ESTIMATING BOTTOM STRESS IN TIDAL BOUNDARY LAYER FROM ACOUSTIC DOPPLER VELOCIMETER DATA

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ABSTRACT: Bed stresses in the bottom boundary layer of the York River estuary, Va., were estimated from 3D near-bottom velocities measured by Acoustic Doppler Velocimeters (ADVs) and also by a profiling array of electromagnetic current meters. By assuming the measurements were made in a constant stress layer, four methods of stress estimation were evaluated using ADVs: (1) direct covariance (COV) measurement; (2) turbulent kinetic energy; (3) inertial dissipation utilizing the Kolmogorov spectrum; and (4) log profile. The four methods yielded similar estimates of frictional velocity \( U_\tau \) based on ADV output from both 14 and 44 cm above bed. All eight estimates of average \( U_\tau \) were consistent with the overall mean of 1.10 cm/s to within the 95% confidence interval for individual burst estimates. The COV method worked slightly better nearer the bed, possibly because of the sensitivity of COV to the upper limit of the constant stress layer. The inertial dissipation method performed marginally well at 14 cm above bed, likely due to sediment induced stratification and insufficient separation of turbulent production and dissipation scales. The log profile method was the most variable and appeared most sensitive to stratification and to the thickness of the constant stress layer. The turbulent kinetic energy method was the most consistent at both heights and appears most promising for further development. Results encourage future use of the ADV in estuarine environments but also favor the simultaneous use of several methods to estimate bottom stress.

INTRODUCTION

Bottom stress and the near-bed structure of turbulence play fundamental roles in the nature of sediment transport in tidal estuaries. Bottom stress controls the erosion and deposition of sediments at the bed as well as their diffusion in the water column. Bottom friction velocity \( U_\tau \) and bed stress \( \tau_b \) are related as \( \tau_b = \rho U_\tau^2 \), where \( \rho \) is the density of water. In the logarithmic portion of a current bottom boundary layer, turbulence and mean flow depend on the friction velocity \( U_\tau \) and the distance from the bed \( z \) (Gross and Nowell 1985), where \( U_\tau \) reflects the velocity scale of turbulence carrying eddies and \( z \) constrains the maximum possible eddy size. A proper estimation of bottom friction has challenged the study of tidal boundary layers [e.g., Green (1992)].

Recent advances in oceanographic instrumentation have greatly expanded the sophistication with which bottom stress can be estimated in an estuary. Until recently, electromagnetic current meters (EMCMs) were among the best instrumentation available for studying the structure of estuarine bottom boundary layers. The EMCMs are sturdy, resistant to fouling, moderately unobtrusive, and reasonably inexpensive. Unfortunately, they also suffer from severe limitations including zero-offset drift, limited frequency response, and relatively large sampling volumes. Nonetheless, careful measurements of velocity structure and turbulent properties associated with bottom stress have been accomplished in estuarine boundary layers with the EMCMs. Techniques applied to near-bed EMCM measurements in estuaries include direct measurement of velocity covariance associated with Reynolds stress [e.g., French and Clifford (1992)], application of the inertial dissipation (ID) method based on the Kolmogorov spectrum [e.g., Green (1992)], and fits to theoretical velocity profiles [e.g., Sanford et al. (1991)].

Within the last few years, acoustical instruments have become increasingly available to estuarine scientists. These instruments, like the EMCMs, are relatively sturdy, resistant to fouling, and increasingly affordable. In addition, acoustic techniques are less intrusive and have the potential to provide sampling frequencies and sampling volumes to finer extremes. A few of the most promising acoustic instruments presently being applied to estuarine bottom boundary layers include the Benthic Acoustic Stress Sensor (BASS), the broadband Acoustic Doppler Current Profiler (ADCP) and lately the Acoustic Doppler Velocimeter (ADV) (field model manufactured by Sontek, San Diego, Calif.). The BASS has been used repeatedly with success in shelf and deep sea environments [e.g., Gross et al. (1992)] and has more recently been used to study estuarine bottom boundary layer processes in the Satilla River in Georgia (T. Gross, personal communication). The Modular Acoustic Velocity Sensor is an improvement of BASS with better cosine response time in vertical velocity and lower noise, and it is able to measure flow within the lowest few millimeters of the bottom boundary layer (S. Williams, personal communication). The broadband ADCP has been used in various configurations to examine turbulence and near-bed velocity structure in several estuaries. Among the most relevant studies are those under way in San Francisco Bay, where the broadband ADCP is being used to resolve fine near-bed current structure and also to examine turbulent structure throughout the water column (Cheng et al. 1999).

