Bed Drag Coefficient Variability under Wind Waves in a Tidal Estuary

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Abstract: In this paper we report the results of a study of the variation of shear stress and the bottom drag coefficient $C_D$ with sea state and currents at a shallow site in San Francisco Bay. We compare shear stresses calculated from turbulent velocity measurements with the model of Styles and Glenn reported in 2000. Although this model was formulated to predict shear stress under ocean swell on the continental shelf, results from our experiments show that it accurately predicts these bottom stress under wind waves in an estuary. Higher up in the water column, the steady wind-driven boundary layer at the free surface overlaps with the steady bottom boundary layer. By calculating the wind stress at the surface and assuming a linear variation of shear between the bed and surface, however, the model can be extended to predict water column shear stresses that agree well with data. Despite the fidelity of the model, an examination of the observed stresses deduced using different wave–turbulence decomposition schemes suggests that wave–turbulence interactions are important, enhancing turbulent shear stresses at wave frequencies.

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Introduction

The erosion, transport, and deposition of sediments play a central role in many geological, biological, geochemical, and ecological processes operant in estuaries (Shrestha and Orlob 1996). Accordingly, engineering studies done to assess how large-scale works such as constructing airport runways on new landfill, restoring tidal wetlands, or changes in the disposal of dredge materials might degrade or improve the environment of the estuary on which these actions would take place often make use of coupled hydrodynamic/sediment transport models that are designed to predict sediment dynamics (Blumberg et al. 1999). A key feature of all these models is the way in which bed friction is modeled, i.e., how it depends on bed roughness, mean currents, and as we discuss in this paper, how it is affected by wind-wave orbital motions (Perlin and Kit 2002).

In most three-dimensional (3D) circulation models, bottom friction is represented by a quadratic drag law based on a bottom drag coefficient $C_D$.

$$\tau_c = \rho C_D |U_c| U_c$$

where $\tau_c$=steady shear stress at the bed; $\rho$=water density; and $U_c$=velocity of the mean current at height $z_r$ (Signell et al. 1990), usually taken as 1 m above the bed (mab). In 3D modeling, $\tau_c$ is then used as a bottom boundary condition on the velocity.

$$\tau_c = -\rho \nu_r \frac{\partial U}{\partial z}$$

where $\nu_r$=eddy viscosity.

It is commonly assumed, and often found to be true in practice (Nezu and Nakagawa 1993; Lueck and Lu 1997), that the well-known law of the wall (e.g., Tennekes and Lumley 1972; Nezu and Rodi 1986)

$$|U_c| = \left[\frac{u_r}{\kappa}\ln\left(\frac{z}{z_0}\right)\right]$$

applies to unsteady estuarine and coastal flows. Here $\kappa\approx0.41$ is von Kármán’s constant; $z$=positive upwards (and zero at the bed); the roughness length $z_0=k_b/30$, where $k_b$=sand grain roughness of the bed (Nikuradse 1932); and the shear velocity is

$$u_r = \sqrt{\frac{\tau_c}{\rho}}$$

Using these relations, $C_D$, referenced to some height $z_r$, can be inferred from the roughness length

$$C_D = \left[\frac{\kappa}{\ln(z_r/z_0)}\right]^2$$

[see, e.g., Gross et al. 1999]. Generally, the value of $C_D$ depends upon bed sediment grain size and bed-form geometry.

However, in the shoals of many estuaries, areas that are often of particular environmental and engineering concern, currents and sediment dynamics can be strongly influenced by wind waves (Schoellhammer 1996). As discussed by Grant and Madsen (1979) and others (e.g., Fredsøe 1984; Perlin and Kit 2002), bot-
bottom drag is enhanced when surface waves are long enough to reach the bed. In this case, a thin, oscillatory, wave boundary layer (also known as a Stokes layer [Kundu 1990]) develops near the bed that is more strongly sheared and experiences phase-dependent stresses that are larger than those operant in the thicker, steady current bottom boundary layer. It is the rectification of the periodic stress that alters the overlying flow.

Grant and Madsen (1979) modeled the inner wave boundary layer using an eddy viscosity based on a phase-dependent shear velocity, \( u_{\text{cuv}} \), defined by the phase-dependent bottom stress (which is due to both waves and currents) to determine velocity structure. Outside the wave layer, Grant and Madsen determine the effect of this inner wave layer on the main part of the boundary layer by requiring the outer flow stress and velocity to match the time averaged stresses and velocities of the inner wavy flow. The result is that enhanced turbulence within the wave layer results in greater drag on the mean flow, an effect that can be modeled as an enhanced apparent roughness. The original Grant–Madsen model has been extended by various writers, most recently Styles and Glenn (2000, hereinafter SG2000), who incorporated the effects of stratification as well as adjusting the assumed profile of eddy viscosity upon which the theory is based.

