Mixing rates across the Gulf Stream, Part 1: On the formation of Eighteen Degree Water

R. Inoue, M. C. Gregg and R. R. Harcourt

Applied Physics Lab, University of Washington, 1013 NE 40th, Seattle, WA 98105-6698, USA

Draft 8.1, Sep 10 2009
Abstract

Microstructure profiles taken in February 2007 across the Gulf Stream (GS) measured the temporal and spatial variability of the intense mixing that forms Eighteen Degree Water (EDW). Strong winds, gusting to 60 kt, and heat fluxes up to 1000 W m$^{-2}$ produced moderate-to-strong mixing in the surface mixed layer and the entrainment zone, as well as in the thermocline. In the limit of a vertically balanced heat budget, EDW formation is driven primarily by surface heat loss to the atmosphere across a region extending O(100) km south from the GS core, where entrainment heat fluxes based on dissipation rates were relatively low, O(10$^{10}$~100) W m$^{-2}$. Near the GS core, much larger entrainment fluxes, O(100~1000) W m$^{-2}$, contribute significantly to cooling the mixed layer, but less so to overall EDW formation due to its smaller volume. Below the mixed layer near the GS, diapycnal diffusivities in the thermocline averaged about O(10$^{-4}$) m$^{2}$ s$^{-1}$, and are approximately 10 times levels previously observed in the GS during other seasons. Banded shear structures in velocity profiles, dominated by shoaling phase coherence and clockwise rotation, indicate that downward-propagating near-inertial waves are responsible for much of this enhanced subsurface mixing.
1. Introduction

Eighteen Degree Water (EDW), the subtropical mode water of the North Atlantic, is observed in the recirculation region of the subtropical gyre (Worthington 1959, 1976; Hanawa and Talley 2001). Up to 300 m thick and weakly stratified, EDW is defined as having a temperature of 17–19°C (Kwon and Riser 2004) and salinity 36.40–36.60 psu (Worthington 1959). EDW is thought to be formed during winter convection south of the Gulf Stream (GS) and subsequently subducted to form the low-potential-vorticity layer between 200 and 500 m in the Sargasso Sea (Worthington 1959; McCartney 1982; Talley and Raymer 1982; Lozier et al. 1995; Peng et al. 2006). Formation and dissipation of EDW may affect climate, owing to the large thermal capacity of the layer (Marshall 2005), providing the rationale for studies of EDW as part of CLIVAR (Climate Variability and Predictability). Due to the severity of winter conditions in the North Atlantic, EDW formation had not been observed before CLIMODE (CLIVAR Mode Water Dynamics Experiment).

In spite of strong vertical shear, past studies report weak diapycnal mixing in the GS. Based on temperature microstructure, Oakey and Elliott (1977) reported a diapycnal diffusivity of $K_T = 3.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ in March 1974, slightly more than the $1.8 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ they observed during October 1972, suggesting seasonal variability. During autumn 1975, Gregg and Sanford (1980) observed $K_T < 5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$, noting the importance of apparent double diffusion signatures on upper and lower boundaries of large slope water intrusions into the GS. Simultaneously, Gargett and Osborn (1981) made the first direct measurements of the turbulent dissipation rate $\varepsilon$ in and north of the GS, reporting $\overline{K_\rho} \sim 5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ (Osborn 1980). Winkel et al. (2002) also report low mixing levels,
Here, we report observations in and around the GS between 7 and 27 February 2007 during a CLIMODE cruise, and discuss enhanced mixing processes and estimate their effect on EDW formation. Section 2 describes the microstructure measurements and relates them to meteorological and hydrographic backgrounds, followed in Section 3 by summaries of the primary sets of observations. Section 4 examines water modification in surface mixed layers, and Section 5 compares mixing rates in the thermocline with existing mixing parameterizations. Results are summarized in Section 6.

2. Measurements and Meteorology

a. Cruise Overview

Working from the R/V *Knorr*, Advanced Microstructure Profilers (AMPs) were dropped at CTD stations across and around the GS (Fig. 1). Velocity and shear were measured with 75- and 150-kHz broadband Acoustic Doppler Current Profilers (ADCPs) sampling 40–700 m with 16-m vertical resolution and 10–200 m with 4-m resolution, respectively. After two test drops, a south-to-north transect (S1) was made across the GS 11–13 February. Late on 13 February the Air-Sea Interaction Spar (ASIS; Graber et al. 2000) was launched in the southern part of the GS, east of S1 (Fig. 1), and became the focus of much subsequent sampling. Attempts to make a second transect farther east (S2)

\[ \overline{K_{\rho}} \approx 2 \times 4 \times 10^{-5} \, \text{m}^2 \, \text{s}^{-1} \], from fine- and microscale shear measurements in the Florida Current. These observations provide background for the CLIMODE measurements presented here, but because they were not made during winter storms, they are not directly relevant to the analysis of EDW formation.
were interrupted repeatedly by heavy weather and terminated when large seas, driven by winds gusting to 30 m s$^{-1}$, tore the AMP winch off the deck, leaving only cable ends.

The Deep AMP, operating to 1000 m, was the principal profiler, but when conditions on deck were too rough the lighter AMPs limited to 300 m were used. Both versions carry two airfoils to measure centimeter-scale velocity fluctuations, Neil Brown conductivity cells, and FP07 thermistors. Nearby CTD casts were used to correct conductivity drift, but this was not possible in some mixed layers, and those data are not used in the analysis.

b. Meteorological Forcing, Mixed and Mixing Layers, and Monin–Obukhov Length

Strong winds and surface cooling occurred throughout the cruise, producing peak wind stresses $\tau = 1.4$ Pa and surface heat fluxes up to 1000 W m$^{-2}$ (Fig. 2). The latter was calculated from short and long wavelength radiation, and latent heat and sensible heat fluxes using the TOGA CORE 3.0 bulk algorithm (Fairall et al. 2003). The surface buoyancy flux $J^b_h$ was obtained from the heat and salinity fluxes; sea surface salinity was measured by the Knorr’s seawater intake, adjusted for a constant bias of about 0.6 psu relative to the CTD.

Depth of the surface mixed layer $h$ was initially estimated by finding where potential temperature was 0.05°C cooler than the surface. Equivalent to a density increase of 0.01 kg m$^{-3}$, temperature changes were used because a drift of the AMP’s Neil Brown conductivity cell rendered small density changes unreliable on some profiles. Because we were limited to one profile per station, some depths were adjusted by eye to compensate
for algorithm outputs that appeared artificially shallow or deep owing to transient structures, likely produced by lateral advection near the GS.

