On the Structure of Turbulence in the Bottom Boundary Layer of the Coastal Ocean

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(Manuscript received 12 September 2002, in final form 15 July 2004)

ABSTRACT

Six sets of particle image velocimetry (PIV) data from the bottom boundary layer of the coastal ocean are examined. The data represent periods when the mean currents are higher, of the same order, and much weaker than the wave-induced motions. The Reynolds numbers based on the Taylor microscale (Re_{λ}) are 300-440 for the high, 68-83 for the moderate, and 14-37 for the weak mean currents. The moderate-weak turbulence levels are typical of the calm weather conditions at the LEO-15 site because of the low velocities and limited range of length scales. The energy spectra display substantial anisotropy at moderate to high wavenumbers and have large bumps at the transition from the inertial to the dissipation range. These bumps have been observed in previous laboratory and atmospheric studies and have been attributed to a bottleneck effect. Spatial bandpass-filtered vorticity distributions demonstrate that this anisotropy is associated with formation of small-scale, horizontal vortical layers. Methods for estimating the dissipation rates are compared, including direct estimates based on all of the gradients available from 2D data, estimates based on gradients of one velocity component, and those obtained from curve fitting to the energy spectrum. The estimates based on vertical gradients of horizontal velocity are higher and show better agreement with the direct results than do those based on horizontal gradients of vertical velocity. Because of the anisotropy and low turbulence levels, a -5/3 line-fit to the energy spectrum leads to mixed results and is especially inadequate at moderate to weak turbulence levels. The 2D velocity and vorticity distributions reveal that the flow in the boundary layer at moderate speeds consists of periods of "gusts" dominated by large vortical structures separated by periods of more quiescent flows. The frequency of these gusts increases with Re_{λ} , and they disappear when the currents are weak. Conditional sampling of the data based on vorticity magnitude shows that the anisotropy at small scales persists regardless of vorticity and that most of the variability associated with the gusts occurs at the low-wave-number ends of the spectra. The dissipation rates, being associated with small-scale structures, do not vary substantially with vorticity magnitude. In stark contrast, almost all the contributions to the Reynolds shear stresses, estimated using structure functions, are made by the high- and intermediate-vorticity-magnitude events. During low vorticity periods the shear stresses are essentially zero. Thus, in times with weak mean flow but with wave orbital motion, the Reynolds stresses are very low. Conditional sampling based on phase in the wave orbital cycle does not show any significant trends.

1. Introduction

Predictions of ocean dynamics, sediment transport, pollutant dispersal, and biological processes require

knowledge on the characteristics of turbulence in the bottom boundary layer. Modeling of the turbulence, whether in the context of Reynolds-averaged Navier Stokes or large-eddy simulations, requires data for development and validation of closure models. In an effort to address some of the relevant issues we have developed, in recent years, a submersible particle image velocimetry (PIV) system for measuring the flow structure and turbulence in the bottom boundary layer of the coastal ocean. PIV measures the instantaneous distributions of two velocity components in a sample area. Thus, PIV provides direct data on the structure of the turbulence without assumptions involving Taylor's hy-

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pothesis. A time series of PIV datasets provides the time evolution of the turbulence. Since PIV produces both spatial and temporal series, it permits separation between waves and turbulence that have different length scales but similar frequency. Early results and several configurations of this system have been reported in Bertuccioli et al. (1999), Doron et al. (2001), and Nimmo Smith et al. (2002).

In this paper we select and compare several datasets from coastal waters with different height above the bottom and varying ratios of the mean flow to the amplitude of wave-induced motion. These datasets are selected to represent conditions of relatively high, intermediate, and weak mean flows. They also represent substantially different turbulent Reynolds numbers. A laboratory dataset of locally isotropic turbulence, with similar Reynolds number but with substantially different length scales, is used for comparison. Spatial spectra are used for addressing questions of isotropy, and various methods for calculating/estimating the dissipation rates are compared. Clear evidence that the turbulence in the coastal bottom boundary layer is anisotropic, particularly at small scales, is provided, leading to questions on the validity of dissipation rates estimates based on one-dimensional spectra and to variations in dissipation estimates based on gradients of one velocity component. Furthermore, the 2D velocity distributions allow us to examine the structure of the flow, enabling, for example, identification of phenomena associated with the anisotropy at small scales. At large scales, we demonstrate the occurrence of intermittent "gusts" with quiescent periods between them. These gusts dominate the Reynolds stresses but have a lesser impact on the dissipation rate. By combining a series of velocity distributions displaced by the local velocity (i.e., by using Taylor's hypothesis), one can gain insight on the flow structure at scales that exceed a single velocity distribution. The results are compared to relevant prior investigations of ocean turbulence. However, because of the vast amount of available data on this subject, we restrict the discussions to the most relevant prior studies for the sake of brevity.

2. Experimental setup, deployments, and data

a. Apparatus

A detailed description of the oceanic PIV system can be found in Nimmo Smith et al. (2002), and only a summary is provided here. The present system has evolved and improved substantially from the original setup described in Bertuccioli et al. (1999) and Doron et al. (2001). It now features two sample areas, higherresolution cameras, a massive data acquisition system, and an extended-range profiling platform. A schematic of the submerged components of the PIV system is shown in Fig. 1. The light source is a dual-head, pulsed, 350 millijoule-per-pulse dye laser, which generates pairs of 2-µs pulses at 594 nm. The laser is located on the surface vessel, and the light is transmitted through two optical fibers to submerged probes, each containing the light-sheet forming optics, which illuminate the two sample areas. The thickness of the light sheets, about 3 mm, and the delay between exposures are set to allow misalignment to the flow of up to $\pm 20^{\circ}$ before losing one of the particle traces (on average).



FIG. 1. Schematic diagram of the submersible PIV sampling platform fully retracted. The inset on the right shows the platform fully extended.

Images are acquired using two 2048 \times 2048, 12 bitsper-pixel CCD cameras (Silicon Mountain Design 4M4), each capable of sampling at up to four frames per second, producing a maximum data rate of 32 MBytes s^{-1} per camera. Each camera and associated light sheet can be aligned independently, in the same or different planes, near each other or apart. In the setup shown in Fig. 1 the two sample areas are located in the same plane, but some of the data have been acquired in perpendicular planes. For each of the cameras, both exposures are recorded on the same frame. A hardwarebased "image shifter" offsets the second exposure by a known fixed amount, which overcomes the directional ambiguity problem of double exposure images. These cameras are about 4 times as sensitive as typical crosscorrelation cameras (where the two exposures are recorded on different frames), but recording both exposures on the same frame reduces the signal-to-noise ratio. The data are recorded on two 240-GByte shipboard data acquisition systems (one for each camera), each comprising six 40-GByte IDE disks, allowing continuous sampling at 0.5 Hz for over 15 h or at faster sampling rates for proportionately shorter periods.