Fredsoe’s (1984) model is similar to that of Grant and Madsen, although he assumed logarithmic profiles rather than solving explicitly for the velocity structure. At the other end of the spectrum, Perlin and Kit (2002) make no assumptions about eddy viscosity profiles, instead using a turbulence closure that solves for the phase-dependent turbulent kinetic energy (TKE) from which phase-dependent eddy viscosities, etc. can be computed. In a similar fashion, Groeneweg and Klopman (1998) more formally evaluated the average effects of waves on mean currents using a \( K-e \) model and generalized Lagrangian mean (GLM) averaging.

Most circulation models use a drag coefficient to parameterize bottom drag (e.g., Casulli and Cattani 1994; Blumberg et al. 1999; Gross et al. 1999). Since these models resolve neither individual waves nor the thin wave bottom boundary layer, the effects of waves are then included via a wave-dependent \( C_D \) or wave-dependent roughness (Davies and Lawrence 1994, 1995; Signell and List 1997; Xing and Davies 2003). Kagan et al. (2003) found that inclusion of ocean-swell-enhanced roughness in a circulation model of Cadiz Bay brought tidal stage and currents predicted by the model into better agreement with observations. All of these studies have found that wave-enhanced roughness can have significant effects on tidal and wind-driven circulation, flushing rates and residence times of contaminants, channel-shoal asymmetry, and sediment transport.

Despite the recent application of Grant and Madsen’s enhanced roughness theory to a variety of circulation models, field experiments to test the validity of this theory have been carried out only under ocean swell on the continental shelf (Caciolone et al. 1994; Drake and Caciolone 1992; Green and McCave 1995; Lacy et al., unpublished), and not under wind waves on the shoals of an estuary. For ocean swell on the continental shelf, the steady surface wind-driven boundary layer and the steady current bottom boundary layer are separated by an inviscid core (Grant and Madsen 1986); these are the conditions for which Grant and Madsen’s model was formulated. The accuracy of the drag coefficients and stresses predicted by models like that of Grant and Madsen have not been tested for shallow flows on the shoals of an estuary in the presence of wind waves, an environment where these two steady boundary layers overlap.

In this paper we investigate the variation of shear stress and \( C_D \) under combined tidal currents and wind waves (generated by the local diurnal sea breeze) on a shoal in South San Francisco Bay. Our results show that the model presented by Styles and Glenn (2000, hereinafter SG2000), while formulated for ocean swell on the continental shelf, is capable of reasonably accurate predictions of the enhancement of shear stress and the bed drag coefficient over estuarine shoals by wind waves.

**Experiments**

To investigate variations of water column shear stresses and the bed drag coefficient under wind waves on the shoals of a tidal estuary, we ran two experiments at Coyote Point in South San Francisco Bay (SSFB, Fig. 1). This location was chosen because of ease of access and because of an interest in understanding circulation and sediment processes in this part of the SSFB in light of proposals to build new runways nearby on fill that would be placed on the local shoals. Our experiments took place in June of 2000 and June–July of 2002. Each experiment was run for 2 weeks to capture a full spring-neap tidal cycle. Coyote Point experiences mixed semidiurnal/diurnal tides, a diurnal sea breeze (in summer), and frequent spilling whitecaps. The bathymetry at the study site is relatively flat, with a depth varying between 1 and 4 m during a near-solstice spring tide. The bed at the study site consists of silts and fine sands (a representative value of \( k_B =0.01 \) cm), and bed forms were not present. Shoreward (south) of the site is a gently sloping sandy beach, which causes incident waves to break as spilling and plunging breakers, and prevents reflections.

**Instruments**

During the first (June 2000) experiment we deployed two SonTek field acoustic Doppler velocimeters (ADVs) mounted on a mast sitting approximately 90 m north of the beach at high water (see Figs. 1 and 2) and in 1 m of water (mlw—mean lower lower water). The ADV measuring volumes were located 20 and 62 mab (centimeters above bed). These instruments sampled three components of velocity at 25 Hz, had an accuracy of ±3 mm/s, and were cabled to computers on the shore for data acquisition. Approximately 200 m further north (and thus further offshore) we deployed an upward-looking 1.5 MHz NorTek high-resolution acoustic Doppler profiler (ADP) mounted on a small gimbaled frame. Operating in a pulse-to-pulse-coherent mode (Lohrmann et al. 1990), it recorded 2-min averages of all three components of velocity in 3 cm bins covering a range of depths from 43 to 133 cmab. Time-averaged measurements of tidal stage and sea state were made with a SeaBird SBE26 absolute pressure sensor (accuracy greater than 1 mm in free surface elevation). An Ocean Sensors OS200 CTD made conductivity and temperature measurements. To record local wind speed and direction, we mounted an anemometer on a 3 m high mast. However, because the shore site where the anemometer was deployed was partially shielded by trees, for our analysis below we used an average of local winds as measured by the anemometer and those recorded nearby at San Francisco International Airport (SFO), located about 2 mi upwind (northwest) of the study site.