Following Brainerd and Gregg (1995), we consider the mixed layer as distinct from the mixing layer $D$ encompassing continuously elevated turbulence extending downward from the surface, here defined to end where $\varepsilon$ decreased to 2% of the average up to the surface. When the mixing layer extended below the mixed layer, the entrainment zone was defined as the difference in their depths. The entrainment heat flux is estimated as

$$J_{q_{\text{ent}}} = -\rho c_p K_\rho \frac{\partial \theta}{\partial z},$$

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where $K_\rho = 0.2 \frac{\varepsilon}{N^2}$ is the usual microstructure-based estimate of diapycnal diffusivity (Osborn 1980), and the temperature gradient and $\varepsilon$ are averages across the entrainment zone. Based on the assumption that dissipation equals local production of turbulence, this is valid provided that the turbulence is generated in shear-driven overturns in the entrainment zone. In some cases small density steps within the mixed layer ended the mixing layer and there was very little entrainment heat flux because the turbulence acted only to homogenize the layer without incorporating water from the thermocline.

Within the mixed layer, the Monin–Obukhov length,

$$L_{MO} = \frac{-u_*^3}{\kappa J_b^0},$$

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measures the depth where the production of turbulent kinetic energy by wind stress equals that by the buoyancy flux, where $u_* = \left(\frac{\tau}{\rho_w}\right)^{1/2}$ is the friction velocity, $\kappa$ is von Karman’s constant, and the negative sign is a convention. Convection dominates turbulent production when $D >> -L_{MO}$, and when $-L_{MO} < 0$ the buoyancy flux suppresses
turbulence. In convective conditions, $L_{MO}$ varied between $-2$ and $-200$ m, sometimes greatly exceeding mixed layer depths.

3. Observations

a. First Cross-Section (S1)

i. South of the Gulf Stream (AMP21184–21190)

South of the GS along S1, mixing layers formed a bowl rising to slightly less than 100 m at both ends and deepening below 300 m in the center, where temperatures were around 19°C, the upper bound of the EDW range (Figs. 3 and 4). Surface heat fluxes remained high, 272–836 W m$^{-2}$, producing Monin–Obukov lengths of $-9$ to $-32$ m, indicating convection. Figure 5 shows an example of this regime, sampled by AMP 21186, revealing the deepest mixing layer we observed. Entrainment fluxes were large, $-131$ W m$^{-2}$. The next profile to the north, AMP 21187 (Fig. 6), sampled a similar mixed layer, but the upper 60 m were capped by strongly stratified warmer water. Because the surface heat flux was positive, i.e., convection, this water must have been advected over the thicker mixed layer, isolating it from direct surface forcing. Consequently, the entrainment heat flux of $-17$ W m$^{-2}$ must have been produced by background shear. The effects of shear across the mixed layer base increased to the north, producing entrainment fluxes of $-44$ W m$^{-2}$ in profile AMP21190.

Also bowl-shaped, the thermocline consisted of weakly stratified EDW containing patches of strong shear, low gradient Richardson numbers, and elevated mixing, e.g., 450 m in AMP21185 (Figs. 3 and 4). Enhanced mixing with low gradient Richardson numbers was also observed just south of the GS (e.g., between 350 and 500 m in
AMP21190). Overall, however, dissipation rates were modest, averaging \((2.1 \pm 0.3) \times 10^{-9}\) W kg\(^{-1}\), and diapycnal diffusivities were \((4.0 \pm 0.6) \times 10^{-5}\) m\(^2\) s\(^{-1}\).

\[\text{ii. Gulf Stream (AMP21191–21193)}\]

Fifty kilometers wide, the GS velocity exceeded 1 m s\(^{-1}\) to 500 m (Fig. 3). Surface buoyancy and heat fluxes were large, the latter 486–876 W m\(^{-2}\). Wind stress remained near 0.25 Pa. Mixed layers were shallower than south of the GS, 27–75 m, owing to the strong stratification in the GS front and possibly to advection of warm water from the west (Fig. 1). The Monin–Obukhov length exceeded the depth of the shallow 27-m mixed layer on the northern side (AMP 21193), generating an entrainment flux of \(-1038\) W m\(^{-2}\), cooling the layer from below faster than it was cooled at the surface. However, the layer temperature, 16.2°C, was already well below the EDW range.

In the thermocline under the GS core (AMP21192), banded structures lying along isopycnals were observed between 300 and 500 m (Fig. 3), similar to downward-propagating near-inertial internal waves in other fronts (Kunze and Sanford 1984; Mied et al.1986; Shcherbina et al. 2003). Some gradient Richardson numbers \(R_i\) were less than 1, even when determined with measurements reduced to 16-m vertical resolution. Beginning at 350 m the phase, \(\tan(\frac{dv}{dz} / \frac{du}{dz})\), of the banded features decreased with depth (Fig. 7), indicating downward propagation for freely-propagating waves (Kunze 1986).

Below the mixed layer on the north side (AMP21193 at –20 km) of the GS, cold-core thermohaline intrusions were observed between 50 and 250 m, similar to those reported by Gregg and Sanford (1980), indicating interleaving across the front (Fig. 8).
The intrusions, however, were not accompanied by elevated dissipation. Deeper in the same profile, turbulence was intense between 450 and 500 m, averaging $1.4 \times 10^{-7}$ W kg$^{-1}$ across continuous density overturns having average Thorpe scales of 3.5 m and $K_\rho = 1.5 \times 10^{-3}$ m$^2$ s$^{-1}$. Dividing the kinetic energy density, 0.19 J kg$^{-1}$, by the dissipation rate gives a lifetime of 30 days, and reduced to 11 days if the geostrophic component of the kinetic energy is excluded.

At a roughly similar location under the Kuroshio, Rainville and Pinkel (2004) obtained a time series showing elevated shear in a region of positive vorticity adjacent to a region of negative vorticity on the inshore side. Finding upward and downward propagation, they speculated that this enhancement occurred where low-frequency internal waves propagating seaward from the shelfbreak were strongly refracted by the ‘wall’ of positive vorticity created by the Kuroshio. Based on shipboard and lowered ADCP measurements, they infer dissipation rates of $(0.5 – 1.5) \times 10^{-8}$ W kg$^{-1}$ using the internal wave scaling by Gregg (1989). We measured $3.9 \times 10^{-8}$ W kg$^{-1}$ between 300 and 500 m, and based on our shipboard ADCP would have inferred $0.8 \times 10^{-8}$ W kg$^{-1}$ (Section 5).

