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Sheared turbulence in a weakly stratified upper ocean

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Abstract

Microstructure, ADCP and CTD profiles taken in the North Atlantic along 53°N under moderate and high winds showed that the median of log-normally distributed kinetic energy dissipation rate ε within the upper mixing layer is 1.5×10^{-7} W/kg and the layer depth, on the average, is ~40 m. Assuming that mixing efficiency γ is a constant ($\gamma = 0.2$), the following scaling is proposed for the normalized eddy diffusivity:

$\hat{K}_\mathrm{b} = K_\mathrm{b}/\kappa u_* z = (1 + Ri/Ri_\mathrm{cr})^{-p} Pr_\mathrm{tr}^{-1}$,

where $K_\mathrm{b} = \gamma\varepsilon/N^2$, N^2 is the squared buoyancy frequency, u_* the surface friction velocity, Ri the local Richardson number, $Pr_\mathrm{tr} = 1 + Ri/Ri_\beta$ the turbulent Prandtl number, $p = 1$ or 2/3, $Ri_\mathrm{cr} = 0.1$ and $Ri_\beta = 0.1$ or 0.05. The power-law function with $p = 1$ relates the asymptotes of $K_\mathrm{b}(Ri)$ to the buoyancy scale $L_N \sim (\varepsilon/N^3)^{1/2}$ at $Ri \gg Ri_\mathrm{cr}$ and to the shear scale $L_\mathrm{Sh} \sim (\varepsilon/\mathrm{Sh}^3)^{1/2}$ at $Ri \ll Ri_\mathrm{cr}$. If $p = 2/3$, the lengthscale $L_R = (\varepsilon/N^2\mathrm{Sh})^{1/2}$ replaces L_N in spectra of ocean microstructure than due to the influence of local shear. This mixing regime corresponds to intermediate Richardson numbers ($\sim 0.25 < Ri < \sim 2$).

Alternatively, if γ is not a constant, but an increasing function of Ri for $0 < Ri < 1$, then $\hat{K}_\mathrm{b}(Ri)$ shows a very weak dependence on Ri for $Ri < 0.25$. Numerical experiments using a one-dimensional q^2–ε model with different K_b parameterizations indicate that the measured mixed-layer depth agrees well with model results when the diffusivity is parameterized with an Ri-dependent γ (the GISS model approach). The modeled dissipation profiles, however, resembled microstructure measurements better if γ is treated as a constant and the proposed formula for $\hat{K}_\mathrm{b}(Ri)$ is used.

Keywords: Turbulence; Dissipation; Diffusivity; Mixing parameterization; Upper ocean; Modeling

1. Introduction

Drift currents and associated wind-induced mixing are important components of upper oceanic dynamics. Depending on the wind stress and variability of buoyancy flux at the sea surface, turbulent mixing in the upper layer can produce vertical transports of kinetic energy, momentum and heat down to the underlying pycnocline or across the sea surface to the atmosphere. The upper ocean in high and mid-latitudes is usually well mixed by sustained moderate and high winds. With the exception of relatively short periods of intense

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convective mixing, the residual mean stratification of the upper layer is weakly stable with an averaged squared buoyancy frequency $\tilde{N}^2 = (g/\rho_0)\partial\bar{\rho}/\partial z$ of about $10^{-6}\mathrm{s}^{-2}$ or less; here ρ_0 is the reference density, $\partial\bar{\rho}/\partial z$ the mean vertical density gradient (z is positive downward), and g the gravity. Small density gradients and large vertical shears $\mathrm{Sh} = \sqrt{(\partial u/\partial z)^2 + (\partial v/\partial z)^2}$ in the upper layer ($\mathrm{Sh}^2 \sim 10^{-5}$–$10^{-4}\mathrm{s}^{-2}$) provide favorable conditions for developing Kelvin–Helmholtz (K–H) instability, and thus local turbulent mixing, when the Richardson number, $Ri = N^2/\mathrm{Sh}^2$, falls below a critical value Ri_{cr}. Traditionally such turbulence is parameterized in terms of exchange coefficients also known as eddy viscosity K_{M} and diffusivity K_{b}. The exchange coefficients are assumed constant (say K_{Mo} and K_{bo}, respectively) for $Ri \ll Ri_{\mathrm{cr}}$, Ri dependent at intermediate Ri whilst decreasing to the molecular values at $Ri \gg Ri_{\mathrm{cr}}$. For locally generated turbulence, this variation of K_{M} and K_{b} can be modeled as