The ADV has a small sampling volume of 85–255 mm³ (a cylinder, 3–9 mm in length and 6 mm in diameter, with distance of 150 mm from the sensor head). The sampling volume is an order of magnitude smaller than that of the EMCM and also significantly smaller than that sampled by the broadband ADCP, allowing the ADV to be less affected by severe attenuation in the variance of vertical velocities. It also extends frequency response, because the maximum useful sampling rate is inversely related to sensor dimension [e.g., Soulsby (1980)]. The ADV has been previously shown to measure successfully 3D fluid velocity at a single point for both uniform and oscillatory water flows (Kraus et al. 1994). The ADV-measured mean flows and turbulences compared favorably with other independent measurements from a laser Doppler velocimeter in a laboratory flume (Lohrmann et al. 1994) and

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Note. Discussion open until November 1, 2000. To extend the closing date one month, a written request must be filed with the ASCE Manager of Journals. The manuscript for this paper was submitted for review and possible publication on July 6, 1999. This paper is part of the Journal of Hydraulic Engineering, Vol. 126, No. 6, June, 2000. ©ASCE, ISSN 0733-9429/00/0006-0399–0406/$8.00 + $5.50 per page. Paper No. 21356.
in an ocean boundary layer (Lohmann et al. 1995), Vougaris and Trowbridge (1998) gave a thorough evaluation of the accuracy of ADV for the turbulence measurement in a laboratory flume. Snyder and Castro (1999) tested ADV in a stratified towing tank and provided the correction scheme. Sukhodolov et al. (1998) used ADV to investigate turbulence structure in a natural open channel flow. Considering the rapid advances occurring in the sophistication and accuracy of readily available velocity sensors such as ADV, the evaluation of observational techniques for estimating bottom stress becomes particularly relevant. With this motivation, we used ADV to investigate various commonly applied techniques available for estimating bottom stress in an estuary to determine its success and consistency.

BACKGROUND

In this study, a velocity vector \( \mathbf{u} = (u, v, w) \) is considered in a Cartesian coordinate system (x, y, z) in which x is in the mean flow direction, y is horizontally across the mean flow direction, and z is the vertical coordinate. The flow is assumed to be composed of mean (overbar) and fluctuating (prime) parts such that \( \mathbf{u} = \overline{\mathbf{u}} + \mathbf{u}' \). Intrinsic in this assumption is the separation of two distinct timescales so that there is no contamination from gravity waves. The flow is further assumed to be neutrally stratified, horizontally homogeneous, and stationary. Under these conditions, a logarithmic velocity profile is observed within the “constant stress” layer where stress within the water column only varies slightly from bottom stress \( \tau_0 \). The log layer forms the transition between the surface layer at the bed, which is influenced by viscosity, and the outer layer, which is influenced by the total thickness of the boundary layer (Tennekes and Lumley 1972). Bottom stresses are estimated from the models based on the prescribed assumptions. In practice, the estimation may be linked to the first moment statistics (mean) such as in the log profile (LP) method or to the second moment statistics such as in the COV or total turbulent kinetic energy (TKE) methods. Spectral characteristics also provide a viable tool such as in the ID method.

