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Wave attenuation in the shallows of San Francisco Bay

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ABSTRACT

Waves propagating over broad, gently-sloped shallows decrease in height due to frictional dissipation at the bed. We quantified wave-height evolution across 7 km of mudflat in San Pablo Bay (northern San Francisco Bay), an environment where tidal mixing prevents the formation of fluid mud. Wave height was measured along a cross-shore transect (elevation range −2 m to +0.45 m MLLW) in winter 2011 and summer 2012. Wave height decreased more than 50% across the transect. The exponential decay coefficient λ was inversely related to depth squared ($\lambda = 6\times10^{-4}h^{-2}$). The physical roughness length scale k_b, estimated from near-bed turbulence measurements, was 3.5×10^{-3} m in winter and 1.1×10^{-2} m in summer. Estimated wave friction factor $\hat{f}_w$ determined from wave-height data suggests that bottom friction dominates dissipation at high Re_w but not at low Re_w. Predictions of near-shore wave height based on offshore wave height and a rough formulation for f_w were quite accurate, with errors about half as great as those based on the smooth formulation for f_w. Researchers often assume that the wave boundary layer is smooth for settings with fine-grained sediments. At this site, use of a smooth f_w results in an underestimate of wave shear stress by a factor of 2 for typical waves and as much as 5 for more energetic waves. It also inadequately captures the effectiveness of the mudflats in protecting the shoreline through wave attenuation.

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1. Introduction

In water depths sufficiently shallow that wave motions reach the bed, wave energy is frictionally dissipated at the sea floor. The resulting reduction in wave height can be significant over broad tidal flats with shallow slope. In San Francisco Bay, as in many coastal plain estuaries, intertidal and subtidal mudflats extend for kilometers between the shoreline and deep central channels. Sediment in these broad shallows is regularly mobilized by the combination of waves and tidal currents (Brand et al., 2010; MacVean and Lacy, 2014; Lacy et al., 2014). The correct characterization of local shear stress is an important component of numerical modeling of sediment resuspension and transport in this environment, and wave shear stress is influenced by the cross-shore evolution of wave height. Wave attenuation over shallows also influences the impact of waves on the shoreline, and thus is an important component of predicting changing conditions at the shore as sea level rises.

The degree of frictional dissipation of waves depends on bottom type and wave energy. On sandy sea floors, bed roughness and drag vary with grain size and bedform morphology. At some sites with cohesive sediments, the formation of fluid mud during and after storms produces dramatic attenuation of waves (e.g. Elgar and Raubenheimer (2008)). The shallows of San Francisco Bay exemplify a third type of setting, in which the bed is fine grained, yet tidally-driven vertical mixing prevents resuspended sediments from forming a thick fluid mud layer (Brand et al., 2010; MacVean and Lacy, 2014). These conditions occur in many estuaries, yet the cross-shore evolution of waves in such settings is not well documented. Previous investigations of wave-height attenuation across sand or mud flats include Möller et al. (1999), Le Hir et al. (2000), Cooper, (2005), and Houser and Hill (2010).

Wave-height attenuation is commonly characterized by an exponential model of decay over distance:

$$a = a_0 e^{-\lambda x} \tag{1}$$

where a is wave-height amplitude, a_0 is incident wave-height amplitude, and x is distance from the incident wave location in the direction of propagation. The attenuation coefficient λ varies with site and wave conditions because of their influence on the processes affecting wave height, including shoaling, refraction, bottom friction, and viscous dissipation.

The influence of bottom friction on waves is characterized by the wave friction factor f_w, which relates maximum wave shear stress τ_w to wave orbital velocity u_b (Jonsson, 1966):

$$\tau_w = \frac{1}{2}\rho f_w u_b^2 \tag{2}$$

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where ρ is water density. For a laminar wave boundary layer it can be shown analytically that f_w is a function of the wave Reynolds number Re_w alone:

$$f_w = 2\ Re_w^{-0.5} \tag{3}$$

where $Re_w = A^2\omega/\nu$, $A = u_bT/2\pi$ is near-bed wave amplitude or semi-excursion, T is wave period, $\omega = 2\pi/T$ is angular frequency, and ν is kinematic viscosity (Jonsson, 1966). In contrast, for a rough-turbulent wave boundary layer f_w is independent of Re_w and varies with relative roughness A/k_b, where k_b is the physical roughness length scale. Several formulae for f_w in rough flows have been empirically developed from laboratory studies, including the widely used expression (Nielsen, 1992):

$$f_w = \exp\left[5.213\left(\frac{k_b}{A}\right)^{0.194} - 5.977\right]. \tag{4}$$

Classification of the wave boundary layer as smooth-laminar, rough-turbulent, or transitional is based on the two nondimensional quantities Re_w and A/k_b. Thresholds between the regimes were determined empirically from comprehensive laboratory studies in the 1960s and 1970s (Jonsson, 1966; Kamphuis, 1975).

This paper describes observations of wave-height evolution across 7 km of mudflat in San Pablo Bay, in northern San Francisco Bay. The goals are to quantify rates of attenuation and investigate the processes and physical factors governing attenuation, to inform numerical modeling of sediment resuspension and transport over mudflats impacted by waves. We characterize the dependency of wave attenuation on water depth, and the variation of the wave friction factor f_w (determined empirically) with wave energy. The physical roughness length scale k_b is estimated from hydrodynamic data. We compare the accuracy of rough or smooth formulations for f_w for predicting the observed wave-height attenuation. Finally, we discuss implications of the results to morphology and shoreline response to sea-level rise.

