A physical model for seismic noise generation from sediment transport in rivers

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Received 3 November 2011; revised 28 December 2011; accepted 28 December 2011; published 28 January 2012.

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1. Introduction

The transport of coarse sediment by rivers sets the pace of landscape evolution by controlling channel morphology and the rates of bedrock incision [e.g., Whipple, 2004]. Moreover, accurate predictions of sediment flux are needed for diverse applications including sedimentation engineering, river restoration, and flood hazard mitigation. Most models for bed load sediment transport are empirical and typically rely on data from flume experiments where sediment flux is at the transport capacity [e.g., Fernandez Luque and van Beek, 1976]. In many mountain streams, sediment flux is under-capacity, however, and is governed by the sediment supply from upstream and neighboring hillslopes, for which no bed load-flux models exist [e.g., Whipple, 2004]. Our inability to accurately model bed load transport stems from a lack of measurements during floods in steep rivers where traditional measurement techniques (e.g., sediment traps) are extremely difficult, if not impossible, to apply.

[1] Measuring sediment flux in rivers remains a significant problem in studies of landscape evolution. Recent studies suggest that observations of seismic noise near rivers can help provide such measurements, but the lack of models linking observed seismic quantities to sediment flux has prevented the method from being used. Here, we develop a forward model to describe the seismic noise induced by the transport of sediment in rivers. The model provides an expression for the power spectral density (PSD) of the Rayleigh waves generated by impulsive impacts from saltating particles which scales linearly with the number of particles of a given size and the square of the linear momentum. After incorporating expressions for the impact velocity and rate of impacts for fluvially transported sediment, we observe that the seismic noise PSD is strongly dependent on the sediment size, such that good constraints on grain size distribution are needed for reliable estimates of sediment flux based on seismic noise observations. The model predictions for the PSD are consistent with recent measurements and, based on these data, a first attempt at inverting seismic noise for the sediment flux is provided. Citation: Tsai, V. C., B. Minchew, M. P. Lamb, and J.-P. Ampuero (2012), A physical model for seismic noise generation from sediment transport in rivers, Geophys. Res. Lett., 39, L02404, doi:10.1029/2011GL050255.

2. Model

Our goal is to derive a mechanistic model for seismic noise generated by river sediment-bed impacts using as simple and generic of parameterizations as possible that still incorporate the relevant first-order physics. The primary idea of this model is that each sediment particle impacts the river bed and creates a force impulse that then excites seismic waves that travel to nearby seismic stations. With many particles, each impacting the bed at random times relative to the others, one can calculate the total noise power spectral density (PSD) observed at a given station. To attempt to make comparisons with the observations discussed in the Introduction, we make model predictions over the frequency range 1 Hz < f < 1000 Hz.

2.1. Seismic Impact Model

In this forward model, it is assumed that individual grains of diameter D each impact the river bed with speed w_i, assumed normal to the bed, where w_i depends on D and other fluvial parameters, as will be discussed in Section 2.2. For a single particle, the impact force can be described by an elastic contact problem, of which the simplest case is perfectly elastic Hertzian contact [Johnson, 1987]. For this case, the maximum force amplitude F_0 and time of contact Δt are given by well known expressions [Johnson, 1987]. For
$D < 2 \text{ m, } w_i > 0.1 \text{ m/s, and a typical elastic modulus} (E = 5 \cdot 10^8 \text{ Pa}), \text{ then } \Delta t \approx 10^{-3} \text{ s, and the impact can be assumed to be instantaneous relative to the frequency range of interest, and impart an impulse equal to} \quad I = \frac{2}{\pi} F_0 \Delta t \approx 2 m w_i, \quad (1)

where \( m = \rho V_p \) is the mass of the particle, \( V_p \) is the particle volume, and \( \rho \) is rock density. If the impact is not perfectly elastic, \( I \) could be as much as a factor of 2 smaller (when perfectly inelastic), in which case \( I \approx mw_0 \). A force history for a single particle’s instantaneous impulse can then be expressed as \( F_i(t) = I \delta(t) \) where \( \delta(t) \) is the Dirac delta function.