LP Method

For fully rough turbulent boundary layers, the logarithmic velocity profile is expressed by the von Kármán-Prandtl equation

\[
\overline{u} = \frac{U_*}{\kappa} \ln \left( \frac{z}{z_0} \right)
\]

(1)

The components of this equation are friction velocity \( U_* \), mean velocity \( \overline{u} \), and height above the bed \( z \); \( z_0 \) is hydraulic roughness—a constant of integration found to increase with bottom roughness—and \( \kappa \) is the von Kármán constant (~0.4). The above equation is based on the assumption of nearly constant stress so that \( \overline{\partial u/\partial z} = U_*/\kappa z \). The direct implementation of the above equation is the use of a profile of mean velocities within a log layer to estimate both \( U_* \) and \( z_0 \). The measured values of \( \overline{u} \) are highly accurate, using either EMCMs or ADVs. But the independent variable \( z \) in the above model may not be fixed under a natural flow condition because of the ambiguity of the bed level. Without sufficient degrees of freedom assured by a vertical array of many sensors, the estimation of \( U_* \) from a profile requires a very high correlation coefficient. In the case of a limited number of sensors, we may need further assumptions such as fixed \( z_0 \).

COV Method

The value of bed stress \( \tau_0 \) can be assumed to be close to that of the near-bed Reynolds stress, \( \tau = \rho \overline{\overline{u}'w'} \) [e.g., Heathershaw and Simpson (1978)]. In a log layer, \( \overline{\overline{u}'w'} \) and \( U_* \) are related by (Tennekes and Lumley 1972)

\[
\frac{-\overline{\overline{u}'w'}}{U_*^2} = 1 - \frac{1}{\kappa U_* z} = 1 - \frac{1}{R}
\]

(2)

For a fully turbulent flow with large Reynolds number, \( R = \kappa U_* z / \nu > 1 \), this effectively becomes \( -\overline{\overline{u}'w'} = U_*^2 \), giving the direct estimation of \( U_* \). Unlike the LP method, the COV method is not dependent on \( z \). This method seems to be ideal because the covariance represents the unbiased estimation of bottom stress. However, as shown later in this paper, there are potential error sources such as tilting of the sensor or contamination from intratidal frequency flows such as secondary flows.

TKE Method

The absolute intensity of velocity fluctuations—variances can also be used to infer bed stress through TKE, \( E = (u'^2 + v'^2 + w'^2)/2 \). Simple linear relationships between turbulent energy and shear have been formulated in turbulence models [e.g., Galperin et al. (1988)]. Soulsby and Dyer (1981) analyzed turbulence data of tidal currents and showed that the average ratio of shear stress to TKE is constant; \( |\sigma| = C_E \). Where C_E (~0.20) is a proportionality constant. Stapleton and Huntley (1995) later adopted \( C_E = 0.19 \), which is the same as the atmospheric value. By assuming linear relationships between TKE and the variances, we can also relate the bottom stress to a variance component such as \( |\sigma| = C_I \). Like the COV method, the TKE method also depends on the second moment statistics and is not subject to the error related to the sensor height \( z \). The method still requires the finding or confirming of universal coefficients \( C_i \) or \( C_z \).

ID Method

For a log layer, a first-order balance between shear production \( P \) and energy dissipation \( \varepsilon \) is a fair assumption [e.g., Tennekes and Lumley (1972)]

\[
-P + \varepsilon = \overline{\overline{u}'w'} \frac{\partial U}{\partial z} + \varepsilon = 0
\]

(3)

Taking \( -\overline{\overline{u}'w'} = U_*^2 \) (the COV method) and \( \overline{\partial u/\partial z} = U_*/\kappa z \) (the LP method), we have

\[
U_* = (\kappa z)^{1/3}
\]

(4)

There exists an inertial subrange over which energy is cascaded from energy-producing low frequencies to energy-dissipating high frequencies [e.g., Champagne et al. (1977)]. The 1D spectrum applicable to the inertial dissipation range has the form

\[
\Phi_i(k) = \alpha_i e^{2/3 k^{-5/3}}
\]

