estimated by fitting in each 1-m segment the theoretical Batchelor [1959] spectrum of the vertical gradient of the temperature fluctuations to the observed spectrum of the vertical gradient of the temperature fluctuations [Oakey, 1982]. We used the maximum likelihood method developed by Ruddick et al. [2000], which has an explicit incorporation of the instrumental noise, and allows an automated rejection of the segments that poorly fit the spectra. The spectrum of the temperature fluctuations gradient was obtained using a fast Fourier transform with a Hamming window. The noise spectrum was modeled under the assumption that the noise is the result of white noise from the thermistor passing through the SCAMP spatial derivative channel, and other theoretical–electronic considerations are already corrected for in the host SCAMP software.

[24] In each segment, the representative \( \epsilon \) was determined from the corresponding optimal value of the Batchelor wave number (\( k_{\text{segment}} \) cpm):

\[
\epsilon_{\text{segment}} = (2\pi k_{\text{segment}})^4 \nu_{\text{segment}} D_T^2
\]

where \( \nu_{\text{segment}} \) (m\(^2\) s\(^{-1}\)) is the arithmetic mean of the molecular kinematic viscosity in the segment. Various examples of 1-m segments that passed the rejection criteria from Ruddick et al. [2000] are presented in Figure S3 in the auxiliary material. A reliable estimation of \( k_{\text{segment}} \) was critical because of the sensitivity of \( \epsilon_{\text{segment}} \) to \( k_{\text{segment}} \) according to (3).

[25] The Batchelor fitting is endorsed by the pioneering studies by Dillon and Caldwell [1980] and Oakey [1982], who concluded that \( \epsilon \) determined indirectly through Batchelor fitting agrees within a factor of 2 with the \( \epsilon \) determined from records of velocity shear. The more recent study performed in a lake by Kocsis et al. [1999] leads to similar conclusions. Nash and Moum [2002] slightly favor the theoretical temperature spectrum proposed by Kraichnan [1968] over the Batchelor [1959] spectrum. However, the physics are clearer in Batchelor’s model than in Kraichnan’s semi-empirical model, and we used the former as the basis of our estimations. Besides, the fitting algorithm used in this work was developed for the Batchelor spectrum and for data of SCAMP [Ruddick et al., 2000]; indeed, fitting to another spectrum would lead to an increase in uncertainty due to the lack of clear rejection criteria.

[26] The MATLAB routines used to compute these turbulent quantities were partly based on the processing software provided by PME with the SCAMP system, version 1.09 (http://www.pme.com/HTML/Docs/Scamp_Home.html).

4. Results and Discussion

4.1. Derived Turbulent Quantities

[27] The temperature eddy diffusivity \( K_T \), the temperature variance dissipation rate \( \chi_T \), and the TKE dissipation rate \( \epsilon \) show a substantial variability, with values that are in the range of reported values in the literature (Figure 5). Over the upper 100 m, the station-averaged values of \( K_T \) (\( \langle K_T \rangle = 10^{-8} - 10^{-9} \) m\(^2\) s\(^{-1}\)) within the mixed layer, \( \langle K_T \rangle = 10^{-7} - 10^{-8} \) m\(^2\) s\(^{-1}\) below the mixed layer) and \( \epsilon \) (\( \langle \epsilon \rangle = 10^{-8} - 10^{-9} \) m\(^2\) s\(^{-3}\) within the mixed layer, \( \langle \epsilon \rangle = 5 \times 10^{-9} - 10^{-7} \) m\(^2\) s\(^{-3}\) below the mixed layer) are low, characteristic of highly stratified water columns measured at midday and in summer. The uncertainty of the turbulent quantities, expressed by the relative standard deviation \( \text{RSD}_{\text{bin}} \) as determined in Appendix A, ranges from \( \sim 80\% \) to \( 200\% \) for \( K_T \), from \( \sim 30 \) to \( 70\% \) for \( \chi_T \), and from \( \sim 3\% \) to \( 10\% \) for \( \epsilon \). A careful validation is difficult due to the inherent variability of turbulent-related quantities and the lack of reported values in the region covered by the cruise.
The values $K_T$ (Figure 6a) peak around the middle of the mixed layer and decrease both toward the surface (where the proximity of the boundary reduces the size of the turbulent eddies) and toward the bottom of the mixed layer (where the increased water column stability reduces the turbulent mixing). Both $\chi_T$ (Figure 6b) and $\varepsilon$ (Figure 6c) display enhanced values near the surface, which may be due to the presence of surface waves. $\chi_T$ profiles appear substantially affected by the density gradients which peak in the pycnocline. The vertical distribution of $\varepsilon$ is more uniform than that of $K_T$ and $\chi_T$. However, the differences of $\varepsilon$ between one cast and another one (e.g., the one 20 min earlier) are considerable. Note that there is a large percentage (around 40\%) of rejected bins for $\varepsilon$ due to a bad fitting to a Batchelor spectrum.