During the second experiment (June–July 2002) we made a more comprehensive set of measurements: Along with 3 SonTek ADVs measuring velocities at 20, 53, and 153 cmab we also deployed a NorTek Vector ADV measuring velocities at 95 cmab. In addition to the SBE26 used previously, we also used a capacitive wave gauge (1 cm accuracy; 1 mm precision) to measure surface...
waves. For wave–turbulence decomposition purposes (see below), the wave gauge was synchronized with the ADVs. In order to get a more complete velocity profile than we had in 2000, the body of the ADP was buried, so the transducer face was nearly flush with the bed, giving us good data starting at 15 cmab and extending to a maximum of 200 cmab (at high tide). Finally, in addition to the OS200 CTD, temperature was also measured by a vertical array of six high-precision (±0.01°C) SeaBird SBE39 temperature loggers. Unfortunately, during the second experiment our anemometer failed and so when needed for analysis, we used wind data from San Francisco International Airport’s (SFO) weather station.

**Observations of Conditions at Coyote Point**

During both experiments, winds were nearly westerly in direction and diurnal in strength, reaching 12 m/s each afternoon [Fig. 3(a)] which corresponds to a stress of approximately 0.2 Pa [Fig. 3(b)]. Waves showed the same diurnal trend, increasing to a maximum height $H_s = 50$ cm and a period, $T_w = 2$ s in the afternoon [Figs. 3(c and d)]. These periods and amplitudes are in good agreement with what might be predicted using formulas presented in the U.S. Army Corps of Engineers Shore Protection Manual (Bricker 2003).

Tidal velocities reached a maximum of 30 cm/s during peak flood with a favorable wind [Fig. 4(a)]. Flood currents were stronger than ebb currents (maximum ebb velocity was about 5 cm/s), and flood usually lasted longer than ebb, due to both the topography of the site and the winds over the site. Due to local bathymetry, the flood tide (coming from the north) hit the study site on the north side of Coyote Point with full force, while during ebb tide, the study site was in the lee of the Point, and was thus in its wake. During the afternoons, near-bed wave-induced orbital velocities averaged 12 cm/s, and was thus comparable in strength to the tidal current although they were nearly orthogonal to the tidal current [Fig. 4(b)]. Bottom stresses (determined as below) ranged from 0 to 0.13 Pa.

![Fig. 1. Coyote Point (South San Francisco Bay, Calif.) field site](image1)

![Fig. 2. Schematic of instrument arrangement (see text for details)](image2)
Fig. 3. Conditions existing in June 2002: (a) wind speed at 10 m measured at San Francisco International airport (SFO); (b) wind stress calculated per Eq. (11) from the SFO winds; (c) significant wave height at the instrument array; (d) dominant wave period at the instrument array; and (e) tidal variations in water depth at the instrument array.

Fig. 4. Velocities and stresses measured in June 2002 by ADV located at $z=20$ cmab: (a) long-shore (—) and cross-shore (—) mean velocity; (b) long-shore (—) and cross-shore (—) wave orbital velocity; and (c) long-shore turbulent shear stress inferred by the phase method (…), the Benilov–Filyushkin method (—), and by the Shaw–Trowbridge method (—).
Analysis Methods

Determination of Shear Stress via Time Series of Point Velocity Measurements

In principle, turbulence data from the vertical array of ADVs can be used to determine \( C_D \) by assuming a constant-stress steady bottom boundary layer and directly calculating Reynolds stresses from the fluctuating velocities. However, because variance associated with the waves is much larger than that associated with turbulence, some form of wave–turbulence decomposition scheme must be used (Jiang and Street 1991; Thais and Magnaudet 1995; Trowbridge 1998). As we will discuss below several approaches are possible and it remains an open question as to which is most appropriate in a given situation.

In a flow with both waves and currents, the instantaneous velocity can be written as

\[
\vec{u} = \vec{U} + \vec{u} + \vec{u}'
\]

where \( \vec{U} \) = mean velocity; \( \vec{u} \) = wave velocity; and \( \vec{u}' \) = turbulent velocity. After Reynolds averaging the mean momentum equation using Eq. (6), the Reynolds stress becomes [see, e.g., Jiang and Street (1991)]

\[
-\frac{\tau}{\rho} = \vec{u}\vec{w}' + \vec{u}'\vec{w} + \vec{u}'\vec{w}'
\]

For irrotational waves (Dean and Dalrymple 1991), the first term on the right-hand side (RHS) of Eq. (7) is zero. Furthermore, when waves and turbulence coexist, the latter can be defined as motions that do not correlate with waves (Jiang and Street 1991; Thais and Magnaudet 1995), so the second and third terms on the RHS of Eq. (7) are also zero. Thus, under these conditions, the Reynolds stress is the same as that which is found for steady flows

\[
-\frac{\tau}{\rho} = \vec{u}'\vec{w}'
\]

As shown by Trowbridge (1998), small uncertainties in instrument orientation or a gently sloping bed can bias velocity measurements such that in practice \( \vec{u}\vec{w}' \) may not be exactly zero.

