MEASUREMENT AND ANALYSIS OF THE ENERGY-CONTAINING EDDIES OF TURBULENT FLOWS IN THE COASTAL OCEAN

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ABSTRACT

Acoustic remote sensing techniques now allow measurement of the three-dimensional velocity field associated with the large-scale eddies of turbulent geophysical flows in the coastal ocean. Such techniques, continuous in time and requiring a minimum of technical supervision, are essential for assessment of turbulent coastal regimes, because of short space and time scales of variability. Algorithms under development should provide estimates of kinetic energy $E$, length scales, and kinetic energy dissipation rate $\epsilon$ of the turbulence, as well as the shear $dU/dz$ of the mean flow. Recent addition of a towed CTD allows a direct measurement of buoyancy flux $\overline{\rho'w'}$, a major goal of ocean microscale measurements over the last two decades. Preliminary data are available to compare this direct measurement with the widely used estimate $\overline{\rho'w'} = 0.2\rho_0 g^{-1}\epsilon$, made from measurements of dissipation rate.

1. AN ACOUSTIC REMOTE SENSING TOOL FOR TURBULENCE RESEARCH

While shipborne acoustic Doppler current profilers (ADCPs) have been widely used for measuring "mean" currents in the surface layers of the ocean, use of a commercial ADCP for turbulence research required modification to both hardware and software. The hardware modification was to rotate one of the four beams of a standard Janus-configuration transducer head to vertical, leaving the other three beams at the normal (30°) slant angle from vertical. When mounted on a ship (Fig. 1), this beam (B3) is closely adjusted to vertical (±0.5°), allowing a direct and unequivocal measurement of vertical velocity $w$. A combination of B4 and B3 provides an estimate of across-ship velocity component $u$, while a combination of B2 and B3 (or of B1 and B3) provides the along-ship component $v$. These horizontal velocity components can be affected by the slant-beam configuration, so this account will mostly use the straightforward measurement of $w$. Direct shipborne measurement of $w$ is possible with incoherent Doppler systems because coastal turbulence is vigorous, and because the inner coastal waters of British Columbia, in which these data were taken, provide low levels of platform motion contamination. If a stable platform can be provided, however, recent development of more accurate coded-pulse Doppler systems suggests that the techniques discussed here will soon be extensible to the deep ocean.
Special acquisition software was written to allow recording of raw (single-ping) beam velocities and acoustic amplitudes. After each ping, the processor associated with a single beam returns time (radial distance)-binned estimates of radial velocity, defined positive when the velocity is towards the transducer, and a measure of the strength of the return signal. In acquisition mode, both fields are recorded for all four beams, while up to four fields can be selected for colour-coding and real-time display. At present, we use amplitude signal only from the vertical beam, in order to locate the bottom (or lack of it) in the velocity records; subsequent processing uses only the water column velocities.

Single-ping velocity data are noisy. Figure 2a is a (poor) rendition of raw data from a turbulent tidal front. [An apology: Grey-scale rendering of signed quantities such as velocity is difficult, but must be attempted when colour graphics are not available. For presentation in this paper, I have chosen to bin the data very coarsely, effectively grey-scale 'contouring' the fields. With such coarse-binning, it is possible to use a symmetric grey-scale that differs only in the textures assigned to the bins nearest zero (center): thus in Figure 2, a maximum (black) that occurs as a progression through light grey (small circles) is a maximum downwards (upwards) w. While this presentation works reasonably well with smooth fields, it does a very poor job of the original noisy raw data in Figure 2a.] The standard technique for reducing the noise level of Doppler velocity estimates is to average values from consecutive pings: Figure 2b illustrates this technique, using an
Figure 2. Grey-scale coded representation of the (signed) field of $w$ as measured by the vertical beam: (a) Raw (single-pong) data, noise standard deviation $\sigma \approx 10$ cm/s. (b) Standard ADCP processing required to produce $\sigma \approx 2$ cm/s (boxcar average over 25 pings) fails to resolve the spatial structure of the turbulent flow in this tidal front. (c) Filtering with a sequential running mean filter yields $\sigma \approx 2$ cm/s with spatial resolution of about 20 m.