The inter-station variability of the turbulent quantities (Figure 7) is a distinctive feature of this study due to the large latitudinal coverage of the cruise. We observe an increase of the station- and column-averaged $\varepsilon$ (in the mixed layer and below the mixed layer) and $K_T$ (below the mixed layer) toward the northern stations. Within the mixed layer, $K_T$ is more affected by the local atmospheric conditions and lacks an inter-station trend. The quantity $\chi_T$, largely affected by thermal stratification, decreases northward. The inter-station variability may be connected to the reported northward weakening of the background stratification (Figure 3), and with the large amount of energy supplied to the ocean in the temperate regions in response to the intermittent passage of storms. Note that the inter-station tendencies are based on relatively few profiles in each station; more experimental evidence is needed to derive meaningful latitudinal trends.

### 4.2. Atmospheric Forcing of the Turbulent Quantities

Column- and station-averaged values of temperature eddy diffusivity $K_T$ in the mixed layer correlate positively with the wind speed (Figure 6 and Figure 7). The best linear least squares fit follows from $\langle K_T^* \rangle_{-\text{MLD}} = 0.001 \langle \mu_{10} \rangle - 0.003$ with a Pearson’s correlation coefficient of 0.58 ($\langle \rangle$ indicates a station-average, $1^*_{-\text{MLD}}$ column-average from MLD to 1 m depth, and $\mu_{10}$ is the wind speed at 10 m height). The slope of the linear regression differs significantly from zero (P-value from the linear regression t-test is 0.01, which is lower than the significance level of 0.05).

Another effect of the atmospheric forcing on $K_T$ is disclosed when analyzing its vertical distribution. $K_T$ profiles show a sharp decrease close to the MLD when the Monin-Obukhov length is relatively large compared to the MLD (Figure 8, station 30a). Conversely, when the Monin-Obukhov length is relatively small, the profiles of $K_T$ show a smoother decrease at the MLD (Figure 8, station 7).

To investigate the relation between the TKE dissipation rate, $\varepsilon$, and the atmospheric forcing, we have assessed the scaling of $\varepsilon$ with the related wind stress similarity variable, $\varepsilon_{\text{vs}}$. We have not assessed the scaling of $\varepsilon$ with the related convective similarity variable because none of the measured profiles presented turbulent production dominated by convection (all the casts had downward net surface buoyancy fluxes). We have not assessed the relation of the temperature variance dissipation rate, $\chi_T$, and the atmospheric forcing because the scaling has proven successful only when the turbulent production was dominated by convection [Brubaker, 1987; Lombardo and Gregg, 1989]. The $\varepsilon_{\text{vs}}$ follows from

$$\varepsilon_{\text{vs}} = \frac{\mu_{10}^*}{-\kappa z}$$

where $z$ is defined positive upward, with $z = 0$ at the sea surface and $\kappa = 0.4$ is again the von Kármán constant. In conditions of dominant wind stress forcing, $\varepsilon_{\text{vs}}$ was found to average around 1 in the mixed layer [Oakey and Elliott, 1982; Soloviev et al., 1988; Anis and Moum, 1995], i.e., the TKE dissipation rate followed the so-called law of the wall with a proportionality constant of 1.

In the STRATPHYT-I cruise, only three stations presented mixed layers dominated by wind mixing, as determined by larger LMO compared to the MLD in all the casts of the station: station 25 (with only one cast and MLD/LMO = 0.6), station 29 (MLD/LMO = 0.8), and station 30a (MLD/LMO = 0.2). Their station-averaged $\varepsilon$ follow the law of the wall with proportionality constants of 0.25, 0.16 and 0.21 respectively. The proportionality constants have been calculated by averaging $\langle \varepsilon_{\text{vs}} \rangle$ throughout the mixed layer and avoiding the first 5 m close to the surface because of possible surface wave breaking contamination. Figure 9a depicts the profiles of scaled TKE dissipation rate for a wind-dominated mixed layer (low MLD/LMO). In this station, $\langle \varepsilon_{\text{vs}} \rangle$ normalizes $\langle \varepsilon \rangle$ to a uniform value of 0.21 in most of the mixed layer (0.2 < $z$/MLD < 0.8). The deviation from the scaling in the first meters of water column may be related to surface wave activity [Soloviev
Figure 9b shows the scaled TKE dissipation rate for a station where the MLD/LMO ratio is relatively high. The corresponding $\langle \varepsilon \rangle$ doesn’t scale with $\langle \varepsilon_s \rangle$ to a uniform value, and $\langle \varepsilon \rangle$ deviates from the law-of-the-wall prediction. Note that mixed layers with relatively strong wind mixing, and thus the wind stress scaling of $\langle \varepsilon \rangle$, occurred in stations northward 45$^\circ$N. It reflects the average atmospheric conditions of high-latitude regions, where wind induced mixing tends to be more important than convective mixing.