Above 400 m, where turbulence was moderate, $\varepsilon = (4.5 \pm 1.0) \times 10^{-9}$ W kg$^{-1}$ and $K_\rho = (3.6 \pm 0.9) \times 10^{-5}$ m$^2$ s$^{-1}$, similar to previous observations during different seasons and calmer weather. Below 400 m, $\varepsilon = (3.10 \pm 0.8) \times 10^{-8}$ W kg$^{-1}$ and $K_\rho = (3.1 \pm 0.9) \times 10^{-4}$ m$^2$ s$^{-1}$, six times the largest diffusivity measured previously in the GS. This larger diffusivity is due partly to the strong mixing patch observed in AMP21193. However, even without this patch, mixing is still high, $\varepsilon = (9.1 \pm 3.0) \times 10^{-9}$ W kg$^{-1}$ and $K_\rho = (8.4 \pm 2.9) \times 10^{-5}$ m$^2$ s$^{-1}$. 

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iii. North of the Gulf Stream (AMP21194–21196)

Mixed layers were shallow at 20–75 m, very turbulent, and dominated by stress production with Monin–Obukhov lengths equaling or greatly exceeding layer depths. As a consequence, the entrainment flux into the shallowest mixed layer was the largest positive value observed, 1174 W m$^{-2}$, warming the layer from below 10 times faster than the surface heat flux cooling. However, with surface temperatures of 10°C and less, there is insufficient heat content below these mixed layers to warm them by entrainment into the EDW range.

Numerous warm-core intrusions between the mixed layer and 250 m were evidence of northward interleaving of water from the GS front. In this depth range, dissipation rates averaged $(9.7 \pm 0.2) \times 10^{-9}$ W kg$^{-1}$ with $K_{\rho} = (5.7 \pm 1.2) \times 10^{-5}$ m$^2$ s$^{-1}$, assuming breaking internal waves produced the turbulence rather than double diffusion. In somewhat less stratified water below the intrusion zone, the average diapycnal diffusivity was $(3.5 \pm 0.6) \times 10^{-5}$ m$^2$ s$^{-1}$, one-tenth that under the GS at the same depths.

b. Near ASIS

Imbedded in the southern part of the GS in an area of negative vorticity, ASIS was also above an intra-thermocline eddy, revealed by Sea-Soar surveys (Leif Thomas, personal communication, 2008). On 17 February, seven AMP profiles (AMP21198–21104) were taken near ASIS. Because of rising winds and building seas, lightweight AMPs were used for three of the drops, but these fully penetrated the mixed layers of temperature 18.9–19.3°C and depth 54–246 m (Fig. 9). Both surface and entrainment
heat fluxes were large, 527–784 W m⁻² and –25 to –1326 W m⁻², respectively. The peak value of entrainment fluxes occurred at a cold, fresh intrusion. Due to the stronger wind stress and sampling region (south of the GS), mixing was generally elevated below the mixing layer and entrainment zone, with $K_\rho$ averaging $(2.1 \pm 0.3) \times 10^{-4}$ m² s⁻¹.

c. Second Cross-Section (S2)

Attempted for five days, and only partially completed owing to the strongest winds during the cruise, S2 sampled active mixed layers. The deepest mixed layer reached 400 m (AMP21209 in Fig. 10), but mixing layers were shallower, indicating that the deep layers were formed by stronger prior forcing. Surface heat fluxes were fairly large, 185–583 W m⁻², and entrainment fluxes spanned the wide range, –548 to 272 W m⁻². In response to higher wind stress, Monin–Obukhov lengths were –200 to 875 m. Owing to the large Monin–Obukhov lengths, mixed layers 79 m deep in AMP21207 could have been dominated by stress-driven turbulence. Below the mixed / mixing layers $K_\rho = (7.4 \pm 0.8) \times 10^{-5}$ m² s⁻¹.

4. Mixed Layer Turbulence and Eighteen Degree Water Formation

a. Mixed Layer Turbulence

To reproduce EDW formation in ocean circulation models, mixed layer evolution must be predicted by means of parameterized turbulent mixing. Here, we test the validity of such parameterizations indirectly, by comparing observed dissipation rates against empirical similarity scalings for atmospherically-driven ocean boundary layers. A companion paper (Inoue et al. 2009) describes a more direct comparison with a nonlocal
mixed layer model. Such empirical scalings and turbulence models have generally been constructed for laterally homogeneous environments, and their validity in strongly baroclinic mixed layers is not well established.

**i. Scaling with Wind Stress and Surface Buoyancy Flux**

When wind stress is the dominant source of upper ocean turbulence, \(-h/L_{MO} < 1\), the dissipation rate \(\varepsilon\) should scale as in wall-bounded layers, with \(\varepsilon = u^3/(\kappa)\).

Conversely, if surface buoyancy flux dominates turbulence production, \(-h/L_{MO} > 1\) and dissipation should scale with the surface buoyancy flux \(J_b^0\). Averaging over data below 5 m in ~60 m deep layers, Lombardo and Gregg (1989) report average empirical ratios of \(\varepsilon/J_b^0 \approx 0.58\), ranging from 0.44 to 0.65 depending on entrainment rates, and mean shear-driven scaling ratios averaging \(\varepsilon/\varepsilon_s \approx 1.76\).

Figures 11a, b plot the layer mean of scaling ratios \(\varepsilon/\varepsilon_s\) and \(\varepsilon/J_b^0\), respectively, omitting contributions from above the uppermost measurement at 16 m. The average is taken down to the lesser of either the mixed layer to depth \(h\), or (for AMP21204 and 21209) the mixing depth \(D\). Of the six stations with \(0 < -h/L_{MO} < 1\), four profiles out of six are consistent with the wind stress scaling (Fig. 11a), but with somewhat lower \(\varepsilon/\varepsilon_s\), 0.91–1.28. Over buoyancy-dominated conditions with \(1 < -h/L_{MO}\) (Fig. 11b), many layer averages are consistent with the range of scaling ratios reported by Lombardo and Gregg (1989). However, several outliers in these scaling comparisons are found, the largest of which lie north of the GS (AMP21195 and 21207) for the wind scaling and
south of the GS (AMP21182) for the buoyancy flux scaling. The high level of dissipation at AMP21195 may stem from its relatively shallow depth \((h=24\text{m})\), while the low level encountered in AMP21182 at the edge of a cold core ring (see Fig. 1) may be due to the buoyant suppression of turbulence by restratification due to lateral advection.