average of 25 pings to reduce noise standard deviation from 10 cm/s to 2 cm/s. With post-processing, this brute strength technique, which severely degrades much of the spatial structure that is present, is easily replaced by a filter that produces the same 2 cm/s standard deviation, but retains spatial structure down to horizontal wavelengths of order 20 m (Fig. 2c).
2. WHY DO WE NEED THE VERTICAL BEAM?

While significant vertical velocities do not guarantee that a flow is turbulent, flows are not turbulent without significant vertical velocities. In survey mode, we may thus look for large vertical velocity as a necessary condition for turbulence. Having found this condition, such flows may be subject to more rigorous scrutiny with regard to characteristics—for example relative "eddy" and internal wave time scales, vertical buoyancy flux, phase between \( w \) and fluctuation density—which we associate with turbulence. Thus accurate measurement of the vertical velocity field is essential to turbulence measurement.

With a standard ADCP, velocity components are calculated under the assumption that the velocity field is uniform over the spread of slant beam pairs (Fig. 3a). If this is the case, the horizontal component \( v \) in the plane of B1 and B2 makes contributions of opposite sign to the beam velocities V1 and V2 in bin b, hence slant beam vertical velocity \( w_s = (V1+V2)/2 \cos 30^\circ \). This slant-beam vertical velocity is shown in Figure 3c, below the field of \( w \) measured directly by the vertical beam (Fig. 3b) for a section of data from a tidal front. The obvious differences between \( w \) and \( w_s \) are caused by the fact that the turbulent field has spatial scales that are comparable to the slant beam spread.

Scatter plots of \( w_s \) vs \( w \) (Fig. 4) show that while \( w_s \approx w \) at shallow depths (a), the correlation decreases with increasing depth (b). By the deepest bins (c), \( w_s \) is essentially uncorrelated with \( w \), although both remain significantly above the noise level, shown in (d). This must be expected to be a normal state of affairs in coastal waters, where the water depth \( H \) sets a maximum outer scale for turbulent eddies (the actual outer scale may be even smaller, because of conditions of shear or stratification). With the 30° angle of the standard slant beam pairs, slant beam separation at depth \( H \) is \( H \), i.e., the scale at or below which we expect turbulent energy to reside. Accurate measurement of the vertical velocity field in coastal areas thus clearly requires the special vertical beam that is part of the DOppler Turbulence system (DOT).
Figure 3. (a) Accurate determination of $w$ from two slant beams (B1 and B2) requires that the velocity field be uniform over the (increasing with depth) horizontal spread between the beams. Fields of (b) $w$ from B3 and (c) $ws$ from B1 and B2 differ considerably in this tidal front, suggesting that this requirement is not met.
Figure 4. Scatter plots of $w_s$, vertical velocity determined from the paired slant beams B1 and B2, versus the "true" $w$ measured from B3, for various depths (a) 23 m, (b) 112 m, and (c) 201 m; (d) is noise level, taken at slack tide in a sheltered location. Near the transducer, the two variables are correlated, but as depth (slant beam separation) increases, $w_s$ and $w$ become increasingly uncorrelated.

3. AN ALBUM OF COASTAL MIXING

With the shipborne, semi-automated system described above, it is possible to survey coastal waters for locations and processes that cause significant turbulence. Our experience is that most intense turbulence is associated in some way with flow geometry such as submarine sills, horizontal channel constrictions, or sharp changes in channel direction. Coastal turbulence varies rapidly in time, since it is driven predominantly by the tides and is clearly modulated on the neap/spring cycle.