The submersible components of the PIV system are mounted on stable seabed platforms, which can be rotated to align the sample areas with the mean flow direction, and extend vertically to sample the flow at different elevations. The flow direction is determined by monitoring the orientation of a vane mounted on the platform using a submersible video camera. Some of the data described in this paper were collected using the original platform, a hydraulically operated scissor-jack, which has a limited traversing vertical range of 1.1 m (Doron et al. 2001). The recent data were collected using a new platform, consisting of a five-stage, doubleacting telescopic hydraulic cylinder, which has a vertical sampling range of 9.75 m. A fully extended platform is illustrated on the right-hand side of Fig. 1. The submersible system also contains a Sea-Bird Electronics SeaCat 19-03 CTD, optical transmission and dissolved oxygen sensors, a ParoScientific Digiquartz Model 6100A precision pressure transducer, an Applied Geomechanics Model 900 biaxial clinometer, a KVH C100 digital compass, and a video microscope for sampling the particle distributions at high magnification.

b. Data analysis procedures

The extraction of the velocity distributions from each image is carried out following the methodology detailed in Roth and Katz (2001), Roth et al. (1999), and Sridhar and Katz (1995) with some modifications to the image enhancement procedures (Nimmo Smith et al. 2002). Constrained by the particle concentration, we used 64×64 pixel interrogation windows and 50% overlap between windows. Thus, each vector map consists of a maximum of 63×63 velocity vectors. The effects of optical distortion and mean out-of-plane motion are corrected by using calibration and geometric arguments, as described in Nimmo Smith et al. (2002). The data quality varies depending on the particle concentration, size, and spatial distribution of the particles and the orientation of the laser sheet relative to the instantaneous flow. Typically, 60%–80% of the vectors satisfy the accuracy criteria of the data analysis software (Roth and Katz 2001). Instantaneous velocity distributions that contained less than 60% vectors are not used during subsequent analyses. Provided that the interrogation area contains 5-10 particle pairs, the uncertainty in velocity is estimated conservatively at 0.3 pixels. With a typical displacement of 20 pixels between exposures, the uncertainty of a single, instantaneous velocity vector is 1.5%. The uncertainty in the instantaneous vorticity and other parameters based on velocity gradients is about an order of magnitude higher ($\sim 20\%$). However, the uncertainty of ensemble-averaged quantities for example, the mean dissipation rate (which involves velocity derivatives squared), improves by about the square root of the number of data points being averaged. Thus, by using at least 1000 instantaneous realizations, the uncertainty in terms involving averaged spatial velocity derivatives decreases to less than 1%.

For calculating the dissipation rate estimates and vorticity distributions, gaps within the velocity data are filled by linear interpolation using the vectors located in the direct neighborhood of a missing vector. We follow a recursive process, which fills the gaps surrounded by the most "good" (directly measured) vectors first. The outermost two strips of vectors in each map are discarded since they have been found to be less reliable. A sample of the resulting instantaneous velocity field, containing 59 \times 59 vectors, is shown in Fig. 2. The instantaneous sample-area mean horizontal and vertical velocities are subtracted from each vector to aid visualization of the turbulent fluctuations.

The Reynolds shear stress, $\overline{-u'w'}$, is estimated using the second-order structure function, $D_{13}(r)$, the covariance of the velocity difference between two points separated by a distance r from each other, that is, [u'(x + r) - u'(x)][w'(x + r) - w'(x)]. This methodology was introduced by Trowbridge (1998) and has already been applied to PIV data in Nimmo Smith et al. (2002). The magnitude of D_{13} asymptotically approaches that of $2\overline{u'w'}$, while being free from contamination by wave-induced motion, when r is larger than the integral scale of the turbulence but is still much smaller than the wavelength of the surface waves. The PIV data enable us to calculate the distribution of $D_{13}(r)$ as a function of the separation distance. We combine and average the data from different points at the same elevation to increase the sample size.

c. Deployments and characterization of the data

Table 1 summarizes the data series that are used for the analysis described in this paper. These data were collected during two field deployments off the eastern seaboard of the United States in 2000 and 2001. To



FIG. 2. Sample instantaneous velocity distribution. The instantaneous sample area mean horizontal and vertical velocity components (shown at the top of the figure) are subtracted from each vector to show the turbulent fluctuations.

characterize the mean flow and relative amplitude of surface waves, we average the velocity over the entire vector map to obtain the instantaneous average streamwise and vertical velocity components (over the 59×59 vectors), U and W, respectively. The mean current is characterized by \overline{U} and \overline{W} , the overall average velocity components (averages of U and W over all distributions), whereas $U_{\rm rms}$ and $W_{\rm rms}$ are their rms values, representing mostly the effect of surface waves, but also turbulence at scales larger than the instantaneous distributions. Of the data collected in 2000, we only present the results of measurements performed on the night of 19-20 May, inside the harbor of refuge at the mouth of Delaware Bay. These series are indicated as runs A and B in Table 1. The flow at this location was characterized by strong tidal currents with surface velocities in excess of 1.5 m s⁻¹ [measured by the ship acoustic Doppler current profiler (ADCP)] and little wave motion. The water depth at the deployment site was 11 m, and the seabed consisted of coarse sand and broken shell without any ripples or other notable bed forms. We recorded data at 3.33 Hz for periods of 5 min (1000 images per camera) at elevations (center of the vector map) of 0.35 and 1.44 m above the bottom.

A second deployment took place in September 2001 (runs C-F in Table 1), about six nautical miles to the southeast of the Longterm Ecosystem Observatory (LEO-15) site off the New Jersey coast at 39°23'37"N, 74°9'32"W. The water was 21 m deep, and the seabed consisted of coarse sand with ripples having wavelength of approximately 50 cm and height of 10 cm. The currents in the region were generally moderate; however, the site was exposed to oceanic swell. The water column was strongly stratified with a sharp thermocline situated about 7 m above the seabed. Data were collected continuously for periods of 20 min, at a sampling rate of either 2 or 3.33 Hz, and at elevations up to 8.5 m above the seabed. In this paper we present data obtained at mean elevations (center of the vector map) of 0.55 and about 2.5 m above the bottom.

Time series of U for the periods detailed in Table 1 are presented in Fig. 3. It can be seen that for the two Delaware Bay sampling periods (Figs. 3a and 3b), the amplitude of wave-induced flow, with a period of about 7 s, is weak in comparison with the mean flow. Largeramplitude, longer-period oscillations ("beating") are also in evidence. In contrast, for the sampling periods at the LEO-15 site shown in Figs. 3c and 3d, the amplitude of the wave-induced velocity is of the same magnitude as the mean flow and at times is large enough to cause reversal of the flow direction. In the extreme, when the mean current is very low, when the tidal flow changes direction (Figs. 3e and 3f), the flow consists almost entirely of oscillatory wave-induced motion. A magnified part of Fig. 3d, whose timing is indicated by dashed lines, showing also the vertical velocity component is presented in Fig. 4. It demonstrates that the oscillatory motion induced by the surface gravity waves, with a period of about 10 s, is well resolved by the sampling at 3.33 Hz.

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Run	Date	Start time (UTC - 5 h)	Site	Size of square sample (cm)	Sampling duration (frames)	Sampling frequency (Hz)	Elevation of center of sample (cm)	\overline{U} (cm s ⁻¹)	$U_{ m rms}$ (cm s ⁻¹)	\overline{W} (cm s ⁻¹)	$W_{ m rms}$ (cm s ⁻¹)
А	20 May 2000	0240:00	Delaware	50.20	1000	3.33	144	38.2	2.8	-0.8	1.4
В	20 May 2000	0304:30	Bay Delaware Bay	50.20	1000	3.33	35	32.6	3.7	-0.3	1.2
С	9 Sep 2001	0604:00	LEO-15	34.65	4000	3.33	257	14.9	4.5	-0.8	0.7
D	9 Sep 2001	0339:30	LEO-15	34.65	4000	3.33	55	7.7	3.3	-0.2	0.3
Е	10 Sep 2001	0159:00	LEO-15	34.65	2400	2.00	245	-0.5	4.0	0.1	0.7
F	10 Sep 2001	0259:00	LEO-15	34.65	2400	2.00	55	0.9	3.2	-0.1	0.1

TABLE 1. Summary statistics of the datasets selected for this paper.



FIG. 3. Time series of the sample area mean horizontal velocity for each of the sampling periods. (a)–(f) Runs A–F, respectively, in Table 1.

