Turbulent mixing in the springtime central Adriatic Sea

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A small set of observations of upper ocean turbulent mixing, stratification and currents was obtained in the central Adriatic Sea in May 2003. Owing to light winds, the surface mixed layer was at most 10 m thick and usually much thinner. The water column below was mostly strongly stratified with partially restratified remnant of previous mixed layers. Mesoscale currents were weak with a significant barotropic component. Weak shear and strong stratification tended to combine to large Richardson numbers. Below a layer of enhanced mixing in the upper 10–20 m, eddy diffusivities were mostly small, ranging from $10^{-6}$ m² s⁻¹ to about $5 \times 10^{-5}$ m² s⁻¹. Much larger values occurred in a few events, however.

Keywords: microstructure, turbulence, Adriatic

1. Introduction

While it appears to be widely accepted that turbulent mixing is an important part of the ocean circulation, it still is the least understood component. And although direct observations of mixing have become increasingly common since oceanic microstructure measurements matured to a «production» tool in the early 1980s, they are still being made far less frequently than conventional hydrographic and current measurements. With this background, it appears worthwhile to report herein one of the first direct observations of ocean mixing in the central Adriatic Sea even though the set of measurements is small as a result of the loss of the core instrumentation less than two days out of port.

Previously published studies of the Adriatic mixing were limited to the determination of eddy diffusivities by fitting some simple models to data. Schmidt (1917) and Gačić (1972) related a one-dimensional heat diffusion equation to vertical profiles of temperature repeatedly taken at a station. Saint-Guily (1965) and Malačić (1991) minimized the difference between an analytical solution of the heat diffusion equation and observed annual cycle
of temperature. On the other hand, Zore-Armanda (1963) combined a theoretical result previously obtained by Jacobsen (1927) with TS diagrams successively recorded at a station. Finally, Supić et al. (1997) and Grbec and Morović (1997) assumed continuity of heat flux across the sea surface and utilized estimates of the air-sea heat flux as well as vertical temperature gradients measured at the sea surface. For the springtime central Adriatic this resulted in widely differing eddy diffusivities of heat: according to Gačić (1972) they range between 8 and $58 \times 10^{-4}$ m$^2$ s$^{-1}$ along the vertical, following Grbec and Morović (1997) they are close to $2 \times 10^{-4}$ m$^2$ s$^{-1}$ at the sea surface.

The present paper simply describes the observations, which are introduced in Section 2 along with our methods for their analysis. Brief characterizations of the oceanic conditions in the central Adriatic during the time of the measurements near the end of May, 2003, follow. These address surface forcing, ocean currents and stratification. Thereafter the oceanic mixing is discussed in Section 4. For better readability of the paper detail of the data reduction and an assessment of uncertainties are relegated to the Appendix.

2. Observations and Flow Variables

The observations analyzed herein were made as part of the international DOLCEVITA project – Dynamics Of Localized Currents and Eddy Variability In The Adriatic. A first cruise within the project had taken place in the northern Adriatic in January/February 2003 under wintertime conditions and repeated bora wind events (Lee et al., 2005). Turbulence observations from this cruise will be analyzed elsewhere. A second set of cruises followed in late spring under very different environmental conditions. Extensive observations of stratification, currents and optical and biological parameters were made from the R/V Knorr. In parallel with the Knorr cruise, we observed turbulent mixing from the R/V G. Dallaporta. Her cruise track is depicted in Fig. 1, and stations are listed in Table 1.