Figure 5 is a sampler of the kind of mixing regimes found in B.C. coastal waters. The depth range of the measurements vary, as marked; the horizontal scale is ~1100 m. In the upper panel (a) is a record taken in mid-winter at a time of minimum water column stratification. The tide floods from left to right over a sharp submarine sill that nearly blocks a tidal channel located in the southern Strait of Georgia. Water descends the downstream side of the sill with vertical velocity near 1 m/s; the subsequent flow exhibits intense fluctuations of vertical velocity far downstream. The centre panel (b) is another situation in which the tide floods from left to right across a sill; this however is a silled, fjord-type inlet, at a time of very strong near-surface density stratification. Whether because of this stratification "cap" or because of the gentler sill relief, dense water from outside the sill is found entering the inlet on the flood as a bottom boundary current, most visible in the vertical velocity field at those places where it accelerates downwards with increases in bottom slope. A final example in Figure 5c shows a turbulent surface jet...
flowing (left to right) out of a narrow and shallow tidal passage. Water exiting the passage is well-mixed and lighter than the deeper water outside, hence flows out at the surface. Abrupt increase in channel width causes rapid shallowing of the jet just outside the channel mouth.

Figure 5. A variety of flows generate turbulence in the coastal ocean; the associated $w$ fields are displayed in grey-scale. Recall that the sign of $w$ maxima (black) may be determined by the surrounding pattern, light grey for downwards, circles for upwards vertical velocity. In all cases the mean horizontal tidal flow is from left to right. (a) weakly stratified flow over a sill: $R=50$ cm/s, (b) strongly stratified inflow to a coastal fjord: $R=20$ cm/s, (c) a "jet" of well-mixed fluid out of a narrow tidal channel: $R=20$ cm/s.
4. ESTIMATION OF TURBULENCE QUANTITIES

What properties of turbulence would we like to know? - turbulent kinetic energy $E$, the rate $\epsilon$ at which it is being dissipated, and the associated vertical fluxes of mass and momentum, are some that spring to mind. The DOT system, augmented by sporadic vertical profiles of density, should offer information in nearly all of these areas.

Turbulent kinetic energy:

The definition of turbulent kinetic energy per unit mass as $E = 1/2 \left( u^2 + v^2 + w^2 \right)$ uses the components $(u, v, w)$ of the turbulent velocity $\mathbf{u}$, itself defined as the (zero-mean) part left after removal of a "mean" velocity $\mathbf{U} = (U, V, W = 0)$ (where $U$ and $V$ are normally assumed to be functions of $z$ only) from the total velocity $\mathbf{u}_T$. Inherent in this so-called Reynolds decomposition of the flow is an appropriate definition of the averaging process that defines the "mean" flow. While the assumption that $W = 0$ seems safe, it is difficult to decide how to form a "mean" horizontal component in situations where the flow is substantially inhomogeneous. The problem is illustrated in the record of Figure 6 which shows (a) the horizontal velocity component $v$ (relative to the ship) along the axis of a tidal channel and (b) $v_b$, the baroclinic part of this field, formed by removing the local depth-average of $v$. At the beginning (left) of this record, $v_b$ has a three-layer structure, with surface and bottom layers moving more rapidly than a mid-depth layer. By the end (right) of this section of record, the structure had changed to bottom-intensified two-layer flow. It is not at all clear what horizontal scale should be chosen for calculating a "mean" horizontal velocity component $V$, nor how that scale should change with time (horizontal distance).

Because of this uncertainty as to the appropriate averaging for the horizontal "mean" components, the cleanest definition of $E$ would seem to be $E_i = 3/2 \langle w^2 \rangle$, where the overbar denotes an averaging length such that $\bar{w} = 0$, and the subscript is a reminder that this is an isotropic estimate, obtained from the vertical velocity component only.