**ii. Combined Wind and Buoyant Scaling**

Lombardo and Gregg (1989) compare observed dissipation profiles with a depth-dependent linear combination of their wind and buoyancy scalings

\[ \varepsilon_{\text{emp}} = 1.76 \varepsilon_s + 0.58 J^0_b \]

and report a mean scaling ratio of \( \frac{\varepsilon}{\varepsilon_{\text{emp}}} \approx 0.87 \). Figure 11c compares layer-averaged dissipation measurements to this combined scaling prediction. In Fig. 12, CLIMODE dissipation profiles are normalized on the three surface forcing scalings of Fig 11 and then averaged together into nondimensional depth bins of \( \Delta(-z/h) = 1/10 \). The overall comparison finds dissipation levels \( \frac{\varepsilon}{\varepsilon_{\text{emp}}} \approx 0.56 \), lower than those reported by Lombardo and Gregg (1989). A companion paper (Inoue et al. 2009) seeks to identify the role of baroclinicity in this reduction of GS mixed layer dissipation levels.

**b. Entrainment Flux**

Large entrainment heat fluxes observed at the base of mixed layer may contribute significantly to EDW formation. Here, we try to identify the driving mechanisms of entrainment and the critical processes that must be understood to predict where and when these fluxes become large.
In standard parameterizations of mixed layer deepening, the entrainment buoyancy flux may depend on atmospheric forcing, or on the entrainment zone. Richardson number $Ri_{\text{ent}} = \Delta b/\Delta U^2$, where $\Delta b$ and $\Delta U$ are buoyancy and velocity differences across $\Delta Z_{\text{ent}}$ (Table 1). If wind stress drives mixing at the mixed layer base ($-h/L_{\text{MO}} < 1$), dissipation in the entrainment zone will scale with the average $\varepsilon_{s,\text{ent}}$ in that depth interval. If instead the dynamics of free convection control entrainment work, a divergence of kinetic energy transport within the entrainment zone is balanced primarily by a combination of dissipation and buoyancy flux. The entrainment zone dissipation rate $\varepsilon_{\text{ent}}$ and buoyancy flux $J^0_{\text{ent}}$ will therefore scale as small fractions of the dominant upper layer balance, $\varepsilon \sim J^0_{\text{ent}}/2$ (Deardorff et al. 1969). Local shear may also drive entrainment when the Richardson number approaches critical levels, $Ri_{\text{ent}} \approx 0.25$, in which case $\varepsilon_{\text{ent}}$ could be different from $\varepsilon_{s,\text{ent}}$.

Figure 13a, b shows the nondimensional ratios of entrainment dissipation on surface stress $\varepsilon_{\text{ent}}/\varepsilon_{s,\text{ent}}$ and on surface buoyancy flux $\varepsilon_{\text{ent}}/J^0_{\text{ent}}$ plotted versus $-h/L_{\text{MO}}$ and indicating $Ri_{\text{ent}}$ values. In the wind-dominated regime with $0 < -h/L_{\text{MO}} < 1$, half the points are consistent with the surface stress scaling $\varepsilon_{\text{ent}} \sim \varepsilon_{s,\text{ent}}$ (Fig. 13a), and these three cases also have near-critical $Ri_{\text{ent}}$ values. In PART 2 (Inoue et al., submitted 2009), TKE production rates for these three cases are also found to have TKE production rates similar to $\varepsilon_{\text{ent}}$ when eddy flux coefficients are predicted as functions of local gradient Richardson number. They are therefore consistent with both the surface-driven scaling and with a local shear-driven scaling. The remaining three outlying wind-driven cases
with $0 < -h/L_{MO} < 1$ (Fig. 13a) have $Ri_{ent} > 1$, and neither wind stress nor local shear mechanisms offer relevant scaling predictions. These outliers may represent restratification and destratification effects due to baroclinicity, or we may not have resolved the shear, thus underestimating $Ri_{ent}$.

In the regime strongly dominated by convection, approximately $-h/L_{MO} > 2.5$, a consistent surface flux scaling $\varepsilon^{ent} \approx 0.1 J_b^0$ prevails among cases with $Ri_{ent} \geq 1$. Most cases with $1 < -h/L_{MO} < 2.5$ are intermediate between these scalings, and most of the outliers from these overall trends in $-h/L_{MO} > 1$ may be due to shear instabilities in the entrainment zone, with $Ri_{ent} < 1$.

The buoyancy flux profile in free convection has a minimum within the entrainment zone, conventionally parameterized for bulk mixed layer models in fixed proportion $\min(J_b)/J_b^0 \approx -0.2$ to the surface flux (Stull 1976; Large et al. 1994). An average over the entrainment zone between $-J_b/J_b^0 \approx 0.2$ near $h$ and $\left|J_b/J_b^0\right| << 0.2$ at $D$ would reduce this by about half, giving a mean $-J_{b}^{ent}/J_{b}^{0} \approx 0.1$. Assuming these predictions, the $-h/L_{MO} \gg 1$ scaling for entrainment zone dissipation (Fig 13b) suggests an apparent mean efficiency of $-J_{b}^{ent}/\varepsilon^{ent} \approx 1$. In the TKE budget $\partial_z \langle \varepsilon \rangle = J_b - \partial_z \langle w'(e + p') \rangle - \varepsilon$ this implies, in cases with small growth $\partial_z \langle \varepsilon \rangle$, a nearly equal partition of losses into dissipation $-\varepsilon$ and buoyancy flux $J_b$ for the incoming kinetic energy gains from transport divergence $-\partial_z \langle w'(e + p') \rangle$. 