2. Methods

2.1. Field data collection

The study was conducted in the shallows of San Pablo Bay, where broad inter- and sub-tidal mudflats extend for more than 10 km from the shoreline to a depth of −3 m relative to mean lower-low water (MLLW). Wave characteristics were measured along a cross-shore transect during two deployments: February 3–March 17, 2011 (winter) and June 2–29, 2012 (summer) (Fig. 1). In both deployments the offshore end of the transect was located approximately 7 km from shore at an elevation of −2 m MLLW, and the landward end of the transect was intertidal (0.45 m MLLW). Average bed slope across the transect was 3.8×10^{-4}. Bed sediments were predominately mud and varied little between seasons or along the transect. In winter the median disaggregated grain size in nine grab samples was 7–11 μm, and the percent sand ranged from 2 to 8%; in summer the median grain size in eight samples was 8–18 μm, with 5–13% sand. Tides in San Pablo Bay are mixed semi-diurnal with a spring-tide range of 2.5 m.

Water-surface elevation and wave height were measured with high-frequency (6 Hz or faster) pressure sensors at all stations (4 in winter, 5 in summer) at intervals ranging from 15 to 60 min. In addition, acoustic Doppler velocimeters (ADVs) collected high-frequency (10 Hz) velocity bursts at three stations in winter and two stations in summer (Table 1). Pressure data were corrected for atmospheric pressure and converted to water depth. Root-mean-square (RMS) wave height H was calculated from the pressure frequency spectra, with correction for attenuation with depth below the water surface. Wave statistics including H, and representative wave period T, direction θ, and bottom orbital velocity u_b were calculated from the spectra of velocity bursts following Madsen (1994) and Wiberg and Sherwood (2008). At stations W2, W3, W4, and S5, we used wave heights calculated from velocity spectra in our analysis, because of irregularities in the high-frequency pressure data. Significant wave height was calculated as $H_s = \sqrt{2}H$.

For examining attenuation, burst frequency was reduced to the lowest frequency in the transect: hourly in winter and every 30 min in summer. In addition, data were limited to bursts with $H > 0.1$ m at the offshore station (W1 or S1) and wave direction between 140° and 200°, measured at W2 or S4. Data from W4 and S5 were only included at times when the local water depth was greater than 0.45 m.

2.2. Determination of roughness length scale k_b

To determine whether the wave boundary layer was rough or smooth we used the flow regimes in the parameter space of A/k_b vs. Re_w defined by Kamphuis (1975), which requires specification of the physical roughness length scale k_b. Bed sediments were fine mud, without organized bedforms; however, it is likely that there was biogenic roughness in the form of irregular depressions and/or benthic organisms (for example, *Ampelisca abdita* tubes were present in some sediment samples). Therefore, rather than estimating k_b from grain size and bedform dimensions, we determined k_b from the hydrodynamic roughness of the current boundary layer z_0.

The relationship between k_b and z_0 depends on conditions in the current boundary layer. For hydraulically rough flows $z_0 = k_b/30$, whereas for hydraulically smooth flows z_0 is a function of the thickness of the viscous sublayer. The current boundary layer can be classified based on the nondimensional roughness $k_* = u_{*c}k_b/\nu$ as smooth for $k_* < 5$, rough for $k_* > 70$, or transitional (Nielsen, 1992). More recent research suggests that the lower threshold for hydraulically rough is $k_* = 25$ (Schulz and Flack, 2007).

Hydrodynamic roughness z_0 was determined from high-frequency velocity data collected near the bed by ADVs. In winter, data were used from ADVs mounted at two elevations above the bed at W3, and one elevation at W2 (Table 2). We also estimated z_0 from data collected by one ADV at S4 and one at S4A in summer. In the ADV burst data, turbulence and waves were decomposed following Bricker and Monismith (2007) as described in MacVean and Lacy (2014), and friction velocity due to currents was calculated as $u_{*c} = \left(\overline{u'w'} + \overline{v'w'}\right)^{0.25}$. Friction velocity at the bed u_{*cb} was determined assuming the vertical distribution of shear stress in the logarithmic layer, $\tau(z) = \tau_b(1 - z/h)$, with $\tau = \rho u_{*c}^2$. The elevation of the velocity measurements z was taken from the distance to the bed measured by the ADV for each burst, so variations during the deployment due to platform settling and other factors were taken into account. z_0 was calculated from the law of the wall:

$$u(z) = \frac{u_{*cb}}{\kappa}\ \ln\left(\frac{z}{z_0}\right) \tag{5}$$

where $\kappa = 0.41$ is the von Karman constant.

Eq. (5) was applied to a restricted set of data, to improve the estimate of z_0. We required $z/h < 0.3$, so that measurements were likely to be within the logarithmic layer. To minimize the influence of waves on z_0, data were restricted to bursts with $u_b < 0.05$ m/s. Data were also chosen to avoid hydraulically smooth conditions. Since k_b (the unknown) is required to determine when the flow is smooth, we made an initial guess of an appropriate restriction of $0.006 < u_{*cb} < 0.012$ m/s, based on the observation that bin-averaged z_0 was relatively constant in all five data sets for this range of u_{*cb}. Log-mean values of z_0 for bursts meeting the restrictions on u_{*cb} and u_b were quite consistent for the three data sets from winter, whereas the two summer estimates were greater (Table 2). We determined separate k_b values for winter and summer, using $z_{0w} = 1.3 \times 10^{-4}$ for winter and $z_{0s} = 3.9 \times 10^{-4}$ for summer.