3. Flow structure

The main advantage of PIV data over those obtained using point measurement techniques is the ability to examine instantaneous 2D cross sections of the flow structure. Figure 5 presents four sample instantaneous velocity and (plane-normal component of) vorticity distributions, characteristic of different flow regimes. Figure 5a shows the flow structure that is typical of the data series with weak mean current, but with waveinduced motion (runs E and F). The flow is dominated



FIG. 4. Time series of the sample area mean horizontal and vertical velocity components for the region bounded by the dashed lines in Fig. 3d, sampled at 3.33 Hz.

by very small (<2-cm diameter) weak eddies, and there are no large-scale vortical structures. Figures 5b, 5c, and 5d, while all from the same sampling period of moderate mean flow and wave motion (runs C and D), are distinctly different. Figure 5b is fairly typical of about 60% of the 20-min data series, showing that the flow is moderately quiescent in parts but contains some small (\sim 4-cm diameter) eddies, which appear singly or in small groups. Intermittenly, groups of larger vortical structures appear within the flow, either as very large (\sim 10-cm diameter) separated vortices (Fig. 5c) or as "strings" (trains) of medium-sized (~6-cm diameter) vortices aligned along lines of intense shear (Fig. 5d). The intermittent groups of intense vortical structures, which we will refer to as gusts, have been found to be very coherent in both time and space. A sample time sequence of velocity-vorticity distributions, showing a group of large vortices (Fig. 5c) being advected by the mean flow through the upstream sample area, is presented in Fig. 6. The centers of the three, well-defined, large vortices are located ~ 16 cm upstream and ~ 8 cm lower than the proceeding one. The very same group of vortices are observed passing through the downstream sample area (not shown) about 8 s later with very little change to their individual or group structure.

One has to extend beyond the scales of a single vector map to observe the large-scale, spatial structure of a gust. Figure 7 shows two approximations of gusts, produced by overlaying extended series of instantaneous vorticity distributions, each offset from the previous by the instantaneous mean streamwise and vertical velocity components. The out-of-plane shift between distributions is estimated from the instantaneous mean outof-plane velocity component (distortion), which is removed from in-plane data (Nimmo Smith et al. 2002). By making each layer partially transparent, the resulting image is an integral of several overlapping distributions. The series of dots indicate the top-left corner of each vorticity distribution, illustrating the amount of offset between frames. The three large vortices of Fig. 6 are clearly visible in the -60 < x < -20 cm range of Fig. 7a. Farther upstream (-100 < x < -60 cm), there is a region of small, intense vortices aligned horizontally beyond which (-120 < x < -100 cm) there are two more large vortices. Overall, within the height range of this dataset, the gust has a streamwise extent in excess of 1 m. This coherent pattern of an extended train of large separated vortices is entirely consistent with a vertical section through a packet of "hairpin" vortices, previously observed in laboratory measurements (Adrian et al. 2000) and numerical simulations (Zhou et al. 1999). Like here, they show that the "heads" of successive hairpin vortices appear below and upstream of the previous ones. Figure 7b is another example of the spatial structure of a gust centering around the frame shown in Fig. 5d. Here, the train of intense medium-sized vortices has a streamwise extent of about 0.8 m, forming along an inclined shear layer. In



FIG. 5. Examples of instantaneous velocity fluctuation superimposed on the vorticity distribution showing typical structures: (a) run F, U = -0.7 cm s⁻¹, W = -0.1 cm s⁻¹; (b) run D, U = 6.9 cm s⁻¹, W = -0.2 cm s⁻¹; (c) run D, U = 11.3 cm s⁻¹, W = -0.2 cm s⁻¹; (d) run D, U = 6.0 cm s⁻¹.

both samples of gust, the regions upstream and downstream of the powerful train of vortices are relatively quiescent with no large-scale vortices present. This phenomenon of gusts separated by quiescent regions persists throughout the moderate mean flow data (runs C and D). Examinations of the high-flow results (runs A and B) reveal the existence of similar trains containing larger and more intense vortices. However, the intervening quiescent periods between gust events become much smaller and completely disappear close to the bed (run B). Conversely, under low mean current conditions (runs E and F) the gusts disappear and, except for a few rare occasions (that will be discussed), the flow does not contain distinct powerful vortices.

4. Spatial turbulent energy spectra

Figure 8a presents mean, one-dimensional energy spectra of $u(E_{11})$ and $w(E_{33})$ integrated along the streamwise (k_1) and vertical (k_3) directions for run D. They are calculated by averaging the spatial spectra from the central seven rows (or columns) of all the frames in the data series. The procedures include linear interpolation to fill the gaps within the data, removal of the mean, linear detrending, and fast Fourier transforms. Unlike our previously published spectra (Doron et al. 2001), we do not use a windowing function. Note that these spectra are calculated from the instantaneous velocity distributions and do not involve use of Tay-



FIG. 6. Time sequence of instantaneous velocity fluctuation and vorticity distributions. The delay between adjacent frames is 0.3 s, and the series runs down the first column and then down the second. The corresponding sample area mean velocity components are shown by open circles in Fig. 4.



FIG. 7. Two sample series of vorticity maps combined to present the flow structure over extended areas. Adjacent frames are offset by the instantaneous mean velocity. Displacement in the cross-stream (y) direction is based on the out-of-plane velocity estimate.

lor's hypothesis. Clearly, there are significant variations between the four spectra with the differences between velocity components (E_{11} vs E_{33}) being significantly larger than the variations associated with the direction of integration. For comparison, Fig. 8b shows the equivalent one-dimensional energy spectra for the sample laboratory data of isotropic grid turbulence, which are calculated using the same procedure. In contrast to the field data (Fig. 8a), these spectra are essentially identical, indicating that, as expected, the turbulence is locally isotropic and our selected procedure does not introduce any bias into the results.

Because of the small effect of the direction of integration on the spectra, Fig. 9 only presents the distributions of $E_{11}(k_1)$ and $(3/4)E_{33}(k_1)$ for all the present cases. The 3/4 coefficient is added since in isotropic turbulence $E_{11}(k_1) = (3/4)E_{33}(k_1)$. Each figure also includes the same line with a -5/3 slope, which helps in comparing the energy levels. In addition, each graph contains an insert containing the distributions of



FIG. 8. Mean turbulent energy spatial spectra of (a) run D and (b) the sample laboratory data.



FIG. 9. Mean spatial energy spectra. Solid lines: $E_{11}(k_1)$; dashed lines: $E_{33}(k_1)$. Inset figures are spectra of $\varepsilon_{LF}^{-2/3} k_1^{5/3} E_{ii}(k_1)$. (a)–(f) Runs A–F, respectively.

 $\varepsilon_{\rm LF}^{-2/3} k_1^{5/3} E_{11}(k_1)$ and $(3/4) \varepsilon_{\rm LF}^{-2/3} k_1^{5/3} E_{33}(k_1)$, where ε is the dissipation rate. For the purpose of plotting these inserts we estimate the dissipation rate based a line fit to $E_{33}(k_1)$ in the range that it has a -5/3 slope. As will be shown in section 5, where we use different methods for estimating ε , because of the high level anisotropy at dissipation scales, this estimate is incorrect.

