

RESEARCH ARTICLE

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Key Points:

- Novel, low-frequency (0.1–2 Hz) signals are recorded using near-stream seismometers
- These signals appear on the horizontal components of seismometers and can be used to determine discharge rates after empirical calibration
- These low-frequency signals are interpreted to arise through local strain and tilting excited by pressure variations on the channel bank

Supporting Information:

- Supporting Information S1
- Data Set S1

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Measuring Mountain River Discharge Using Seismographs Emplaced Within the Hyporheic Zone

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Abstract Flow and sediment transport dynamics in fluvial systems play critical roles in shaping river morphology, in the design and use of riverine infrastructure, and in the broader management of watersheds. However, these properties are often difficult to measure comprehensively. Previous work has suggested the use of proximal seismic signals resulting from flow and bed load transport to construct more complete records of these fluvial processes. We investigate a small (184 km²; < 20 m³/s), snowmelt-fed mountain river in the Northern Colorado Rocky Mountains during May–August 2015 to capture peak runoff with colocated measurements of discharge and seismic noise. Three-component seismometers were placed in close proximity to the channel bank (~1 m) within the hyporheic zone (at times submerged beneath the water table). We recorded a broad spectrum of seismic signals excited by discharge, including novel, low-frequency (0.1–2 Hz) signals observed predominantly on the horizontal components. The characteristics of these low-frequency signals are not consistent with an elastically propagating seismic wave. We instead infer that they likely arise from the sensor tilting in response to viscoelastic deformation occurring near the channel and propose large-scale turbulent eddies as a forcing mechanism. Calibrating horizontal seismic power with hydrograph flow rates over the course of a rainstorm for individual sensors, we demonstrate that these unique signals can be used to accurately estimate river discharge with simple regressions. This technique shows promise for augmenting seismic monitoring of rivers by enabling discharge rates to be estimated from outside the channel using easily deployed and noninvasive seismic instrumentation.

Plain Language Summary We deploy a small array of seismometers in close proximity to a small mountain river in Northern Colorado to record seismic signals in conjunction with peak snowmelt runoff during the summer of 2015. The seismic instruments are colocated with measurements of discharge (in channel pressure transducer), suspended sediment, and precipitation. After a short calibration period with discharge (here a rainstorm over which large variations in discharge occurred), we found that accurate discharge rates could be obtained solely through signals recorded on the horizontal components of the seismometers. The signals likely arise from the seismic sensor physically tilting as it is forced by pressure pulses on the stream bank generated by turbulent water flow. Determining discharge rates in this manner may usefully complement seismic arrays designed to continuously monitor sediment transport in fluvial systems.

1. Introduction

High flow and sediment discharge episodes are fundamental to the geomorphic evolution of fluvial systems in both natural and human-altered environments, and may be increasing in intensity and frequency in some regions (e.g., Foulds et al., 2014; Naylor et al., 2017). Due to spatial-temporal variability of flow, sediment flux, and the hazards associated with in-channel measurements, in situ measurements of flow and sediment transport may be unable to provide a comprehensive record, especially in extreme flood conditions or where such measurements may alter the physical channel (Bunte & Abt, 2005) and adversely affect or even destroy instrumentation. Quantitative understanding of the relationships between river discharge, sediment transport, and excitation of seismic signals has the potential to usefully augment field sampling of fluvial systems with a noninvasive method.

Multiple environmental processes on Earth's surface contribute to the local and global seismic wavefield (e.g., Larose et al., 2015). The first study using seismic signals to detect discharge and bed load transport occurred in the early 1990s and was carried out using a vertical component geophone deployed directly in the channel and recording on magnetic tape (Govi et al., 1993). Advances in seismic instrumentation and processing

techniques have spurred renewed interest in the field of fluvial seismology in the past decade. In contrast to this early work, these recent studies have performed analysis primarily in the frequency domain and have begun rapidly developing the observational (e.g., Bartholomaeus et al., 2015; Burtin et al., 2016, 2008, 2011; Hsu et al., 2011; Roth et al., 2016; Schmandt et al., 2013, 2017) and modeling (Gimbert et al., 2014; Tsai et al., 2012) foundation necessary to interpret riverside seismic observations for discharge and bed load transport.

