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CORRESPONDENCE

Comment on "Using an ADCP to Estimate Turbulent Kinetic Energy Dissipation Rate in Sheltered Coastal Waters"

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ABSTRACT

Greene et al. revisit the suggestion that the turbulent kinetic energy dissipation rate could be estimated through a "large-eddy estimate," employing acoustic measurements of velocity fields associated with the largest energy-containing scales of ocean turbulence. While the large-eddy estimate as originally proposed used vertical velocity and a vertical eddy length scale, Greene et al. chose instead to substitute a horizontal length scale for the latter. This comment argues that combining a horizontal scale for length with a vertical velocity scale produces a large-eddy estimate of the dissipation rate that is accurate only if the *energy-containing* eddies are isotropic, and that this condition is highly unlikely in naturally occurring ocean turbulence, subject as it is to influences of stratification, vertical shear, and/or the presence of horizontal boundaries. The problem is documented using data from a large-eddy simulation of Langmuir supercells.

1. Comment

The paper of Greene et al. (2015) revisits a suggestion that acoustic measurements of velocity fields associated with the largest scales of ocean turbulence could be used to estimate a small-scale parameter, the turbulent kinetic energy dissipation rate ε, through a "large-eddy estimate,"

$$\varepsilon_l = C_\varepsilon \frac{u_l^2}{\tau_l} = C_\varepsilon \frac{u_l^3}{l}, \tag{1}$$

where u_l, τ_l, and l are typical velocity, time, and length scales of the largest energy-containing eddies of the turbulence (Gargett 1994, 1999). It is widely accepted (Pope 2000) that these energy-containing scales are almost always anisotropic, since they bear the imprint of the instability process(es) that supply them with energy and/or that of nearby boundaries. This is true regardless of whether the Reynolds number is sufficiently high to produce *local* isotropy (Kolmogoroff 1941), that is, an inertial subrange of isotropic eddies between these energy-containing scales and the dissipation scales that irreversibly remove their energy. In the ocean, stable stratification, vertical shear, and/or horizontal boundaries all conspire to produce anisotropy of energy-containing scales such that a characteristic vertical length scale L_v is normally smaller than a characteristic horizontal length scale L_h. From continuity, $u/w \sim L_h/L_v$, that is, characteristic turbulent vertical velocity w is smaller than characteristic horizontal velocity u by the length scale ratio L_v/L_h. Thus, when energy-containing eddies are anisotropic, a characteristic time scale of their decay through irreversible transfer to dissipation scales,

$$\tau_l \sim u/L_h \sim w/L_v, \tag{2}$$

must be calculated with the appropriate combination of velocity/length scales, that is, as either $\tau_l = \tau_h = u/L_h$ or $\tau_l = \tau_v = w/L_v$. Gargett (1994, 1999) used acoustic measurements of vertical velocity w' to calculate a characteristic vertical velocity $w = \langle w'^2 \rangle_l^{1/2}$ as the rms value of w' averaged ($\langle \rangle_l$) over $l = L_v$, a large-eddy vertical length scale determined by a zero-crossing algorithm. The time scale associated with these choices is $\tau_l = \tau_v = w/L_v$.

For reasons that are not explained, the authors of the paper under discussion did not investigate the algorithm as originally suggested but instead choose to compute and use a *horizontal* length scale $l = L_h$ with a velocity scale defined as the rms value of w' over L_h; that is, they use an implicit decay time scale $\tau = w/L_h$ that is neither of the appropriate time scales in Eq. (2). Provided that

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the *energy-containing* scales of the turbulence are isotropic, there will be no difference between this definition and those of Eq. (2). Comparing spectra from a vertical beam and one canted 60° from vertical, Green et al. (2015, p. 325) argue that "close agreement between the spectra of the two components suggests that the flow is isotropic" at the horizontal scale of 30 m used for their computation of L_h, hence that "the horizontal correlation scale is a reasonable surrogate for the vertical," that is, that the turbulence *is* isotropic. However, while 30 m may be within an inertial subrange for some of the values of ε reported, some of the observed spectra do not exhibit clear and/or extensive inertial subranges (Greene et al. 2015, their Fig. 5). Moreover, the location within an inertial subrange merely suggests *local* isotropy, that is, isotropy at scales small enough that the turbulence has become independent of the larger scales at which energy resides. In the same vein, the referenced observations of Gargett et al. (1984) show only that *local* isotropy is guaranteed if a buoyancy Reynolds number is sufficiently large: in that study, measurements of energy-containing scales were contaminated by vehicle motion, hence nothing could be deduced about their degree of isotropy.

