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Daniel D. Sternlicht  
*University of California - San Diego*

Christian de Moustier  
*University of California - San Diego*

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Time-dependent seafloor acoustic backscatter (10–100 kHz)\textsuperscript{a)

Daniel D. Sternlicht\textsuperscript{b) and Christian P. de Moustier\textsuperscript{c)}

Marine Physical Laboratory, Scripps Institution of Oceanography, University of California at San Diego, La Jolla, California 92039-0205

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A time-dependent model of the acoustic intensity backscattered by the seafloor is described and compared with data from a calibrated, vertically oriented, echo-sounder operating at 33 and 93 kHz. The model incorporates the characteristics of the echo-sounder and transmitted pulse, and the water column spreading and absorption losses. Scattering from the water–sediment interface is predicted using Helmholtz–Kirchhoff theory, parametrized by the mean grain size, the coherent reflection coefficient, and the strength and exponent of a power-law roughness spectrum. The composite roughness approach of Jackson \textit{et al.} [J. Acoust. Soc. Am. \textbf{79}, 1410–1422 (1986)], modified for the finite duration of the transmitted signal, is used to predict backscatter from subbottom inhomogeneities. It depends on the sediment’s volume scattering and attenuation coefficients, as well as the interface characteristics governing sound transmission into the sediment. Estimation of model parameters (mean grain size, roughness spectrum strength and exponent, volume scattering coefficient) reveals ambiguous ranges for the two spectral components. Analyses of model outputs and of physical measurements reported in the literature yield practical constraints on roughness spectrum parameter settings appropriate for echo-envelope-based sediment classification procedures. © 2003 Acoustical Society of America. [DOI: 10.1121/1.1608018]

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I. INTRODUCTION

In the typical bottom echo-sounding geometry, a sound pulse is transmitted by a sonar system and the time of arrival of the echo provides a measure of the altitude of the sonar above the bottom. However, the shape and duration of the echo are often very different from the original pulse and these distortions contain information about the seafloor acoustic backscattering process as well as the relief and geoacoustic properties of the bottom. By comparing a time-dependent physical model of the acoustic intensity backscattered by the bottom with data from a calibrated echo-sounder operating at 33 and 93 kHz, we intend to infer physical characteristics of the bottom such as the roughness of the water–sediment interface and the mean grain size of the sediment, and geoacoustic properties such as the sediment’s volume scattering coefficient and its acoustic impedance.

The bottom echo intensity envelope model described in this paper is a temporal implementation of the SONAR equation,\textsuperscript{1,2} based on acoustic backscatter models described in Jackson \textit{et al.},\textsuperscript{3} and analytical tools developed by de Moustier and Alexandrou\textsuperscript{4} for modeling seafloor echoes measured with multibeam seafloor mapping sonars. No new theories are presented; rather, a specific combination of physical acoustic backscatter models, geoacoustic parameters, and echo processing techniques is used to estimate the shape and intensity of the average bottom echo envelope measured with simple calibrated echo-sounders. Similar temporal models described by others include: (a) time-dependent representations of surface and volume scattering originally formulated in “Physics of Sound in the Sea;”\textsuperscript{1} (b) temporal seafloor scattering models for swath-mapping sonars developed in Morgera,\textsuperscript{5} Morgera and Sankar,\textsuperscript{6} and de Moustier and Alexandrou;\textsuperscript{4} (c) comprehensive treatment of the time dependence of signals scattered by rough surfaces given by Berry,\textsuperscript{7,8} Berry and Blackwell,\textsuperscript{9} and Haines and Langston;\textsuperscript{10} (d) and, most specifically, average echo envelope models by Nesbitt,\textsuperscript{11} Jackson and Nesbitt,\textsuperscript{12} de Moustier and Alexandrou,\textsuperscript{4} and Pouliquen and Lurton,\textsuperscript{13,14} based on combining energy backscattered from the sediment surface and subbottom. Our approach differs from those presented in Refs. 4, 11–14 in its attempt to model and match absolute sound-pressure echo levels measured with a fully calibrated sonar system. Most of the model parameters used are common to Refs. 1–14, with some variations due to choice of physical scattering models detailed herein; however, we propose a specific combination of these parameters in a numerical implementation that incorporates the digitized transmitted waveform, and formulations for reducing the number of geometric dimensions during synthesis of the average backscatter echo envelope. The attempt to obtain absolute backscatter levels related to measurable geoacoustic parameters sets this work apart from mostly phenomenological sediment characterization approaches used in commercial systems,\textsuperscript{15,16} which rely on nonparametric classifiers to separate the various substrates, and require independent identification of the substrate by video or core sampling.

At the acoustic wavelengths of interest here (respectively, 4.5 and 1.6 cm at 33 and 93 kHz) the generalized Rayleigh criterion for diffuse scattering of a monochromatic
sound wave, with wavelength $\lambda_o$ and wave number $k_o = 2\pi/\lambda_o$, on a surface with rms height deviation $\sigma$ about a mean plane

$$2k_o\sigma \gg 1,$$  \hspace{1cm} (1)

implies that the rms height deviations $\sigma$ must exceed 1 cm. This is satisfied in the data presented here for sandy and silty substrates in San Diego Bay, and we can use a modeling approach based entirely on incoherent scattering. Also, at these frequencies, penetration of the bottom is limited to the first few meters; therefore, assumptions of bottom homogeneity are more likely to be valid. Based on the assumptions that interface scattering dominates the return at normal incidence and volume scattering dominates at oblique incidence, temporal separation of these components is achieved by informed selection of transducer beamwidth and orientation.

The model incorporates the characteristics and geometry of the echo-sounder: the transducer’s beam pattern, its altitude, its tilt with respect to vertical, the characteristics of the transmitted sound pulse, and the roll and pitch angles of the platform to which the transducer is mounted; and environmental factors: spherical spreading and absorption losses as the signal propagates through the water column, backscattering of the signal at the water–sediment interface, and by inhomogeneities in the sediment volume.

The mean grain size, defined as $(M_d = -\log_2 D_p)$ where $D_p$ represents the mean particle diameter in millimeters, is an important component of the sediment’s compressional sound speed, saturated bulk density, and attenuation constant, and is of particular interest as it roughly correlates to these parameters through a set of experimentally derived linear regression equations developed by Hamilton$^{17}$ and refined by others.$^{18,19}$

Following Jackson et al.,$^3$ Nesbitt,$^{11}$ Nesbitt and Jackson,$^{12}$ de Moustier and Alexandrou,$^4$ and Pouliquen and Lurton,$^5$ the acoustic backscattering at the water–sediment interface and in the sediment volume are modeled and computed independently, then summed to estimate the overall echo intensity measured by the echo-sounder (Fig. 1). Formally, the time-dependent intensity of the bottom backscattered acoustic signal $I(t)$ measured at the transducer’s face equals the sum of the intensity $I_i(t)$ backscattered at the water–sediment interface and the intensity $I_v(t)$ backscattered from the sediment volume

$$I(t) = I_i(t) + I_v(t).$$ \hspace{1cm} (2)

This paper provides a detailed description of the model and of its sensitivity to changes in its parameters. Data examples are presented with the echo alignment and ensemble averaging processes required to compare measured and modeled echo envelopes. Then, a metric is developed to quantify the closest fit between model and data from which seafloor geoacoustic parameters can be inferred, and the potential for ambiguous results is discussed. The implementation of this model for automatic geoacoustic parameter estimation is the subject of another paper by the authors.$^{20}$

II. MODEL

This section describes the components of Eq. (2), which define the temporal model of high-frequency acoustic seafloor backscatter.

A. Sediment interface characteristics

The analytical form of the signal component backscattered by the water–sediment interface is based on a solution of the Helmholtz diffraction integral for monochromatic sound waves. It uses the Kirchhoff approximation to express the pressure field at planes that are locally tangent to the interface, and the reflection coefficient, $R(\theta_i)$ at angle of incidence $\theta_i$, is assumed constant at each point on the interface and equal to the normal incidence reflection coefficient $R_n = R(0)$. The reflection coefficient is a function of the ratios of sediment saturated bulk density over water mass density ($\rho$) and sediment sound speed over water sound speed ($v$)

$$R(\theta_i) = \frac{\rho v \cos(\theta_i) \left[ 1 - (\cos(\theta_i))^2 \right]^{1/2}}{\rho v \cos(\theta_i) \left[ 1 - (\sin(\theta_i))^2 \right]^{1/2}}.$$ \hspace{1cm} (3)

Given that we are concerned with acoustic backscatter measurements near normal incidence, shadowing and multiple scattering effects are neglected.