(5)

where \( \Phi_i(k) \) is spectral density of \( i \)th velocity component at the wave number \( k \); and \( \alpha_i = 1 \)D Kolmogorov constant. In locally isotropic turbulence, \( \alpha_1 \sim 0.51 \) and \( \alpha_2 = \alpha_3 = 4/3\alpha_1 \sim 0.69 \) (Green 1992). Combining the above two equations, we have

\[
U_* = (\kappa z)^{1/3} \left( \frac{\Phi_i(k)k^{5/3}}{\alpha_i} \right)^{1/2}
\]

(6)

Near the bed, vertical velocities are less contaminated by waves than are horizontal velocities, and the fluctuations \( w' \) are more likely to be due to turbulence (Stapleton and Huntley 1995). Thus in this study, we systematically applied the ID method only to vertical velocities. For the velocity data ob-
tained in a time domain, we need to translate frequency spectra into wave-number spectra. With the condition of $kb_{\omega_m}(k)/u^2 << 1$, Taylor’s “frozen turbulence” hypothesis can be applied assuming $kb_{\omega_m}(k) = f\phi_{\omega_m}(f)$ with $k = 2\pi f/\bar{u}$, where $f$ is frequency [e.g., Huntley (1988)]. Then (6) becomes

$$U_s = \left( \frac{2\pi \kappa}{\bar{u}} \right)^{1/3} \left( \phi_{\omega_m}(f) \frac{f^{5/3}}{\alpha_3} \right)^{1/2}$$

(7)

The inertial subrange was determined as the range of frequency in which $f^{5/3}\phi_{\omega_m}(f)$ is constant. In practice, the average $f^{5/3}\phi_{\omega_m}(f)$ values around the maximum value were used.

**DATA ACQUISITION AND ANALYSES**

We deployed a bottom boundary layer measuring tripod over a 10-day period in April 1996, in a secondary channel of the lower York River estuary (Fig. 1) as part of a larger study of biological mediation of material and transport processes in tidal estuaries. The mean depth was about 5 m and the bottom sediments consisted of soft mud with little sand and slightly more silt than clay (Dellapenna et al. 1998). The tripod was equipped with two ADV field models and five EMCMs (diameter = 2.5 cm) to measure flow and five optical backscatter sensors (OBSs) to measure the suspended sediment concentration profile (Fig. 2). A pressure sensor was also used to record water surface level variation. The lowermost two EMCMs and OBSs were mounted at the same heights as the two ADVs’ sampling elevations at $z = 14$ and 44 cm above bed (cmab).

The elevation of the lower ADV was confirmed during each burst by way of the ADVs’ acoustic altimetry feature. The upper three EMCMs and OBSs were located at 74, 104, and 134 cmab, respectively. For ADVs, the intraburst record length was 700 s with a 5-Hz sampling frequency. For the EMCMs, the OBSs, and the pressure sensor, the sampling frequency was 1 Hz with a record length of 1,024 s. The burst interval for all sensors was 1 h.

For each burst, flow measurements by each ADV were transformed to a mean flow coordinate system through $u = [R]u$, where $u = (u_s, v_s, w_s)$ is the flow vector in an orthogonal coordinate $(x_s, y_s, z_s)$ taken by the ADV and $[R]$ is a coordinate rotation matrix. The coordinate rotation involves first a rotation of $\beta$ around the $z_s$-axis followed by a rotation of $\theta$ around the new $y$-axis

$$[R] = \begin{bmatrix} \cos \theta \cos \beta & \cos \theta \sin \beta & \sin \theta \\ -\sin \beta & \cos \beta & 0 \\ -\sin \theta \cos \beta & -\sin \theta \sin \beta & \cos \theta \end{bmatrix}$$

(8)

The rotation angles $\beta$ and $\theta$ were calculated by assuming $\bar{v} = 0$ and $\bar{w} = 0$, respectively. Except for the flow near the slack water, $\beta$ is aligned along the major axis of the channel ($135^\circ$ and $-45^\circ$) and typically changes with the rate of $<5^\circ$/h (1°/burst), which verifies negligible veering within a burst, and in turn, justifies the use of burst-average values for the calculation of $\beta$ (Fig. 3). Unlike $\beta$, $\theta$ distribution is less organized, showing ebb-flood variation at 14 cmab and more scattered distribution during ebb at 44 cmab (Fig. 4). The sources of variations of $\theta$ may include the tilt of the sensor relative to the direction of gravitational acceleration as well as the existence of a secondary flow or wakes. Since the focus of this study is on bottom stress rather than sources of secondary flow, we will not adopt any assumption about the vertical rotation.