<table>
<thead>
<tr>
<th>Station</th>
<th>Time (UTC)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>CTD drops</th>
<th>SWAMP drops</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>27-May-03, 0829–1154</td>
<td>43°32.1’N</td>
<td>14°02.6’E</td>
<td>1</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>27-May-03, 1708–1727</td>
<td>43°04.9’N</td>
<td>15°04.7’E</td>
<td>0</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>27-May-03, 2204–2230</td>
<td>43°15.4’N</td>
<td>14°14.6’E</td>
<td>0</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>28-May-03, 0224–0306</td>
<td>43°25.0’N</td>
<td>15°26.6’E</td>
<td>177</td>
<td>0</td>
<td>10</td>
</tr>
<tr>
<td>5</td>
<td>28-May-03, 0753–0830</td>
<td>43°43.8’N</td>
<td>15°08.1’E</td>
<td>96</td>
<td>8</td>
<td>0</td>
</tr>
<tr>
<td>6</td>
<td>28-May-03, 1154–1224</td>
<td>43°34.3’N</td>
<td>14°57.8’E</td>
<td>101</td>
<td>0</td>
<td>7</td>
</tr>
</tbody>
</table>
The stratification was observed on station with a regular SeaBird 911+ CTD and with the CTD component of the Shallow Water Microstructure Profiler (SWAMP; Peters, 1997), which also employed SeaBird sensors. Ocean currents were observed on station and while steaming in between stations at 6 kn with a RD Instruments 300-kHz Workhorse acoustic Doppler current profiler (ADCP) towed over the stern of the *Dallaporta*. While drifting on station we took multiple SWAMP or CTD »drops«.

Stratification and current data are evaluated in terms of temperature \( T \), potential temperature \( \theta \), salinity \( S \), density \( \rho \), potential density \( \sigma \), buoyancy frequency \( N \), east \( u \) and north \( v \) velocity components or speed \( V \) and direction \( \Phi \), shear \( V_z = \sqrt{(\partial u/\partial z)^2 + (\partial v/\partial z)^2} \), and gradient Richardson number \( \text{Ri} = N^2 V_z^{-2} \).

The microstructure section of SWAMP employed two shear probes, a fast FP07 thermistor, and a SeaBird dual needle micro-conductivity probe, all sampled at 256 Hz. As detailed in the Appendix, these measurements are analyzed in terms of the viscous dissipation rate \( \varepsilon \) and the thermal dissipation rate \( \chi \) (Gregg, 1987). Following the Osborn-Cox method (Osborn and Cox, 1972), an eddy diffusivity of heat can be estimated from \( \chi \) and the vertical temperature gradient:

\[
K_h = \frac{1}{2} \chi \left( \frac{\partial T}{\partial z} \right)^{-2}.
\] (1)

Winters and D’Asaro (1996) show that the Osborn-Cox method is rigorously valid in a volume-average sense. In the following we use \( K_h \) as the primary measure of the »intensity« of turbulent mixing. Similarly, following Osborn (1980), an eddy diffusivity of mass can be estimated from \( \varepsilon \) and \( N^2 \) assuming a constant flux Richardson number,

\[
K_\rho = 0.2 \varepsilon N^{-2}.
\] (2)

The value of 0.2 for the mixing efficiency is customary but subject to uncertainty and variability. This mixing efficiency and the Osborn method in general are further discussed, e.g., by Gregg (1987), Yamazaki and Osborn (1993), Moum (1996), Smyth et al. (2001), and Baumert and Peters (2000). Peters et al. (1988) show that \( K_\rho \) and \( K_h \) track each other closely over a wide range from weak to strong mixing.

As an additional turbulence characteristic we present the buoyancy Reynolds number, \( Re_b = \varepsilon / (\nu N^2) \), where \( \nu \) is the kinematic viscosity. \( Re_b \) has been used as an indicator of the »activity« of turbulence. Following Gargett et al. (1984), Gregg (1987), and Rohr et al. (1988), active mixing sustaining a vertical buoyancy flux requires \( Re_b \geq 20 \), turbulence becomes isotropic for \( Re_b \geq 200 \) and exhibits fully universal character at \( Re_b \geq 4 \times 10^4 \). We note that \( Re_b \) is nei-
ther an ordinary flow Reynolds number nor a 1:1 substitute for a turbulent Reynolds number (Gregg, 1987; Peters et al., 1995).

Auxiliary data used herein include meteorological time series routinely collected at Šibenik and moored current time series taken at Stations A and B indicated in Fig. 1. The latter measurements were performed with bottom-mounted 300-kHz ADCPs within the framework of the East Adriatic Coastal Experiment (EACE; Orlić, 2003).