The relief statistics of the sediment interface are assumed to be isotropic and described by Gaussian-distributed height deviations ($\xi$) about a mean plane, with rms height $\sigma$, and by a 2D energy density spectrum. Since $\xi$ is usually not stationary, and since the choice of a reference surface is
somewhat arbitrary, it is useful to consider the height difference between points on the surface separated by a fixed distance (horizontal vector \( \mathbf{r} \)). This height difference is a locally stationary random variable whose mean-squared value is the structure function \( D(\mathbf{r}) \)

\[
D(\mathbf{r}) = \langle (\zeta(\mathbf{r} + \mathbf{r}_0) - \zeta(\mathbf{r}_0))^2 \rangle.
\] (4)

As described in Appendix B of Ref. 21, the relief energy density spectrum \( W(\mathbf{k}) \) for the 2D spatial wave vector \( \mathbf{k} \) of magnitude \( k \), and the structure function are related by

\[
D(\mathbf{r}) = 2 \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} (1 - \cos(\mathbf{k} \cdot \mathbf{r})) W(\mathbf{k}) d^2 k.
\] (5)

Power-law relief spectra have been measured over a wide range of spatial wavelengths, from geographic scales (meters–kilometers)\( ^{22-24} \) to centimeter scales commensurate with high-frequency acoustic wavelengths.\( ^{25-28} \) The relief energy density spectrum \( W(\mathbf{k}) \) is formulated as\( ^{3,19} \)

\[
W(\mathbf{k}) = w_2 k^{-\gamma},
\] (6)

where the spectral strength \( w_2 \) has units of length to the power \((4 - \gamma)\), whereas the spectral exponent \( \gamma \) is unitless. For a reference length \( h_0 \), \( W(\mathbf{k}) = w_2(h_0 k)^{-\gamma} \). This removes dimensionality from the power term, and gives \( w_2 \) the dimensionality of \( h_0^\gamma \). In this work \( h_0 = 1 \text{ cm} \), and \( w_2 \) is expressed in \text{cm}^{-1}.

Spectral analyses of centimeter-scale topography have yielded 2D spectral exponents in the range \( 2 < \gamma < 4 \) (Refs. 25–28). In the absence of measurements, a mean value of \( \gamma = 3.25 \) can be assumed for most bottom types.\( ^{19} \) In practice, \( \gamma \) and \( w_2 \) are determined by straight-line regression fit to the 2D spectra plotted on log–log scale, over spatial wave numbers spanning roughly an order of magnitude above and below the acoustic wave number.

Integration of Eq. (5), after substituting in Eq. (6), yields a simple expression for the structure function\( ^{3} \)

\[
D(\mathbf{r}) = C_\xi^2 r^{2\gamma},
\] (7)

with the structure constant \( C_\xi \) given by

\[
C_\xi^2 = \left[ 2 \pi w_2 \Gamma(2 - \alpha) 2^{-2\alpha} \right] \left[ \alpha(1 - \alpha) \Gamma'(1 + \alpha) \right],
\] (8)

where \( \Gamma \) is the gamma function, and \( \alpha = (\gamma/2) - 1 \) is constrained between 0 and 1.

Following Jackson et al.\(^ {3} \), the bandlimited power-law relief spectrum is separated into large- and small-scale roughness components with the boundary defined by a cutoff spatial wave number \( k_c \)

\[
k_c = \left( \frac{\gamma - 2}{8 \pi w_2 k_0^2 \cos^2 \theta_i} \right)^{1/(2 - \gamma)}.
\] (9)

Within the bandlimited large-scale relief, we shall use also the mean-square slope \( s^2 \) to compute sound energy transmission into the sediment, and the mean-square curvature \( R_c^2 \) for the Kirchhoff criterion

\[
R_c^2 = \frac{2 \pi k_c^4}{2 \pi w_2 k_c^2} \int_0^{k_c} W(\mathbf{k}) k^2 dk = \frac{2 \pi w_2 k_c^{4 - \gamma}}{4 - \gamma}.
\] (11)

B. Sediment volume characteristics

At 10–100 kHz, penetration of water-saturated sediments is limited to the first several meters, and the bottom can be modeled as a lossy fluid with acoustic scattering due to fluctuations in the density and refraction indices of the medium. In addition, the intensity of a plane monochromatic sound wave entering the sediment is attenuated with distance. The corresponding attenuation coefficient \( \alpha_b \) in dB/m is usually frequency dependent

\[
\alpha_b = \kappa_p \mu_d^m,
\] (12)

where \( \kappa_p \) is an attenuation constant expressed in dB/m/kHz, and \( f_p \) is the acoustic frequency in kHz. For unconsolidated ocean sediments of the type presented here, we follow Hamilton’s observations of \( m = 1 \) and employ his regression equations\(^ {29} \) relating \( \kappa_p \) to the mean grain size \( (M_0) \).

Ivakin and Lysanov\(^ {30} \) described a sediment volume backscatter model which includes the dissipative effects of a lossy medium and the transmission characteristics of a randomly rough interface. They used bottom slope variations to compute the energy transmitted through the interface. Jackson et al.\(^ {3} \) combined these ideas with Stockhausen’s\(^ {31} \) scheme for subbottom acoustic scattering below a flat surface, by integrating transmission terms over a Gaussian distribution of incidence angles whose mean value was determined by the rms slope \( (s) \). This approach is a variant of the composite roughness scattering model described by McDaniel and Gorman.\(^ {32} \) Here, we modify the composite roughness treatment to account for the finite duration of the transmitted signal.

C. Angle-dependent seafloor acoustic backscatter

Following the results of Jackson et al.\(^ {3} \), the angular-dependent backscatter coefficient \( (s_i) \) at the water–sediment interface is given by

\[
s_i(\theta_i) = \left\{ \begin{array}{ll}
R_c^2 (8 \pi \cos^2 (\theta_i) \sin^2 (\theta_i))^{-1} \int_0^\infty \exp(-qu^2) \\
\times J_o(u)u du & \theta_i > 0 \\
R_c^2 (8 \pi \alpha)^{-1} C_\xi^{-2\alpha} (2k_c^2 (\alpha - 1)/2)^{1/2} & \theta_i = 0,
\end{array} \right.
\] (13)

with

\[
q = \cos^2 (\theta_i) \sin^{-2\alpha} (\theta_i) C_\xi^{2\alpha - 1} \mu_d^{2(1 - \alpha)}.
\] (14)

In this formulation, \( J_o \) is the zeroth-order Bessel function of the first kind, and the normal incidence term is as derived in Ref. 4.

In the sediment volume, we use the unitless acoustic backscatter cross section per unit area per unit solid angle \( s_v(\theta_i) \), defined by

\[
s_v(\theta_i) = \sigma_2 \nu_i(\theta_i)/A, \quad \text{with} \quad A = 2/(10 \log_{10} e),
\] (15)

where \( \nu_i(\theta_i) \) accounts for the two-way transmission losses at the water–sediment interface with large-scale roughness hav-
ing small rms slope. This expression is identical to the large-scale volume scattering cross section described in Jackson et al.,
where $\sigma_2 = \sigma_v / \alpha_b$ is termed the volume parameter, and $\sigma_v$ is the volume scattering coefficient.

Stockhausen\textsuperscript{31} gave an angle-dependent expression for two-way transmission through a flat water–sediment interface in terms of the plane-wave coherent reflection coefficient $R$ and the sound-speed ratio $v$ at the interface

$$V_i(\theta_i) = [1 - R^2(\theta_i)]^2 \cos^2(\theta_i)[1 - (v \sin(\theta_i))^2]^{-1/2}.$$  \hspace{1cm} (16)

However, at 10–100 kHz very few water–sediment interfaces are likely to appear flat, and the effects of interface roughness must be considered in the two-way transmission through the interface. This is done by considering the distribution of incidence angles expected for large-scale roughness around normal incidence with respect to the local surface is $\theta_i - \vartheta$.