$$K_{\mathrm{M}} \sim K_{\mathrm{Mo}}(1+\beta Ri)^{-m} \text{ and } K_{\mathrm{b}} \sim K_{\mathrm{bo}}(1+\beta_{\mathrm{b}} Ri)^{-n}, \tag{1}$$

where m, n, β, and β_{b} are constants. Some commonly used variables are, $m = 1/2$, $\beta = 10$, $n = 3/2$, $\beta_{\mathrm{b}} = 10/3$ (Munk and Anderson, 1948); $n = 1$, $\beta_{\mathrm{b}} = 5$ (Monin and Yaglom, 1975); $m = 2$, $n = 3$, $\beta = \beta_{\mathrm{b}} = 5$ (Pacanowski and Philander, 1981, hereafter PP-81); $m = 1.5$, $n = 2.5$, $\beta = \beta_{\mathrm{b}} = 5$, for $Ri > 0.25$ (Peters et al., 1988, hereafter PGT-88), $n = 3/2$, $\beta_{\mathrm{b}} = 10$, (Pelegri and Csanady, 1994); and $m = 1/2$, $n = 3/2$, $\beta = 10$, $\beta_{\mathrm{b}} = 20$ (Paka et al., 1999). Based on the direct measurements of microstructure in the ocean, Soloviev et al. (2001) recently reported $K_{\mathrm{M}} = K_{\mathrm{b}} \sim K_0(1 - Ri/Ri_{\mathrm{cr}})$ for $Ri < Ri_{\mathrm{cr}} = 0.25$, which signifies a very weak dependence on Ri at $Ri \leqslant 0.1$. A similar result has been found for atmospheric stratified flows (Monti et al., 2002). On the other hand, PGT-88 showed an extremely steep variation $K_{\mathrm{b}} \sim Ri^{-9.6}$ for $Ri < Ri_{\mathrm{cr}}$, but obviously, on physical grounds, the diffusivities are expected to asymptote to their non-stratified values as $Ri \to 0$. The discrepancy between different results, however, is curious as it points to the unreliability of existing scaling for $Ri \ll Ri_{\mathrm{cr}}$. We believe that the difficulties of N^2 calculations in quasi-homogeneous layers, the limited bandwidth of airfoil sensors that are used to measure small-scale shear in highly turbulent zones and insufficient averaging of K_{b} and Ri are major contributors to the high uncertainty of $K_{\mathrm{b}}(Ri)$ at small Ri (see Appendix B for details). In addition, possible deviation of mixing efficiency γ (at subcritical Ri) from the assumed constant value $\gamma = 0.2$ (Oakey, 1982) used for K_{b} calculations is also an important contributor to the mystery (e.g., Strang and Fernando, 2001a, b; Smyth et al., 2001; Canuto et al., 2001; Fringer and Street, 2003).

In this paper, new microstructure measurements of the dissipation rate ε, thermohaline, and velocity structure in the upper layer of North Atlantic are presented to further support the notion of relatively weak growth of K_{b} at $Ri \ll Ri_{\mathrm{cr}} \approx 0.1$ (Paka et al., 1999). Note that the dependence of eddy viscosity K_{M} on Ri is believed to be weaker than of that of K_{b}, but the critical Richardson number seems to play the same role in the demarcation of weakly and strongly stratified regimes for both cases. Recent atmospheric studies (e.g., Monti et al., 2002) indicate that K_{M} may even grow at $Ri > Ri_{\mathrm{cr}}$ possibly because of the increase of momentum transfer by internal waves in stably stratified boundary layers (Lee et al., 2005); a parameterization based on this result has produced better predictions in mesoscale atmospheric models. Direct measurements of vertical momentum flux in the marine environment remain technically challenging (Moum, 1998), and thus understanding of the role of momentum transfer in oceans has made only limited progress.

The data to be discussed herein (Section 2) were obtained during the 9th cruise of R./V. *Akademik Ioffe* (P.P. Shirshov Institute of Oceanology, Russian Academy of Sciences) along 53°N in April 2001 (Tereschenkov et al., 2002), and they have already been utilized to study the response of mixed-layer depth to atmospheric forcing (Lozovatsky et al., 2005). The present work focuses on statistics of the dissipation rate ε, Ri, K_{b}, and the buoyancy Reynolds number R_{b} (Section 3) and on the dependence of local K_{b} on Ri within the upper boundary layer (Section 4). Here, a new version of $K_{\mathrm{b}}(Ri)$ scaling is proposed and we also compared the processed data with several widely used mixing parameterizations. The effects of an Ri-dependent γ on the behavior of $K_{\mathrm{b}}(Ri)$, which greatly impacts relevant numerical modeling, are illustrated and discussed in Sections 5 and 6. The main results are summarized in Section 7.