As is evident from Fig. 9, in all cases the turbulence is nearly isotropic at low wavenumbers ($k_1 < 40$ rad m⁻¹) but becomes increasingly anisotropic with increasing wavenumber in the high wavenumber range (50 rad m⁻¹ < $k_1 < 500$ rad m⁻¹). This highwavenumber anisotropy is most evident when the mean flow is weakest and the wave-induced oscillatory motion is dominant (runs E and F). Also, the anisotropy diminishes slightly with increasing elevation from the bed, at least for the LEO-15 data (runs C and E in comparison with runs D and F, respectively). In five of the six cases (A-E) the vertical velocity spectra have a range of wavenumbers of almost a decade with a slope close to -5/3. Conversely, the horizontal velocity spectra do not exhibit a range with -5/3 slope. Instead, they appear to have a "bump" with a maximum in the $k_1 =$ $100-300 \text{ rad m}^{-1}$ range. The magnitudes of these bumps are emphasized in the inserts showing the compensated spectra that have a linear vertical axis. They show that smaller but clear bumps exist also in the vertical velocity spectra, especially for the cases with low mean velocity (E and F). The existence of spectral bumps has been observed in several previous high Reynolds number turbulence measurements in the laboratory (e.g., Saddoughi and Veeravalli 1994; Saddoughi 1997) and in the atmosphere (e.g., Champagne et al. 1977). Recently, Kang et al. (2003) included extra parameters in the functional form of the spectra that they measured to account for the shape of the bumps. However, the present bumps appear to be larger than the previous results. We have also seen small bumps in previous ocean PIV measurements (Doron et al. 2001), but our laboratory PIV data of locally isotropic turbulence (Fig. 8b and, e.g., Liu et al. 1999) do not show such bumps. Upon seeing the bumps we carefully checked the data analysis procedures to ensure that they did not affect the shape of the spectra. In addition to testing the procedures using laboratory data (Fig. 8b), we also implemented several modifications to the analysis techniques, such as the choice of windowing functions (or lack of), interpolation, FFT length, and so on. We found that varying the procedures had very little effect on the shape of the bumps. Similarly, identical bumps were also found in truncated spectra calculated from data at the center of the sample areas, eliminating as a possible source the effects of out-of-plane motion, which has no effect at the center of the images (Nimmo Smith et al. 2002).

There have been a few attempts to explain the formation of the spectral bumps theoretically (e.g., Falkovich 1994; Lohse and Muller-Groeling 1995). They are attributed to a phenomenon defined as a "bottleneck" that occurs at the transition between the inertial and dissipative range of the turbulence spectrum. Qualitatively, this phenomenon is a result of viscous suppression of high wavenumber (small) eddies, which makes the transfer of energy from larger scales less efficient, causing a "pileup" of energy in eddies with scales at the transition between inertial to viscous range. Analysis of the time evolution of the correlation function (the energy spectrum is the Fourier transform of the pair correlation function), performed by Falkovich (1994), indeed shows that the occurrence of this phenomenon is an inherent effect of viscous suppression of small scale eddies.

The puzzling question is what affects the magnitude of the bumps and why are they more pronounced in some cases and hardly noticed in others (e.g., Gargett et al. 1984). Unfortunately, the answer to this question requires knowledge of the time evolution of the pair correlation function, which depends on higher-level correlations, that is, on nonlinear interactions between eddies of different scales. Thus, the spectrum depends on the entire time history and dynamical balances of the turbulence (i.e., production, dissipation, buoyancy flux, diffusion, advection). For example, it depends on whether the flow generates new turbulence locally (at the scale of energy containing eddies) or the turbulence is a decaying residue from production that occurred a while ago somewhere else. The differences between the present spectra may be a result of variations in their time history. When the current is relatively strong (Figs. 9a,b), the bump is less evident. Conversely, when the

local current is weak (Figs. 9e,f), the turbulence is not produced locally and has already decayed in part before reaching the sample area. Examination of the instantaneous distributions of cases E and F (e.g., Fig. 5a) reveal that the flow does not contain distinct large-scale structures, unlike the A–D cases (e.g., Figs. 5b–d) and almost all typical laboratory boundary layer flows. Finally, note that in the present data the bumps are much more evident in the streamwise velocity distribution, but that is not always the case. In the data of Saddoughi and Veeravalli (1994), for example, the bump is larger in the wall-normal component. We do not know how to explain the difference.

5. Dissipation-rate estimates

In Doron et al. (2001) we have shown that, even in cases with substantial turbulence anisotropy, integration of the dissipation spectrum and curve-fitting of a line with a -5/3 slope to the energy spectrum (in the domain that appears to have this slope) gives reasonable estimates to the dissipation rate ε . The existence of spectral bumps leads to questions and uncertainty in the estimates of ε based on the curve fit, although Saddoughi and Veeravalli (1994) show that at high Reynolds numbers (much higher than the present range, as shown later in this section) a curve fit in the wavenumber range that has a -5/3 slope still gives reasonable estimates. Since the spectra of the vertical velocity appear to have a more extended "inertial" range, we use $E_{33}(k_1)$ to estimate the dissipation rate, assuming that (Tennekes and Lumley 1972)

$$E_{33}(k_1) = \frac{3}{4} \frac{18}{55} 1.6\varepsilon_{\rm LF}^{2/3} k_1^{-5/3} \tag{1}$$

and denoting this estimated dissipation as ε_{LF} . A least squares fit is used in the range that appears to have the correct slope, and the results are summarized in Table 2.

Having data that extend to wavenumbers in the dissipation/viscous range, as the spectra clearly show, enables us to obtain estimates for the dissipation directly from the spatial derivatives of the velocity. Following the methodology of Doron et al. (2001), we utilize all the available measured velocity gradients, use the continuity equation to calculate $\partial u_2/\partial x_2$, and estimate the out-of-plane cross gradients as averages of the in-plane gradients. The resulting "direct" estimate of the dissipation rate, ε_D , is

$$\varepsilon_D = 3\nu \left[\left(\frac{\partial u}{\partial x} \right)^2 + \left(\frac{\partial w}{\partial z} \right)^2 + \left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial w}{\partial x} \right)^2 + 2 \left(\frac{\partial u}{\partial z} \frac{\partial w}{\partial x} \right) + \frac{2}{3} \left(\frac{\partial u}{\partial x} \frac{\partial w}{\partial z} \right) \right].$$
(2)

A center-differencing method is used for calculating the derivatives. Values of the direct estimate averaged over

TABLE 2. Dissipation rate estimates, Kolmogorov scale, and subgrid-scale dissipation for runs A–F and reference laboratory data. Note that for the ocean data the multiplier of dissipation rates is 10^{-7} , whereas for the laboratory data it is 10^{-4} .

Run	$\begin{array}{c} \overline{\epsilon}_{\rm LF} \times 10^7 \\ (m^2 \ s^{-3}) \end{array}$	$\frac{\overline{\epsilon}_D \times 10^7}{(\text{m}^2 \text{ s}^{-3})}$	η (mm)	δ (mm)		$\overline{arepsilon}_{\partial z} imes 10^7 \ (\mathrm{m^2~s^{-3}})$	$ \overline{ \overline{ \epsilon}_{SG} \left(8 \delta \right) \times 10^7 } \\ (m^2 s^{-3}) $
A	217	109	0.71	7.84	44.6	100	30.8
В	307	127	0.68	7.84	58.9	117	185
С	7.65	14.7	1.17	5.41	6.74	12.7	0.59
D	5.68	15.0	1.16	5.41	6.87	13.1	1.10
E	1.92	8.78	1.33	5.41	3.52	7.65	0.09
F	1.47	8.20	1.35	5.41	2.98	7.18	-0.05
$Lab \times 10^4$	24.2	8.47	0.24	1.56	7.91	8.23	8.78

all the data (all points and all vector maps—denoted by overbars) are presented in Table 2. Note that the vector spacing, δ , in our PIV data is still larger than the Kolmogorov scale, η , estimated using

$$\eta = (\nu^3 / \overline{\varepsilon_D})^{1/4}, \tag{3}$$

where ν is the kinematic viscosity of the water (= 1.4×10^{-6} m² s⁻¹), shown in Table 2. In runs A and B, our resolution is 10η , in runs C and D it is about 5η , and in runs E and F it is 4η . Consequently, we underestimate the dissipation rate. Since $\partial u/\partial z \propto k\sqrt{E(k)}$ and assuming a $k^{-5/3}$ slope we obtain $\partial u/\partial z \propto k^{1/6}$. Substituting the present values suggests that we underestimate the actual dissipation by 45%, 30%, and 26% for cases A–B, C–D, and E–F, respectively.