In analyzing our data, we used three methods of wave–turbulence decomposition to remove the waves from our turbulence data. The first was that of Benilov and Filyushkin (1970), in which motions that correlate with displacement of the free surface are considered to be due to the waves. The second was that of Trowbridge (1998); and Shaw and Trowbridge (2001), which uses two velocity measurements spaced farther apart than the largest turbulence scale [approximately \( \frac{1}{2} \) the water depth, according to Shaw and Trowbridge (2001)], but are well within one surface wave wavelength of each other. Motions that correlate between the sensors are waves, while motions that are uncorrelated are defined as turbulence. The third method, which we call the “Phase Lag” method, is described in Bricker (2003). Assuming equilibrium turbulence and no wave–turbulence interaction, in this method the phase lag between the \( u \) and \( w \) components of surface waves is used to interpolate the magnitude of turbulence under the wave peak within the inertial subrange of the spectral domain, which is otherwise removed. In essence, in this approach, one removes not only the waves, but also the local enhancement of turbulence at or near wave frequencies (see Lumley and Terray 1983).

After applying the various wave decompositions, we calculated \( \vec{u}'\vec{w}' \) using blocks of \( 2^{13} \) samples (about \( 5\frac{1}{4} \) min of data). This period was chosen because it is significantly longer than the wave period, yet shorter than the time over which the sea state or tidal regime would change. We assumed that the lowest ADV (\( z = 20 \) cmab) measurements came from within the constant-stress inner layer of the steady bottom boundary, and thus that values of \( \vec{u}'\vec{w}' \) measured there are equivalent to the bottom stress exerted on the mean current. Lastly, using this value of \( \vec{u}'\vec{w}' \), \( C_D \) can be found in the usual fashion from Eqs. (1) and (8).

Determination of Theoretical Shear Stress via SG2000 and the Overlap Method

Shear stress was also calculated at the bed via SG2000, which assumes bottom-boundary-layer turbulence only. Stress was then calculated at the height of each ADV via the overlap method (see below), which takes into account the wind-driven surface boundary layer. To use SG2000, we needed to supply the model with the following inputs: \( \phi_b \), the angle between waves and the mean current; \( U_r \), a reference current a height \( z_r \) above the bed; \( u_b \) and \( A_p \), the near-bottom wave orbital velocity and excursion, respectively; and \( k_p \), the physical roughness scale. The reference height \( z_r \) was taken as the height above the bed of each ADV. The other input parameters were determined via the methods below.

Fine-grain sand dominates the near-shore subtidal zone at Coyote Point. For typical values of \( u_b \) at Coyote Point, a physical roughness of \( k_p = 0.01 \) cm, appropriate for fine sand, means that the flow was hydraulically smooth. As a result, the equivalent \( z_0 \) for a hydraulically smooth bed is \( z_0 = 0.11 \nu / u_b \), resulting in \( z_0 = 0.001 \) cm. Rather than use a variable value of \( z_0 \), we used this constant value of \( z_0 \) as an input to SG2000.

Maximum wave-induced near-bed orbital velocity \( u_b \), orbital excursion \( A_p \), and the angle between waves and currents \( \phi_b \) were determined spectrally from linear wave theory (Dean and Dalrymple 1991) using ADV and SBE26 data. With all these parameters determined, SG2000 then calculated the steady shear stress near the bed, and Eq. (1) was used to determine the drag coefficient based on this shear stress.

Overlap with Surface Wind-Driven Layer—The Overlap Method

Since the wind-driven surface boundary layer can overlap the bottom boundary layer in shallow flows, we further modified the approach given by SG2000 by assuming a linear variation in stress between the bed and the surface. Neglecting nonlinear accelerations and assuming pressure to be hydrostatic, the mean momentum equation in the \( x \) direction is

\[
\frac{\partial U}{\partial t} = -\frac{\rho}{\partial x} + \frac{1}{\rho} \frac{\partial \tau}{\partial z}
\]

where \( \eta = \) free surface deflection. The pressure term has no depth dependence, and the unsteady term is observed to vary far more with time than with depth (Fig. 5). Therefore, to a first approximation

\[
\frac{\partial \tau}{\partial z} = f(t)
\]

i.e., we can make the approximation that stress varies linearly with depth. From a practical standpoint, a linear variation is the
most complex one we can predict, given knowledge only of stresses at the bed and the surface.