[7] Given a force history \( F(t) \) at location \( x_0 \), the ground velocity \( \dddot{u}(t) \) at location \( x \) is given in the frequency domain by

\[ \tilde{u}(f, x) = 2\pi f F(f, x_0) G(f, x; x_0) \quad (2) \]

where \( F(f) = \mathcal{F}[F(t)] \) is the Fourier transform of \( F(t) \), and \( G(f) \) is the displacement Green’s function. Since \( w_0 \) is vertically incident, Rayleigh waves are expected to be the dominant waves excited [e.g., Sanchez-Sesma et al., 2011]. For a horizontally homogeneous medium, and assuming an approximate Rayleigh-wave sensitivity that decays with depth proportional to \( e^{-rz} \), we can follow Aki and Richards [2002] and approximate the amplitude of the Rayleigh-wave Green’s function as

\[ |G(f, x; x_0)| \approx \frac{k}{8\rho_0 v_w} \sqrt{\frac{2}{\pi fr}} e^{-\eta f/(v_w)0}, \quad (3) \]

where \( v_w \) is the Rayleigh-wave phase velocity, \( v_w \) is group velocity, \( k = 2\pi f v_w \) is the angular wavenumber, \( r = |x - x_0| \) is the source-station distance, and \( Q \) is the (dimensionless) quality factor.

[8] To use equation (3), we must have estimates of the frequency-dependent \( v_w \), \( v_r \), and \( Q \). For \( v_w \) and \( v_r \), we use values of average shear wave speed for a typical generic rock site (and in the frequency range of interest) given as

\[ v_w = v_0 (z_0/z)\alpha, \quad (4) \]

where \( v_0 = 2206 \text{ m/s, } z_0 = 1000 \text{ m, and } \alpha = 0.272 \) [Boore and Joyner, 1997]. Using these values, and approximations as above, one can solve for \( v_w \) as

\[ v_w = v_0 (f/f_0)^{-\xi}, \quad (5) \]

where \( v_0 = [(2\pi z_0 \xi)/(\alpha)]^{1/(1-\alpha)} \xi \Gamma(1 + \alpha) \) and \( \Gamma(x) \) is the gamma function (see auxiliary material). With values as given above, \( v_0 = 1295 \text{ m/s, } f_0 = 1 \text{ Hz, } \xi = 0.374 \). Furthermore, under the same approximations, it can be shown that \( v_w = v_w/(1 + \xi) = 0.73 v_w \). On the other hand, \( Q \) is typically assumed to be of the form of \( Q = Q_0(f/f_0)^\eta \) [e.g., Erickson et al., 2004]. Anderson and Hough [1984] suggest that for the relatively high frequencies and shallow depths of interest, \( Q_0 = 20 \) and \( \eta = 0 \) are reasonable approximations, and we use these values throughout, despite known spatial variations in \( Q \) [e.g., Campbell, 2009].

[9] To describe the rate of impacts, we define \( n(D) \) to be the number of particles with grain size \( D \), per unit length of river per unit depth. If the average time between impacts of each particle is \( t_i(D) \), then \( \int n \cdot t_i dD \) is the total rate of impacts per unit length of river. In Section 2.2, we relate \( n(t) \) to sediment flux.

[10] In the simplest version of the model, we assume that all impacts occur randomly spaced in time, i.e. \( F(t) = I \sum \delta(t - t_i) \), with random \( t_i \) and \( N \) being the number of impacts. For random \( t_i \) one can show that \( F(f) = F_0 \sqrt{N} \approx I \sqrt{N} \), so that the sum of impacts does not affect the shape of the force spectrum, and the force amplitude squared grows linearly with \( N \). We can now express the power spectral density (PSD) of a station’s velocity time series (per unit grain size \( D \)) to be

\[ P_v(f; D) = \int \frac{\pi^2 f^3 m^2 v_0^2}{t_i^2} G_0^2 e^{-x^2/2} \chi(\beta), \quad (7) \]

where the integral is along the full length of river \( R \), and \( \dddot{u}(t) \) is the ground velocity due to \( F(t) \).