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This larger efficiency ratio $-J^\text{ent}_b / \epsilon^\text{ent} \approx 1$ is consistent with turbulence budgets from Large Eddy Simulations (LES) showing $\epsilon / J^0_b$ around 0.2 at depth $h$, averaging to around 0.1 between $h$ and $D$ (see case “R2” in Fig. 11b of Mironov et al. 2000). These model predictions are therefore consistent with the $\epsilon^\text{ent} \approx 0.1J^0_b$ observations for convective cases with $R^\text{ent}_i \geq 1$. However, this agreement in $\epsilon / J^0_b$ ratios implies entrainment zone efficiency levels $-J^\text{ent}_b / \epsilon^\text{ent}$ approaching unity at and below the entrainment flux minimum for free convection in a similar rotational regime. Even after accounting for entrainment zone averaging that could reduce the ratio in half, the mean efficiency in these simulations is still larger than the empirical efficiency

$$\Gamma = -J_b / \epsilon \equiv 0.2 \text{ for shear-driven mixing (Osborn 1980). The other possibility is that free convection in this baroclinic environment does not conform to the predicted minimum value of the flux ratio in the entrainment zone, } \min(J_b) / J^0_b \equiv -0.2. \text{ If we retain the assumption that } \Gamma \equiv 0.2, \text{ entrainment dynamics would appear similar to non-penetrative convection (i.e., convective adjustment), where the buoyancy change } \Delta b \text{ is negligible across the entrainment zone and } \min(J_b) / J^0_b \equiv (-\Delta b \partial / h)_b / J^0_b \equiv 0. \text{ Under local balance assumptions, the mixing efficiency may also be diagnosed empirically as } \Gamma_x = \chi_\theta N^2 / 2\epsilon \Theta^2 \text{ using observed thermal variance dissipation } \chi_\theta \text{ (Oakey 1982). This is calculated by fitting a Batchelor spectrum as a function of } \chi_\theta \text{ at the given level of } \epsilon^\text{ent} \text{ (Batchelor 1959; Grant et al. 1968) to the entrainment zone temperature gradient spectrum } k^2 \Phi^\text{entrain}_\theta, \text{ or three times the measured vertical component } k^2 \Phi^\text{entrain}_\theta = 3\Phi^\text{entrain}_\theta \text{ under assumptions of isotropy. The fit is restricted to a wavenumber} \text{.} \quad 16$
range not affected by finestructure at low $k$ or instrument noise at high $k$. This produced $\chi_\theta$ estimates ranging from $O(10^{-5})$ to $O(10^{-9})$ K$^2$ s$^{-1}$. However, inadequate resolution of temperature microstructure due to the fast falling speed of instruments could result in underestimating $\chi_\theta$ as well as $\Gamma_x$. This diagnosed mixing efficiency (Fig. 13c) shows consistently low values $\Gamma_x \approx 0.1$ for stations strongly dominated by convective forcing with $-h/L_{MO} > 10$ and $Ri^{ent} > 1$, implying that entrainment buoyancy fluxes could be very low at $O(-0.01)$ times the surface buoyancy flux in this free convective extreme (Fig. 13b). Efficiency estimates in the less convectively controlled parameter range $1 < -h/L_{MO} < 10$ are relatively high and vary widely over $0.02 < \Gamma_x < 0.6$ but average 0.18, close to levels expected for locally balanced shear production. Some of these results at $1 < -h/L_{MO} < 10$ could be consistent with the LES results of Mironov et al. (2000), for which we estimate from variance budget plots (their Fig. 12b) that $\Gamma_x$ in the entrainment zone averages 0.25–0.5 for free convection. However, $\Gamma_x = 0.1$ efficiency estimates are too low to be consistent either with the LES variance and TKE budgets, or with expectations for locally balanced shear turbulence. Microstructure measurements with a slow falling speed may be necessary to resolve these issues empirically. Spectral analysis of numerical simulations may also provide further guidance on methods for determining $\Gamma_x$ in the entrainment zone.

c. Mixed Layer Cooling

To examine the relative effects of surface and entrainment heat fluxes on mixed layer cooling in EDW formation, Fig. 14 plots the ratio $|J_{q}^{ent}|/J_{q}^{0}$, where $J_{q}^{ent}$ is first
computed using $\Gamma = 0.2$, and also using $\Gamma = 1$ for the strongly convective cases, 

$-h/L_{MO} >> 1$. Heat flux ratios $|J_q^{\text{ent}}|/J_q^0$ are high for $-h/L_{MO} < 1$, when wind-forced turbulent entrainment drives layer cooling. North of the GS, all stations show a warming of the mixed layer by entrainment because of the temperature inversion immediately below. In the GS and ASIS sections there are several stations with $|J_q^{\text{ent}}| > J_q^0$ and $\Gamma = 0.2$ at $1 < -h/L_{MO} < 10$, indicating a significant role for entrainment in layer cooling at these low $Ri^{\text{ent}}$ locations. In the strongly convective regime ($-h/L_{MO} > 10$), computing $|J_q^{\text{ent}}|/J_q^0$ with $\Gamma = 0.2$ gives ratios below 10%, except for one low $Ri_{\text{ent}}$ case.

If entrainment heat fluxes for $-h/L_{MO} > 2.5$ are recomputed with the $\Gamma = 1$ efficiency indicated above, cooling rates for $-h/L_{MO} > 10$ cases south of the GS remain dominated by surface fluxes, but ratios for several profiles with $1 < -h/L_{MO} < 10$ may become $O(1)$.

In a bulk mixed layer model, the net rate of temperature change in the mixed layer in its advected reference frame due to surface and entrainment fluxes is

$$\frac{d\Theta}{dt} = \frac{1}{\rho_c \rho} \frac{dJ_q}{dz} = -\frac{J_q^0 - J_q^h}{\rho_c \rho h^{\text{ent}}},$$

where $J_q^h$ is the flux at $h^{\text{ent}}$ (top of the entrainment zone $Z^{\text{ent}}$ in Table 1, usually $= h$).

For wind stress or shear-driven entrainment, $J_q^h$ is estimated by [1], and for convectively driven cases the same assumptions do not bear significantly on [3], because $J_q^0 >> J_q^0 >> J_q^h$. Plotted in Fig. 15, this approximate cooling rate varied from $-0.03$ to $-0.0007^\circ C \cdot hr^{-1}$. Within 40 km of the GS, where surface temperatures exceed $18^\circ C$, rates of $-0.01^\circ C \cdot hr^{-1}$ are typical, and estimates for these stations imply that 3–12 days would
be required to cool mixed layers by 1°C. Such cooling rates were observed to persist for long enough to produce EDW, and in many of these cases, \(O(1)\) heat flux ratios (Fig. 14) indicate that entrainment fluxes significantly enhanced cooling rates.