Turbulent kinetic energy dissipation rate $\epsilon$:

Also of interest is the rate at which mean flow energy is being removed to dissipation scales by the action of the turbulence. The possibility of remote measurement of this quantity has its roots in the work of Batchelor and Townsend (1948), who showed that the large scale eddies of turbulence lose their energy to the turbulent energy cascade (Kolmogoroff, 1941) within at most a few eddy turnover times. Since energy that enters the cascade is delivered to dissipation scales, this means that

$$\epsilon \sim \frac{rw^2}{\tau} \sim \frac{rw^3}{\ell}$$  \hspace{1cm} (1)
Figure 6. (a) Field of horizontal velocity \( v \) relative to the ship (determined from the fore-aft slant beam pair B1,B2) as the ship moves along the axis of a tidal channel. Variations in ship speed and/or the barotropic field are removed in (b) the baroclinic field \( v_b = v - <v> \) where \(<v>\) is the (local) depth-averaged value. The strongly inhomogeneous nature of the horizontal flow makes calculation of horizontal turbulent velocity components difficult.

where \( \tau = \ell / rw \) is the turnover time of an eddy of scale \( \ell \) and rms turbulent vertical velocity \( rw \). This is only a scale relationship, leaving an unknown constant to be determined. Direct measurements of \( \epsilon, rw, \) and \( \ell \) from the atmospheric boundary layer have confirmed the relation (1) above, and suggest that the constant involved is between 3 and 5 (Wamser and Müller, 1977).

Thus for both \( E_1 \) and \( \epsilon \) estimates, it is necessary to derive values for \( rw \), an rms velocity typical of the energy-containing eddies of the turbulent field; for \( \epsilon \), we need in addition a value for the characteristic length scale of such eddies. Meteorologists identify the turbulent length scale \( \ell \) as the location of the peak of a spectrum of vertical velocity as a function of horizontal wavenumber, the turbulent velocity scale \( rw \) as the square root of the spectral integral, a procedure that makes sense in view of the long and homogeneous records that can be obtained from meteorological towers. Unfortunately, the marked inhomogeneity of the turbulent fields in coastal waters means that "a" wavelength doesn't remain constant over the large number of wavelengths necessary for its determination by
remain constant over the large number of wavelengths necessary for its determination by such a Fourier technique. Wavelet analysis (Farge, this volume) may offer a more sophisticated means of determining local wavelength and energy values, but for now, I have used a very simple algorithm, shown schematically in Figure 7. The curve is that of vertical velocity $w$, measured at constant depth (bin), as a function of horizontal distance $x$: horizontal dashed lines denote $\pm \sigma$, one standard deviation of the measurement noise level about the zero mean. Starting with a point (say that marked by the open circle) where $|w| > \sigma$, the algorithm searches for locations of the nearest preceding and following points with $|w| > \sigma$ but of the opposite sign (respectively $P$ and $F$ in Fig. 7). The distance $L$ between these points is taken as a local estimate of a half-wavelength. The average of $w$ over $L$, denoted $aw$, is similarly considered to be the average of $w$ over a half-wavelength. One then moves to point $F$ and repeats the process, resulting in new estimates $L'$ and $aw'$. These local estimates are assigned to the region over which they are calculated; in the (usually small) regions of overlap, the first (in space/time) estimates are arbitrarily chosen. Figure 8b shows the field of $aw$ that results when this algorithm is applied to the tidal front data of Figure 8a.

Assuming that the other half-wavelength exists (although not necessarily in the plane of measurements), the values of $aw$ are converted to a corresponding root-mean-square value ($rw$) by the scaling factor (1.11) appropriate for a pure sinusoid, then used with the length scale estimate $\ell = 2L$ (not shown) to form the estimate of $\varepsilon$, $e2 \equiv |1.11 \, aw|^3 / 2L$, which is shown in logarithmic form in Figure 8c. Note that this estimate of $rw$ can also be used in the estimate $E_i = 3 / 2(\bar{w}^2) = 3 / 2 \, rw^2$ of turbulent kinetic energy.