[11] In order to approximate equation (6) analytically, we assume an infinitely long and straight river whose closest point is \( r_0 \) from the seismic station (see Figure 1a). Substituting \( F_0(f) \), equation (1) and equation (3) into equation (2) we arrive at

\[ P_v(f; D) = \frac{\pi^2 f^3 m^2 v_0^2}{t_i^2} G_0^2 e^{-x^2/2} \chi(\beta), \quad (7) \]

where

\[ \chi(\beta) = \int_{-\infty}^{\infty} \frac{1}{\sqrt{1 + y^2}} e^{-\beta \sqrt{1 + y^2}} dy \quad (8) \]

and \( \beta = 2\pi r_0 (1 + \xi) f^\xi e^{-\eta/(v_0 Q_0 t_i)} \) are dimensionless. As shown in the auxiliary material, \( \chi(\beta) \) can be approximated as

\[ \chi(\beta) \approx 2\log \left( 1 + \frac{\beta}{1} \right) e^{-\beta} + 1 - e^{-\beta} e^{-\beta^2/2 \beta} \quad (9) \]

It may be noted that equation (7) scales with frequency roughly as \( f^{4.9} \exp[-2\pi r_0 f^{1.4} Q_0 v_0 \eta] \).

[12] Finally, the total PSD, \( P_v(f) \), is given as an integral over the grain size distribution

\[ P_v(f) = \int_D P_v(f; D) dD. \quad (10) \]

Using equation (9) in equation (7) and substituting into equation (10) then yields an algebraic expression for the total PSD, \( P_v(f) \), as a function of frequency \( f \), grain size distribution and other model parameters.

[13] The random impact model discussed above can be made more realistic by including the correlated impacts of the same particle. If \( N_c \) hops of a single particle occur before the impact time becomes significantly different from an integer multiple of the timescale between impacts, \( t_i \), then the forcing

\[ F_{N_c}(t) = \sum_{j=0}^{N_c-1} \delta(t - j t_i) \quad (11) \]
that characterized these processes in the context of bedrock incision [e.g., Sklar and Dietrich, 2004; Turowski et al., 2007; Lamb et al., 2008a]; in particular, we use the model of Lamb et al. [2008a] because it explicitly solves for particle fall velocity. Owing to the strong dependency of seismic energy on particle size and impact velocity (i.e., equation (7) scales as \( \rho v^2 \)), we focus on seismic energy generated from saltating particles alone and neglect particles that are rolling or sliding along the bed, particles suspended in the flow, and viscous damping of particles impacts by the fluid [cf. Lamb et al., 2008a]. The rate of particle impacts per unit channel length (for a given grain size) can be calculated from

\[
\frac{n}{\ell_b} = \frac{C_1 W q_b \bar{w}_s}{V_p U_b H_b} \tag{13}
\]

where \( W \) is the average channel-bed width, \( q_{bd} \) is the volumetric sediment flux per unit grain size \( D \) per unit channel width traveling as bed load, \( U_b \) is the vertically-averaged streamwise particle velocity and \( H_b \) is the bed load layer height, \( \bar{w}_s \) is the depth-averaged particle settling velocity, and \( V_p \) is the particle volume. \( C_1 = 2/3 \) accounts for the fact that the total time between impacts should also include the particle ejection or rise time as well as the fall time [Sklar and Dietrich, 2004].