If we assume that the slopes are Gaussian distributed about the horizontal plane, with small rms slopes ($s < 0.1$), then the expression for large-scale interface roughness, $V_l$, is the average of the flat surface coherent reflection coefficients at each planar slope facet of the rough surface

$$V_l(\theta_i) = \frac{1}{\sqrt{\pi s}} \int_0^\infty V_i(\theta_i - \vartheta) \exp\left(-\frac{\vartheta^2}{s^2}\right) d\vartheta.$$  \hspace{1cm} (17)

Angle-dependent backscatter curves ($S(\theta_i)$) for sand ($M_\phi = 2$), silt ($M_\phi = 4$), and clay ($M_\phi = 7$) substrates are computed for acoustic frequencies of 33 and 93 kHz, and plotted in Fig. 2, where generic values for the sediment geometric parameters are correlated to mean grain size by the relationships described in Appendix A. These plots illustrate the dominance of interface scattering around normal incidence, giving way to volume scattering for angles exceeding 5–10 degrees. However, sand substrates exhibit a more gentle curve slope, whereas clay substrates exhibit large interface scattering strengths about normal incidence, falling off sharply as $\theta_i$ increases. As larger particles are more likely than fines to settle in high-energy hydrodynamic environments, the relief energy density spectrum of coarse-grain sediments (small $M_\phi$) typically has more energy (higher spectral strength $w_2$) than that of fine-grain sediments. An increase in $w_2$ causes a commensurate increase in the cutoff spatial wave number $k_c$ [Eq. (9)]; hence, higher spatial frequencies are included in the theoretical large-scale surface roughness. It follows that the estimated rms slope ($s$) of the interface increases, yielding more backscattered energy at higher angles and a backscatter curve with a gentler slope. While increasing roughness, relative to the acoustic wavelength, reduces the normal incidence component, this trend is typically offset by a commensurate rise in the impedance contrast ($\rho v$), resulting in $S(0)$ values which are similar for each of the three substrates.

D. Time-dependent sediment interface backscatter

Consider a monostatic transducer at altitude ($a$) which emits a pulse of duration $\tau_p$ seconds, with an intensity time series $I_x(t), 0 \leq t \leq \tau_p$. The energy in the pulse propagates as a spherical shell with sound speed $v_w$. The intersection of this shell with the bottom initially takes the shape of a disk, changing to that of an annulus. For a level surface, the pulse impacts nadir at time $a/v_w$, and the annulus diverges from this point of origin. The bottom projection of the pulse at some time $t > a/v_w + \tau_p$ is illustrated in Fig. 3.

The length of the pulse in water is $\nu_w \tau_p$, and its leading and trailing edges make angles $\theta_a$ and $\theta_t$, respectively, with nadir. The average echo intensity envelope $I_x(t)$ of the backscattered signal begins at the time of bottom detect, $t_{bd}$, $= 2a/v_w$. The backscattered energy in the scenario of Fig. 3

![FIG. 2. Angular dependent seafloor acoustic backscatter at 33 and 93 kHz. Top row: total backscatter strength $S = 10 \log_{10}(S_i + S_v)$ over silt, and component backscatter strengths $S_i = 10 \log_{10}(S_i)$ for the interface term [Eq. (13)], and $S_v = 10 \log_{10}(S_v)$ for the volume term [Eq. (15)]. Bottom row: total backscatter strength $S$ over sand, silt, and clay.](image-url)
reaches the receiver at \( t = t_{bd} / \cos \theta_i \). The elemental area \( dA \) is located at incidence angle \( \theta_i \), azimuthal angle \( \phi \), and range \( R = a/\cos(\theta_i) \) such that

\[
dA = R^2 \sin(\theta_i) d\theta_i d\phi.
\]

Similar to sonar equation formulations described inRefs. 4, 11, 13, and 33, the total acoustic intensity field backscattered by the interface and received at the transducer is evaluated over the angular sectors \( \theta_1 \leq \theta_i \leq \theta_2 \) and \( 0 \leq \phi < 2\pi \) by convolving the transmitted waveform \( I_x(t) \) with the bottom backscattering coefficient \( s_i(\theta_i) \) bound by the transducer beam pattern \( b(\theta_i, \phi) \)

\[
I_x(t) = \int_{\phi=0}^{2\pi} \int_{\theta_1}^{\theta_2} I_x(t - t_{bd} \frac{2R}{v_w} ) b(\theta_i, \phi) \frac{1}{R^4} \frac{\cos^2 \frac{\theta_i}{2}}{\cos(\theta_i)} dA,
\]

where

\[
\theta_1(t) = \begin{cases} \cos^{-1} \left( \frac{2a}{v_w (t - \tau_p)} \right) & t - \tau_p \geq t_{bd} \\ 0 & t - \tau_p < t_{bd} \end{cases}
\]

and

\[
\theta_2(t) = \begin{cases} \cos^{-1} \left( \frac{2a}{v_w t} \right) & t \geq t_{bd} \\ 0 & t < t_{bd} \end{cases}
\]

and terms in the denominator of Eq. (19) account for spherical spreading and absorption in the water column during the round-trip travel of the pulse between the transducer and the seafloor. \( \alpha_w \) is the frequency-dependent absorption coefficient in water. In the ideal case of a perfectly rectangular transmit pulse, the transmitted intensity \( I_x(t - [2R/v_w]) \) may be replaced by the constant \( I_x \) and moved outside the integral.

For sediment interface scattering we use the Kirchhoff approximation with the power-law roughness spectrum to calculate the bottom backscattering coefficient \( s_i(\theta_i) \). This approach has a frequency dependence based on power-law seafloor roughness spectra. This is in contrast to the geometrical optics approach which relies on the high-frequency limit of the Kirchhoff approximation, as described by Beckmann and Spizzichino and Brekhovskikh and

![FIG. 3. Geometric representation of the elemental area \( dA \) and volume \( dV \) used to compute the time-dependent echo intensity envelope.](image3)

![FIG. 4. Integration of backscattered energy from an elemental volume tube: (a) Volume scattering geometry, where scatterers within the tube are ionized from distances \( l_1 \) to \( l_2 \); (b) illustration of transmitted pulse and juxtaposition with region described in (a).](image4)

Lysanov, and used in Pouliquen and Lurton as a frequency-independent interface scattering component. This high-frequency limit does not exist in the power-law spectrum formulation chosen here. Although each approach has been used successfully in prior work, as we chose the frequency-dependent path because the power-law spectrum is linked to the relief statistics [Eq. (4)] of the types of sediments considered here.

Discrete formulation and implementation of our time-dependent sediment interface backscatter model is described in Appendix B 1.

E. Time-dependent sediment volume backscatter

Single scattering is assumed and the statistics governing subbottom inhomogeneities are assumed to be isotropic and homogeneous, leading to a constant sediment volume scattering coefficient \( \sigma_v \).

Figure 4(a) shows a closeup of the elemental surface \( dA \) at time \( t > (R/v_w) + \tau_p \), where a portion of the incident energy has refracted into the sediment, and the edges of the pulse have propagation distances \( (l_1, l_2) \), referenced from the point of entry into the sediment.

We express the acoustic scattering from the sediment volume as a convolution of the transmit signal \( I_x(t) \) with the volume scattering characteristics along the propagation path, taking into account the effects of transducer directivity and of round-trip spherical spreading and absorption in the water column. At time \( t \), the total sediment volume backscattering
field \( I_x(t) \) is the sum of all the contributions from volume tubes of cross section \( dA \) that converge at the transducer’s face. It is expressed in integral form, over the angular intervals \( 0 \leq \theta \leq \theta_2 \) and \( 0 \leq \psi \leq 2\pi \), as

\[
I_x(t) = \int_{\theta=0}^{2\pi} \int_{\theta=0}^{\theta_2} \sigma_v b^4(\theta, \psi) V_I(\theta) R^4 10^{\alpha_v R/5} d\theta d\psi.
\]

\[ \times \left\{ \int_0^{l_2(t)} \int_{l_1(t)}^{l_2(t)} v_p \frac{1-l_1(t)}{v_b} e^{-2\beta_v l} \right\} dA, \quad (21) \]

with \( \theta_2(t) \) defined in Eq. (20), and

\[
l_1(t) = \begin{cases} 
\frac{t}{2} - \frac{R}{v_w} & t \geq 2 \frac{R}{v_w} + \tau_p \\
0 & t < 2 \frac{R}{v_w} + \tau_p
\end{cases}
\]

and

\[
l_2(t) = \begin{cases} 
\frac{t}{2} - \frac{R}{v_w} & t \geq 2R/v_w \\
0 & t < 2R/v_w
\end{cases}
\]

Equation (21) is valid from normal incidence to the critical angle.