In general, although the range of river systems studied using seismic observations has primarily focused on larger fluvial systems associated with high levels of bed load transport, these studies suggest that bed load transport tends to excite high-frequency (> 15 Hz) power (Roth et al., 2016; Schmandt et al., 2013, 2017), while turbulent flow and discharge modulates lower frequency seismic power (between ~ 1 and 10 Hz) (Bartholomaeus et al., 2015; Burtin et al., 2011; Schmandt et al., 2013, 2017). These observations of bed load generating higher-frequency energy than turbulent flow at a given seismic station are reflected in the physical modeling of seismic noise generated through stresses induced by fluid flow over a coarse riverbed proposed by Gimbert et al. (2014). One of the key predictions of this model is that the peak frequencies excited by bed load transport and turbulent flow become increasingly separated as the seismometer is placed farther away from the channel. Thus, these two process may be identified independently by taking measurements at various distances from the channel. However, this technique may not be feasible for smaller fluvial systems, which do not produce strong enough signals to be observable at large distances from the channel.

Expanding on this work, Roth et al. (2016) were able to constrain bed load transport rates using only calibrated seismic data, precipitation rates, and discharge measurements in the foothills of the Swiss Alps. As turbulent flow and bed load transport occupied nearly identical frequency bands at the stations located in close proximity (~ 2 – 5 m from the channel bank) to the stream, independent discharge data were particularly valuable for separating the contribution of turbulent flow to the seismic wavefield from that of bed load transport. In widespread practice, discharge measurements typically are made via in-channel measurements of stage height (e.g., using pressure transducers), which must be emplaced and removed seasonally in alpine regions to avoid damage from freezing. The feasibility of obtaining stage height directly from high-frequency (> 1 Hz) seismic observations of turbulent flow was also proposed by Gimbert et al. (2014) and could thus usefully augment other measurements designed to monitor flow and sediment transport. However, using high-frequency seismic noise to determine stage height (and thus discharge) in this context has several limitations, including (1) a model that assumes the sensors are emplaced at distances from the channel much greater than the channel width. At these larger distances (10–100 m) from the channel, turbulent flow signals from smaller fluvial systems are often not observed (Barrière et al., 2015; Govi et al., 1993). (2) At sensors located in close proximity to the stream, the frequency band excited by turbulent flow may be dominated or heavily influenced by bed load transport (Barrière et al., 2015; Gimbert et al., 2014; Roth et al., 2016; Tsai et al., 2012). (3) Above ~ 10 Hz, anthropogenic or wind noise contamination can overwhelm the dominant turbulent flow signals even in relatively rural environments (Anthony et al., 2015; Barrière et al., 2015; Roth et al., 2016; Schmandt et al., 2017; Withers et al., 1996). (4) The amplitude of seismic signals generated by turbulent flow is strongly dependent on the relative roughness of the riverbed (Gimbert et al., 2014), which may change appreciably following episodes of elevated discharge and associated bed load transport.

Here we present a case study in which intermediate frequency (0.33–2 Hz) seismic signals recorded within the hyporheic zone are used to accurately determine discharge rates on a small mountain stream across a runoff season using empirical calibrations determined from a single, 2 week high discharge event. By utilizing these lower frequency signals, measured ~ 1 m from the channel, the above issues associated with determining discharge rates from higher-frequency signals attributed to turbulent flow are largely avoided. These signals are not generated by elastically propagating seismic waves and may instead arise from instrument tilt as the sensor directly responds to local viscoelastic deformation due to meter-scale eddy formation and turbulent flow forces applied near the bank. Thus, use of these low-frequency tilt signals recorded on seismometers directly buried adjacent to the channel offer potential for a complementary, noninvasive method of estimating discharge rates that is likely insensitive to bed load transport and anthropogenic noise contamination.