3. Surface Forcing, Currents and Stratification

a. Surface Forcing

Hourly wind vectors measured at Šibenik before and during the Dallaporta cruise are shown in Fig. 2a. There were two wind episodes during the interval considered, a stronger one on 21–23 May and a weaker on 28–29 May. Both were characterized by a considerable along-basin wind variability, and, in particular, operational mesoscale meteorological model run by the Hydrometeorological Institute of the Republic of Croatia revealed that during the second episode winds were stronger in the vicinity of Šibenik and EACE stations, weaker in between – close to our stations. However, no wind data were available for our measurement area. The first of the wind episodes was related to a cyclone which swept over the Adriatic on 21 May, the second
to the along-Adriatic air-pressure gradient which was established by 28 May. Between the two episodes the winds were weak, dominated by coastal breezes which developed in an almost uniform air-pressure field. Some precipitation was recorded at Šibenik on 20–21 May and again on 27–28 May, whereas air temperatures varied between 15 and 30 °C – with a considerable diurnal signal being visible in the data.

Figure 2. Hourly wind data taken at Šibenik (a) and hourly current shear ($V_z$ as defined in the text) recorded at Station A with a 2-m vertical resolution (b). Fig. 1 shows the locations of these measurements.
b. Stratification, Shear and Richardson Numbers

Corresponding to a surface forcing characterized by weak winds over our measurement area, the upper 70–100 m of the Adriatic were mostly strongly stratified as shown in station summary profile plots of Figs. 3–8. These depict the mean from all drops taken at any one station. The deepest observed surface mixed layers of about 10 m occurred in Station 3 (Fig. 5), which was taken at night, and hence presumably under conditions of oceanic heat loss and mixed layer convection. Otherwise, the surface mixed layer was at most a few meters deep. However, remnants of deeper previous mixed layers, now partially restratified, occurred in Stations 1–3 (Figs. 3a–5a), the deepest with 30 m in Station 2. The strongest stratification observed with $N > 20$ cph (cycles per hour) occurred at shallow depths at the bottom of active or previous mixed layers. Otherwise typical values of $N$ decreased from near 10 cph to 4 cph with depth over the top 70 m (Figs. 3b–8b). $TS$ relationships tended to be complex owing to a highly variable $S(z)$, $z$ being depth and numerically close to pressure ($p$) in dbar. Inversions in $T(z)$ were less prominent than those of $S(z)$ but occurred frequently nevertheless.

Our test station, Station 1 (Fig. 3 and Table 1), shows a highly mixed layer below 58 m depth. Our operations not having been established fully, we do not know the exact water depth at Station 1, and we also lack high-resolu-

![Figure 3. Mean Station 1 profiles: stratification (a) $\theta$, $S$, $\sigma_\theta$ and (b) buoyancy frequency ($N$, shaded) and temperature gradient; (c) thermal dissipation rate ($\chi$, with bootstrap confidence limits shaded) and uncorrected $\chi$ (dash-dotted, as explained in the text); (d) eddy diffusivity of heat ($K_h$, with bootstrap confidence bounds shaded); (e) number of sample: drops (solid) and number of $\chi$ data per 5-m bin (dash-dotted).]
tion bathymetric data from the area. However, the descending SWAMP had not hit bottom at 77 m, its maximum depth during the drop. A depth greater than 77 m contrasts with the coarse bottom topography depicted in Fig. 1, which would indicate a water depth near 65 m. This suggests that the deep mixed layer of Station 1 exists in a local deep hole in the sea floor filled with a remnant of North Adriatic Dense Water formed during the previous winter (Vilibić et al., 2004).

Figure 4. Mean Station 2 profiles: as Fig. 3 except (c) viscous dissipation rate (ε with bootstrap confidence bounds shaded) and buoyancy Reynolds number (Re_b, dash-dotted, light); (d) χ as in Fig. 3c; (e) K_h as in Fig. 3d with eddy diffusivity of mass added (K_m, dash-dotted, pink/light); (f) as Fig. 3e with number of samples of ε added (dashed).