FIG. 1. Location Map of Study Area; Pod Was Deployed in Secondary Channel of Tidal York River Estuary with Mean Water Depth of about 5 m

FIG. 2. Deployment Scheme: (a) Tripod Orientation; (b) Sensor Arrangement

FIG. 3. Distribution of Veering Angle $\beta$ as Function of Mean Current Speed: Positive Current Represents Ebb and Negative Current Represents Flood; Plus Sign and Circle Represent 14 and 44 cmab, Respectively
The second moment statistics utilized by the COV and TKE methods are subject to sampling errors. The confidence limit for a sample mean $X$ becomes $\bar{X} \pm t_{N,N} s_x / \sqrt{N}$ where $s_x$ is sample standard deviation, $N$ is number of independent samples, and $t_{N,N}$ is the Student’s $t$ value of probability $\alpha$ with $N$ degrees of freedom [e.g., Snedecor and Cochran (1967)]. In this study, $N$ is considered as the number of independent events within intraburst samples. We took the autocorrelation timescale as an indicator of the independent event. The autocorrelations of turbulent vertical velocity $w$ and covariance $u'w'$ decrease relatively fast compared to those of horizontal velocity $u'$ (Fig. 7). By taking 2.5 s as an event timescale of $w'$ over a 700-s intraburst record, we have 280 degrees of freedom.

The standard deviation of covariance $s_{uw}$ is about 2.5 and 3.5 times its mean value at 14 and 44 cmab, respectively (Fig. 8). Thus, the 95% confidence interval with 280 degrees of freedom becomes about ±30% at 14 cmab and about ±40% at 44 cmab, respectively. This translates to a confidence interval on individual $U_u$ estimates using the COV method of about ±17% and ±23%, respectively. Observed covariances at 14 cmab are generally higher than those at 44 cmab (Fig. 9).

Standard deviations of turbulent kinetic energy $s_E$ were about 1.1 times $E$ both at 14 and 44 cmab (Fig. 10). With 280 degrees of freedom, the 95% confidence interval becomes about ±13%, which translates to an uncertainty of approxi-
bottom shear stress, the value of $C_1$ in the TKE equation $|\tau| = C_1 E$ is estimated here to be $\sim 0.21$ by best fit.

The mean variance terms are ordered $\overline{u^2} > \overline{v^2} > \overline{w^2}$ as in a typical shear-generated turbulence (Fig. 12). Relative to $E$, each variance showed a linear relationship as follows: $\overline{u^2} \approx 1.2E$; $\overline{v^2} \approx 0.5E$; and $\overline{w^2} \approx 0.25E$. The variance and covariance measurements are subject to the errors associated with unbiased, identical but uncorrelated Doppler noise level, which varies among sensors (Lohrmann et al. 1995). It is known that the instrument noise errors associated with vertical velocity variances are smaller than the instrument noise errors for horizontal velocity variances by at least an order of magnitude. A sensible variation of the TKE method is thus to utilize only vertical velocity fluctuations. Assuming the COV method to be unbiased, a best fit then gives $C_2 \sim 0.9$ for $|\tau| = C_2 E^{7/2}$.

The 95% confidence interval of $\phi_{uu}(f)$, based on a chi-square distribution with 203 degrees of freedom ($= 3.71T_s/M$ where total number of sample $T_s = 3,500$ samples/burst, and the lag $M = 64$ samples for Parzen window) is between $-16$ and $+14\%$. Assuming no bias in the mean current $\bar{u}$ and negligible error in sensor height $z$, (7) gives approximately $\pm 15\%$ error in $U_*$ estimation with a 95% confidence interval. Fig. 13 shows the estimated friction velocities from the two sensors. Observed $U_*$ at 14 cmab are about 25% higher than those at 44 cmab.