2. Field Site and Data

The South Fork of the Cache la Poudre River is a small mountain stream that originates in the Mummy Range of Rocky Mountain National Park, Colorado, and has a drainage area of 184 km^2 (Ryan, 2007). The study site



Figure 1. (a) Field site along a straight reach of the South Fork Cache la Poudre River, Northern Colorado. Seismic stations SF01–SF03 are shown along with infrasound sensors locations (blue circles, sensor 1 colocated with seismic stations, person for scale next to SF03 label). (b) Locations of rain gauge, pressure transducer, and turbidity meter. (c) Three-component Sercel L28 (4.5 Hz lower corner frequency) geophones were placed ~1 m from the channel and at water table depth just prior to rising limb conditions in mid-May. GPS-time-stamped data were recorded on solar powered Reftek RT-130 data loggers. This reach of the river runs nearly north-to-south, so that the north component of the seismometer is very nearly parallel, and the east component very nearly perpendicular, to flow, which is to the right (north) in Figures 1a and 1b.

(Figure 1a) features a straight, nearly north-south flowing reach with a ~1% bed slope, a riffle channel morphology, and an 11 m bankfull channel width corresponding to a discharge of ~10 m³/s. The site was previously selected by the U.S. Forest Service (USFS) to monitor changes in suspended sediment load following the 2012 High Park fire. Although strong flooding in the Colorado Front Range during September of 2013 led to exceptionally high discharge rates which scoured the bank (e.g., Gochis et al., 2015; Rathburn et al., 2017), repeat surveys from 2013 to 2015 have shown this reach to be stable (including during smaller out of bank discharge events such as recorded here) following the 2013 flood. The channel bed is composed of gravel and cobble ($d_{50} \sim 70$ mm), which is largely stable (occasional sliding and rolling visually observed at peak runoff) except during strong floods such as the 2013 event. To facilitate monitoring efforts, the USFS installed a pressure transducer, calibrated turbidity sensors, and a rain gauge at the study site to enable continuous (10 min sampling interval) collection of discharge, suspended sediment, and precipitation data from 2013 to 2015 (Figure 1b). Stage height, as measured by the pressure transducer, is converted to discharge by a single rating curve (Figure S1 in the supporting information) attained by measuring both channel geometry and flow velocity multiple times over the

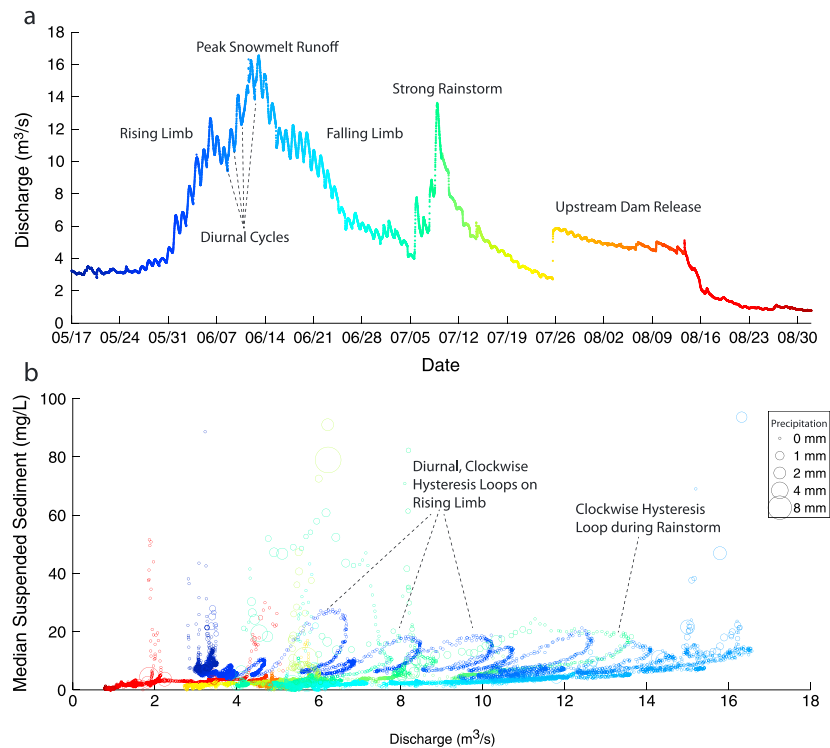


Figure 2. (a) Hydrograph spanning the entire study period, as determined by pressure transducer measurements of stage height and rating curve (Figure S1). The hydrograph is color coded by time (as indicated on the x axis) and has significant events leading to changes in discharge labeled. (b) Suspended sediment concentration versus discharge (sampled every 10 min), identically color coded by time. Clockwise hysteresis loops are observed both on the rising limb of snowmelt runoff and during the July rainstorm event. Marker size corresponds to colocated precipitation measurements during each 10 min sampling period.