Each component of the wind stress at the surface takes the form

\[ \tau_{\text{wind}} = \rho_{\text{air}} C_{D,\text{wind}} U_1^2 \]  

resulting in a shear velocity at the top of the water column of

\[ u_{*\text{wind}} = \sqrt{\tau_{\text{wind}} \rho_{\text{water}}} \]  

with \( C_{D,\text{wind}} \) determined from Yelland and Taylor’s (1996) empirical relations. The stress throughout the water column was then assumed to vary linearly from its (vector) value at the bed \( (u_{*b}) \) to its (vector) value at the surface \( (u_{*\text{wind}}) \). Finally, in our “overlap method,” the predicted stress at the height of each ADV is calculated using a linear interpolation between the calculated bottom stress and calculated surface stress.

Discussion

**Effect of Waves on Bottom Stress**

In what follows, we assume that the observed stresses are best represented by stresses determined via the phase method. In Fig. 6 we plot the June 2002 data; the June 2000 data is essentially identical and is not shown. We note that near bottom stresses calculated via Shaw and Trowbridge’s method (Fig. 6(a)) agree well with those calculated by the phase method. We also note that because the surface stress has little effect on the stress near the bottom, predictions of the bottom stress made using only SG2000 are essentially identical to those made using the more complete overlap method (Figs. 6(b and c)). In general either prediction based on SG2000 gives a reasonable prediction of the effects of waves on bottom stress. However, by comparison, the constant \( C_D \) case, at least if based on a physically plausible value of \( C_D \), consistently underestimates the stress. Clearly, in terms of use in a numerical model for cases where winds and hence waves are

**Fig. 5.** Acceleration measured by the ADP in June 2002 at \( z =100 \text{ cmab} \) as a function of the acceleration at \( z=60 \text{ cmab} \). The line shown marks the case where the two accelerations are equal.

**Fig. 6.** Stresses at 20 cmab in June 2002 as functions of stresses inferred using the phase lag method (“Phase”): (a) stresses inferred from ADV data using the Shaw and Trowbridge method (“ST”); (b) stresses computed from wave and mean flow conditions using the model of Styles and Glenn (“SG”); (c) stresses computed from wave and mean flows using the overlap method based on Styles and Glenn and a linear variation of stress over depth; and (d) stresses computed using a constant value of \( C_D \) and measured mean currents.
Fig. 7. Measured bottom drag coefficient ($C_D$) and predictions of bottom drag coefficient based on the SG model (---) and on the overlap method (---), all as functions of the ratio of mean velocity to wave orbital velocity. The measured values include error bars based on observed variability of stresses for each wave condition.

relatively repeatable (for example, in the case of diurnal sea breezes), $C_D$ could be adjusted upwards in the calibration process to match observation as was done by, e.g., Gross et al. (1999).

The difference between the constant $C_D$ case and the observations can be clearly seen in Fig. 7, where we have plotted $C_D$ as a function of the ratio of the mean velocity to maximum orbital velocity [as was done by Fredsøe (1984) or Perlin and Kit (2002)]. It is clear that bottom stresses we measured were significantly affected by waves, as Coyote Point is shallow enough for waves to “feel” the seabed. Shear stresses obtained through the ADVs, and the model of SG2000, all agree in trend. All drag coefficients converge to a value comparable to the canonical drag coefficient at 1 m of 0.0025 (Dronkers 1964) in the limit where mean current velocity is much greater than the maximum near-bed wave-induced orbital velocity. All methods also reveal an increase in $C_D$ of an order of magnitude over the canonical value when mean current velocity is much less than the near-bed orbital velocity. Results from the 2000 experiment were identical. We note that this variation in $C_D$ is due to the use of a drag law based entirely on the mean current, i.e., Eq. (1). Were we to use a drag law that explicitly included wave motions (cf. Hearn et al. 2001), $C_D$ would not need to be as large.

While in general the comparison of SG2000 to the observations is quite good, stresses calculated by SG2000 asymptote for strong currents to a value calculated using the canonical value of $C_D$. However, the stresses calculated from the ADV data are smaller. Therefore it appears that even in the current-dominated case, the water column cannot be modeled as a constant-stress bottom boundary layer, since the stress reduces to nearly zero at the surface on a calm day. Stresses calculated via SG2000 assume the measurements to be within a constant-stress bottom boundary layer only, thus one should expect the model to overestimate the shear stress observed at points a finite height above the bed (as much as 20% of the depth at low tide) during times of minimal wind.