[15] The depth-averaged bed load velocity and layer height are given as empirical expressions by Sklar and Dietrich [2004] derived from several different bed load studies. The best fit relationships are

\[
U_b = 1.56 \sqrt{R g D \left( \frac{\tau_s}{\tau_c} \right)^{0.56}}, \quad U_b \leq U \tag{14}
\]

and

\[
H_b = 1.44 D \left( \frac{\tau_s}{\tau_c} \right)^{0.50}, \quad H_b \leq H \tag{15}
\]

where \( R = (\rho_s - \rho) \rho_s g D \approx 2700 \text{ kg/m}^3 \) and \( \rho_s = 1000 \text{ kg/m}^3 \) are the sediment and fluid densities, respectively, \( g = 9.8 \text{ m/s}^2 \) is the acceleration due to gravity, \( \tau_s = u^* (R g D) \), \( u^* \) is the bed shear velocity, and \( H \) is the total flow depth. \( U \) is the depth-averaged flow velocity calculated as \( U = 8.1 u^*(H/k_s)^{1/6} \) [Parker, 1991], where \( k_s = 3 D_{50} \) [e.g., Kamphuis, 1974], and the critical value of the Shields stress \( (\tau_c) \) is the value of \( \tau_s \) at the threshold of particle motion found from Lamb et al. [2008b] for the median grain size \( D_{50} \). For other grain sizes, we calculate \( \tau_c = \tau_{c,50}(D/D_{50})^{-\gamma} \), where \( \gamma \approx 0.9 \) [Parker, 1990]. The bed shear velocity is calculated assuming steady and uniform flow as \( u^* = \sqrt{gH \sin \theta} \), where \( \theta \) is the channel-bed slope angle.

[16] The particle impact velocity normal to the bed can be calculated from a balance between the forces of gravity and drag for spherical particles [Lamb et al., 2008a] as

\[
w_i(H_b) = w_{st} \cos \theta \sqrt{1 - \exp[-H_b]}, \tag{16}
\]

where \( w_{st} = \sqrt{4 R g D / (3 C_d)} \) is the terminal settling velocity, and \( H_b \equiv 3 C_d \rho_s H_b (2 \rho_s \cos \theta) \). The drag coefficient \( C_d \) depends on the particle Reynolds number and grain shape, and we calculate \( C_d \) from the empirical formula of Dietrich...
assumed here in rivers with large scale bedrock roughness where oblique bed impacts may cause impact velocities to scale with flow velocity \( U \) rather than settling velocity \( w_s \) [e.g., Johnson and Whipple, 2010]. If true, this would cause equation (7) to scale approximately with \( D^2 \) rather than \( D^3 \), and non-vertical impact would also cause a higher fraction of Love waves to be generated than assumed.

[21] Using representative values from Lavé and Avouac [2001], Attal and Lavé [2006] and Burtin et al. [2008], we obtain PSDs as function of frequency \( f \) as shown in Figure 1 for 2 choices of \( r_0 \) and 3 choices of \( D \). In these plots, it is assumed that all particles are of median size \( D = D_{50} \) and that \( q_b = q_{bd} \) describes the flux of these particles. We predict the general spectral features of such PSDs to be similar for any steady impact model where the frequencies of interest are less than one over the impact time (i.e., \( f < 1/\Delta t \)). As shown, the modification introduced by equation (12) creates a modulation in the PSDs with frequency spacing \( \Delta f = 1/\Delta t \), revealing a potential seismic signature of sediment size independent of sediment flux, but has no effect on a smoothed version of the PSDs. Compared with the PSDs in Figure 6 of Burtin et al. [2008], the model is able to predict some aspects of the observations, including the general peak around \( \approx 7 \) Hz, the sharper increase of the PSDs to this peak compared to the more gradual decrease at higher frequencies, and the higher PSD values at high frequencies (up to \( 15\)–\( 20 \) Hz) for stations that are closer to the Trisuli River. The modulation introduced by equation (12) may also be observed by Burtin et al. [2008], and could potentially be used as a constraint on the grain size distribution.