course of the rising and falling limb of spring snowmelt runoff. Additionally, bed load was periodically measured with a hand-held sampler throughout the study period and found to consist primarily of coarse-grained sand and fine gravel ($d_{50} = 2$ mm at peak runoff) that moved sporadically during high discharge events over the study period.

From mid-May to early September we operated three 3-component seismographs utilizing Sercel L28 geophones (corner frequency 4.5 Hz) deployed 1 m from the river edge at 12 m, 23 m, and 35 m downstream from the stage height pressure transducer (Figure 1a). Each sensor was emplaced ~ 0.3 m deep and within 10 cm of the water table during moderate (~ 3.2 m³/s) early-season flow conditions (Figure 1c). Note that SF03 is located immediately downstream and in close proximity to several large boulders in the channel (Figure 1a), which has important implications for the study results. Until early August, stations SF02 and SF03 were each supplemented with a 15 m aperture array of three ground-mounted infrasound sensors with a corner frequency of 0.05 Hz (Marcillo et al., 2012). SF01 recorded at 250 Hz, while stations SF02 and SF03 recorded at 125 Hz (sampling rate was reduced to accommodate the additional three channels of infrasound data). All stations experienced a 4 to 5 day data gap in mid-June during which the data storage cards on each data logger became full prior to scheduled service. Additionally, SF01 lost 7 days of data in early July due to corrupted storage.

During the study period, discharge varied between 1 and 16 m³/s, including strong modulation from the rising and falling limbs of spring snowmelt in June, an intense rainstorm in early July, and an upstream dam release on 26 July that more than doubled discharge over the course of an hour (Figure 2a). The dam release lasted for nearly 3 weeks and maintained elevated levels (5–6 m³/s) of discharge during this time period before closure of the dam dropped flow rates to ~ 1 m³/s in late August. Diurnal cycles in discharge are pronounced during the rising and falling limb as snowmelt runoff is increased during daylight hours. Peak discharge during both snowmelt runoff and the rainstorm event caused minor out-of-bank

flooding (discharge greater than bankfull $\sim 10 \text{ m}^3/\text{s}$), and completely submerged the geophones beneath the water table.

Suspended sediment loads were generally $< 10 \text{ mg/L}$ but increased up to $\sim 100 \text{ mg/L}$ following rain events and increases in discharge (Figure 2b). The spring snowmelt cycle shows general clockwise hysteresis between suspended sediment and discharge with higher levels of suspended sediment observed on the rising limb of the hydrograph. Clockwise hysteresis loops (defined here as higher levels of a given quantity on the rising limb than on the falling limb with respect to discharge) are typical and arise from entrainment of available sediment stored within the bed and near-channel environments, transported, and subsequently quickly depleted during higher flows (e.g., Seeger et al., 2004). We additionally observe clockwise hysteresis loops accompanying each diurnal pulse of increased discharge on the rising limb and during the 7–10 July rainstorm event.

3. Spectral Analysis

To study the seismoacoustic wavefield in response to changing discharge conditions, we perform our analysis in the frequency domain using moving time windows. Seismic and infrasound data channels were converted to 20 min window acceleration or pressure power spectral density (PSD; in decibels relative to $1 \text{ m}^2/\text{s}^4/\text{Hz}$ for seismic and to $1 \text{ Pa}^2/\text{Hz}$ for infrasound data) estimates with 50% temporal overlap using the Incorporated Research Institutions for Seismology (IRIS) Noise Took Kit (NTK) (Bahavar et al., 2013; Incorporated Research Institutions for Seismology Data Management Center, 2014). After the power spectrum is computed, the NTK removes nominal instrument responses (provided by the instrument manufacturer and available from the IRIS Nominal Response Library) through spectral division (e.g., Scherbaum, 2013). To avoid amplifying instrument self-noise at frequencies far below the 4.5 Hz geophone corner, a band-pass deconvolution filter was applied between 0.15 and 0.05 Hz. Thus, the reported PSD levels at frequencies below 0.15 Hz are accordingly reduced in amplitude (Figure 3).