Overall, comparison of predictions of bottom stresses with observations for both experiments shows that SG2000 predicts stresses better than one can do using a constant drag coefficient. Table 1 presents the normalized mean square error (Bevington and Robinson 1992) between two-week-long time series of shear stresses obtained via theory versus observations. Even without considering wave state, these values reinforce the view that SG2000 predicts stresses better than does a constant drag coefficient. In situations where wave-induced orbital velocities were larger than mean currents (Table 2), the disparity with the constant $C_D$ case is even greater. This agreement between theory and observation shows us that, despite the fact that SG2000’s model was developed to predict enhanced roughness on the continental shelf under ocean swell, it can also be effectively applied to the shallows of an estuary under the action of wind waves.

**Overlap of Steady Surface- and Bottom-Boundary Layers**

While SG2000’s predictions of stress agree with wave–turbulence decomposed near-bed (20 cmab) ADV data, as one moves higher in the water column, the quality of the prediction (which neglects surface stress) often underestimates the actual stress (Fig. 8). The reason for this is the overlap of the steady wind-sheared surface boundary layer with the steady bottom boundary layer (driven by both wind and tides). SG2000’s model was formulated for the bed of the continental shelf, on which shear stress is constant and originates from bed friction only. In our data, however, strong waves were accompanied by strong winds. During wind events, instruments high in the water column were affected by the shear stress and turbulence generated at the free surface as well as at the bed. Thus, very close to the bottom, bed shear stress is dominant, and thus predictions of the SG2000 model agree well with observations. Further up in the water column, though, the shear and turbulence in the free surface boundary layer becomes progressively stronger, and the model inaccurately predicts shear. The stress obtained via assuming an overlap of the bed and surface boundary layers, however, agrees well with wave–turbulence decomposed stress at all elevations [Figs. 8(a, c, and e)]. To better show the quality of the prediction, we have plotted trend lines of

**Table 1. Normalized Mean Square Error ($\varepsilon^2$) between Two-Week Time Series of Shear Stresses Observed via the Phase Lag Decomposition Method, and Shear Stress Predicted via the Overlap Method, SG2000, or a Constant $C_D$**

<table>
<thead>
<tr>
<th>Instrument</th>
<th>cm above bed</th>
<th>$\varepsilon^2$ phase overlap</th>
<th>$\varepsilon^2$ phase SG</th>
<th>$\varepsilon^2$ phase constant</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coyote lower ADV</td>
<td>20</td>
<td>0.54</td>
<td>0.54</td>
<td>1.03</td>
</tr>
<tr>
<td>Coyote2 lower ADV</td>
<td>20</td>
<td>0.33</td>
<td>0.38</td>
<td>0.60</td>
</tr>
<tr>
<td>Coyote upper ADV</td>
<td>62</td>
<td>0.41</td>
<td>0.57</td>
<td>0.98</td>
</tr>
<tr>
<td>Coyote2 vector</td>
<td>95</td>
<td>0.73</td>
<td>0.91</td>
<td>0.92</td>
</tr>
<tr>
<td>Coyote2 upper ADV</td>
<td>153</td>
<td>0.79</td>
<td>1.12</td>
<td>1.50</td>
</tr>
</tbody>
</table>

**Table 2.** Same as Table 1, Except Only Considering Times When the Maximum Wave-Induced Orbital Velocity Exceeded the Mean Current Considered in the Error Computation

<table>
<thead>
<tr>
<th>Instrument</th>
<th>cm above bed</th>
<th>$\varepsilon^2$ phase overlap</th>
<th>$\varepsilon^2$ phase SG</th>
<th>$\varepsilon^2$ phase constant</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coyote lower ADV</td>
<td>20</td>
<td>0.84</td>
<td>0.85</td>
<td>1.68</td>
</tr>
<tr>
<td>Coyote2 lower ADV</td>
<td>20</td>
<td>0.49</td>
<td>0.61</td>
<td>0.81</td>
</tr>
<tr>
<td>Coyote upper ADV</td>
<td>62</td>
<td>0.43</td>
<td>0.76</td>
<td>1.43</td>
</tr>
<tr>
<td>Coyote2 vector</td>
<td>95</td>
<td>1.13</td>
<td>1.12</td>
<td>2.31</td>
</tr>
<tr>
<td>Coyote2 upper ADV</td>
<td>153</td>
<td>1.03</td>
<td>1.37</td>
<td>2.61</td>
</tr>
</tbody>
</table>
the stresses at 153, 95, and 20 cmab from the 2002 experiment along with the corresponding predictions of the overlap method based on SG2000 (Fig. 9). Excepting spikes in the observed stresses, most likely associated with instrument noise, the overall agreement is excellent, lending support to the conclusion that the SG2000 model and the simple dynamics embodied in the linear stress distribution are reasonably accurate.

Wave–Turbulence Interaction

Stresses determined at 20 and 62 cmab were relatively independent of the method used to do wave–turbulence decomposition. However, closer to the water surface, the phase lag method consistently deduces stresses that are as much as an order of magnitude smaller than that calculated via the methods of Benilov and Filyushkin or Shaw and Trowbridge (Fig. 10). This result should be contrasted with the result that at all elevations stress determined by the phase lag method agrees remarkably well with that determined by the overlap method.