Spherical spreading within the elemental tubes is considered negligible, and the assumption of statistical volume homogeneity allows placement of \( \sigma_v \) outside the bracketed convolution integral. The integrand includes a spatial representation of \( I_x(t) \), tempered by the absorption characteristics of the sediment. The round-trip attenuation between scattering center \( l \) and the water–sediment interface is expressed with the exponential intensity attenuation rate \( \beta_v \), which has units of power nepers per unit distance.

In model computations, the transmitted waveform is digitized and the bracketed integrand of Eq. (21) is evaluated numerically at each time \( t \). If one assumes a perfectly rectangular pulse, \( I_x \) may be moved outside of the integral, and the remaining expression evaluates to an attenuation length \( L(t) \), which we define as

\[
L(t) = \int_{l_1(t)}^{l_2(t)} e^{-2\beta_v l} dl = \frac{1}{2\beta_v} \left( e^{-2\beta_v l_1(t)} - e^{-2\beta_v l_2(t)} \right).
\]

Equation (22) is valid from normal incidence to the critical angle.

F. Influence of seafloor macro-roughness

The model described so far predicts the average echo intensity envelope measured with a monostatic transducer aimed at the bottom. The model includes the ratio of scattered to incident energy as the pulse traverses the water–sediment interface and penetrates the substrate. Volume scattering is treated in three dimensions and scattering at the interface is essentially two-dimensional because the ensemble of scattering elements \( dA \) are aligned along a mean flat horizontal plane at distance \( a \) from the transducer (Fig. 3). Significant deviation of bottom relief from this mean plane at the scale of a beam footprint may result in elongation of the echo, as illustrated in Fig. 5. Temporal stretching of the echo due to large-scale bottom roughness is incorporated by convolving model computations based on the small-scale relief with the height distribution of the interface relief as described by several authors.\(^9\)\(^{10}\)\(^{13}\)\(^{38}\)\(^{39}\) The specifics of our discrete implementation of this macro-roughness effect are described in Appendix B 4.
III. DATA

With a calibrated echo-sounder, bottom echoes were collected over substrates in San Diego Bay ranging from clay to sand. The acoustic frequencies $\sim 33$ and $93 \text{ kHz}$, transducer orientations (maximum response axis at $0^\circ$–$15^\circ$ incidence), and $-3 \text{ dB}$ beamwidths ($10^\circ$–$21^\circ$) of the system are consistent with the scattering theory incorporated in the temporal model. The transducer was elevated to a specified angle from nadir in the roll plane. Angles of pitch and roll were digitized for each ping and, combined with knowledge of local bathymetry, used to determine the incidence angle ($\theta_I$) of the transducer’s maximum response axis.

The waterfall and raster plots of Fig. 6(a) depict 100 consecutive bottom echoes from the San Diego Bay silt substrate measured at $33 \text{ kHz}$, with $\theta_I = 2^\circ$. With 5 pings per second at a vessel speed of 1 kn, a small bottom patch would typically be sampled 60 times over the $-3 \text{ dB}$ footprint of the beam; thus, spatial overlap of echoes between consecutive pings is greater than 98%.

Acoustic wavelengths at frequencies greater than $10 \text{ kHz}$ are generally small compared to the large-scale relief of the water–sediment interface, resulting in mostly incoherent scattering of the incident acoustic energy [Eq. (1)]. As seen in Fig. 6(a), bottom echoes are incoherent, varying significantly in amplitude and shape as the sonar translates longitudinally above the interface. Because of this variability, echoes must be treated stochastically.

The measured bottom echo consists of a pulsed CW signal modulated by the scattering process. Envelope detection of this signal yields an rms pressure time series, $p(t)$, expressed in units of pascals (Pa), and represented by the discrete sequence $p[n]$ when sampled with a period $\tau_e$. For comparison with the temporal model, an ensemble of $M$ contiguous returns is characterized by the average echo sequence $p_{a}[n]$ with $N$ samples, and a mean altitude ($\bar{a}$) representing the transducer–bottom distance along a vector normal to the water–sediment interface

$$p_a[n] = \frac{1}{M} \sum_{m=0}^{M-1} p[m,n], \quad n = 0, 1, ..., N - 1. \quad (24)$$

As we only have knowledge of the signal at amplitudes above the noise floor, $p_a[n]$ is truncated at both ends by application of a threshold minimum, $P_{th}$. The leading edge index corresponding to $P_{th}$ identifies the time of bottom detect ($t_{bd}$), allowing for calculation of the mean altitude

$$\bar{a} = \frac{v_a t_{bd}}{2}, \quad (25)$$

where straight-path association of $\bar{a}$ and $t_{bd}$ is a reasonable assumption for the modest transducer elevation angles char-
acterizing the measurements. This value of $\bar{\alpha}$ is used for generating the echo envelope model.

In the following sections the ensemble size ($M$) necessary for computing the average echo envelope is discussed, and alignment techniques compensating for vessel heave and topography variations are described.

A. Ensemble size

The left-hand-side plots of Fig. 7 represent the peak rms pressure (top) and the corresponding elevation angles measured with a clinometer for the 100 sequential echoes of Fig. 6(a). The energy density spectra displayed to the right indicate that most of the energy in the peak rms pressure sequence occurs at short spatial periods, and is minimal around 1.3 m—where the energy in the elevation angle spectrum is maximum. Hence, under stable survey conditions, changes in sensor attitude which are small relative to the beamwidth (standard deviation of 0.2° vs -3-dB beamwidth of 10°) have a negligible effect on the bottom echo statistics.

From analysis of a number of such data sets, we concluded that averaging over an echo ensemble covering the footprint of the -6-dB width of the beam (8.3 m in this scenario) provides adequate statistical representation. For the San Diego Bay measurements, $M = 100$ pings meets or exceeds this criterion.

B. Echo envelope alignment

Envelope averaging should be performed on echoes that have been aligned in time, thus removing the effects of transducer heave and of small depth variations over consecutive pings. In this section, we compare the effectiveness of two common alignment techniques based on (1) the threshold minimum ($P_{th}$) and (2) the echo peak amplitude, and we introduce a third method which exploits phase information in the echo’s discrete Fourier transform.

Echo alignment typically relies on tracking and indexing a temporal feature. The echoes within the ensemble are then shifted in time to line up on that feature. For the two-dimensional matrix $p[m,n]$ of Eq. (24), an alignment index ($j_m$) takes on integer values ($0 \leq j_m \leq N - 1$). A mean alignment index, $j$, defined by

$$j = \frac{1}{M} \sum_{m=0}^{M-1} j_m,$$

leads to a delay $d_m = j_m - j$ for each of the $M$ echoes. This yields the aligned array $p[m,(n-d_m)]$ that is substituted for $p[m,n]$ in (24) to compute the average echo sequence $p_{av}[n]$.

The temporal features presented here for comparison are labeled on the canonical echo envelope illustration of Fig. 8. This signal is characterized by a well-defined initial rise and peak amplitude, followed by a slow decay. The threshold minimum and peak amplitude indices associated with the times $T_t$ and $T_p$ are identified by serial search through the sequence, $p[n]$.

For echoes with poorly defined temporal features we employ $T_g$, which is determined by a process analogous to calculating the signal’s group delay. This method provides an alignment index based on energy contributions spanning the entire length of the return, rather than on a single temporal

FIG. 7. Amplitude and spectra of peak rms pressure and corresponding transducer elevation angle measured along survey track.

FIG. 8. Signal features used for echo envelope alignment offsets: threshold $T_t$ (rising edge), $T_p$ (peak), $T_g$ (group delay).
The panels of Figs. 6

The weighted phase

Echoes situated near the beginning of the time window have

where "arg" refers to the phase of the bracketed quantity.

responding to the

and

$\phi_n = 2\pi n / N - \pi$, $-\pi < \phi_n < \pi$.