We estimate the formation rate of EDW by assuming that constant cooling rates lower surface temperature by \(\Delta \theta = -1^\circ C\) from 19°C to 18°C both within the GS core and to the south:

\[
\left(\Delta \theta \times \frac{d\theta}{dt}\right)^{-1} \times h_{\text{ent}} \times L_{\text{along}} \times L_{\text{across}} = \left(-\frac{J_{q}^{0} - J_{q}^{\text{ent}}}{\Delta \theta \times h_{\text{ent}} \times \rho \times C_{p}}\right) \times h_{\text{ent}} \times L_{\text{along}} \times L_{\text{across}}.
\]

The along-stream dimension \(L_{\text{along}} \approx 1000\text{km}\) is assumed to be the same for both the 10-km-wide GS core with \(h_{\text{ent}} \approx 50\text{m}\) and the 100-km-wide formation region extending to the south with \(h_{\text{ent}} \approx 300\text{m}\). Using cooling rates of \(d\theta/dt = -0.01^\circ C\text{ hr}^{-1}\) in the core and \(-0.001^\circ C\text{ hr}^{-1}\) to the south (Fig. 15) puts the production rate estimates at 1.4 and 8.4 Sv, respectively. While strong entrainment fluxes may contribute significantly to \(d\theta/dt\) in the GS core, much more EDW is formed to the south, where rates are controlled by surface heat fluxes. If mesoscale or submesoscale eddies carry a substantial lateral flux of heat southward from the GS core, then the role of GS core entrainment could be considerably more significant. Greater cooling rates within the core would speed EDW formation rates in the deeper layer to the south.

5. Mixing in the Thermocline

Internal waves at the Garrett–Munk (Garrett and Munk, 1975) background level produce a diapycnal diffusivity of 5.2 x 10^{-6} \text{m}^2\text{s}^{-1} (Gregg 1989). By comparison,
diffusivity levels from $\varepsilon$ were elevated near the GS in S1 and ASIS profiles, averaging $1.7 - 3.0 \times 10^{-4} \text{ m}^2 \text{s}^{-1}$, more than 30 times the GM mixing level.

As expected from the elevated diffusivities, there were many places in the thermocline where shear variance was well above the GM level (Fig. 16). Shear variance was calculated by a first differentiation with correction for the loss of variance due to the differentiation (Gregg and Sanford 1988),

$$Sh_{obs}^2 = c[(\Delta u/\Delta z)^2 + (\Delta v/\Delta z)^2], \quad \text{with} \quad c = \frac{\int_0^{1/16} \phi_{Sh}^{GM} dk}{\int_0^{1/16} \phi_{Sh}^{GM} \text{sinc}^2(k\Delta z)dk},$$

where $\Delta z$ is 16 m for the 75-kHz ADCP. Further north and south of the GS, vertically averaged observed shear variance is up to four times the GM level, and near the GS, by up to seven times. As inferred from the banded shear structure, geostrophic shear is not the main component of high shear below 200 m.

a. Internal Wave Scaling

Based on the rate of energy transfer to high wavenumbers and breaking by wave–wave interactions within the internal wave field (Henyey et al. 1986), this scaling compares variances of shear and strain with GM levels to predict how much $\varepsilon$ is elevated above a background level (Gregg 1989). Following Winkel et al. (2002), the total observed shear is used for the predictions, including the geostrophic shear in the GS. It is corrected after Polzin et al. (2002) for ADCP filtering. Following Gargett (1990), Polzin et al. (1995), and Gregg et al. (2003), we use mean observed shear spectra $Sh_{obs}^2 = \left\{ \phi_{Sh}^{obs} \right\}$ to rescale dissipation from a reference state with $\varepsilon_{ref}^0 = 6.7 \times 10^{-10} \text{ W kg}^{-1}$, the GM shear
level \( Sh_{GM}^2 = \langle \phi_{Sh}^{GM} \rangle \), and buoyancy frequency \( N_0 = 0.0052 \text{ s}^{-1} \). The predicted dissipation of internal wave energy

\[
\varepsilon_{IW} = \varepsilon_{IW}^0 \left( \frac{N}{N_0} \right)^2 \left( \frac{\phi_{Sh}^{obs}}{\langle \phi_{Sh}^{GM} \rangle} \right)^2 f(R_{\omega})
\]

includes a factor

\[
f(R_{\omega}) = \frac{1 + 1/R_{\omega}}{4/3} \left( \frac{2}{R_{\omega} - 1} \right)^{1/2},
\]

to account for the dominance of near-inertial waves near fronts as a function of the ratio \( R_{\omega} \) of shear variance to strain variance. Strain variance is approximated from deviations of the density profile from a linear fit. Where noise levels permit, \([6]\) is determined using a cutoff wave number \( k_c \), for which

\[
\int_{1/L_z}^{k_c} \phi_{Sh}^{obs} \, dk = \int_{0}^{0.1} \phi_{Sh}^{GM} \, dk,
\]

\[
\varepsilon_{IW} = \varepsilon_{IW}^0 \left( \frac{N}{N_0} \right)^2 \left( \frac{0.1}{k_c} \right)^2 f(R_{\omega}).
\]

Dissipation predictions \([6–8]\) were applied in subsurface data segments of length \( L_z = 192 \text{ m} \) with roughly uniform stratification. Where noise level constraints \( (k_c \leq 0.02 \text{ cpm}) \) did not admit solutions in the wavenumber formulation \([8]\), \( (0.1/k_c)^2 \) in \([8b]\) was approximated by \( \left( \frac{\langle \phi_{Sh}^{obs} \rangle}{\langle \phi_{Sh}^{GM} \rangle} \right)^2 \), truncating the observed spectrum at \( k_c = 0.02 \text{ cpm} \).

Although accuracy is limited by the low (16 m) resolution of the 75-kHz ADCP, dissipation predictions agree well with observed levels (Fig. 17). Several components of this calculation are detailed in Fig. 18. Nine of twenty-one predictions (Fig. 17) lie more than a factor of one or two from observations. Near-inertial internal waves with high shear, high shear–strain ratio, and clockwise motion dominate (Fig. 18). Underestimates
may be attributed to inadequate resolution of shear (e.g., AMP21198b), or possibly to a change in internal wave-breaking dynamics in some cases (AMP21193) with strongly near-inertial signatures in $\left( \phi_{Sh}^{\text{obs}} \right) / \left( \phi_{Sh}^{\text{GM}} \right)$ and $R_{ow}$. Overestimates may be due to inaccurate $R_{ow}$ values that were low, near the GM level ($R_{ow} = 3$), in conjunction with strongly banded shear features (e.g., $R_{ow} = 3$ in AMP21191). Clarifying the effects of geostrophic shear, internal wave strain, and ADCP resolution may improve this parameterization and our understanding of the internal wave field near the GS.