Similar to the widely used PQLX PSD-estimating software (McNamara & Boaz, 2011; McNamara & Buland, 2004), the NTK employs Welch's section averaging method (e.g., Oppenheim & Schaffer, 1975). We subdivide each 20 min data window into 15 subsegments of 50% overlap and estimate a PSD in 1/8-octave period bins and 1 dB power increments. However, in contrast to the default 1-octave smoothing window used in PQLX, the user can specify the degree of PSD smoothing within the NTK. We used a smaller, 1/4-octave smoothing window to minimize smearing of power across period bins and attain greater frequency resolution. To facilitate direct comparison with discharge and suspended sediment measurements, 50% overlapping 20 min PSD estimates were converted to local time within the U.S. Mountain Daylight Time zone. For comparative analysis, we time align our fluvial and power spectral density/spectrogram data using the average spectrum of two 20 min 50% overlapping time sections covering the 30 min time period that is centered on the corresponding (10 min) fluvial data segment. This averaging scheme was necessitated by differences in how the PSDs and fluvial data are time stamped and enables more accurate temporal alignment between the two measurements.

Acceleration PSDs are a standard format for referencing noise at a seismic station to both global models of background noise (e.g., Peterson, 1993) and self-noise models of seismic instrumentation (e.g., Ringler & Hutt, 2010). For subsequent study of total power in a given frequency band, these acceleration PSDs are integrated to ground velocity, given that absolute velocity power has a physical interpretation in terms of localized kinetic energy at the seismometer (e.g., Aster et al., 2008). We note that differential acceleration and velocity PSDs normalized to the same reference period will be identical because acceleration PSD estimates are simply velocity PSDs scaled by $(2\pi f)^2$, where f is the frequency at which the power density is estimated, and this frequency-dependent scaling cancels when ratios are calculated.

4. Results

4.1. Infrasound Spectral Analysis

Infrasound measurements were dominated by quasi-diurnal, 0.1–3 Hz noise bursts that were often 180° out of phase with diurnal discharge variations (Figure S2). These signals showed a red spectrum, varied in duration from 10 to 16 h, and typically began in middle-late morning and subsided late in the evening. The spectral characteristics and timing of these events are consistent with wind-generated noise (e.g., Withers et al., 1996), a well known and common issue for infrasound sensors that are not equipped with strong wind filters