(28)

The weighted phase $\phi$ for the energy in the echo envelope is
determined by

$$\phi = \arg \left\{ \sum_{n=0}^{N-1} |p[n]|^2 \right\} ,$$

(29)

where "arg" refers to the phase of the bracketed quantity. Echoes situated near the beginning of the time window have a negative $\phi$, whereas echoes situated toward the end of the window exhibit a positive $\phi$. The alignment index ($j_m$) corresponding to the $m$th ping in an ensemble is calculated as

$$j_m = \left( \frac{\phi + \pi}{2\pi} \right) (N-1) ,$$

(30)

yielding $T_g = j_m \tau_e$. We refer to this method as group delay alignment, as expanding Eq. (29) reveals a formula similar to the phase slope of the discrete Fourier transform evaluated at the first two Fourier coefficients.

The effects of applying alignment techniques to an ensemble of measured echoes are illustrated in Figs. 6 and 9. The panels of Figs. 6(b)–(d) represent the adjusted envelopes $p\{m,(n-d_m)\}$ resulting from peak, threshold, and group delay alignments of the data in Fig. 6(a). In the top panels of Fig. 9, average echo envelopes created from these data segments are plotted alongside temporal model computations for which bottom characteristics were determined from grab samples and video images. The signal to error ratio (S/E) quantifies the model-data fits, and is evaluated as follows:

1. Truncate the leading and trailing edges of the temporal model simulation, $\hat{p}_a[n]$, by applying the threshold minimum $P_{th}$.
2. Align $p_a[n]$ and $\hat{p}_a[n]$ along their respective $P_{th}$ leading edge indices.

(3) Truncate the tail end of the longer of the two time series such that $p_a[n]$ and $\hat{p}_a[n]$ are of equal length.
4. Define $n_1$ and $n_2$ as initial and final indices for both waveforms and calculate S/E according to

$$S/E = \frac{\sum_{n=n_1}^{n_2} p_a[n] - \hat{p}_a[n]}{\sum_{n=n_1}^{n_2} |p_a[n]|} ,$$

(31)

for which a high value represents a "good" match of model with data.

This figure of merit is independent of signal scale and length, and thus provides a convenient comparison of matching results across data sets.

As seen in Fig. 9(a), averaging "raw" data may result in a distorted rising edge—quantified by the low S/E ratio of 11 dB. Alignment by peak tracking [Figs. 6(b), 9(b)] yields a poor representation of the echo. In general, alignments based on signal enhancement techniques (e.g., peak tracking and matched filters) induce vertical disproportions, unsuitable for echo envelope matching.

Bottom echoes from substrates whose relief is small compared to the acoustic wavelength exhibit consistent temporal energy distributions, particularly when measurements are conducted near normal incidence. In these situations, stacking and averaging via minimum threshold [Fig. 6(c)] preserves the integrity of the echo's rising edge, as demonstrated by the 26-dB signal to error match of Fig. 9(c).

In comparison to the other alignment techniques, group delay alignment [Fig. 6(d)] yields a more symmetric distribution of signal energy about the alignment index [Fig. 9(d)], and is less likely to trigger on an early blip or anomalous peak. Threshold alignment may be ineffectual in high-noise environments, or when signal shapes are highly variable—for example, when echoes are measured from substrates which are extremely rough relative to the acoustic wavelength. Under these conditions, group delay alignment may yield average echoes which are more consistent with theoretical predictions. This is illustrated by the 22-dB model–data match of Fig. 9(h), where the data were collected at 93 kHz and oblique incidence.

Based on the above, we have relied on threshold mini-

FIG. 9. Comparison of temporal model output (dashed line), given a silt substrate at normal incidence for 33 kHz, and at oblique incidence for 93 kHz, with stacked and averaged echoes (solid lines) for the various alignment techniques. From Sternlicht and de Moustier (Ref. 43).
forth, the sediment–water mass and density ratios data for measurements at 33 kHz over a silt substrate. Hence-

density and sound speed.

A. Sensitivity of echo shape to model parameters

The influence of each parameter \( (M_\phi, \gamma, w_2, \sigma_v) \) on echo shape is determined by comparing recorded data to it-
erations of the temporal model. We start in Fig. 11(c) with the closest match between the average echo from another set of 100 sequential echoes measured over a silt bottom, and a model echo with best-fit parameters yielding an S/E of 29.3

dB. Figure 11(d) illustrates the angular dependence curve for a substrate with these parameters: \( S = 10 \log_{10}(s_i + s_{ul}) \), where interface and volume scattering coefficients, \( s_i \) and \( s_{ul} \), are described by Eqs. (13) and (15), respectively. All other parameters being held constant, each of the four parameters is increased [Figs. 11(a), (b)] or decreased [Figs. 11(e), (f)] relative to its best-fit value to evaluate its effect on the model output.

1. Grain size \( (M_\phi) \) influence on echo envelope

An increase in \( M_\phi \) is accompanied by a lower signal amplitude [Fig. 11(a)]. This is due to the smaller impedance contrast \( (\rho v) \), hence the lower backscatter strength \( (S) \), predicted for fine-grain sediments. The higher peak amplitude observed for lower \( M_\phi \) is due primarily to higher impedance contrast [Fig. 11(e)]. The increase of energy in the tail of the backscattered signal seen in Fig. 11(a) is explained by the decrease in \( (\kappa_p) \) predicted for fine-grain, water-saturated sediments. These theoretical relationships are supported by the observation that bottom echoes from sand substrates typi-
cally exhibit large peak amplitudes (high reflection coefficients), whereas echoes from fine-grain sediments are char-
acterized by long tails indicative of greater subbottom penetration.

2. Relief spectrum \( (\gamma, w_2) \) influence on echo envelope

Changes in either the spectral exponent \( \gamma \) or the spectral strength \( w_2 \) produce similar changes in the character of the backscattered echo. Their respective effects on the interface backscatter coefficient \( (s_i) \) depend on the range of spectral parameters considered. For the granular sediments studied in this work, the ranges likely to be encountered are \( (3.0 \leq \gamma \leq 3.6) \) and \( (0.0002 \leq w_2 \leq 0.01) \). Within these bound-
aries, increasing \( \gamma \) or \( w_2 \) has the effect of increasing the cutoff spatial wave number \( k_c \) [Eq. (9)]; hence, higher spatial frequencies are included in the theoretical large-scale surface roughness. Likewise, the estimated rms slope \( (s) \) of the bot-

FIG. 10. Example of bottom type aliasing. 33 kHz data (solid line): San Diego Bay silt site. Model (dashed line): \( \theta_3=8^\circ \) and "optimum" parameter combinations: (a) “correct” solution: \( M_\phi=4.76, \gamma=3.26, w_2=0.0012, \sigma_v = 0.091\): S/E=27.6 dB. (b) “aliased” solution: \( M_\phi=4.64, \gamma=3.00, w_2 = 0.0031, \sigma_v = 0.111\): S/E=31.4 dB.

IV. DATA–MODEL COMPARISON

Envelope detection, followed by the alignment and av-
eraging steps described in Sec. III, yields an average rms pressure sequence, \( p_i[n] \), whereas model computations yield an echo intensity. We chose to compare model and data as rms pressure sequences because the nonlinear conversion of the measured pressure signal in Pa to a power signal in W/m\(^2\) introduced complications in the echo alignment procedures and in the matching operations. Therefore, a temporal model estimate \( \hat{p}_i[n] \) is generated with specified mean altitude and sediment geoacoustic parameters

\[
\hat{p}_i[n] = \sqrt{\rho_w \nu_w \hat{I}_i[n]},
\]

where \( \hat{I}_i[n] \) represents the intensity of the computed echo envelope [Eq. (B8)] and \( \rho_w \) and \( \nu_w \) correspond to seawater density and sound speed.

Figure 10 shows comparisons of model and averaged data for measurements at 33 kHz over a silt substrate. Hence-
forth, the sediment–water mass and density ratios \( (\rho, \nu) \) and the sediment attenuation constant \( (\kappa_p) \) are correlated to the mean grain size parameter \( (M_\phi) \) through the linear regres-
sion equations described in Appendix A. The interface con-

tom increases such that more energy is scattered at higher incidence angles, and proportionally less energy is scattered in the vicinity of the normal to the surface.