b. Diffusivities Parameterized by Gradient Richardson Number

Subsurface parameterizations of turbulent diffusivities based on the gradient Richardson number $Ri_g$ are applied below the mixed layer in many numerical ocean models, including those estimating EDW production as part of CLIMODE. One common version, appearing in both the K-Profile Parameterization (KPP) of Large et al. (1994) and in variants of the higher order turbulence closures (e.g., Kantha and Clayson 1994), takes the form

$$K_{\rho}^{Ri} = K_{\rho}^{0} + K_{\rho}^{IW} \quad Ri_g < 0$$
$$K_{\rho}^{Ri} = K_{\rho}^{0} \left[ 1 - \left( Ri_g / Ri_0 \right)^{\frac{3}{2}} \right] + K_{\rho}^{IW} \quad 0 < Ri_g < Ri_0 .$$
$$K_{\rho}^{Ri} = K_{\rho}^{IW} \quad Ri_0 < Ri_g$$

While both model versions use $K_{\rho}^{0} = 5 \times 10^{-3}$ m$^2$ s$^{-1}$ and $Ri_0 = 0.7$, KPP uses $K_{\rho}^{IW} = 10^{-5}$ m$^2$ s$^{-1}$ (twice the GM level of $K_{\rho}$) for the background level set by internal waves, and Kantha and Clayson (1994) use $K_{\rho}^{IW} = 5 \times 10^{-5}$ m$^2$ s$^{-1}$ (10 times GM $K_{\rho}$). Empirical contributions in the $Ri$-dependent regime $0 < Ri_g < Ri_0$ are admittedly of uncertain
remedy when \([9]\) is invoked in regional to global ocean models where grid-resolved \(R_i\) has little, if anything, to do with the observed internal wave field responsible for this mixing. On the other hand, observations from the \(R_i\)-independent regime above \(R_i^0\) may be of more general utility in setting background \(K_{\rho}^{\text{IW}}\) levels.

The comparison of \([9]\) and \(K_{\rho}\) levels from \(\varepsilon\) (Fig. 19) shows that diffusivities set by the background level \(K_{\rho}^{\text{IW}}\) when \(R_i > 0.7\) lie between 1–2 times the Kantha and Clayson (1994) level of \(5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}\), or 10 to 20 times the GM level of \(K_{\rho}\). In contrast, comparisons at \(R_i < 0.7\) show that \([9]\) overestimates \(K_{\rho}\), possibly because the parameterization is tuned to give more reasonable levels, on average, in models where \(R_i\) of the internal wave field is not resolved. The comparison also suggests that if the wave field is resolved, that a resolution-dependent adjustment in the transition point \(R_i^0\) may also be necessary. More reasonable predictions are obtained (Fig. 19) using \(K_{\rho}^{\text{IW}} = 8.3 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}\) and \(K_{\rho}^0 = 10^{-3} \text{ m}^2 \text{ s}^{-1}\), but in models that do not resolve the internal wave field measured here, adopting the higher value of \(K_{\rho}^0\) may not improve predictive skill. On the other hand, increasing \(K_{\rho}^{\text{IW}}\) tenfold is roughly consistent with the elevated shear observed in and near the GS during February 2007 and can be used in OGCMs, but it should not be applied to other places and times.

6. Summary

From 7 to 27 February 2007 microstructure measurements in and near the GS sampled production of EDW during severe wintertime conditions. Strong winds with
peak wind stresses of 1.4 Pa, and surface cooling with heat fluxes up to 1000 W m\(^{-2}\), provided a rare opportunity to study intense mixing near the baroclinic front. In this cruise, a south-to-north transect (S1) was made across the GS (GS). Then, measurements were taken following ASIS in the southern part of the GS, east of S1. A second transect farther east (S2) was attempted but was repeatedly interrupted by heavy weather. Our main results are summarized as follows.

- Surface mixing layers 300 m deep were observed south of the GS, where the lateral structure of the thermocline formed a weakly stratified bowl. On the northern side of the bowl, surface forcing produced strong turbulence in the entrainment zone. The accompanying heat flux was comparable to that at the surface north of the GS where cold water intruded beneath the mixed layer. Farther south, entrainment fluxes decreased as mixed layers deepened. Cooling of mixed layer water into the EDW temperature range was dominated by O(100) W m\(^{-2}\) surface heat fluxes with a much larger impact than the O(10) W m\(^{-2}\) entrainment heat fluxes.

- Some higher levels of mixed layer entrainment flux were found close to the GS core, but the surface area of this region was too small for it to have been a major source of EDW formation, most of which appears to form in the much larger area to the south. Our estimates of formation rate from surface and entrainment heat fluxes suggest 1.4 Sv in the GS core, and 8.4 Sv in a region to the south, not including any contributions from lateral mixing processes. This estimate represents the rate of EDW formation over just a few weeks of winter, but is
similar to the annual formation rate of 4–5 Sv estimated by Kwon and Riser (2004).

- Patches of enhanced thermocline mixing were observed in and near the GS. Average vertical diffusivity beneath the surface boundary layer was about $10^{-4}$ m$^2$ s$^{-1}$, approximately 10 times higher than previous observations around the GS. ADCP vertical shear showed horizontally coherent positive and negative bands with shoaling phase structures, indicative of downward-propagating near-inertial waves. Some of those contained low Richardson numbers and mixing patches. Shear squared was up to 10 times larger than the canonical internal wave fields (Garrett and Munk 1975). Although the contribution of enhanced thermocline mixing to EDW formation was small, notable relationships between the proximity to the GS, the internal wave field intensity, and elevated levels of dissipation, point to a potentially significant pathway for the forward cascade of large-scale and wind-driven energy to the microscale.

- The internal wave scaling of $\varepsilon$ and the parameterization of diffusivity by Richardson number were also compared with these thermocline observations. The internal wave scaling of dissipation was realistic in most of the cases near the GS. However, it failed to reproduce the elevated $\varepsilon$ and enhanced mixing just north of the GS core, possibly due to ADCP resolution or perhaps because wave–wave interaction assumptions do not hold. In the thermocline, the subsurface diffusivity model of KPP, based on gradient Richardson number, was found to under predict diffusivity at high Richardson numbers and over predict them at low values. We
proposed adjusting the background diffusivity to $K^{\rho}_{\nu} = 8.3 \times 10^{-5}$ m$^2$ s$^{-1}$ in this environment of enhanced internal wave shear.