The dependence of the inferred stress on the wave decomposition method suggests that while wave–turbulence interaction may be negligible near the bed, it is more important near the surface where wave-induced orbital velocities, and the strain field they generate, intensify. The phase lag assumes no interaction between waves and turbulence, an assumption that may not be valid in the upper water column, where shear-generated turbulence is stretched by the wave-induced strain field (Teixeira and Belcher 2002) his modulated turbulence therefore overlaps with the wave field in the spectral domain, yet is still considered turbulence by Benilov and Filyushkin’s and Shaw and Trowbridge’s methods of wave–turbulence decomposition (Thais and Magnaudet 1995). In contrast, any spectrally local enhancement of turbulence by waves is rejected as a wave by the phase lag method.

Using a simplified form of rapid distortion theory (RDT), Monismith and Magnaudet (1998) showed that this type of wave–turbulence interaction is essentially described by the same model that predicts Langmuir circulations, i.e., the interaction of the Stokes drift with the vorticity field of the mean flow. Since wave motions increase in strength further up in the water column, the degree to which turbulence is strained is also enhanced closer to the surface, and the disparity between the phase lag method and the others grows larger.

The importance of wave strains on turbulence can be gauged by the rapidity parameter $R$, which following the methods of RDT (Townsend 1976), is defined as the ratio of wave-induced strain to turbulence-induced strain (Monismith and Magnaudet 1998)

$$R = \frac{\left( \frac{\partial u}{\partial z} \right)_{\text{wave}}}{\left( \frac{\partial u}{\partial z} \right)_{\text{turbulence}}}$$  (13)

When $R \ll 1$, turbulence-induced strain is much stronger than wave-induced strain, and thus wave strain has little effect on the turbulence. It is in this region that the assumptions of the phase lag method are valid. In contrast, when $R > 1$, wave-induced strain field strongly modulates the turbulence field (cf., Teixeira and Belcher 2002). In this case, because the phase lag method does not account for any periodic variation in turbulence intensity, etc., any additional periodic turbulent stress is attributed to a wave bias instead of turbulence, potentially causing the phase lag method to underestimate turbulent stresses. However, even in this
Fig. 9. Measured (—) and predicted (○) stresses during June 2002 at (a) \( z=153 \), (b) \( z=95 \), and (c) \( z=20 \) cm above the bed.

Fig. 10. Stresses inferred using different methods at several heights above the bed: \( z=153 \) cm above the bed June 2002 (a) Benilov–Filyushkin (“Benilov”) method and (b) Shaw–Trowbridge (“ST”); \( z=95 \) cm above the bed June 2002 (c) Benilov method; \( z=62 \) cm above the bed June 2000 (d) ST method; \( z=20 \) cm above the bed June 2002 (e) Benilov method and (f) ST method. In all cases we have used stresses inferred via the phase lag method as the independent variable.
case, the phase lag method still accounts for any rectified effects of the periodic wave straining on the turbulence.

From linear wave theory (Dean and Dalrymple 1991), the wave-induced strain can be shown to be

$$\left( \frac{\partial u}{\partial z} \right)_{\text{wave}} = \frac{H_s}{2} \frac{k_w}{T_w} \left( \frac{\sinh(kz)}{\sinh(kh)} \right)$$

(14)

where $H_s$=wave amplitude and $T_w$=wave period. The turbulent strain scales as

$$\left( \frac{\partial u}{\partial z} \right)_{\text{turbulence}} \sim \frac{u_*}{\kappa z}$$

(15)

For the conditions at Coyote Point, $H_s \sim 50$ cm, $T_w \sim 2$ s, $2\pi u/k = \lambda \sim 10$ m, $h \sim 2$ m, and $u_* \sim 0.01$ m/s. At $z=20$ cmab, this results in $R=0.05$ whereas $t_z=1$ mab, $R=5.0$. Thus we predict that close to the bed, turbulence is not affected by the wave-induced strain field (Fredsøe 1984), and the phase lag method separates wave and turbulent stresses very well. Higher up in the water column, however, wave-induced strain is as strong as turbulent strain, and the phase lag method attributes any coherent modulation of the turbulent stress by wave strain to waves instead of turbulence.