For higher values of $\gamma$ or $w_2$, a gentler slope in the angular response is observed in Fig. 11, with a slow initial decay of backscatter strength from normal to higher incidence angles. Here, the backscatter strength at normal incidence is reduced, and there is a smaller dynamic range between $0^\circ$ and $15^\circ$. This is seen in the extreme for $w_2=0.01$, where the volume component dominates. Figure 11 shows that an increase in $g$ or $w_2$ lowers the peak amplitude of the bottom echo, and retards the rise and fall time of the interface scattering component, evidenced by the smearing of the original pulse shape. The shapes of these bottom echoes and angular response curves are characteristic of rough, hard substrates. However, the reduction in backscatter strength and peak echo amplitude observed in the theoretical plots of Figs. 11 are typically compensated by the large reflection coefficients characteristic of these substrates.

Decreased spectral exponent $\gamma$ and spectral strength $w_2$ produce a steeper angular response near nadir, characterized by a fast initial decay of backscatter strength from normal to higher incidence angles [Figs. 11(e)–(f)]. The backscatter strength at normal incidence increases, and a large dynamic range is evident between $0^\circ$ and $15^\circ$. The model envelopes of Fig. 11(e) show an increase in amplitude, and sharper rise and decay times.

3. Volume scattering ($\sigma_v$) influence on echo envelope

The contribution of subbottom scattering to the received echo is largely determined by the volume scattering coefficient $\sigma_v$, which affects energy levels in the tail of the signal. If $\sigma_v$ is set to zero, the volume component of backscatter is absent. Once the bottom projection of the transmit pulse migrates out of the transducer’s main beam, the signal amplitude becomes extremely small, as illustrated in Fig. 11. Doubling $\sigma_v$ from its best-fit value yields a corresponding increase of 3 dB in $P_{\text{rms}}$ and $S$ [Figs. 11(a)–(b)].

B. Ambiguities—relief spectrum constraints

The echo envelope model presented here is a function of measurable bottom characteristics, and may be used to predict echo shapes and amplitudes from well-characterized sediments. However, a more useful (and more difficult) application is that of fitting the model to data for sediment classification purposes; that is, extracting parameter combinations which are meaningful, unique, and most importantly,
Unambiguous identification of a best fit is optimized when each of the fitting parameters has a unique effect on the model’s output.

To first order, the grain size parameter \( M_f \) controls the simulated echo’s peak amplitude, whereas the volume parameter \( s_v \) controls the energy in the signal’s tail. The roughness parameters, \( g \) and \( w_2 \), control the width and rise time of the signal’s peak, but do so in similar fashion as discussed in Sec. IV A 2. The competing effects of these two parameters may lead to several solutions which qualify as “good” model–data fits.

Relief spectrum parameters derived from model–data comparisons using four-parameter \(( M_f, \gamma, w_2, \sigma_v \) unconstrained global model–data optimization techniques (described in Ref. 20) produced substantially different \(( \gamma, w_2 \) combinations. This situation is illustrated by the model–data comparisons of Fig. 10, in which the better fit (Fig. 10(b)) represents a misleading solution according to criteria established below.

For insight into how relief spectrum parameters relate to bottom type, published relief spectra are summarized in Table I. Measured spectral parameters are listed by bottom type in order of increasing \( M_f \), and a vertical space separates “sands” from “fines.” The 1D relief spectrum is published in many of the references; the 2D analogs listed are calculated using assumptions of isotropy. Inspection of the table shows the large diversity of spectral parameters measured for sands and fine-grain sediments. To identify trends in the direct roughness measurements, a scatter plot of \(( \gamma, w_2 \) for the published spectra is presented in Fig. 12. Plots illustrating how \( g \) and \( w_2 \) vary with grain size \( M_f \) are presented in Fig. 13.

Ignoring outliers and redundant measurements, the data in these graphs are a subset of the values listed in Table I—distinguished by the sand symbols (○) and fines symbols (*) listed in the far right column. The shaded \( \gamma \) and \( w_2 \) regions (Figs. 12,13) are those which we infer to be characteristic of sand and silts, with boundaries specified in Table II. This grouping of spectral parameters makes physical sense. Large particles are more likely than fines to settle in high-energy environments; hence, sand substrates should exhibit more energy at lower spatial wavelengths (i.e., higher \( w_2 \)) than their fine-grain counterparts.

**TABLE I.** Published 2D relief spectrum parameters. Comments: ACS=Anisotropic Across-Strike, AAS =Anisotropic Along-Strike. Notation: ~ estimated value, → graphics offset.

<table>
<thead>
<tr>
<th>Substrate</th>
<th>Mean grain size (PHI)</th>
<th>Spectral exponent ( \gamma )</th>
<th>Spectral strength ( w_2 ) (cm(^3))</th>
<th>Source (Ref. no.)</th>
<th>Comments</th>
<th>Symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse sand</td>
<td></td>
<td>0.2~</td>
<td>3.05</td>
<td>0.00027</td>
<td>26</td>
<td>○</td>
</tr>
<tr>
<td>Coarse sand</td>
<td></td>
<td>0.5</td>
<td>2.47</td>
<td>0.00654</td>
<td>41</td>
<td>Shell hash</td>
</tr>
<tr>
<td>Coarse sand</td>
<td></td>
<td>0.5~</td>
<td>3.46</td>
<td>0.00473</td>
<td>26</td>
<td>ACS storm gen ripples ○</td>
</tr>
<tr>
<td>Coarse sand</td>
<td></td>
<td>0.8</td>
<td>3.12</td>
<td>0.00849</td>
<td>28</td>
<td>Shell hash ○</td>
</tr>
<tr>
<td>Medium sand</td>
<td></td>
<td>1.0</td>
<td>3.0</td>
<td>0.004</td>
<td>25</td>
<td>Shell Fragments ○</td>
</tr>
<tr>
<td>Medium sand</td>
<td></td>
<td>1.5~</td>
<td>3.29</td>
<td>0.00046</td>
<td>26</td>
<td>ACS ○</td>
</tr>
<tr>
<td>Medium sand</td>
<td></td>
<td>1.5~</td>
<td>2.33</td>
<td>0.00038</td>
<td>26</td>
<td>AAS ○</td>
</tr>
<tr>
<td>Fine sand</td>
<td></td>
<td>2.0</td>
<td>3.0</td>
<td>0.003</td>
<td>25</td>
<td>○</td>
</tr>
<tr>
<td>Fine sand</td>
<td></td>
<td>2.5</td>
<td>2.92</td>
<td>0.00616</td>
<td>42</td>
<td>○</td>
</tr>
<tr>
<td>Fine sand</td>
<td></td>
<td>2.5~</td>
<td>3.72</td>
<td>0.00043</td>
<td>26</td>
<td>Sand dollar smoothing ○</td>
</tr>
<tr>
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<td>2.75~</td>
<td>3.17</td>
<td>0.00555</td>
<td>26</td>
<td>○</td>
</tr>
<tr>
<td>Fine sand</td>
<td></td>
<td>3.0~</td>
<td>3.50</td>
<td>0.00081</td>
<td>26</td>
<td>○</td>
</tr>
<tr>
<td>Very fine sand</td>
<td></td>
<td>3.0</td>
<td>3.3</td>
<td>0.174</td>
<td>25</td>
<td>Dense live shellfish ○</td>
</tr>
<tr>
<td>Very fine sand</td>
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<td>3.67</td>
<td>0.00422</td>
<td>26</td>
<td>ACS ○</td>
</tr>
<tr>
<td>Very fine sand</td>
<td></td>
<td>3.0</td>
<td>3.92</td>
<td>0.00598</td>
<td>26</td>
<td>AAS ○</td>
</tr>
<tr>
<td>Silty sand</td>
<td></td>
<td>4.8</td>
<td>3.5</td>
<td>0.0046</td>
<td>25</td>
<td>Bimodal clay/sand-gravel *</td>
</tr>
<tr>
<td>Mud</td>
<td></td>
<td>5.5</td>
<td>3.18</td>
<td>0.00318</td>
<td>26</td>
<td>AAS poststorm ○</td>
</tr>
<tr>
<td>Silt</td>
<td></td>
<td>6.4</td>
<td>3.65</td>
<td>0.000846</td>
<td>27</td>
<td>Prestorm ○</td>
</tr>
<tr>
<td>Silt</td>
<td></td>
<td>6.4</td>
<td>3.73</td>
<td>0.000826</td>
<td>27</td>
<td>AAS poststorm ○</td>
</tr>
<tr>
<td>Silt</td>
<td></td>
<td>6.4</td>
<td>3.38</td>
<td>0.000912</td>
<td>27</td>
<td>Prestorm ◾→6.6</td>
</tr>
<tr>
<td>Sand–silt–clay</td>
<td></td>
<td>6.5</td>
<td>3.29</td>
<td>0.0122</td>
<td>28</td>
<td>○</td>
</tr>
<tr>
<td>Silty clay</td>
<td></td>
<td>9.9</td>
<td>3.42</td>
<td>0.00231</td>
<td>28</td>
<td>Methane bubbles ○</td>
</tr>
</tbody>
</table>