Acknowledgments

The National Science Foundation funded this work with grant OCE-0424779. In addition, Harcourt was supported by NSF grant OCE-054948 and ONR grant N00014-08-1-0446. We are indebted to Jack Miller, Dave Winkel, Steve Bayer and Andrew Cookson for working through trying conditions on deck and to Terry Joyce, Chief Scientist, and our other shipmates on the R/V Knorr for their skill and determination to wring data from the North Atlantic, braving winds gusting to 60 kt.
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Gargett, A. E. 1990. Do we really know how to scale the turbulent kinetic energy dissipation rate $\varepsilon$ due to breaking of oceanic internal waves? J. Geophys. Res., 95, 15971-15974.


Table 1. Mixed layer properties. AMP profile number, group name, depth range $Z_{ent}$ of entrainment zone, mixed layer depth $h$, mixing layer depth $D$, Monin-Obukhov length $-L_{MO}$, mixed layer temperature $\theta$, surface heat flux $J_q^0$, surface buoyancy flux $J_b^0$, wind stress, $\tau$, entrainment shear $Sh_{ent}^2$, buoyancy frequency squared $N_{ent}^2$, Richardson number $R_{i ent}$, dissipation rate $\varepsilon_{ent}$ and heat flux $J_{Q_{ent}}$. If two mixed layers are vertically stacked, entrainment values are given for each interface and the one used in Section 4 is in bold.

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Figure 1. Sea surface temperature (SST) and velocity calculated from surface slope on 13 February 2007. Locations of all AMP profiles are superimposed. Though SST changed in the Gulf Stream, the position and intensity of the stream remained similar in subsequent satellite images during the cruise.
Figure 2. a) Wind stress, b) air (thick) and sea surface (thin) temperatures, c) heat (thick) and buoyancy (thin) fluxes, and d) Monin–Obukhov length $-L_{MO}$ and mixing layer depth (*). Inverted triangles along the top mark times of AMP drops.
Figure 3. Background observations during S1: a) Current speed ($\sqrt{u^2 + v^2}$), b) vertical along-stream shear, as 16-m first-differences, back-rotated in time to the observations at 0 km, c) shear squared, d) buoyancy frequency squared $N^2$ over 16 m, and e) gradient Richardson number. The dashed line is mixed layer depth $h$, and the dotted line is mixing layer depth, $D$. Contours of potential temperature at 1°C intervals (thick) are overlaid on all panels, supplemented by thin lines every 0.2°C between 18 and 19°C. Horizontal distance is relative to the maximum vertically-integrated speed.
Figure 4. Mixing observations during S1: a) wind stress (solid line) and surface buoyancy flux (dashed), b) 10-m scale turbulent dissipation rate, and c) diapycnal diffusivity.
Figure 5. AMP21186: Vertical profiles within boundary layer. a) Potential temperature and dissipation rate, b) $\epsilon/J_b^0$ (thick) and $\epsilon/\epsilon_s$ (thin), c) velocities, $u$ (thick) and $v$ (thin) from 75 kHz, and d) shear squared.
Figure 6. Same as Fig. 5 but for AMP21187.
Figure 7. Summary of AMP21192, through the core of the Gulf Stream, at 0 km in stream coordinates. a) Potential temperature and kinetic energy dissipation rate averaged over 10 m, b) vertical diffusivities, c) shear (thick line) and buoyancy frequency (thin line) squared, d) phase, and e) Thorpe displacement.
Figure 8. Same as Fig. 7 but for AMP21193, on the north side of the Gulf Stream, at –20 km in stream coordinates.
Figure 9. a) 2-m averaged potential temperature in ASIS profiles $\theta$ offset by 1°C and b) 10-m averaged turbulent kinetic energy dissipation rate.
Figure 10. S2 profiles in the same format as Fig. 9. Profiles are shown ordered from north (AMP21207) to south (AMP21205).
Figure 11. Mean mixed layer dissipation scaling based on a) wind stress, b) convection and c) combined surface forcing, indicating the empirical levels of Lombardo and Gregg (1989). Diamonds show stations from the north of GS, squares are those in the GS, stars are ASIS stations, and circles are south of the GS. Error bars show 95% confidence intervals from bootstrap methods (Efron and Gong 1983).
Figure 12. Mean scaled dissipation profiles based on the (a) wind-driven, (b) surface buoyancy flux-driven, and (c) combined empirical scaling of Lombardo and Gregg (1989). Bootstrap 95% confidence intervals are indicated by thin lines.
Figure 13. Dissipation rate in entrainment layer normalized by a) wind stress and b) surface buoyancy flux. c) Mixing efficiency $\Gamma = \chi_\theta N^2 / 2 \Theta^2$ under local balance assumptions. Color bar shows magnitudes of $Ri_{\text{ent}}$. Symbols are as in Fig. 11.
Figure 14. Entrainment heat flux normalized by surface heat flux. Crosses indicate $J_q^\text{entr} > 0$, and the entrainment flux warms the mixed layer. Symbols are as in Fig. 11.
Figure 15. The net rate of temperature change in the mixed layer due to surface and entrainment fluxes. Negative values are plotted positively as * to show net heating rates. Symbols are as in Fig. 11.
Figure 16. Shear squared: a) $\log_{10}[S_{\text{obs}}^2/(S_{\text{geo}}^2+S_{\text{GM}}^2)]$ with potential density contours; b) averaged shear squared with blue dots representing observed values and red squares GM, and the dashed line is geostrophic shear squared; c) average of ratios in thermocline shown in a).
Figure 17. Internal wave scaling prediction from [8]. Number is AMP drop number.
Figure 18. a) Averaged dissipation rate. Observed values are shown by filed circles with 95% confidence interval and crosses are predicted values. b) Shear to strain variance ratio. GM value is shown in dashed line. c) Ratio of observed shear variance to GM level. d) Ratio of clockwise to counter-clockwise motion. The x-axis is drop number (86 for AMP21186); a and b represent shallow and deep layers, respectively.
Figure 19. Vertical diffusivities below the mixed layer entrainment zone, as a function of Richardson number at 16-m scale from 75-kHz ADCP (circles) and at 4-m scale from 150-kHz ADCP (squares). Averages are taken over data segments of 50 samples for both scales. Error bars are 95% confidence limit. Solid line is the $\text{Ri}$-dependent subsurface diffusivity parameterization of KPP. Dashed line is with adjusted parameters $K_\rho^{\text{m}} = 8.3 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ and $K_\rho^0 = 10^{-3} \text{ m}^2 \text{ s}^{-1}$ in (9).