Notwithstanding the good agreement between the prediction of SG2000 and our measurements, these observations of wave–turbulence interaction are important in light of the theoretical underpinnings of models like that of SG2000 or Perlin and Kit (2002). For example, Perlin and Kit assume that wave shear contributes instantaneously to TKE production. In essence, their formulation includes waves as part of the mean flow despite the fact that in frequency space they overlap considerably with the frequencies at which energetic turbulence is found. In contrast, in the RDT view of wave–turbulence interaction, the wave strain modulates the turbulence field at wave frequencies such that the production term can be positive or negative depending on wave phase (Texeira and Belcher 2002), an effect observed in laboratory experiments by Pidgeon (1999). The Grant–Madsen style models like SG2000 (also Fredsøe 1984) are simpler in principle, only assuming a wave enhancement in the bottom stress via a bottom friction factor. However, their model implicitly assumes no wave–turbulence interaction since it assumes that the eddy viscosity that acts on the mean flow also acts on the waves. The importance of this neglect of the rectified effects of wave straining on the turbulence has yet to be quantified.

Conclusions

Our study shows that wind waves and the overlap of bottom and surface boundary layers clearly have a large effect on shear stress and the drag coefficient on the shoals of an estuary. The enhanced steady shear stresses and drag coefficients predicted by SG2000 agreed well with near-bottom (20 cm high) observations of shear stress in the steady bottom boundary layer under wind waves on shoals. Further up in the water column, however, SG2000 inaccurately estimated the shear stress because the steady wind-driven free surface boundary layer overlapped with the steady bottom boundary layer. At these higher locations, shear stress in the free surface boundary layer was as important as that in the bottom boundary layer, even in the absence of wind. In this case it was necessary to calculate the shear stress at both the bed and the surface, and then to assume a linear variation between these two throughout the water column. Shear stresses obtained in this way agreed well with data.

Water column shear stresses were underestimated by SG2000 when the wave-induced strain field was strong. In this situation, wave–turbulence interaction stretched turbulence and enhanced Reynolds stresses above those predicted by bottom boundary layer theory. This effect had greater influence higher in the water column, where the wave-induced strain field was stronger, and no influence near the bed.

Overall our results show that surface waves should be explicitly included in circulation models that are used to predict flows and sediment dynamics in shallow estuaries. As discussed in Perlin and Kit (2002), this requires also including some form of wave model, either one that is empirical such as the formulas given in the U.S. Army Corps of Engineers Shore Protection Manual or one that is based on wave dynamics such as SWAN [Simulating Waves Nearshore—see, e.g., Booij et al. (1999)]. In either case, from the standpoint of engineering practice, it appears that a first-order description of the effects of surface waves affect on bottom stress can be had through use of the model described in SG2000. Nonetheless, it appears that improvement in our understanding of how waves and turbulence interact in shallow estuarine flows would improve our ability to predict those flows.

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Notation

The following symbols are used in this paper:

- $A_b$ = near-bed orbital excursion;
- $C_D$ = bottom drag coefficient;
- $C_D,\text{wind}$ = drag coefficient for wind blowing over water;
- $f(t)$ = general function of time;
- $g$ = gravitational acceleration;
- $H_s$ = significant wave height;
- $h$ = water column depth;
- $k$ = wave number;
- $k_p$ = physical roughness length;
- $l$ = turbulence length scale;
- $R$ = rapidity;
- $r^2$ = square of correlation coefficient;
- $t$ = time;
- $T_w$ = wave period;
- $U_{50}$ = wind velocity at 10 m above free surface;
- $U_m$ = mean flow velocity;
- $u$ = total instantaneous horizontal velocity;
- $u_b$ = near-bed wave-induced orbital velocity;

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\[ u_{tb} = \text{shear velocity near the bed, above the wave bottom boundary layer}; \]
\[ u_c = \text{shear velocity above the wave bottom boundary layer}; \]
\[ u_{cw} = \text{maximum shear velocity inside the wave bottom boundary layer}; \]
\[ u_{wind} = \text{shear velocity in water at free surface}; \]
\[ \tilde{\nu} = \text{wave-induced orbital velocity in the horizontal}; \]
\[ u^* = \text{turbulent velocity fluctuation in the horizontal}; \]
\[ V = \text{mean transverse velocity}; \]
\[ U_h = \text{near-bed wave-induced orbital velocity in \( y \) direction}; \]
\[ w = \text{total instantaneous vertical velocity}; \]
\[ \tilde{w} = \text{wave-induced orbital velocity in the vertical}; \]
\[ w' = \text{turbulent velocity fluctuation in the vertical}; \]
\[ x, y = \text{horizontal axes}; \]
\[ z = \text{vertical coordinate, increases from 0 at bed}; \]
\[ z_0 = \text{reference height}; \]
\[ \lambda_{wave} = \text{surface wave wavelength}; \]
\[ \nu' = \text{eddy viscosity above the wave bottom boundary layer}; \]
\[ \rho = \text{density of water}; \]
\[ \rho_{air} = \text{density of air}; \]
\[ \tau_{\tau}, \tau_{\tau} = \text{steady-current shear stress}; \]
\[ \tau_{wind} = \text{wind stress at free surface}; \]
\[ \phi_e = \text{angle between waves and mean current}. \]

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