**FIG. 12.** Published relief spectrum parameters: \( \gamma \) vs \( w_2 \) (Table I).
The exceptions to this rule are disconcerting and warrant closer examination. As evident in Fig. 13, a progressive reversal of the spectral parameters takes place and is most clearly observed for \((M_f=2.5, 2.75, 3.0)\), where large \(g\) values are associated with small \(w_2\) values and, conversely, large \(w_2\) values are associated with small \(g\) values. In Fig. 14, the 2D relief spectra for these three grain sizes are plotted in log–log scale to demonstrate the potential for ambiguities in the linear representation of the power-law relief spectrum, where the spectral “slope” is equal to \(g\), and \(w_2\) represents the energy at the intercept \((k=1)\).

For the relief spectrum parameters \((\gamma, w_2)\) reported in the literature (Table I): first, the 1D spectra of an ensemble of fine-scale bottom profiles are averaged; the 1D slope and intercept of the average spectrum is determined through linear regression; then, assuming bottom isotropy, the 1D spectral parameters are converted to 2D spectral parameters. The average spectra can be quite noisy, as illustrated by relief spectrum plots presented by Briggs,\(^{26}\) so a degree of ambiguity is likely in these estimates, which may explain the wide range of spectral parameters reported in the literature. We have observed similar ambiguities in unconstrained optimization of model parameters from our acoustic data, which suggests that spectral parameter estimates from topographic and acoustic data share similar degrees of uncertainty.\(^{40}\)

For the San Diego Bay data that we have collected and processed, fewer ambiguous results were obtained when the \((\gamma, w_2)\) constraints proposed in Table II were enforced during four-parameter model–data matching procedures. This led to the development of a two-stage optimization procedure where for the first stage \(\gamma\) is set to 3.25, the generic value\(^{19}\) which falls conveniently within the overlap of the regions proposed for sand and fines. For the second stage, \(\gamma\) is set to a value dependent on the initial estimate of mean grain size, \(\hat{M}_f\). Model–data matching tests using \(\gamma\) constants within the proposed boundaries led us to replace Table II with the simple rule

\[
\begin{align*}
&\text{If } M_f < 4.0 \quad \gamma = 3.0, \\
&\text{If } M_f \geq 4.0 \quad \gamma = 3.3.
\end{align*}
\]

Subsequently global optimizations were carried out over \((M_f, w_2, \sigma_p)\), resulting in convergence to unique and sensible solutions.\(^{20}\)

V. SUMMARY AND CONCLUSIONS

A time-dependent model of the acoustic intensity backscattered by the seafloor was described and compared with data from a calibrated, vertically oriented, echo-sounder operating at 33 and 93 kHz. The model incorporates the characteristics and geometry of the echo-sounder: the transducer’s beam pattern, altitude off the bottom, tilt with respect to vertical, and the characteristics of the transmitted sound pulse; and environmental factors: spherical spreading and absorption losses, backscattering of the signal at the water–sediment interface, and by inhomogeneities in the sediment volume.

Scattering from the interface is predicted using Helmholtz–Kirchhoff theory, with physical parameters consisting of the strength \((w_2)\) and exponent \((\gamma)\) of a power law defining the 2D interface roughness spectrum, the ratio of sediment saturated bulk density over water mass density \((\rho)\),

\[
\begin{align*}
&\text{If } M_f < 4.0 \quad \gamma = 3.0, \\
&\text{If } M_f \geq 4.0 \quad \gamma = 3.3.
\end{align*}
\]

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Scattering from the interface is predicted using Helmholtz–Kirchhoff theory, with physical parameters consisting of the strength \((w_2)\) and exponent \((\gamma)\) of a power law defining the 2D interface roughness spectrum, the ratio of sediment saturated bulk density over water mass density \((\rho)\),
and the ratio of sediment sound speed over water sound speed ($v$). A modified version of the composite roughness model, accounting for the finite duration of the transmitted signal, predicts the energy backscattered from the sediment volume. This contribution to the measured intensity field is determined by the subbottom’s volume scattering and attenuation coefficients ($\sigma_v$ and $k_p$, respectively), as well as the water–sediment interface characteristics governing sound transmission into the sediment. Temporal stretching of the echo due to large-scale bottom roughness is incorporated by convolving model computations with the height distribution of the interface relief.

Envelope averaging is performed on echoes aligned in time, where alignments along threshold minima are applied to data with well-defined energy distributions, and offsets calculated from the phase slope of the envelope spectra are applied otherwise. The resulting rms pressure of the measured backscattered signal is compared with estimates of echo intensity converted to pressure [Eq. (31)]. Favorable comparisons of model and data were achieved by correlating the acoustic impedance and attenuation properties of the bottom with measures of its mean grain size ($M_\phi$). Estimation of parameters ($M_\phi$, $\gamma$, $w_z$, $\sigma_v$) from fitting the model to data reveal ambiguous ranges for the two spectral parameters. However, analyses of model outputs and of physical measurements reported in the literature yield practical constraints on the roughness parameter spectrum. These constraints are used to estimate geoacoustic parameters through an optimized echo envelope matching technique described in a companion paper. The equations used in this work relating geoacoustic model parameters to particle size distribution ($-1 \leq M_\phi \leq 9$) are adapted from the APL-UW High-Frequency Ocean Environmental Acoustics Models Handbook. The difference between our equations and Ref. 19 is that, in place of the volume parameter ($\sigma_v$), we establish a separate particle size mapping to the volume scattering coefficient ($\sigma_v$).

### A. Limitations

The echo envelope model assumes isotropic and Gaussian-distributed bottom relief, sediment homogeneity in the upper few meters, as well as sediment–water impedance ratios ($\rho_p$), and sediment acoustic attenuation constants ($k_p$) consistent with mean empirical measures. Radical departure from these conditions may invalidate model results. Bottom samples and video images indicated that the sediments surveyed in San Diego Bay generally met these standards. However, model computations may not produce reliable results for complicated biogenic or anisotropic sediments.

Limitations on interface curvature dictated by the Kirchhoff approximation also restrict modeling to sediments having a large rms radius of curvature relative to the acoustic wavelength. This excludes extremely rough (rocky) substrates, or operation at high frequencies (>100 kHz). To evaluate the usefulness of the temporal model for characterizing a broader range of substrates, measurements at a variety of well-characterized sites will be necessary.