Anisotropy of energy-containing scales with $L_v < L_h$ is the most likely oceanic norm, so in general $\varepsilon_h < \varepsilon_v$ and a choice of horizontal rather than vertical length scale does matter. How *much* can it matter? To illustrate, I consider Langmuir supercells (Gargett et al. 2004; Gargett and Wells 2007), full-depth Langmuir circulations that are the energy-containing eddies of turbulence in shallow coastal seas during conditions driven predominantly by a wave-induced Langmuir vortex force (Gargett and Grosch 2014). Well-resolved (Gargett et al. 2008, 2009) observations of the large-eddy structure of this particular form of turbulence show that $\lambda = 2L_h \sim 6H \sim 6L_v$, where H is the water column depth and the factor of 2 accounts for the pair of large-scale counterrotating vortices ("large eddies") that make up a complete Langmuir cell of horizontal scale λ. In this case $\varepsilon_h \sim \varepsilon_v/3$, a bias that cannot be considered small. Moreover, this is the bias that would be obtained with a measurement path normal to the roughly wind-aligned long axis of the vortex pair; hence, this is very much a best-case estimate. Because Langmuir circulations are much larger in the downwind direction than the crosswind direction, any measurement path other than crosswind would result in even larger horizontal scale, hence even smaller ε_h estimates.

What evidence do we have for applicability of a large-eddy model for dissipation rate in this case (apart from the fact that it is unreasonable to assume that the method works for some large eddies but not others)?

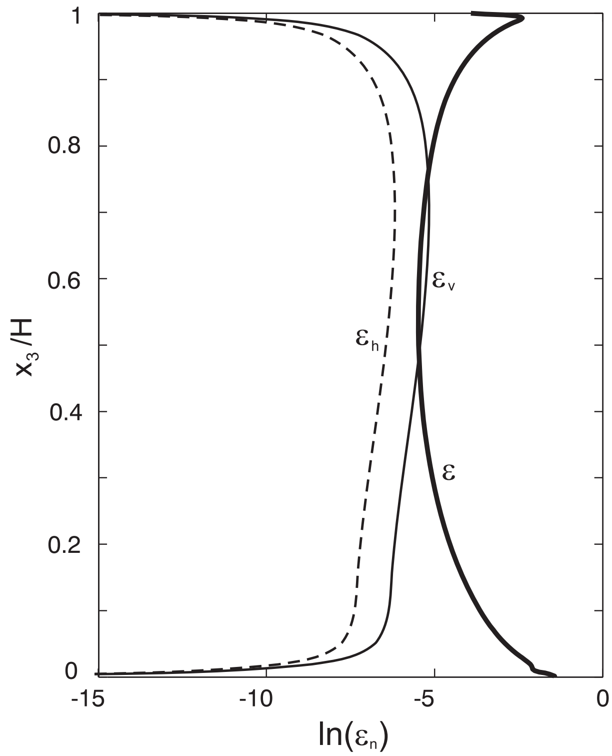

FIG. 1. Profiles as a function of scaled height above the bottom of $\ln(\varepsilon_n)$, the natural logarithm of turbulent kinetic energy dissipation rate normalized with H and the surface stress velocity used to force an LES of Langmuir supercells. Estimates are determined both directly from LES results (heavy line) and as large-eddy estimates ε_v and ε_h made with the same rms vertical velocity (averaged over both horizontal dimensions) but different length scales, $L_v = H$ (light line) and $L_h = 3H$ (dashed line), respectively: both estimates use calibration constant $C_\varepsilon = 1$ (Gargett 1999). The dashed line is a best-case scenario for the ε_h estimate, obtained when the horizontal transect is normal to the long axis of the supercells: because of the substantial anisotropy of Langmuir circulations, any other path would yield larger L_h, hence even smaller ε_h.

Profiler-based dissipation measurements are not available for comparison. Even if profiler operations were possible in the wind and sea-state conditions under which supercells exist, impacts of sediment particles that are resuspended and distributed throughout the water column during these events would most likely irreversibly contaminate shear probe measurements. Available evidence comes instead from a large-eddy simulation (LES) model (Tejada-Martínez and Grosch 2007) that, when run with forcing parameters matched to observational conditions, produces results closely resembling all available observational features of Langmuir supercells (Gargett and Wells 2007). Figure 1 shows profiles of three dissipation rate estimates (normalized with water depth H and the surface stress velocity used to force the LES). The "direct" estimate is the sum of the viscous dissipation rate of the part of the flow resolved by the

LES and the dissipation rate associated with the subgrid-scale closure: results are insensitive to the exact form used for this closure (Tejada-Martínez and Grosch 2007). Both large-eddy estimates use the value of $C_\varepsilon = 1.0$ determined by Gargett (1999). The light line is a large-eddy estimate made with characteristic length scale equal to the vertical length scale $L_v = H$ of the cells and characteristic velocity $\langle w'^2 \rangle^{1/2}$, where angle brackets denote an average of fluctuating vertical velocity over horizontal dimensions and time. This estimate would not differ greatly if averaging were performed over the vertical extent of individual cells, as in Gargett (1999), because ε is relatively uniform over H. These two estimates are seen to agree well, save in near-boundary layers where the effects of the smaller Reynolds number of the LES are observed. The large-eddy estimate that would result from use of L_h measured in the crosswind direction (dashed line) is noticeably inferior, and it would be even more so if the value of $C_\varepsilon = 0.35$ suggested by Greene et al. (2015) were used. Finally, as remarked above, any other path angle would result in larger values of L_h, hence even smaller values of ε_h.