We estimated $U_*$ by the LP method with the EMCMs. Fig.

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**FIG. 8.** Standard Deviations of Covariance $s_{uw}$ (Vertical Axis) versus Mean Covariance $\overline{uw}$ (Horizontal Axis) for Each Burst (Plus Sign and Circle Represent Measurement Heights of 14 and 44 cmab, Respectively)

**FIG. 9.** Covariances $\overline{uw}$ from 44 cmab (Vertical Axis) versus Those from 14 cmab (Horizontal Axis) for Each Burst (Error Bar Represents 95% Confidence Interval at $\overline{uw} = -1.5 \text{cm}^2/\text{s}$)

**FIG. 10.** Standard Deviations of Turbulent Kinetic Energy $s_E$ (Vertical Axis) versus Mean Turbulent Kinetic Energy (Horizontal Axis) for Each Burst (Plus Sign and Circle Represent Measurement Heights of 14 and 44 cmab, Respectively)

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mately $\pm 7\%$ in $U_*$ estimation. Thus the sampling errors in the TKE method are significantly less than those associated with the COV method. In contrast to the Reynolds stress, $E$ was nearly identical at 14 and 44 cmab (Fig. 11). Assuming the COV measurement at 14 cmab is an unbiased estimate of

**FIG. 11.** Turbulent Kinetic Energy $E$ from ADV at 44 cmab (Vertical Axis) versus $E$ at 14 cmab (Horizontal Axis) for Each Burst (Error Bar Represents 95% Confidence Intervals at 12.5 cm$^2$/s)

**FIG. 12.** Mean Variances (Vertical Axis) versus Turbulent Kinetic Energy $E$ (Horizontal Axis) at 14 cmab for Each Burst (Plus Sign, Open Circle, and Open Square Represent $\overline{u^2}$, $\overline{v^2}$, and $\overline{w^2}$, Respectively)

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14 demonstrates the increased $U_*$ estimates that result from EMCM sensors positioned higher off the bottom. This suggests the log layer assumption $\partial U_*/\partial z = U_*/\kappa z$ becomes less valid farther away from the bed. It is very likely that the EMCMs at 74, 104, and 134 cmab may often have been outside the constant stress layer. To apply the LP method to ADV measurements of mean velocity at only two heights, we chose $z_0$ to be the single roughness scale for all the flood tides that gave the minimum difference between estimates of $U_*$ from the two ADV sensors. A single value for bottom roughness is commonly used in modeling tidal flow [e.g., Kuo et al. (1997)] and in analyzing boundary layer observations [e.g., Cheng et al. (1999)]. Following this approach, we obtained $z_0 \approx 10^{-3}$ cm. Errors for each pair of $U_*$ values were estimated by averaging the ratio of the standard deviation to the mean. The average ratio of about 2% with 1 degree of freedom gives $\pm 20\%$ error in $U_*$ estimation with a 95% confidence interval.

Table 1 compares the various estimates of shear velocity at each of the two ADV sensors along with the corresponding uncertainties. The ID method applied at 44 cmab and the COV method applied at 14 cmab gave average $U_*$ estimates of 1.16 cm/s, which is comparable to 1.15 and 1.16 cm/s given by the LP method. At 14 cmab, the ID method estimates were the lowest (average 0.95 cm/s). At 44 cmab, the COV method estimates were the lowest (average 1.0 cm/s). The TKE estimates (using the literature value of $C_l = 0.19$) were consistently low with averages of 1.09 cm/s at 14 cmab and 1.11 cm/s at 44 cmab. The average scatter of the TKE estimates between the two sensors was only 5%. The COV method at 44 cmab relative to 14 cmab gave lower estimates with an average scatter of 15%, and the LP method resulted in higher estimates at 44 cmab with an average scatter of 22%. Compared to the TKE estimates, the variability of the COV method was about 10%. The variability of the ID method was about 15%, and the variability of the LP method was about 15–20%.

\[ \frac{-u'^2}{U_*^2} = 1 - \frac{z}{\delta} \]  

(9)

Most of the time during the experiment period, the observed Reynolds stresses were lower at 44 cmab than at 14 cmab (Fig. 9). This implies that the sensor at 44 cmab was near the top of the log layer. The limited applicability of the log layer assumption is also suggested by an examination of the log-fit of the EMCMs, which shows increasing $U_*$ with the addition of more sensors away from the bed (Fig. 14).