Since typically used oceanographic instruments, for example, airfoil probes, estimate the dissipation based on one (or at most two) velocity gradient(s), assuming isotropy,

$$\varepsilon_{\partial x} = \frac{15}{2} \nu \left(\frac{\partial w}{\partial x}\right)^2 \text{ and } \varepsilon_{\partial z} = \frac{15}{2} \nu \left(\frac{\partial u}{\partial z}\right)^2.$$
 (4)

Table 2 also presents averaged data that would be obtained from such measurements. The effect of coarse spatial resolution on the data also applies to these results.

PIV data can also be used for estimating the "subgrid-scale dissipation" or energy flux (Liu et al. 1994). In large-eddy simulations, the Navier Stokes equations are filtered spatially to give

$$\frac{\partial \tilde{u}_i}{\partial t} + \tilde{u}_j \frac{\partial \tilde{u}_i}{\partial x_j} = -\frac{\partial}{\partial x_j} \left(\frac{\tilde{p}}{\rho} \,\delta_{ij} + \tau_{ij} \right) + \nu \frac{\partial^2 \tilde{u}_i}{\partial x_j \partial x_j}, \quad (5)$$

where the tilde indicates spatial filtering over a scale Δ and τ_{ii} is the subgrid-scale stress tensor

$$\tau_{ij} = \widetilde{u_i u_j} - \widetilde{u}_i \widetilde{u}_j, \tag{6}$$

introduced as a result of the spatial filtering. When the spatial filter is in the inertial subrange of isotropic homogeneous turbulence, the mean ensemble-averaged subgrid-scale dissipation, ε_{SG} , is (almost) equal to the total viscous dissipation rate (Lilly 1967). Thus, another estimate for the dissipation rate is

$$\overline{\varepsilon_{\rm SG}} = \overline{-\tau_{ij}\tilde{S}_{ij}},\tag{7}$$

where $\tilde{S}_{ij} = 0.5(\partial \tilde{u}_i/\partial x_j + \partial \tilde{u}_j/\partial x_i)$ is the filtered strain rate and overbar indicates ensemble average. This agreement has been observed in ocean data with a mean current significantly larger than the amplitude of wave-induced motion (Doron et al. 2001). However, Piomelli et al. (1991) show that near the wall of channel flows at low Reynolds numbers, the mean ε_{SG} is small and can even be negative.

To evaluate ε_{SG} from PIV data one needs to make assumptions on the missing out-of-plane components. Following Liu et al. (1994), we assume that the missing terms involving products of shear stresses and strains are equal to the available measured values ($\tau_{13}\tilde{S}_{13}$), and that the missing normal stress terms are equal to averages of the available normal stress components; that is, $\tau_{22} = 0.5(\tau_{11} + \tau_{33})$. The missing normal strain term can be determined using the continuity equation; that is, \tilde{S}_{22} = $-(\tilde{S}_{11} + \tilde{S}_{33})$. The resulting estimate for the subgridscale (SGS) dissipation is

$$\varepsilon_{\rm SG} = -\tau_{ij}\tilde{S}_{ij} \approx -\frac{1}{2}(\tau_{11}\tilde{S}_{11} + \tau_{33}\tilde{S}_{33} - \tau_{11}\tilde{S}_{33} - \tau_{33}\tilde{S}_{11} + 12\tau_{13}\tilde{S}_{13}). \tag{8}$$

This expression deviates slightly from the one used in Doron et al. (2001) because of differences in the methods used for estimating the unknown, out-of-plane contributions. The differences are not substantial since the dominant term still involves the shear stress. Note that, unlike viscous dissipation, the instantaneous ε_{SG} can be either positive or negative. A positive value indicates flux of energy from large to small scales whereas a negative value indicates "backscatter" of energy from small to large scales.

Sample instantaneous distributions comparing ε_D , $\varepsilon_{\partial x}$, and $\varepsilon_{\partial z}$ of the same data (the velocity field shown in Fig. 5c) are presented in Figs. 10a–c, respectively. Clearly there are significant spatial variations, although the dissipation peaks are associated with the large coherent vortices in all three cases. However, when the averaged values for the entire datasets are compared (see Table 2), $\overline{\varepsilon_D}$ and $\overline{\varepsilon_{\partial z}}$ are consistently very close to each other (supporting the validity of results that would



FIG. 10. Sample instantaneous distributions of dissipation rate for the velocity field shown in Fig. 5c: (a) ε_{D} , (b) ε_{ax} , (c) ε_{az} , and (d) ε_{SG} . For (a)–(c), the contour levels are (0.1, 0.3162, 1, 3.162) × 10⁻⁵ m² s⁻³. For (d), the contour levels are at intervals of 4 × 10⁻⁶ m² s⁻³, negative values are shaded gray, and the transition is the zero level contour.

be obtained from vertical profiling), whereas $\overline{\varepsilon_{ax}}$ is consistently 55%–64% smaller than $\overline{\varepsilon_D}$. As illustrated in Fig. 11, which shows a time series of the dissipation rates averaged over the instantaneous distributions using the run D data, these trends are consistent for the entire data; that is, they are not caused by specific events. The other time series have similar trends. As a reference, Table 2 also provides laboratory data of locally isotropic turbulence generated using four symmetrically positioned rotating grids (Liu et al. 1999; Friedman and Katz 2002). For the laboratory data $\overline{\varepsilon_D}$, $\overline{\varepsilon_{dx}}$ and $\overline{\varepsilon_{dz}}$ are essentially the same. Thus, the consistently lower values of $\overline{\varepsilon_{dx}}$ seem to be associated with the anisotropy of the oceanic boundary layer turbulence.

Figure 10d shows the instantaneous distribution of ε_{SG} for the same velocity field calculated using an 8 δ

box filter. Regions of both positive and negative SGS dissipation are clearly evident. The most intense positive and negative SGS dissipation peaks appear regularly in close proximity to the large-scale coherent vortices. This trend indicates that the evolution of the large vortices involves both forward and backscatter of energy. The ensemble averaged values of SGS dissipation, shown in Table 2, are somewhat surprising (at least at first glance). For the high flow-rate series (runs A and B), $\overline{\varepsilon_{SG}}$ is of the same order as $\overline{\varepsilon_D}$, especially for the near-bed data series (run B), which is consistent with our previous field observations (Doron et al. 2001) and laboratory observations (Liu et al. 1994). However, when the mean flow is moderate (runs C and D) or weak (runs E and F), $\overline{\epsilon_{SG}}$ is one or even two orders of magnitude smaller than $\overline{\varepsilon_D}$. These low values are con-



FIG. 11. Time series of the sample area mean dissipation rates for run D. Solid line: ε_D , dashed line: ε_{ax} , and dotted line: ε_{az} .

sistent with the results of direct numerical simulations of boundary layer (channel) flows at low Reynolds numbers (Piomelli et al. 1991), where the values of $\overline{\varepsilon_{SG}}$ have been found to be very low and in part negative, especially near the wall. Since typical eddy viscosity models for the SGS stresses attempt to predict the correct magnitude of energy flux to subgrid scales, the trends of $\overline{\varepsilon_{SG}}$ and the marked differences from $\overline{\varepsilon_D}$ in the moderate and weak flow conditions have significant implications on modeling these types of flow using LES.

The trends of $\overline{\epsilon_{LF}}$ also differ significantly from the direct estimates, illustrating the uncertainty associated with using curve fitting in anisotropic turbulence. For the high-mean-flow runs (A and B) and for the isotropic turbulence laboratory data the curve-fitted results are 2–3 times $\overline{\epsilon_D}$, consistent with the fact that $\overline{\epsilon_D}$ is calculated using underresolved data. Note that $\overline{\epsilon_D} < \overline{\epsilon_{LF}}$ also in the Doron et al. (2001) data but not by the same ratio. The trends are reversed in runs C–F. When the mean flow is comparable to the wave amplitude (C and D), $\overline{\epsilon_D}$ is 2–3 times $\overline{\epsilon_{LF}}$ and, when the mean flow is very low (E and F), $\overline{\epsilon_D}$ is 4.5–6.5 times $\overline{\epsilon_{LF}}$. Considering that $\overline{\epsilon_D}$ is underresolved, the results indicate that dissipation estimates based on curve fitting to the spectra for cases C–F lead to wrong results.