Figure 5. Mean Station 3 profiles: see Fig. 4.
Stratification and shear combined indicate a potential for turbulent mixing. All individual estimates of $N$ from the Dallaporta CTD and SWAMP have been replotted in Fig. 9, which shows great scatter in the upper 25 m and a decreasing trend with increasing depth. Because of the spatial separation the values of $N$ can only in a very loose sense be compared with hourly shear values from the A and B EACE moorings shown in Fig. 10. Fig. 10 covers the time interval extending from 8 h UTC on 27 May to 12 h UTC on 28 May, i.e. simultaneously with our shipborne measurements. At Station A there was a maximum of shear close to the surface and an increase at depths
greater than 50 m, at Station B maximum occurred between 20 and 30 m. Obviously, the two stations were subjected to different current regimes, and that experienced by the Dallaporta was probably yet different. Nevertheless, we note that, while the Dallaporta N dropped below 5 cph at depths exceeding 60 m, some higher shear data occurred at mooring A.

A closer look at the time series of shear at Station A (Fig. 2) is revealing. Larger values were observed only close to the sea surface during the wind episodes, and at depths of 10–20 m between 24 and 27 May when the wind influence was unimportant. It is interesting to notice that the position of the shear maximum was oscillating during calm weather conditions – obviously due to the presence of internal waves. Their period is, however, difficult to determine from available data. Similar results were obtained for Station B (not shown), except that the shear was generally smaller and its maximum during the calm weather was located deeper and was oscillating with greater amplitudes.

Co-located shipborne measurements of shear and stratification indicate that gradient Richardson numbers generally stayed well above the linear stability threshold of 1/4 (Figs. 6b–8b). However, in the only two station profiles available there is a tendency for Ri(z) to decrease with depth, especially in Station 6. In interpreting these Richardson numbers it has to be kept in mind that the vertical resolution of Ri(z) is low owing to a bin size of 4 m in

Figure 7. Mean Station 5 profiles, stratification from the SeaBird CTD: as Fig. 6a-d.
the ADCP velocity measurements, the derivation of shear through differencing, and the averaging of a large number of ADCP pings. Hence the observed $Ri > 1/4$ does not exclude actual flow instability on unresolved small vertical scales, e.g. in relationship to smallscale internal waves.

c. Mesoscale Currents

The ocean currents observed from the Dallaporta in the depth range of 12–28 m are depicted in Fig. 11, which shows mesoscale variability of rather weak currents not exceeding 0.15 m s$^{-1}$. The velocity profiles of Stations 4–6 (Figs. 6c–8c) exhibit significant barotropic flow as shown by the limited depth variability of the current direction. Hence Fig. 11 is representative of

Figure 8. Mean Station 6 profiles: see Fig. 6.
Figure 9. Individual stratification data computed over 2-m intervals from all Dallaporta CTD and SWAMP drops.

Figure 10. Hourly current shear for EACE moorings A and B and the time interval between 27-May-03 0800 UTC and 28-May-03 1200 UTC. Shear is defined in Section 2 and converted to cycles per hour for comparison with $N$. 
the depth-averaged flow. The dominance of the barotropic over the baroclinic flow component is consistent with the finding of relatively weak shear as discussed above.

4. Mixing

The turbulent mixing characterized by \( \varepsilon, \chi, K_h \) and \( K_p \) varied greatly with depth and at any one depth in our small data set as shown in Figs. 3–8. When all estimates of \( K_h \) are plotted together (Fig. 12), a simple pattern emerges. Disregarding depths shallower than 20 m and one large \( K_h \) at 30 m to be discussed below, values of \( K_h \) were all below \( 10^{-4} \) m\(^2\) s\(^{-1}\), many below \( 10^{-5} \) m\(^2\) s\(^{-1}\), and some as small as \( 10^{-6} \) m\(^2\) s\(^{-1}\). Relatively large eddy diffusivities partly exceeding \( 10^{-4} \) m\(^2\) s\(^{-1}\) occurred at shallow depths of 10–25 m; \( K_h \) stayed below \( 3 \times 10^{-5} \) m\(^2\) s\(^{-1}\) at 30–40 m, and there was a general trend toward an increase of \( K_h \) with depth further below. This pattern of the depth variability of \( K_h \) corresponds with the elevated shear in the upper 20 m at EACE mooring A (Fig. 10A), and the small \( N \) below about 60 m in Fig. 9.