Stable density stratification associated with suspended sediment is likely to affect the velocity profile very near the bed in energetic tidal estuaries containing abundant fine sediments (Sheng and Villaret 1989). Unlike stratification associated with thermohaline effects, sediment-induced stable stratification tends to increase in intensity toward the bed. Near the bed, the suppressed turbulent energy due to stratification would be translated as the reduced velocity scale, which in turn would give reduced eddy viscosity if the length scale were the only function of height from the bed. With a given stress, the reduced eddy viscosity would be compensated by the increased velocity shear. Thus as one approaches the bed, velocity shear may continue to be enhanced by the effects of suspended sediment throughout the constant stress layer. Although the form of the profile very near the bed may still resemble (1), a best-fit log-profile in the presence of significant sediment-induced stratification will overestimate the actual value of $U_*$ (Wright

**TABLE 1. Summary Statistics for $U_*$ Estimation Methods**

<table>
<thead>
<tr>
<th>Summary</th>
<th>TKE</th>
<th>ID</th>
<th>COV</th>
<th>LP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average $U_*$ (cm/s) for all floods with $U &gt; 10$ cm/s</td>
<td>1.09</td>
<td>0.95</td>
<td>1.16</td>
<td>1.16</td>
</tr>
<tr>
<td>95% confidence interval for individual burst estimates (%)</td>
<td>±7</td>
<td>±15</td>
<td>±17</td>
<td>±20</td>
</tr>
<tr>
<td>Average absolute difference from the estimate at 14 cmab (%)</td>
<td>—</td>
<td>15</td>
<td>10</td>
<td>15</td>
</tr>
<tr>
<td>Average absolute difference from TKE estimation (%)</td>
<td>—</td>
<td>—</td>
<td>14</td>
<td>15</td>
</tr>
</tbody>
</table>

**FIG. 13. Friction Velocity Estimated by ID Method from ADV at 44 cmab (Vertical Axis) versus 14 cmab (Horizontal Axis) for Each Burst (Error Bar Represents 95% Confidence Intervals at 1.25 cm/s)**

**FIG. 14. Estimated Bottom Friction Velocities Using LP Method: Horizontal Axis Is for $U_*$ Estimated from EMCM at 14 cmab; Vertical Axis Is for $U_*$ Estimated from EMCMs at 44 cmab (Plus Sign) and 74 cmab (Circle)**
et al. 1999). The simple log-fit of mean velocities generally gave the largest estimate of \( U_* \), a tendency that is consistent with the effects of stable stratification associated with suspended sediment.

With increased sediment concentration near the bed, a balance between production and dissipation in (4) for the ID method is also no longer entirely valid. Accounting for an additional sink term due to stratification induced by suspended sediment will result in a higher \( U_* \) value. The limit of the increase in \( \tau \) associated with stratification is about 1/4 of production [e.g., Tennekes and Lumley (1972)]. Then, the resulting \( U_* \) value will be increased by a maximum of 10%. This would increase the average \( U_* \) from the ID method at 14 cmab from 0.95 cm/s to as much as 1.05 cm/s, bringing the estimate closer to those derived from the other methods. Nevertheless, the ID method using the first-order assumption of a production-dissipation balance is considered to work reasonably well from 0.95 cm/s to as much as 1.05 cm/s, bringing the estimate would increase the average production [e.g., Tennekes and Lumley (1972)]. Then, the result-

\[ U_* = \left( \frac{U^3 \cdot \mathcal{R}_z \cdot \mathcal{V}}{k^2} \right)^{1/4} \]

Here \( z_c \) = critical height where \( U_* = \tilde{U}_* \), which is estimated from (7). \( \mathcal{R}_z \) is on the order of 1.000. By taking \( \mathcal{R}_z = 1.000 \) (with \( z_c = 35 \) cm), the average \( U_* \) estimate at 14 cmab from the ID method was increased by about 25%. This indicates that the height of 14 cmab may have been too close to the bottom to ensure the full separation of production and dissipation.