How much one may trust such an estimate of $\varepsilon$ can be determined by comparing it with values determined directly, by integration of the spectrum of small-scale shear measured in situ. Vertical profiles of such direct measurements of $\varepsilon$ were taken at the two locations.
Figure 8. (a) Measured field of $w$ in a tidal front. (b) Associated field of $aw$ derived using the algorithm depicted in Figure 7. $aw$ and $L$ (not shown) can be used to form a field of $e_2$, estimated turbulent kinetic energy dissipation rate, shown as log($e_2$) in the grey-scale presentation of (c). The vertical lines in (c) denote the launch times of a turbulence microprofiler (operated by Dr. J. Moum, Oregon State University) making direct measurements of $\varepsilon$: maximum profile depths are marked by arrows.
marked in Figure 8c by Jim Moum of Oregon State University. Figure 9(a,b) compares the direct profiler measurement (log $\epsilon$, light line) with the indirect estimate log ($e_2$) for each profile. The heavy line is the logarithm of the average value of $e_2$ over $\pm 10$ pings surrounding the launch of the profiler; the points give some idea of the spread of individual estimates within these 21 pings. The agreement between the two estimates is remarkably good for profile 65. In the subsequent profile, which went somewhat closer to the bottom (about 300 m at both profile locations), we see a defect which tends to recur in many such comparisons; namely a tendency for $e_2$ to underestimate $\epsilon$ near both the surface and bottom boundaries of the flow. This may indicate the need to modify the definition of turbulent length scale $\ell$. Hunt, Stretch and Britter (1988) suggest an alternate form, which tends toward the type of internal scale determined here when the flow is far from boundaries but toward the distance $z$ to the nearest boundary when $z$ is less than this inner scale. Indeed, in measurements taken in the ocean surface layer, Agrawal and Hwang (1991) demonstrate good correspondence between directly measured $\epsilon$ and $(rw)^{3/4}$, with $\ell = z$. Such a modification to $\ell$, causing length scales to decrease, hence $e_2$ to increase near boundaries, would act to correct the discrepancies seen in Figure 9b.

As shown in Figure 9c, however, there are profiles in which there remain very large and unsystematic differences between direct measurements and indirect estimates. Indeed, given the high turbulent intensities and spatial/temporal inhomogeneities characteristic of these flows, this seems scarcely surprising. Consider that the profiler is launched from the stern of the ship, at which time and location the $w$ field is assumed "known" from the Doppler. Thereafter the profiler, falling vertically, can be advected horizontally by the local ambient flow, so does not necessarily remain at this geographic launch position. Even if it were to remain there, the flow field may change in the time taken for the profile (typically 4-5 minutes for a profile to 300 m). Various checks for the likelihood of time change can be devised, using the fact that the fore/aft slant beams allow two measurements of $\nu$ that are separated in time, but this is merely an effort to avoid a statistical problem, that of estimating the degree of agreement (or disagreement) necessary before a remote measurement of a non-stationary and inhomogeneous field can be considered "proven" by a relatively sparse set of ground-truth measurements.

Vertical buoyancy (mass) flux:

Part of the reason one might like a remote technique for $\epsilon$ is because for the last decade, oceanographers have obtained what they often really want, the vertical buoyancy flux $\rho' \nu w$, from what they are able to get from microstructure profiler measurements, namely $\epsilon$, and a model (Osborn, 1980) which suggests that under certain assumptions,

$$\overline{\rho' \nu w} = \frac{R_f}{1 - R_f} \rho_0 g^{-1} \epsilon$$

(2)
where $R_f$, the flux Richardson number, is the ratio of buoyancy sink to shear source terms in the turbulent kinetic energy equation. Oceanographers add the further assumption that $R_f \approx 0.2$, resulting in an estimate of buoyancy flux as a constant fraction of the measured turbulent kinetic energy dissipation rate $\epsilon$. If correct, this model means that a remote measurement of $\epsilon$ would correspond to a remote measurement of buoyancy flux. However, the model has rarely been checked by comparison with direct flux measurements, as these are extremely difficult to make in the ocean environment. The small amount of evidence which does exist (Yamazaki and Osborn, 1993) suggests that $R_f$ is either not constant, or else considerably smaller than 0.2. Vertical turbulent fluxes (or equivalently, turbulent diffusivities) are important products of oceanic microstructure.
measurements; it would be nice to know the circumstances (if any) under which such dissipation-based estimates are accurate, hence remote measurement of buoyancy flux would be possible.