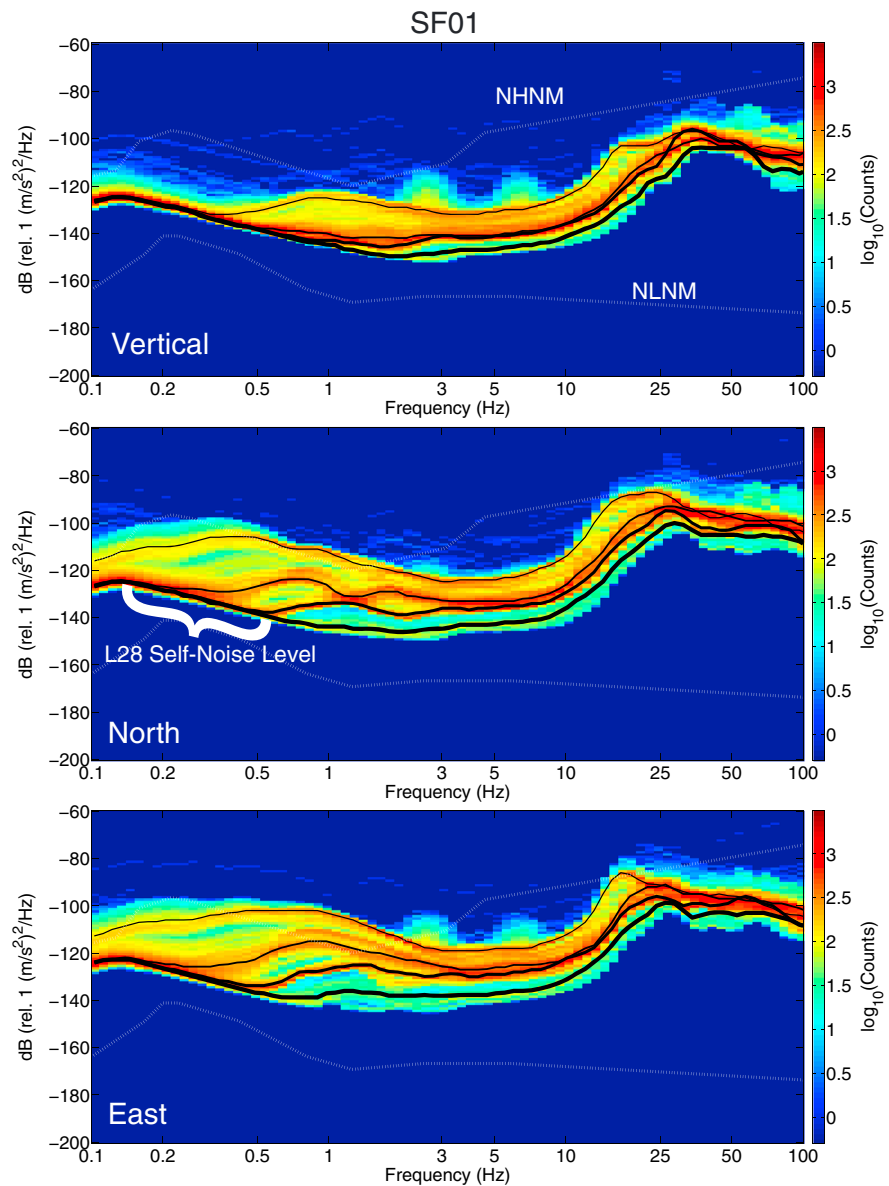


Figure 3. Acceleration power spectral density (PSD) probability density functions (PDFs) for each component of SF01 over the duration of the study period. The black lines correspond to median PSDs for different discharge conditions with thinner lines indicating modal features during higher discharge. From low-to-high discharge the lines represent 1–2 m³/s (late August conditions), 3–4 m³/s (May conditions), 4.5–6 m³/s (dam release), and > 10 m³/s (max discharge). The white dashed lines are the new high and low noise models (NHNM and NLNM; Peterson, 1993). During minimal discharge conditions, the self-noise levels of the L28 geophone define the PSD at frequencies below 0.1–0.5 Hz on all components. At maximal discharge, both horizontal components are above instrumentation self-noise levels to at least 0.1 Hz.

(e.g., Hedlin & Alcoverro, 2005). Recent experimental work utilizing an adjustable hydraulic jump has suggested supercritical flow in fluvial systems, such as that occurring in whitewater rapids and waterfalls, may be necessary for flowing water to produce strong infrasound signals (Johnson et al., 2006; Ronan et al., 2017). During our study period, the dominant flow regime is subcritical and the generation of infrasound signals was clearly limited. We were thus unable to reliably identify acoustic signals originating from fluvial processes above background noise levels and suggest that infrasound sensors without wind filters, even when deployed at ground level, are likely unsuitable for monitoring acoustic signals originating from smaller river systems. Here we primarily use the infrasound signals to identify wind events and characterize their influence on the seismic noise environment.