More specifically, the smallest eddy diffusivities appeared in Stations 1 (Fig. 3) and 6 (Fig. 8). All stations had comparatively large \( K_h \) at shallow depths, and all stations except 2 and 3 exhibit an increase in \( K_h \) with depth toward 60 m. The increase with depth of \( \chi \) and \( K_h \) below 40 m in Station 6 (Fig. 8) is easy to interpret. This enhancement of mixing parallels an in-
crease with depth of shear and a corresponding decrease in $Ri$, which dropped below 1 below 50 m depth. This is the only case where the resolved shear and $Ri$ appear to be related to the variations in mixing. In contrast, the strong increase in $\varepsilon$, $\chi$, $K_h$, and $K_p$ from 30 m depth to 10 m depth in Station 4 (Fig. 6) was paralleled by virtually constant shear and $Ri$.

The most eye-catching feature is the maximum of all turbulence variables near 30 m depth in Station 3 (Fig. 5) with $\varepsilon \sim 10^{-6}$ W kg$^{-1}$ and $\chi \sim 10^{-5}$ K$^2$ s$^{-1}$. While this feature appears smeared out in the vertical in the station averages, individual drops show it to be a sharply defined, 4 m thick internal mixed layer (Fig. 13). The large measured dissipation rate and the pronounced thermal activity demonstrate that this was an actively mixing layer. As the stratification inside this layer was very close to neutral, the eddy diffusivity is undefined inside the mixed layer, and the corresponding data points at 30 m depth of Station 3 in Figs. 5e and 12 should be ignored.

The internal mixing layer of Station 3 is reminiscent of similar features found elsewhere in the oceans, some being attributed to shear-induced, persistent mixing related to internal near-inertial waves (Gregg et al., 1986; Hebert and Moum, 1994). As the ADCP had not yet been set up during Station 3, we are unable to examine the velocity structure associated with the internal mixed layer and to probe for the presence of near-inertial waves. However, the mixing time scale associated with the event is surprisingly short, of the order of a buoyancy period. Following Gregg (1987), the change

![Figure 12. Eddy diffusivity of heat from all stations. The data point marked by «?» corresponds to the mixed layer depicted in Fig. 13, where the Osborn-Cox method is not applicable.](image)
in potential energy density corresponding to creating a mixed layer of thickness \( h \) in a fluid of stratification \( N^2 \) is \( \Delta E_p = N^2 h^2 / 12 \) [J kg\(^{-1}\)]. Fig. 5 indicates \( N = 6 \) cph and \( h = 4 \) m. Equating the buoyancy flux that created the change in potential energy with the observed \( \varepsilon = 7 \times 10^{-7} \) W kg\(^{-1}\) times a mixing efficiency of 0.2 leads to a mixing time scale of \( \Delta E_p / (0.2 \varepsilon) \approx 10^3 \) s, or 17 minutes compared to a buoyancy period of 10 min.

Figs. 4c, 5c and 6e show estimates of the buoyancy Reynolds number. Most of the \( Re_b \) values are below 100 with few data reaching or exceeding 1000. This indicates moderate »activity« for the most part.

The graphs of \( \chi, \varepsilon \) and \( K_h \) in Figs. 3–8 show statistical confidence limits computed by bootstrapping (Efron and Gong, 1983). These outline the effect on the mean of short-term variability within the observations. They are not related to longer-term averages.