Although the example of Langmuir supercells was useful because of the complete information available from an observationally validated LES, the degree of anisotropy of the large eddies and consequent effects on resulting large-eddy computations of dissipation rate are not unique. In Kelvin–Helmholtz instability (KHI), the most likely mechanism producing turbulence in stratified shear flows, the direct numerical simulations (DNS) of Smyth et al. (2001) exhibit primary instabilities (billows) with $L_{\mathrm{KHI}} \sim (2-3)L_v$, where L_{KHI} is the horizontal length scale of the billows in direction of the mean sheared current and L_v is the vertical scale of the billows: this ratio is roughly maintained during vortex pairing. Moreover, the resolution required to approach oceanographically relevant Reynolds numbers requires DNS computations of KHI to severely restrict the horizontal spanwise (cross current) scale [Smyth et al. (2001) explore the spanwise scale to a maximum of $\sim L_{\mathrm{KHI}}/2$, while higher Reynolds numbers restrict Salehipour et al. (2015) to $L_{\mathrm{KHI}}/4$]; thus, these computations cannot address the question of large-eddy scale in the spanwise direction. However, in nature, as dramatically evidenced by roll clouds in the atmosphere (e.g., Holt 1998, his Fig. 1), the spanwise scale can be much larger than L_{KHI}. Reviewing both laboratory measurements and field evidence, Thorpe (2002) finds that during KHI, the *minimum* average length of the billow crests in the spanwise direction is about 4 times L_{KHI}, hence about 10 times L_v. Thus, as with Langmuir circulations, the (observationally unknown) direction of travel across a field of KHI structures may influence the value determined for *the* horizontal length scale.

2. Discussion and conclusions

The main purpose of this comment is a general warning against using a *horizontal* length scale $l = L_h$ together with a characteristic *vertical* velocity scale in a large-eddy estimate of the turbulent dissipation rate *without* supplying an independent calibration by concurrent microprofiler measurements. This necessity for accompanying microscale measurements when using τ as a decay time scale unfortunately negates one of the main reasons for the original proposal of the large-eddy technique, which was to eventually eliminate the costs and labor that are (still) involved in operating microscale profilers.

Returning to use of $l = L_v$, hence τ_v, automatically ensures against an unknown degree of anisotropy of the energy-containing eddies. For use in calculating τ_v, L_v can be determined by a zero-crossing algorithm (Gargett 1999), by a correlation method similar to that used by Greene et al. (2015) for L_h, or in stratified conditions as $L_b \equiv w/N$ (Moum 1996). Use of L_v also eliminates two additional constraints inherent to the use of L_h, namely, that the turbulence must fulfill the frozen field condition necessary to convert temporal measurements to length scales and that ship speed must remain effectively constant for the length of time used to calculate L_h: both conditions are unnecessary when using (nearly instantaneous) measurements of L_v.

Finally, as part of a discussion of estimation of ε via a large-eddy technique, it is interesting to consider the origin of the present use of ε as a measure of turbulence in the ocean. For more than three decades, sea-state constraints meant that ocean turbulence was measured via freely falling profilers divorced from the motion of the sea surface. The response of profilers to internal flows on scales similar to their own meant that only the very smallest scales of turbulence, the dissipation scales, could be reliably observed. The lack of alternatives has allowed measurements of these smallest scales to become enshrined as *the* measurement of ocean turbulence. However, direct information about the connection between turbulence and the processes generating it is lost during the inertial cascade of energy to dissipation scales. As a result, we generally have only inferred properties of these forcing processes, limiting our ability to understand the connections between forcing and mixing in the ocean. As reported in the paper under discussion, advances in Doppler technology over the past 15–20 years have indeed allowed large-eddy estimates of ε to be made at levels much lower than the original ones of Gargett (1994, 1999). It seems even more important, however, that these technological advances in acoustic measurements and in deployment strategies are beginning to allow us to measure the large eddies of turbulence directly and thus observe, rather than infer, the connections between forcings and features of the resulting turbulent structures (Gargett and Grosch 2014). These connections are essential to understanding the properties and the influence of turbulence in the ocean, as well as potential changes in them, as large-scale parameters such as forcings, buoyancy, and shear evolve under climate change.

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