[22] To explore the dependence of equation (7) on fluvial parameters, in Figure 2 we plot \( P_c \) for fixed \( r_0 = 600 \) m at \( f = 7 \) Hz (near the peak of the PSD) but with variable \( H, \theta, D \) and \( q_b \). For fixed \( q_b \) the PSDs have the somewhat unintuitive feature that they decrease both with increasing slope (\( \theta \)) and increasing flow depth (\( H \)) (solid lines of Figure 2a). This results from a larger hop height and velocity, which reduces the impact rate (equation (13)) in the fluvial framework used. However, \( q_b \) may increase with both \( \theta \) and \( H \); for example, if we set \( q_b = q_{bc} \), then the PSDs increase with both \( \theta \) and \( H \) (dashed lines of Figure 2a).

[23] So far, expressions have been evaluated and plotted for a single grain size, \( D \). As noted above, the approximate \( D^2 \) dependence of \( P_c \) implies that larger grain sizes have a disproportionately larger effect on the seismic signal compared to smaller grain sizes. For comparisons with observations, then, it is important to use a grain size distribution that is realistic. Thus, instead of using a typical log-normal distribution of grain sizes, which has an unrealistically long tail at large \( D \), we introduce a new log-‘raised cosine’ distribution, which has almost the same shape as a log-normal distribution but has a cut-off at both large and small \( D \) (see Figure 3a). The raised cosine distribution is defined by

\[
\hat{p}(x; \mu, s) = \frac{1}{2s} \left[ 1 + \cos \pi \left( \frac{x - \mu}{s} \right) \right], \quad -s < x - \mu < s
\]  

and \( \hat{p}(x; \mu, s) = 0 \) otherwise, and has equivalent mean, median, and variance as a normal distribution \( N(\mu, \sigma_g^2) \) if we choose \( s = \sigma_g / \sqrt{1/3 - 2/\pi^2} \) (where \( \mu \) and \( \sigma_g \) are the mean and standard deviation of the normal distribution). To best fit
the data of Attal and Lavé [2006] for the region of interest, we choose $D_{50} = 0.15$ m and $s_g = 0.52$ (dimensionless, see auxiliary material). The resulting log-'raised cosine’ distribution $p(D) \equiv \hat{p}(\log(D) ; \log(D_{50}), s_g)$/D as well as the $P_v$ resulting from this grain size distribution are plotted in Figure 3a. For this example, the peak in $P_v$ occurs at $D = 0.34$ m and corresponds to $D_{94}$, the 94th percentile grain size. Integrating $P_v$ over all $D$ results in the total PSD, which in this case is $P_v_T = 10^{-150.3}$ dB. While we expect $s_g$ to be realistic, in Figure 3b, we show the sensitivity of the dominant grain size to variations in $s_g$. As shown, the dominant grain size is typically far above the median grain size $D_{50}$ except when $s_g$ is unrealistically small (<0.3).

Finally, we make a preliminary attempt at inverting the observations of Burtin et al. [2008] for the total bed load flux $q_b$, without attempting to calibrate the model, and assuming the average estimated and measured fluvial and seismic parameters as before. As stated previously, this inversion relies heavily on accurate knowledge of grain size distribution, which is lacking, and despite other poorly constrained parameters, this is likely the cause of the largest uncertainty. Given the scaling of $P_v$ with $D^3q_bD \equiv D^3pq_b$, and the dominant $D$ being close to $D_{94}$, if the grain size distribution changes with flow discharge through the monsoon season then the seismic PSD approximately constrains $D_{94}$ rather than $q_b$ alone. It is also unknown what fraction of the observed seismic noise may be attributable to water flow noise (or other environmental sources), but a large portion of the seismic signal is likely due to bed load [Burtin et al., 2008]. Assuming the grain size distribution is unchanged (as given above), and assuming the full seismic signal is due to bed load, $q_b$ as inverted from the seismic data (see Figure 3c) seems to scale approximately with $q_b/5$ but has the clear hysteresis inherent in the Burtin et al. [2008] data. The predicted hysteresis in sediment flux as a function of discharge is supported by measurements of Gabet et al. [2008] in a neighboring river, and likely occurs as sediment supply from hillslopes is depleted near the end of the wet season. These observations are consistent with a
supply-limited river, where $g_b < g_{bc}$ (Figure 3c), resulting in bedrock that is partially exposed [Sklar and Dietrich, 2004] and susceptible to rapid erosion [Lavé and Avouac, 2001]. The calculations presented here show the feasibility of such an inversion and we expect the framework described here to be useful to constrain sediment flux from seismic observations at other rivers.