5. Conclusions

In late May 2003 winds over the central eastern Adriatic were light, mixed layers were shallow, stratification was mostly strong, shear weak, and Richardson numbers were large. Below some near-surface enhancement and above weaker enhancement toward 60 m depth, turbulent mixing in a »mid-depth« range of 30–40 m was mostly weak, characterized by eddy diffusivities ranging from \( 10^{-6} \) m\(^2\) s\(^{-1}\) to \( 5 \times 10^{-5} \) m\(^2\) s\(^{-1}\). Much more intense mixing occurred in rare events, with eddy diffusivities being close to the literature values for the area.
The event-related maxima imply that a much larger set of observations than presented here is needed to characterize climatological averages of mixing rates and their variability. The finding also implies that mechanisms creating events of intense mixing deserve attention given that mean mixing rates may be largely determined by relatively few energetic events. Gregg (1987), for example, discusses the approximately lognormal statistics of $c_{101}$ and $c_{99}$ in the ocean and, in mixing events, distinguishes between highly transient «wisps» and longer-lasting «billows», the latter often being related to the instability of persistent shear set up by near-inertial waves. We further note that current numerical circulation models do not resolve processes that create intense mixing events, such as near-inertial internal waves. In addition, turbulence closure schemes typically incorporated in numerical models are ignorant of the internal wave field. If models of the Adriatic circulation need to employ realistic levels of mixing, a renewed interest in observing oceanic mixing and its generating processes may well be required. We would be happy if this limited study could stimulate such interest.

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References


Appendix: Detail of Data Reduction and Uncertainties

A Seabird 911+ CTD was used on Stations 1 and 5 in a quasi-free fall mode with buoyancy attached to the CTD frame. The sensors of this »regular« CTD as well as the SeaBird sensors of SWAMP were calibrated by the manufacturer. However, the conductivity sensor of SWAMP had drifted while the instrument was in storage, and thus we added a constant 0.3 psu to the SWAMP CTD salinity. After this, the 911+ CTD and SWAMP TS relationships of Station 1 matched each other. In consequence of the sensor drift the SWAMP CTD salinity may carry a systematic bias of about \(\pm 0.03\) psu.

The SWAMP thermistor and micro-conductivity sensors were calibrated linearly against their CTD counterparts. A single calibration holds for the thermistor for all drops with an accuracy of \(\pm 0.03\) °C and an RMS electronic noise of about 0.003 K (see Fig. 13a). In contrast, the micro-conductivity had to be individually calibrated for each drop, but even so the micro-C drifted considerably during some drop segments. Unfortunately, the pump of the SWAMP SeaBird TC duct was not turned on during Station 3, and thus stratification information had to be taken from the micro-\(T\) and micro-\(C\) sensors. The corresponding salinity, lacking proper calibration for each drop, is rather uncertain and thus not shown in Fig. 5a. Fortunately, depth-variations of \(S\) were quite small such that \(\sigma_0(z)\) and \(N(z)\) of Station 3 are still acceptably accurate. For this study the accuracy of vertical stratification gradients is important while small biases in the stratification variables are not crucial.
The evaluation of the microscale shear data in terms of $\varepsilon$ follows Peters (1997) and a long line of sources cited therein. It may suffice to mention here that these data are highly sensitive to mechanical noise and that noise and signal are separated in the spectral domain. For this study, additional routines were written that identify segments of elevated signal in the auxiliary accelerations sensor of SWAMP, which correspond to elevated mechanical noise, and flag the corresponding $\varepsilon$ data. The data being analyzed herein were gathered while the operations of our small science party on the Dal-laporta were still in the mode of setting up, shaking down, and gaining experience. It has to be understood on this background that $\varepsilon$ data from Stations 1 and 6 suffer from mechanical noise such that they are not shown in Figs. 3 and 8. With this, this paper relies principally on $\chi$ rate in quantifying turbulent mixing.