[34] The proportionality constants determined in the relation $\varepsilon/\varepsilon_s$ during the STRATIPHYT-I cruise are lower than those reported at other oceanic locations. The majority of the reported empirical constants are based on measurements during nighttime [Soloviev et al., 1988; Lombardo and Gregg, 1989; Anis and Moum, 1995], during winter or spring [Lozovatsky et al., 2005], or in the continental shelf [Oakey and Elliott, 1982]. Accordingly, these studies measured higher TKE dissipation rates than those measured in STRATIPHYT-I, and explain the lower proportionality constants found here. The time shift between $u_10$ and $\varepsilon$ may also have contributed to a decrease in our determined empirical constants [Lozovatsky et al., 2005]. Our estimated empirical constants agree with those derived from the few
daytime summer profiles analyzed by Anis and Moum [1995] and Soloviev et al. [1988].

5. Summary and Conclusion

[35] In this study we have presented a novel data set of temperature eddy diffusivities, KT, temperature variance dissipation rates, χT, and indirectly derived TKE dissipation rates, ε, in 15 ocean CTD stations from Las Palmas de Gran Canaria (27°55’N, 15°22’W) to Reykjavik (64°6’N, 21°50’W), via Cork (51°47’N, 8°16’W). The data was obtained with a commercial free-fall microstructure profiler, deployed continuously from 11 h to 15 h, during July August 2009, in the upper ~100 m of the ocean.

[36] Depth-binned and station-averaged temperature, salinity, and density profiles show the expected variation from a warmer and saltier water column in the low-latitude stations to a colder and fresher water column in the subpolar stations. The derived buoyancy frequencies, N, present also an expected change from strongly stratified water columns to weakly stratified water columns. Measured meteorological parameters indicate mixed layers dominated by wind mixing at three stations north of 45°N. Mixed layer depths (MLDs) range from 20 to 45 m, and the results presented here are representative of a tropical to subpolar Atlantic transect in the summer and during daytime.

[37] The derived temperature eddy diffusivities, KT, display higher values in the mixed layer (⟨KT⟩: 10^{-6} – 10^{-5} m^2 s^{-1}), with both a decrease toward the surface and toward the MLD. Below the thermocline, KT tends to be uniform, with ⟨KT⟩ ranging 10^{-7} m^2 s^{-1} to 10^{-6} m^2 s^{-1}. The temperature variance dissipation rates, χT, present peak values in the first 10 m of the water column and at the base of the mixed layer, before reaching station-averages of 10^{-10} °C^2 s^{-1} to 10^{-7} °C^2 s^{-1} at around 100 m depth. The TKE dissipation rates, ε, present more uniform depth profiles ranging from 5 × 10^{-9} to 10^{-6} m^2 s^{-3}. The values are in the lower range of reported turbulence levels in the upper ocean, and are representative of strongly stratified waters at midday in summer.

Figure 7. Station- and column-averaged turbulent quantities in the mixed layer and below the mixed layer, together with the station-averaged wind speeds. Shown are (a) wind speed at 10 m height u_10, (b) temperature eddy diffusivity KT, (c) temperature variance dissipation rate χT, and (d) TKE dissipation rate ε. Note that station 25 is not considered because it had only one profile. The linear fitting of the column- and station-averaged turbulent quantities with the distance traveled by the ship is added (d denotes the distance in nautical miles, p denotes the P-value from the linear regression t-test). The slope of the linear regression differs significantly from 0 when p is lower than the significance level, chosen here as 0.05. The angle brackets indicate a station-average, and the notation 1 indicates a column-average. The standard deviation of the station-averaged wind speeds is depicted in Figure 7a. The standard deviation of the values depicted in Figures 7b–7d is presented in Table S1 in the auxiliary material.
Inter-station tendencies derived here are based on limited profiles; further experimental evidence and theoretical analysis is needed to be able to extract a meaningful latitudinal gradient of the turbulent quantities.

Column- and station-averaged $K_T$ in the mixed layer correlate positively with the wind speed $\langle K_T \rangle_{\text{MLD}} = 0.001 \langle u_{10} \rangle - 0.003$. In the stations with mixed layers dominated by wind mixing, as determined by larger Monin-Obukhov length compared to the MLD in all the casts of the station, TKE dissipation rates scale with the wind stress similarity variable $\varepsilon/\varepsilon_s$ in most of the mixed layer ($0.2 < -z/\text{MLD} < 0.8$). Scaling factors range from 0.16 to 0.21, substantially lower than the commonly reported value of 1 [Oakey and Elliott, 1982]. The low levels of TKE dissipation rates measured in this cruise, linked to midday and summer conditions of our measurements, may explain the low scaling factors.