[25] Acknowledgments. The authors thank J.-P. Avouac, L. Bollinger, and J. Lavé for helpful comments, and thank J. Johnson and L. Sklar for thoughtful reviews. This research was partially supported by NSF grant EAR0922199 to MPL.

[26] The Editor thanks Joel Johnson and Leonard Sklar for their assistance in evaluating this paper.

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M. P. Lamb, Division of Geological and Planetary Sciences, California Institute of Technology, 1200 E. California Blvd., Pasadena, CA 91125, USA.
1 Evaluation of Phase Velocity

With shear wave speed as given in Eq. (4) of the main text as $v_s = v_0(z/z_0)^\alpha$, and assuming Rayleigh-wave sensitivity that decays with depth proportional to $e^{-kz}$, then the Rayleigh-wave phase velocity can be approximately expressed as

$$v_c \approx \frac{\int_0^\infty v_s(z)e^{-kz}dz}{\int_0^\infty e^{-kz}dz} = \frac{k v_0}{z_0^\alpha} \int_0^\infty z^\alpha e^{-kz}dz = \frac{v_0 \Gamma(1 + \alpha)}{z_0^\alpha k^{\alpha}}. \tag{1}$$

Substituting $k = 2\pi f/v_c$, one can then solve for $v_c$ as

$$v_c = \left[\frac{v_0 \Gamma(1 + \alpha)}{z_0^\alpha (2\pi f)^\alpha}\right]^{1/(1-\alpha)} = \left[\frac{v_0 \Gamma(1 + \alpha)}{(2\pi z_0 f_0)^\alpha}\right]^{1/(1-\alpha)} \left[\frac{f}{f_0}\right]^{-\alpha/(1-\alpha)} \tag{2}$$

2 Approximation of $\chi(\beta)$

$\chi(\beta)$ is defined in Eq. (8) of the main text as

$$\chi(\beta) \equiv \int_{-\infty}^{\infty} \frac{1}{\sqrt{1+s^2}} e^{-\beta \sqrt{1+s^2}} ds. \tag{3}$$

To approximately evaluate this integral, we consider the limits $\beta \ll 1$ and $\beta \gg 1$. For $\beta \ll 1$, $\chi(\beta)$ can be approximated as

$$\chi(\beta) \approx 2 \int_0^1 \frac{1}{\sqrt{1+s^2}} ds + 2 \int_1^{\infty} \frac{1}{s} e^{-\beta s} ds = 2 \sinh^{-1} 1 + 2 \Gamma(0, \beta), \tag{4}$$
where $\Gamma(0, \beta)$ is the incomplete gamma function. $\Gamma(0, \beta)$ can be approximated as $\Gamma(0, \beta) \approx e^{-\beta} \log(1 + 1/\beta)$. For $\beta \ll 1$, $\Gamma(0, \beta) \gg \sinh^{-1} 1$, so we finally have
\[ \chi(\beta) \approx 2e^{-\beta} \log(1 + 1/\beta), \beta \ll 1. \tag{5} \]

On the other hand, when $\beta \gg 1$, $\chi(\beta)$ can be approximated as
\[ \chi(\beta) \approx 2 \int_0^\infty e^{-\beta(1+s^2/2)} ds = e^{-\beta} \sqrt{2\pi/\beta}, \beta \gg 1. \tag{6} \]