Based on the estimated dissipation one can characterize the turbulence using the Taylor microscale Reynolds number

$$\operatorname{Re}_{\lambda}^{w} = \frac{w'\lambda}{\nu}; \lambda^{w} \approx w' \left(\frac{15\nu}{\overline{\varepsilon}}\right)^{1/2} \quad \text{and} \\ \operatorname{Re}_{\lambda}^{u} = \frac{u'\lambda}{\nu}, \lambda^{u} \approx u' \left(\frac{15\nu}{\overline{\varepsilon}}\right)^{1/2}.$$
(9)

To estimate u' and w', the rms values of velocity fluctuations, without being contaminated by waves, one can use the second-order structure function, a method introduced by Trowbridge (1998) and implemented using PIV data in Nimmo Smith et al. (2002). The results are summarized in Table 3. In runs A and B Re_{λ} is in the 300-400 range, which would qualify as high turbulence level, yet the dissipation rate is still almost two orders of magnitude lower than the laboratory conditions. The turbulent Reynolds numbers of the LEO-15 site are 68-83 when the mean flow is moderate and 14-27 when the mean current is weak. Thus, the turbulence in the coastal bottom boundary layer under typical calm weather conditions can have a very low Taylor microscale Reynolds number. The causes for the low Re_{λ} seem to involve both the confined length scales near the bottom boundary layer and the low velocities. This conclusion would not apply farther away from the boundary if the low velocities are compensated by very large scales or if the current speed is higher. As discussed in Pope (2000), the present moderate and low flow cases fall in the range of Re_{λ} where many of the assumptions of universality of the energy spectrum become invalid.

Run	А	В	С	D	Е	F	Lab
$u' (cm s^{-1})$	1.82	2.20	0.55	0.55	0.33	0.28	6.1*
w' (cm s ⁻¹)	1.80	1.86	0.55	0.50	0.26	0.20	4.3*
λ^{u} (cm)	2.5	2.8	2.1	2.1	1.6	1.4	0.57**
λ^{w} (cm)	2.5	2.4	2.1	1.9	1.3	1.0	0.40**
$\operatorname{Re}_{\lambda}^{u}$	325	440	83	83	37	28	248
$\operatorname{Re}_{\lambda}^{\hat{w}}$	321	318	83	68	24	14	122
$\partial \overline{U}/\partial z$ (s ⁻¹)	0.040	0.132	0.017	0.027	0.007	0.004	0.175
$S^* = \hat{S}(\nu/\varepsilon)^{1/2}$	0.014	0.044	0.017	0.026	0.009	0.005	0.004

TABLE 3. Rms values of turbulent velocity fluctuations estimated using second-order structure functions.

* Measured directly.

** Utilizes $\overline{\varepsilon_{LF}}$, because of effect of discrepancy in length scales on direct estimates.

These low Re_{λ} explain the low values of $\overline{\varepsilon_{SG}}$ in comparison with $\overline{\varepsilon_D}$ and their agreement with trends observed in direct numerical simulations at low Reynolds numbers. A future paper will focus on the implication of the low $\overline{\varepsilon_{SG}}$ on subgrid-scale modeling in LES of oceanic flows.

Table 3 also provides the distributions of overall mean shear, \overline{S} , and the ratio between the Kolmogorov time scale and the time scale of the mean flow, $S^* = \overline{S}(\nu/\overline{\epsilon})^{1/2}$. For all present cases, S^* is very small, satisfying the Corsin–Uberoi condition (Corsin 1958; Uberoi 1957; Saddoughi and Veeravalli 1994) for local isotropy in shear flows. Thus, the mean shear in the bottom boundary layer is not a major contributor to the observed substantial anisotropy.

6. Conditional sampling

a. Effects of wave-induced motion on the turbulence spectra

Conditional sampling based on different criteria has been used for characterizing their effects on the turbulence in the bottom boundary layer. First, data within each series were sampled according to their phase in the wave-induced oscillatory motion. The time series of the instantaneous sample area mean horizontal velocity (Fig. 3) were used to group individual frames into those where the oscillatory motion was maximum, minimum, accelerating, and decelerating. The spectral analysis described in the preceding section was then repeated on each of these four groups and for each of the data series. Figure 12 shows, as an example, the conditionally sampled mean spatial spectra of $E_{11}(k_1)$ and $(3/4)E_{33}(k_1)$, based on the phase of the wave induced motion for run



FIG. 12. Mean spatial energy spectra for run D conditionally sampled based on the phase of the orbital wave horizontal velocity. Solid lines: $E_{11}(k_1)$, dashed lines: $E_{33}(k_1)$, crosses: minimum flow, dots: accelerating flow, squares: maximum flow, and triangles: decelerating flow.

D. As is evident, for both $E_{11}(k_1)$ and $(3/4)E_{33}(k_1)$, the four spectra overlie each other with the small variations at low wavenumbers attributable to the smaller database contributing to each mean spectra. The conditionally sampled velocity spectra for all other data series (not shown) also show this same trend. Clearly, the phase within the wave-induced motion cycle has insignificant impact on the turbulence statistics, at least at the present range of elevations and flow (bottom) conditions. Very close to the bottom, one might expect the turbulence to be linked to the wave-induced motion, presumably due to phase-dependent variations in the shedding of eddies from bedforms.

This observation that the wave-induced motion has very little impact on the turbulence, even a short distance from the seabed, is consistent with estimates of the ratio of the time scales of the turbulence and the wave-induced strain rate. For the small-scale turbulence, this ratio can be defined as $S_{Sw}^* = S_{11}^w (\nu/\overline{\epsilon})^{1/2}$, where S_{11}^w is the peak wave-induced strain, which can be estimated from a modeled velocity field for 10-s waves, the typical period of the present data. At the integral scale, the ratio can be defined as $S_{Lw}^* = S_{11}^w K/\overline{\epsilon}$, where *K*, the turbulent kinetic energy, is estimated from our 2D data as $K = (3/4)(u'^2 + w'^2)$ (i.e., by assuming that the missing out-of-plane component is an average of the in-plane components). Even using an overestimate for S_{11}^w , the values of S_{Sw}^* and S_{Lw}^* are at least an order of magnitude below the threshold (~ 1) for rapid straining of the turbulence to be a factor.

b. Effect of gusts on the spectra and Reynolds stresses

The second criterion for conditional sampling focuses on the observed appearance of intermittent gusts. We have explored a variety of criteria to identify the gusts and obtained the most reliable results using the mean vorticity magnitude. Figure 13 compares time series of the instantaneous mean sample area vorticity magnitude for the three near-bed data series. All three sets are normalized by the mean vorticity magnitude in run D ($|\omega_D| = 0.17 \text{ s}^{-1}$), enabling comparisons among runs. Clearly, the vorticity magnitude in the high flow series (run B, Fig. 13a) is substantially higher than the typical values at moderate (run D, Figure 13b) and low flow (run F, also Fig. 13b) series, consistent with our observation that the high flow series consist almost entirely of large vortical structures. The variability within the high flow series is due to the passage of particularly intense vortices through the sample area. The moderate flow series consists of a low-level background with intermittent high vorticity events (examples indicated by X). The two events in the 500-600-s range correspond to the advection of the structures shown in Fig. 7 through the sample area. The very low background level of the low flow data series features only a single high-vorticity-magnitude event, indicated by Y in Fig. 13b. Examination of the relevant velocity distributions



FIG. 13. Time series of sample area mean vorticity magnitude. The data are normalized by the ensemble mean vorticity magnitude of run D. (a) Run B. (b) Black: run D; light gray: run F. Example "gust" events are indicated by X. The obvious "spike" in run F (Y) is probably a fish wake.

reveals that this event consists of a pair of counterrotating vortices, leading us to suspect that it is most probably associated with the wake of a fish or squid. Both were observed in the region by the other submerged video cameras. Consequently, the 70 affected frames have been omitted from all of the analyses.