### ACKNOWLEDGMENTS

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### APPENDIX A: GENERIC VALUES FOR GEOACOUSTIC MODEL PARAMETERS

The equations used in this work relating geoacoustic model parameters to particle size distribution ($-1 \leq M_\phi \leq 9$) are adapted from the APL-UW High-Frequency Ocean Environmental Acoustics Models Handbook. The difference between our equations and Ref. 19 is that, in place of the volume parameter ($\sigma_v$), we establish a separate particle size mapping to the volume scattering coefficient ($\sigma_v$).
APPENDIX B: IMPLEMENTATION OF TIME DEPENDENT BACKSCATTER

1. Sediment interface backscatter

Equation (19) is computed numerically using a discrete representation in which the received signal is calculated at intervals of $\tau_s$ seconds and indexed by $n$ such that $t_n = n\tau_s$. Isotropy is assumed in order to reduce the two-dimensional surface integral to a one-dimensional function of the angle of incidence $\theta_i$. Then, the seafloor may be divided into a series of $J$ concentric annuli, with indices $j$. The area $A[j]$ of each annulus, with range $R[j] = a\cos(\theta_i)$ to its geometric center and radii $r_1[j]$ and $r_2[j]$, is calculated by

$$A[j] = \pi(r_2^2[j] - r_1^2[j]).$$

(B1)

Formulation of $r_1[j]$ and $r_2[j]$ depends on the strategy used to partition the insonifed area. The indexed angles $\theta_{ij}$ may have equal increments in $\theta_i$ or they may be calculated via equal increments in the ring radius, $r$. The former paradigm is used here because it provides finer angular resolution close to normal incidence.

If the change in range from the near to the far edge of $A[j]$ is sufficiently small compared to the length of the pulse, the energy received from $A[j]$ at time index $n$ can be expressed in a form similar to the integrand of Eq. (19) and summed over the elemental annuli, bound by indices $j_{[n]} \leq j \leq j_{[n]}$, which overlap the leading and trailing edges of the pulse projection at time index $n$

$$I[n] = \sum_{j_j=n}^{j_{[n]}} I_2(n \tau_n - \frac{2R[j]}{v_w}) \frac{\alpha_{n}}{\text{att}[j]} B_m[j],$$

(B2)

where $s_1[j]$ is the backscatter coefficient, $\alpha_{n}$ accounts for the transmission loss in the water column

$$\text{att}[j] = R^4[j] \left| \frac{\alpha_n}{\text{att}[j]} \right|^s,$$

(B3)

and $B_m[j]$ is the discrete implementation of the mean directivity function $B_m(\theta_i)$

$$B_m(\theta_i) = \frac{1}{2\pi} \int_0^{2\pi} b^4(\theta_i, \varphi) d\psi.$$  

(B4)

When the transmit waveform is represented by a sampled sequence $I_1(n \tau_n)$, the value of $I(n \tau_n - \frac{2R[j]}{v_w})$ may be determined by interpolation or rounding.

2. Sediment volume backscatter

The discrete representation of Eq. (21) that we use for numerical computations requires an array for the volume tube [Fig. 4(b)]. An $M$ element array of penetration lengths $l_m$ defines volume cell centers, where $l_1$ represents the cell at the surface, and $l_M$ the cell at the substrate's skin depth, $d_s$, that we define as the depth at which the signal strength drops by 10 dB

$$d_s = \frac{10 + 10 \log_{10}(1 - R_{min}^2)}{a_b},$$

(B5)

where for a specific sediment type, $R_{min}$ is the minimum angle-dependent reflection coefficient. The width of each cell, $\delta l$, is chosen to reflect the resolution of the time-sampled transmit pulse. Here, we chose $\delta l = v_t \tau_s / 2$, which corresponds to the round-trip path length of the signal, per sample period.

Insonification of the volume continues for some time after the trailing edge of the pulse has passed over the seafloor patch. Thus, within practical constraints dictated by the skin depth, sediment volume contributions $I_v[n]$ are calculated between nadir ($j = 1$) and the annulus overlapping the leading edge ($j = j_{[n]}$)

$$I_v[n] = \sum_{j=1}^{j_{[n]}} \frac{\alpha_{n} B_m[j]}{\text{att}[j]} V[j] A[j],$$

$$\times \sum_{m=m_{[j,n]}}^{m_{[j,n]}} I_3(\tau_p - \frac{l_m - l_{m_{[j,n]}}}{v_b}) L[m],$$

(B6)

where $m_{[j,n]}$ and $m_{[j,n]}$ index the volume cells which overlap the leading and trailing edges of the pulse at $n \tau_s / 2$. As before, when the transmit waveform is represented by a sampled time series $I_1(n \tau_n)$, the value of $I_3(\tau_p - \frac{l_m - l_{m_{[j,n]}}}{v_b})$ may be determined by interpolation or rounding. This function is defined solely within the interval for which the pulse insonifies the volume annulus $j$ bound by the surface and propagation length $d_s$.

The attenuation length $L[m]$ [the discrete version of Eq. (23)], is an array with elements

$$L[m] = \frac{1}{\lambda} \left( e^{-A_{ms}(l_m - \delta l/2)} - e^{-A_{ms}(l_m + \delta l/2)} \right),$$

(B7)

3. Echo signal intensity

The pressure contributions to the total field that are backscattered by the water–sediment interface and by subbottom volume inhomogeneities are assumed to be uncorrelated, and the total intensity received at time $n \tau$ is expressed with a simple addition of these contributions

$$I_t[n] = I_t[n] + I_v[n],$$

(B8)

where for large-scale roughness, $I_t[n]$ is computed with Eq. (B2) and $I_v[n]$ is computed with Eq. (B6).

4. Seafloor macro-roughness

Like Berry and Blackwell,9 Haines and Langston,10 Ogilvy,35,36 and Pouliquen and Lutron,13 we account for seafloor macro-roughness by convolving the “smooth surface” response with a roughness response. Within the length $r_f$ of the footprint of the −3-dB width of the beam pattern along the interface, the macro-roughness corresponds to an altitude change $\zeta$, such that $\zeta = \zeta - \bar{a}$ is a zero-mean Gaussian random variable with variance $\sigma_{\zeta}^2$. Assuming that a power-law relief spectrum is valid for the macro-roughness at length scales of order $r_f$, such that mean-square height differences across the footprint increase with increasing footprint size, a measure of $\sigma_{\bar{a}}^2$ is given by the structure function [Eq. (4)] evaluated at $r_f$

$$\sigma_{\bar{a}}^2 = C_{\bar{a}} r_f^{\alpha_a}.$$ 

(B9)
over which the macro-roughness is considered, and then we sum over $(N_s)$ increments above and below the mean plane $\bar{a}$ where $a_s$ is predicted to be about

$$\Delta a = \nu_a \tau_s \cos(\theta_o)/2,$$

and we sum over $(N_s)$ increments above and below the mean plane $\bar{a}$

$$N_s = \text{int}[b \sigma_a/\Delta a],$$

where $b$ is a fractional constant specifying the percentage of the distribution included in the computation, and “int” signifies rounding to the nearest integer value. We use an empirically derived $b=1.28$ (specifying 80% of the distribution) for fine-grain sediments (muds), and $b=0.5$ (40% of the distribution) for coarse-grain sediments (sand). The discrete implementation of the effects of macro-roughness is then expressed as

$$I_m[n] = \left\{ \begin{array}{ll} \sum_{-N_s}^{N_s} I[n \tau_s - i \tau_s, a_s[i]]P_s[i] & n = (\bar{a} - b \sigma_a)/\Delta a \\ 0 & \text{otherwise} \end{array} \right.,$$

where $a_s[i] = \bar{a} + i \Delta a$ represents the height for model index $i$

and the discrete probability mass function $P_s[i]$ is calculated by integrating the normal distribution between $(\xi_1 = a_s[i] - \Delta a/2)$ and $(\xi_2 = a_s[i] + \Delta a/2)$ such that

$$P_s[i] = \Phi \left( \frac{\xi_2 - \bar{a}}{\sigma_a} \right) - \Phi \left( \frac{\xi_1 - \bar{a}}{\sigma_a} \right),$$

where $\Phi$ is the distribution function of a normal random variable. For conservation of energy, $P_s[i]$ must be normalized to 1, such that

$$P_s[i] = 1 - \sum_{-N_s}^{N_s} P_s[i] / 2N_s + 1.$$

Macro-roughness lengthens the rise and decay times of the echo envelope and it reduces the maximum signal strength. In the model, these distortive effects are expected to be more pronounced for sand substrates whose macro-roughness standard deviation $\sigma_a$ is predicted to be about three times as large as that of clay (11.5 cm vs. 4.4 cm). A weakness of this approach is that when comparing the model with data, differing values of the macro-roughness parameter $(b)$ are required for sand and fine-grain sediments. For unambiguous estimates of sediment characteristics from bottom echo envelopes, a more sophisticated approach to modeling the effects of macro-roughness will eventually be required.

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