Smoothly transitioning between Eq. (5) and (6) with an exponential weighting then results in
\[ \chi(\beta) \approx 2 \log \left(1 + \frac{1}{\beta}\right) e^{-2\beta} + (1 - e^{-\beta}) e^{-\beta} \sqrt{\frac{2\pi}{\beta}}. \tag{7} \]

### 3 Estimation of Local Water Depth

For given slope $S$, channel width $W$, and channel depth $H$, the total water flux $Q = WHU$, where average velocity $U$ is given in the main text and can be written as $U \propto H^{2/3} S^{1/2}$ so that $Q \propto WH^{5/3} S^{1/2}$. Thus,
\[ H_2 = H_1 \left(\frac{Q_2 W_1 S_1^{1/2}}{Q_1 W_2 S_2^{1/2}}\right)^{3/5}, \tag{8} \]

where subscripts refer to different locations. Since $W$ and $S$ can be estimated from imagery, and Lave and Avouac [2001] provide estimates of $Q$ along the Trisuli, we can estimate $H_2$ near the seismic stations of interest relative to the $H$ measured in the town of Betrawati. Our estimates are that $Q_2/Q_1 \approx 800$ m$^3$/s/1000 m$^3$/s $\approx 0.8$, $W_2/W_1 \approx 35$ m/70 m $\approx 0.5$, and $S_2/S_1 \approx 0.025/0.010 \approx 2.5$ so that $H_2 \approx 1.0H_1$. This means that the water level records at Betrawati can be used, without modification, as estimates of the water levels near the seismic stations of interest.

### 4 Estimation of Grain Size Distribution

To estimate the grain size distribution of the Trisuli River close to the seismic stations of interest, we assume that the distribution is similar to that of the
nearby Marsyandi River at a location with similar drainage area and slope. Based on the descriptions of the two rivers in Plate 6 of Lave and Avouac [2001], we find that a location just north of the main central thrust (MCT) along the Marsyandi has similar slope (2.5°) and $Q_{10} \approx 10^3 \text{ m}^3/\text{s}$ as the region of the Trisuli of interest, and we choose this location as representative. Attal and Lave [2006] provide average grain size distributions on the Marsyandi both upstream and downstream of the MCT, so we take an average of these two zones as representative. For this average, the distribution of grain sizes larger than 5 cm is approximately 17%, 38%, 39%, 6% and 0% in bins of 5-8 cm, 8-16 cm, 16-32 cm, 32-64 cm, and > 64 cm, respectively. Performing a best-fit to these data using the log-‘raised cosine’ distribution discussed in the main text, we obtain a median grain size $D_{50} = 0.15 \text{ m}$ and an equivalent normal standard deviation of $\sigma_g = 0.525$ (i.e., $s = 1.45$). This best-fitting model results in a grain size distribution of 12%, 42%, 38%, 8% and 0% for the same grain size bins.

As stated in the main text, we choose to use the log-‘raised cosine’ distribution rather than the more commonly used log-normal distribution because the log-normal distribution has an unrealistically long tail at large (and small) grain sizes. Since our model is quite sensitive to the largest grain sizes (with an approximate $D^3$ dependence of $P_v$), having a realistic tail at the high end of the grain size distribution is therefore important for the model prediction. To provide a sense for how different the log-‘raised cosine’ and log-normal distributions are, we note that the log-normal distribution plotted in Fig. 3a of the main text has a grain size distribution with 10%, 43%, 38%, 7% and 0.3% in the same grain size bins described above. Thus, while the distributions are very similar (and fit the measured grain size distribution nearly equally as well), the 0.3% at very large grain sizes (> 64 cm) would result in significant seismic power predicted for those grains (from the sensitivity of the model to large grain sizes) despite the very small percentage, and would therefore bias our prediction somewhat. For any model (like the one presented here) that is sensitive to the tails of a grain size distribution that is known to be bounded, we recommend use of the log-‘raised cosine’ over the log-normal distribution.