Probability distributions of the instantaneous (vector map) mean and fourth-order moment of the vorticity magnitude for each of the six data series are shown in Figs. 14a and 14b, respectively. The fourth-order moment are presented because they emphasize the high vorticity "tails" associated with gust events. There is very little difference between the two low flow distributions, except for the aforementioned anomalous tail with high values in the near-bed data, which is particularly obvious in the distribution of the fourth-order moment. The two distributions of the moderate flow series overlie one another for the most part, and both have large positive tails associated with the gust events. However, as the distributions of fourth-order moment highlight, the frequency of high gust events is slightly higher near the bottom. In contrast, the two high flow distributions show marked differences from each other, in addition to being substantially offset from the other distributions. The positive tail of the near-bed data series (run B) is much larger than that of the data ob-



FIG. 14. PDFs of (a) the sample area mean vorticity magnitude and (b) the fourth-order central moment of the vorticity magnitude. Open diamonds: run A, closed diamonds: run B, open circles: run C, closed circles: run D, open triangles: run E, and closed triangles: run F. All PDFs have been normalized by the appropriate ensemble mean value of run D.

tained higher in the water column (run A), indicating that the near-bed flow contains many more large-scale vortical structures.

For the subsequent analyses, we have opted to differentiate the flow into categories of high, intermediate, and low instantaneous mean vorticity magnitude. The thresholds are arbitrarily set at

$$\overline{|\omega_y|} \pm \frac{1}{2} \sigma_{|\omega_y}^2$$

where $|\omega_y|$ is the mean vorticity magnitude of an entire data series (over all points and all measurements), and $\sigma_{|\omega_y|}^2$ is its variance. Figure 15 shows the conditionally sampled mean spatial velocity spectra. In



FIG. 15. Mean velocity spectra, conditionally sampled based on the vorticity magnitude. (a)–(f) Runs A–F, respectively. Open symbols: $E_{11}(k_1)$, closed symbols: $E_{33}(k_1)$, squares: high vorticity, circles: intermediate vorticity, and triangles: low vorticity. The solid line has a slope of -5/3 and is located in the same position within each graph.

all cases, the turbulence remains anisotropic at high wavenumbers, regardless of the vorticity magnitude. The spectral bumps still remain in evidence, although they become a bit less obvious in the high vorticity spectra due to an increase in energy level at low wavenumbers (e.g., Fig. 15c). Differences in trends associated with vorticity magnitude occur mostly at low wavenumbers. During high vorticity events the spectra of streamwise and vertical velocity components converge at low wavenumbers and, conversely, the turbulence becomes more anisotropic during periods of low vorticity. The convergence of spectra at low wavenumbers is consistent with the intermittent passage of large, quite circular vortices during the high vorticity periods.

Conditionally sampled vorticity spectra are shown in Fig. 16. These spectra are calculated directly from the instantaneous vorticity distributions, again using the central seven rows (and columns) to increase the sample size. We provide results of integration in the horizontal (k_1) and vertical (k_2) directions, which, unlike the velocity spectra, differ significantly from each other. Conditional sampling affects the vorticity spectra only at low wavenumbers ($k_i < 100$), where the amplitude increases with vorticity magnitude. This trend is consistent with the observed intermittent passage of large vortices during gust periods. Conversely, there is very little variability at high wavenumbers. The strength of small scale eddies during gust periods remains at the same level as those existing in the background weak turbulence that comes into view between gusts.

However, there is a striking difference between the horizontal (k_1) and vertical (k_3) spectra, which is particularly evident in the low flow data series (runs E and F). There is a clear peak in $\Phi(k_3)$ at intermediate wavenumbers $(k_3 \approx 200 \text{ rad m}^{-1})$, while $\Phi(k_1)$ has a broader maximum at lower wavenumbers ($k_1 < 80 \text{ rad m}^{-1}$). Thus, there is significantly greater vertical variability at intermediate scales compared to the horizontal direction. The physical manifestation of this phenomenon is illustrated in Fig. 17 by low-pass and bandpass filtering the vorticity distribution of the velocity field shown in Fig. 5a, which is typical of this data series (run F). The data are calculated using "box filters" on the same velocity distribution with sizes corresponding to k = 127and k = 381 rad m⁻¹. The first operation provides the low-pass filtered velocity (k < 127), and the difference between them provides the bandpass filtered vorticity (127 < k < 381). Clearly, the large-scale vorticity structure (Fig. 17a) has little obvious orientation. In contrast, the intermediate-scale structures (Fig. 17b) appear to be organized in layers, that are only slightly inclined to the horizontal direction. The same phenomenon seems to recur frequently throughout the samples of instantaneous realizations that we have examined. Consequently, the vertical vorticity spectra $[\Phi(k_3)]$ are more energetic than the horizontal spectra $[\Phi(k_1)]$ at intermediate scales, but the difference between them diminishes at small scales. Thus, the background lowlevel turbulence, which dominates the low-flow conditions, and appears between gusts at intermediate and high flows, consists of slightly inclined, almost horizontal series of thin shear layers.

The conditionally sampled, mean direct dissipation rates are shown in Table 4. Unlike the large differences (up to an order of magnitude) between data series, the variability associated with vorticity magnitude is smaller than 40%. Since dissipation is dominated by small-scale eddies, this trend is entirely consistent with the small differences at high wavenumbers between the conditionally sampled velocity spectra. Clearly, the intermittent passage of large-scale vortical structures has a weak effect on the local dissipation rate, which is dominated by the more uniform distributions of the background small-scale turbulence. In stark contrast to this finding are the conditionally sampled Reynolds shear stresses, or the distributions of the second-order structure function, shown in Fig. 18. Clearly, there are marked differences both between data series as well as between the conditionally sampled distributions within each series. Under low mean flow conditions (run E and F) D_{13} (r) remains effectively zero; that is, the Reynolds shear stress is zero, irrespective of vorticity magnitude (which is always low). When the flow is strong and near to the seabed (run B), all of distributions continue to increase together with increasing separation, indicating that the integral scale is larger than the largest measurable separation within a single sample area. Here, again, there is no difference between the conditionally sampled distributions, showing that all states of the flow (high, intermediate, and low vorticity magnitude) contribute to the stress. This trend is consistent with the observation that the velocity distributions in run B contain a near-continuous supply of large-scale vortices. Conversely, farther from the seabed (run A), when the gust events are already separated by quiescent periods, there are significant differences between the conditionally sampled structure functions. Here, the contributions to the stress are made almost entirely by the high and intermediate vorticity magnitude events, while the low vorticity regions do not generate any shear stress. In the moderate flow data series (runs C and D) only the high vorticity magnitude events contribute significantly to the shear stress. Due to the low frequency of gusts at moderate flows (Figs. 13 and 14), essentially all events with large vortex structures are classified as the high vorticity events. Thus, shear stresses, and as a result shear production, are generated only in flow domains containing large vortex structures, that is, during periods of gusts. Dissipation, on the other hand, occurs continuously, and increases only slightly during high vorticity periods.