As far as we know, the data set presented here constitutes the largest “synoptic” quasi-meridional transect of microstructure measurements in the Atlantic. The values reported here should be used with caution, they are linked to midday conditions during the northeast Atlantic summer. In other seasons and other times of the day, the mean levels of turbulence may be higher, and also the MLD may be different. Our results hopefully will be important for providing a picture of the changes of turbulent quantities in a meridional transect with a decreasing background stratification. This data may be useful for ocean-biochemistry.
models to better understand changes in nutrient and phytoplankton distributions under different stratification scenarios.

**Appendix A: Averaging Procedure and Determination of Uncertainty**

[41] Due to the intermittent character of turbulence, a single profile can give misleading results. For that reason, the obtained profiles of turbulent quantities along the cruise are depth-binned and station-averaged. Station-averages are obtained using the arithmetic mean. In this work we refer to the hat in \( \bar{m} \) as the operation of trimming-smoothening-sharpening and filtering (also referred to as TSSF) on the quantity \( m \); the overbar in \( \bar{m} \) refers to the operation of trimming-smoothening-sharpening-filtering and also depth-binning (also referred to as TSSFB) on the quantity \( m \). The angle brackets refer to the operation of station-averaging, and the notation \( \langle \cdot \rangle \) refers to the operation of column-averaging.

[42] Data-binning is performed in segments of 1 m for all depths. In each segment, the TSSF data values are replaced by the segment central value, which corresponds directly to depths. In each segment, the TSSF data values are replaced by the arithmetic mean in the case of temperature \( T \), salinity \( S \), and density \( \rho \). In particular, the depth-binned values of the buoyancy frequency, \( N \), are determined by first estimating the vertical gradient of the mean (background) density, \( \partial \rho / \partial z \), obtained from the best linear least squares fit in the 1 m depth bin (~700 scans), using a 40 scans-moving average on the TSSF density versus TSSF depth. In the case of the temperature eddy diffusivity, \( K_T \), the depth-binned value is related to \( \partial T / \partial z \), obtained from the best linear least squares fit in the 1-m bin, using a 40 scans-moving average on the TSSF and sorted temperature versus the TSSF depth. In the case of the temperature variance dissipation rate, \( \chi_T \), the segment central value is related to \( \partial^2 T / \partial t \), which is obtained from the arithmetic mean of \( \partial^2 T / \partial t \). Finally, the TKE dissipation rate, \( \varepsilon \), is estimated directly for 1 m segments, and depth binning is not required.

[43] The uncertainty in a depth-binned parameter is quantified through the relative standard deviation, \( RSD_{bin} \) (%), of the bin-central value, where \( RSD_{bin} = 100 \sigma_{bin}/\mu_{bin} \) (\( \sigma_{bin} \) is the standard deviation and \( \mu_{bin} \) the arithmetic mean in the 1-m bin of the TSSF values). The computed \( RSD_{bin} \) increases slightly with the uncertainty derived from the instrumental error. In particular, the \( RSD_{bin} \) of \( T, C \), and \( \rho \) is directly estimated as 100 \( \sigma_{bin}/\mu_{bin} \). The \( RSD_{bin} \) of \( N \) is estimated from the residuals in the least squares procedure to obtain the central value of the bin. In the case of \( K_T, \chi_T \), and \( \varepsilon \), as \( RSD_{bin} \) we consider the upper limit resulting from the sum of the uncertainties of the elements in their formulae:

\[
RSD_{bin}(K_T) = RSD_{bin}(\chi_T) + 2 \cdot RSD_{bin}(\partial^2 T / \partial z^2) \]

\[
RSD_{bin}(\chi_T) = 2 \cdot RSD_{bin}(\partial T / \partial z) \]

\[
RSD_{bin}(\varepsilon) = 4 \cdot RSD_{bin}(K_B) \]

The depth-binned uncertainty of \( \partial T / \partial z \) is computed from the standard deviation and the mean of \( (T_{max} - T_{min})/(\Sigma T_{max} - \Sigma T_{min}) \) in a 40 scans-moving average. The depth-binned uncertainty of the Batchelor wave number \( K_B \) is determined from the goodness of the fit of the theoretical spectrum of the vertical gradient of the temperature fluctuations to the observed spectrum as described by Ruddick et al. [2000]. Additionally, the \( RSD_{bin} \) of \( K_T, \chi_T \), and \( \varepsilon \) may increase when the uncertainties due to the assumption of a steady homogeneous isotropic turbulence and to the Taylor [1938] hypothesis of “frozen turbulence” are taken into account.

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