With the addition of a towed CTD with finescale resolution (Ocean Sensors), it has proven possible to make statistically significant measurements of buoyancy flux using the DOT system. The CTD is towed at constant depth, just in front of the vertical beam of the Doppler, for long periods. Figure 10 shows the CTD measurement of density, and the associated time series of $w$ measured in the Doppler bin that includes the CTD tow depth, over about three hours. Below is an enlargement of a small section of the record (taking care to preserve phase, the density field has been high-pass filtered to remove the very largest scales of variation in water properties). Buoyancy flux will be a positive quantity if on average downward(upward) vertical velocities carry lighter/heavier water.

![Figure 10](image)

Figure 10. The top panel shows time series of CTD density (light line), along with $w$ (dark line) from the Doppler bin within which the CTD was towed. Before calculating fluxes, the density time series is high-pass filtered (preserving phase) to remove the variance associated with large-scale water mass change: the enlargement shows filtered density and $w$ over one of the interval lengths used in the flux calculation.

Figure 11 shows the direct flux estimates, formed by breaking the $\rho'w$ records into pieces of fixed length=spts, then forming $\langle (\rho' - \overline{\rho'}) (w - \overline{w}) \rangle$ where the average is over spts. Error bars are calculated from the variance of such estimates over the number(spts) of different starting points, and an estimate of the number of independent values determined from the
number of zero-crossings of \( w \). The points in Figure 11 are the accompanying estimates of the buoyancy flux made using (2) above with \( R_f/(1-R_f) = 0.2 \) (\( \varepsilon \) values were taken from the Oregon State profiler measurements over a range of 6 m centered on the CTD tow depth: courtesy of Jim Moun). While there is encouraging general agreement, i.e., values tend to be high where the direct flux measurement is large and positive, low when the direct measurement is not statistically different form zero, we face (again) the problem of how best to average "point" estimates from the profiler for comparison with a more broadly based determination from the towed measurement.

![Graph of buoyancy flux](image)

Figure 11. Direct calculation of buoyancy flux (solid line) with estimated error bars (dashed lines) over consecutive 400-point blocks of the time series shown in Figure 10. Circles are indirect estimates of the flux, using profiler “point” measurements of \( \varepsilon \) and the formula \( 0.2 \rho g \varepsilon \).

**CONCLUSIONS**

It is now possible to make measurements of the vertical velocity field in turbulent coastal flows, using a modified ADCP system. This allows us to site-survey for turbulence and, once found, to investigate its spatial and temporal variability. From the \( w \) field measurement, it will be possible to estimate turbulent kinetic energy \( E \) and possibly its dissipation rate \( \varepsilon \). Addition of a towed CTD allows direct measurement of buoyancy flux: if the model (2) relating buoyancy flux to \( \varepsilon \) can be validated, remote measurement of \( \varepsilon \) would be equivalent to remote measurement of buoyancy flux, probably the feature of turbulent flows that is of the greatest importance to coastal applications.

Are results from the coastal ocean likely to be valid when translated to offshore oceans? From the data presented here, velocities characteristic of turbulence in the coastal ocean are clearly much higher than those we expect offshore. However, coastal stratification is
also much larger: the combination makes the coastal ocean less different from that offshore than one might think. The lower offshore signal level poses some challenges, but if a stable platform can be provided, the increased accuracy available with the newer coded-pulse sonars should allow this type of measurement to be made offshore as well: one foresees applications in studies of surface and bottom boundary layers in particular. It is my hope that the techniques discussed here will eventually prove as useful in the offshore environment as they are in the coastal ocean.

REFERENCES