7. Summary and concluding remarks

Six datasets obtained using PIV measurements (and one reference laboratory set) are used for examining



FIG. 16. Mean vorticity spectra conditionally sampled on the vorticity magnitude. (a)–(f) Runs A–F, respectively. Open symbols: $\Phi(k_1)$, closed symbols: $\Phi(k_3)$, squares: high vorticity, circles: intermediate vorticity, and triangles: low vorticity.

the structure and properties of turbulence in the bottom boundary layer of the coastal ocean. The ocean datasets are selected to represent conditions of high, moderate, and weak Taylor microscale turbulent Reynolds numbers, close to the bottom (in the 20–70-cm range) and far from it (~2.5 m). The corresponding Re_{λ} are 300–440, 68–83, and 14–37. Accordingly, the mean currents are significantly higher, of the same order, and much weaker than the wave-induced motions. The present moderate–weak conditions are typical to the LEO-15 site, indicating that typical turbulence in the coastal bottom boundary layer has a very low Taylor microscale Reynolds number. Although Re_{λ} of the laboratory reference data falls within the range of the oceanic data, the dissipation rate and length scales are substantially different. Furthermore, the laboratory



FIG. 17. Sample instantaneous distributions of vorticity calculated from filtering the velocity field shown in Fig. 5a: (a) k < 127 rad m⁻¹ low-pass filter, contoured at intervals of 0.015 s⁻¹; (b) 127 rad m⁻¹ < k < 381 rad m⁻¹ bandpass filter, contoured at intervals of 0.05 s⁻¹. In both figures, negative values are shaded gray and the transition is the zero level contour.

flow conditions can be classified as locally isotropic turbulence with a clearly identified inertial range, whereas the turbulence in the bottom boundary layer is anisotropic, particularly at small scales, for all of the present test conditions.

Examination of instantaneous distributions of velocity and vorticity show that, when the mean flow is weak, the flow consists solely of small-scale turbulence. At times of moderate flow these similarly quiescent periods are interspersed by intermitted groups of largescale vortical structures. These structures appear to be similar to the characteristic packets of hairpin vortices observed in the laboratory (e.g., Adrian et al. 2000) and in numerical simulations (Zhou et al. 1999). When the flow rate is high, the flows become dominated by these large-scale vortices, particularly close to the seabed. Studies by Nimmo Smith et al. (1999) have highlighted the importance of depth-scale coherent structures to the dynamics of coastal waters, particularly in the distribution of sediment and the dispersion of surface pollutants. It is possible that these structures may have grown from hairpinlike groups similar to those observed in the present measurements.

The first striking characteristic of the present ocean turbulence data is the level of anisotropy. Consistent with typical laboratory boundary layers, and without exception, all present spectra of the streamwise velocity component are higher than those of the vertical component. This anisotropy holds irrespective of the wave-number component; that is, $E_{11}(k_1) > (3/4)E_{33}(k_1)$ and $(3/4)E_{11}(k_3) > E_{33}(k_3)$. All of the spectra, but especially the streamwise component, have large bumps at the transition from the inertial to the dissipation range that

increase in size with decreasing Re_{λ} . These bumps have been seen before in some laboratory and atmospheric data (e.g., Saddoughi and Veeravalli 1994; Saddoughi 1997; Champagne et al. 1977). Theoretical analyses (e.g., Falkovich 1994) have shown that they are caused by a bottleneck effect as the slope of the energy spectrum changes from the inertial to dissipation levels. We have also observed smaller but clear bumps in previous oceanic near-bottom measurements performed in the New York Bight (Doron et al. 2001). On the other hand, they do not appear in measurements performed in the ocean away from boundaries (Gargett et al. 1984). Unfortunately, there are as yet no useful tools (theoretical or empirical) to determine or predict their occurrence and magnitudes as they depend on the time history of the turbulence. Furthermore, since the values of Re_{λ} in the moderate to weak flow conditions are significantly lower than 100, many of the universal behavior characteristics of high Reynolds number turbulence in the inertial range would not hold, even if the turbulence was isotropic (Pope 2000).

Several methods for estimating the dissipation rates

TABLE 4. "Direct" dissipation estimates, $\overline{\epsilon_D} \times 10^7$ (m² s⁻³), conditionally sampled based on the vorticity magnitude.

Run	High	Intermediate	Low
А	123	121	116
В	159	126	107
С	17.6	14.7	12.6
D	17.4	15.0	13.0
E	11.1	8.79	6.71
F	10.4	8.06	6.06



FIG. 18. Distributions of $-1/2D_{13}$ as a function of horizontal separation (r_1). (a)–(f) Runs A–F, respectively. Squares: high vorticity; circles: intermediate vorticity; triangles: low vorticity.

are examined, including "direct" estimates that rely on all the available data (ε_D), estimates based on gradients of one velocity component ($\varepsilon_{\partial x}$ and $\varepsilon_{\partial z}$), estimates based on curve fitting to the energy spectrum (ε_{LF}) in a range with -5/3 slope, and estimates of the SGS dissipation (ε_{SG}). Since the vector spacing in the PIV data is larger than the Kolmogorov scale, estimates based on velocity gradients are lower than the actual values. Although the instantaneous distributions vary, the averaged values of ε_D and $\varepsilon_{\partial z}$ agree and follow the same trends. The magnitudes of $\varepsilon_{\partial x}$ are typically 50% smaller but they still follow the same trends as ε_D . Estimates of dissipation based on line-fit to the energy spectrum lead to mixed results. At low Re_{λ}, the streamwise velocity spectra do not have a wavenumber range with -5/3slope, whereas under moderate and high Reynolds numbers the range is very limited. The vertical components have identifiable ranges with a -5/3 slope. Dissipation estimates based on curve fits to the vertical component in the high Re_{λ} case are higher than ε_D , consistent with the measurements being underresolved. However, in the moderate and weak Re_{λ} cases, ε_{LF} is 1/2 and 1/4 of ε_D , respectively. This trend and the associated substantial anisotropy raise serious questions on the validity of estimating the dissipation rate based on a curve fit to the energy spectrum, even if part of one spectrum appears to have a wavenumber range with -5/3 slope.

In the high Re_{λ} conditions $\overline{\epsilon_{SG}}$ is close in magnitude to the viscous dissipation, consistent with the laboratory data [and the results of Doron et al. (2001)]. However, in the moderate and weak conditions the SGS dissipation is much smaller, by more than an order of magnitude, than the viscous dissipation rate. Such discrepancies should not be surprising since the low turbulence level and anisotropy imply that the underlying assumptions are invalid. Substantial discrepancies between SGS dissipation and viscous dissipation have also been observed in direct numerical simulations (DNS) of a boundary layer in a channel flow (Piomelli et al. 1991).

Conditional sampling of the data, based upon the phase of the wave-induced motion, shows very little dependence of the turbulence spectra on the phase within the wave cycle. This result is consistent with the fact that the time scales associated with the turbulence are substantially shorter than those associated with the wave-induced strain. In contrast, conditional sampling based on the vorticity magnitude highlights the importance of the large-scale coherent structures. We find that large-scale coherent vortical structures are key to the transfer of energy from the mean flow into the turbulence. The Reynolds shear stresses (and as a result the shear production) are generated only in flow domains containing large vortex structures, that is, during periods of gusts. When the large structures are not present, that is, when the mean flow is weak or during quiescent periods of moderate flow, the shear stresses are essentially zero. Dissipation, on the other hand, occurs continuously, and increases only slightly during high vorticity periods.

While the data presented in this paper are limited to calm weather conditions, they can still be considered typical of coastal waters with weak to moderate currents. It is obvious that further observations are required to expand our knowledge base to more extreme conditions, as well as to sites with stronger currents and with different bottom topography. Further observations are also required to fully investigate the origin, dynamics, and impact of coherent vortical structures, including their role in mixing and transport processes near the seabed.

Acknowledgments. This project has been funded in part by NSF under Grant OCE9819012 and in part by

ONR under Grant N00014-95-1-0215. We are grateful to Y. Ronzhes and S. King for their technical expertise in development of the equipment; the captain and crew of the RV *Cape Henlopen*; and D. R. Baldwin, L. Luznik, W. Zhu, A. Fricova, and T. Kramer for their invaluable assistance during deployments.

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