The reliability of turbulence measurement is limited by record length and sampling frequency. Soulsby (1980) described low frequency losses due to an inadequate record length and high frequency losses due to either an inadequate sampling rate or spatial averaging by the sensor. For the ADV, the error caused by spatial averaging is not considered significant. The normalized standard error associated with Reynolds' stress, based on Soulsby’s (1980) analysis, is

\[ \text{Err}_{w,w} = \left( \frac{2 \cdot u' \cdot w' \cdot \sigma_{u'w'}^2}{\tau \cdot u' \cdot w'} \right)^{1/2} \]

Here \( L_* \) = integral length scale (\( \bar{u} \cdot t_\tau \)) normalized by the measuring height \( z \), where \( \bar{u} \) = mean flow and \( t_\tau \) = Eulerian time-scale, which is the integral of autocorrelation of variance of \( u' \cdot w' \). Normalized minimum wave number \( k_{\text{min}} = 2 \pi z / (T \cdot \bar{u}) \) is dependent on the record length \( T \). From \( L_* k_{\text{min}} = 2 \pi t_\tau / T \), we have

\[ \text{Err}_{w,w} = 2 \left( \frac{u' \cdot w' \cdot t_\tau}{u' \cdot w'} \right)^{1/2} \]

The random error was the minimum at peak flows associated with the smallest \( t_\tau \) (=0.25 s). It also appears that \( t_\tau \) increases with sensor height from the bed. A typical peak flow showed \( \text{Err}_{w,w} \sim 3.7\% \) and \( \text{Err}_{u,w} \sim 3.9\% \) at 14 and 44 cmab, respectively. For the variances, the random error is reduced to 2.8\% at both elevations. Soulsby gives the high frequency energy loss in vertical velocity as 2.1 \( (U_* \cdot z / \nu)^{1/2} \) where \( \nu \) (=0.014 cm²/s) is the kinematic viscosity. For a flow with \( U_* = 1 \) cm/
error. Regarding individual methods for estimating bottom stress, we conclude the following:

- The COV method is considered to give unbiased estimates of bottom stress as long as the sensor is close enough to the bed to be within the constant stress layer but simultaneously sufficiently far to avoid problems associated with velocity shear within the sampling volume. Contamination of the COV estimates may also result from tilting of the sensor or from secondary flows.
- The ID method may require the sensor location to be somewhat farther away from the bed ($z \approx z_0$) to have an adequate separation of production and dissipation scales. Also, stratification associated with suspended sediment may cause the simple assumption of a balance between production and dissipation of turbulent energy to underestimate $U_*^2$ by up to 10%.
- The TKE method is considered to be the most consistent and exhibits the least variability but still requires more study to determine the relationships between TKE and bottom stress. This study favored the best-fit constant between TKE and COV to be 0.21 near the bed, compared with 0.19 in the literature.
- The LP method generally gave the largest estimate of $U_*$, a tendency that is consistent with the effects of sediment-induced stratification. The estimates were lowered when a fixed $z_0$ was assumed. But the variability was still the highest among the methods tested.

ACKNOWLEDGMENTS

This study was supported by the Office of Naval Research through Grant N00014-93-1-0986 and Grant N00014-95-1-0391. This is Contribution No. 2283 from the Virginia Institute of Marine Science. R. A. Gammisch, F. Farmer, T. Nelson, and W. Reisner assisted in field work. The writers are grateful to anonymous reviewers for their constructive criticism.

APPENDIX I. REFERENCES