Thermal dissipation rates are relatively insensitive to mechanical noise as they respond only to sensor displacement rather than to acceleration. For the determination of $\chi$, noise was separated from signal in the spectral domain, a noise floor invariant in time being established by studying many temperature gradient spectra. In these, the noise floor follows from white noise modified by the various filters employed in the analog electronic section of SWAMP, principally an analog differentiator and an anti-aliasing filter. Temperature gradient signals were used to a maximum of 60 Hz, and the spectrum was corrected for the temporal response of the thermistor following Gregg (1999) and Nash et al. (1999), with further reference to Lueck et al. (1977), Gregg and Meagher (1980), and Fleury and Lueck (1994). The correction assumes a double pole response $H(f) = [1+(f/f_c)^2]^2$, where $f$ is the frequency and the nominal cut-off frequency $f_c$ marks the 6 db point of the response function. We assume $f_c = 32w^{0.32}$ with fallrate $w$ in m s$^{-1}$. With a fallrate of 0.5 m s$^{-1}$, the 6 db point of the temporal response is at 26 Hz, and thus spectral corrections are mostly large. The corresponding time constant is 6 ms. As the instrument and its sensors were lost at sea, it was not possible to experimentally determine $\tau$ for the individual thermistor used after the cruise, and thus the principal source of uncertainty in $\chi$ is rooted in the unknown temporal thermistor response.

In order to further explore this effect we conducted a sensitivity study. Noting that the fallrate of SWAMP of just under 0.5 m s$^{-1}$ varied little, we determined $\chi$ and $K_h$ with $f_c = 21.3w^{0.32}$, which, following experience with FP07 thermistors, characterizes the slow end of plausible thermistor responses. The corresponding $\chi$ are larger than their counterparts depicted in Figs. 3–8 by factors that increase with $\chi$ itself, typically being smaller than 50% but rising to factors of 2–3 at large $\chi$.

At small $\chi$ temperature gradient spectra tended to drop into the noise without resolving the diffusive cut-off. We corrected $\chi$ for unresolved temperature variance on the basis of the Batchelor (1959) spectrum. The magnitude of this correction was generally small and significant only in some segments.
of small $\chi$, e.g. near 40 m depth in Fig. 8d. Figs. 3–8 show both corrected and uncorrected $\chi$. In cases of weak mixing such as this in Fig. 8d, the actual spectra tended not to follow the Batchelor (1959) shape, a finding not unanticipated (Gargett, 1985). The uncorrected $\chi$ provides a lower bound for the true $\chi$. Hence, in the case of Station 6, 33–42 m depth, the true $K_h$ may have been smaller than the value plotted in Fig. 8e by factors of 3–5.

Above, we state that the maxima of $K_h$ and $K_p$ at 30 m depth in Fig. 5e should be ignored because the vertical temperature and density gradients vanish inside the internal mixed layer observed at that site. This statement emphasizes limitations in the local applicability of the Osborn-Cox method, equation (1), and the Osborn method, equation (2). Respectively, the mean vertical temperature gradient and mean vertical density gradient, i.e. $N^2$, need to be well-defined. The Osborn-Cox method can fail in regions of complex $TS$ variations and weak and variable $\partial T/\partial z$. The large excess of $K_h$ over $K_p$ at 20–25 m in Station 4 (Fig. 6g) is explained by such condition.

SAŽETAK

**Turbulentno miješanje u srednjem Jadranu tijekom proljeća**

**Hartmut Peters i Mirko Orlić**

U svibnju 2003. godine prikupljen je u srednjem Jadranu nevelik niz podataka kojim je dokumentirano turbulentno miješanje u površinskom sloju te stratifikacija i strujanje. Budući da su vjetrovi bili slabi, površinski izmiješani sloj dosizao je najviše do dubine od 10 m, često i manje. Vodeni stupac ispod tog sloja bio je uglavnom jako stratificiran, uz prisutnost ostataka ranijih izmiješanih slojeva koji su dijelom ponovno stratificirani. Struje su bile malih brzina na srednjoj skali, s time da je barotropna komponenta bila značajna. Slabo smicanje i dobro izražena stratifikacija doprinosili su velikim iznosima Richardsonovog broja. Ispod sloja debljine 10–20 m u kojem je miješanje bilo pojačano, koeficijenti turbulentnog miješanja uglavnom su bili malih iznosa, između $10^{-6}$ m$^2$ s$^{-1}$ i $5 \times 10^{-6}$ m$^2$ s$^{-1}$. Međutim, u nekoliko slučajeva opažene znatno veće vrijednosti.

**Ključne riječi:** mikrostruktura, turbulencija, Jadran

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