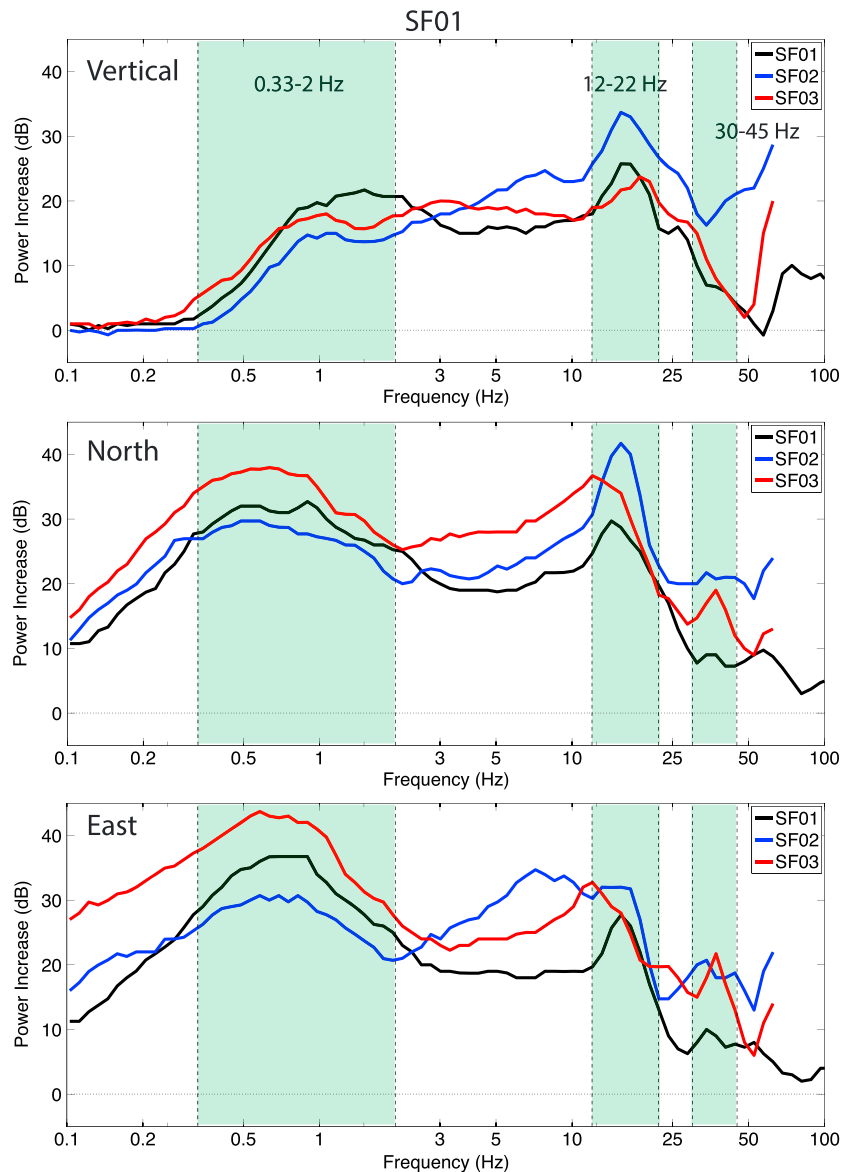


Figure 4. Ratios (difference in decibels) between acceleration power spectral densities (PSDs) obtained during maximal ($> 10 \text{ m}^3/\text{s}$) and minimal ($1\text{--}2 \text{ m}^3/\text{s}$) discharge conditions for each component of each station in this study. Highlighted are three distinct bands that appear particularly excited by discharge and are discussed throughout the study.

4.2. Seismic Response to Changes in Discharge

Spectral analysis of ground motion reveals a clear wideband ($\sim 0.1\text{--}45 \text{ Hz}$) seismic response to river discharge on both the horizontal and vertical components across the observational period (Figures 3–5 and S3–S7). For this study, we define a reference background seismic noise environment for each component of each sensor as the median PSD recorded during times of minimal discharge ($1.0\text{--}2.0 \text{ m}^3/\text{s}$) during late August (thick black curve in Figure 3). Consistent with previous fluvial seismology studies where the sensor is emplaced in close proximity ($< 10 \text{ m}$) to the channel (Barrière et al., 2015; Roth et al., 2016), the background ambient noise field for these stations on all components is dominated by high-frequency ($> 25 \text{ Hz}$) energy that decays by $\sim 35 \text{ dB}$ between 25 and 10 Hz (Figures 3, S3, and S4). However, power levels within this high-frequency peak vary significantly between individual stations (up to 15 dB at $\sim 50 \text{ Hz}$) with the highest levels recorded on SF03 (the station immediately downstream of the boulders). In contrast, background power levels between 1 and 25 Hz are quite similar between stations with maximum interstation variations not exceeding 7 dB (Figure S5). Below $\sim 0.5 \text{ Hz}$ on horizontal and $\sim 1 \text{ Hz}$ on

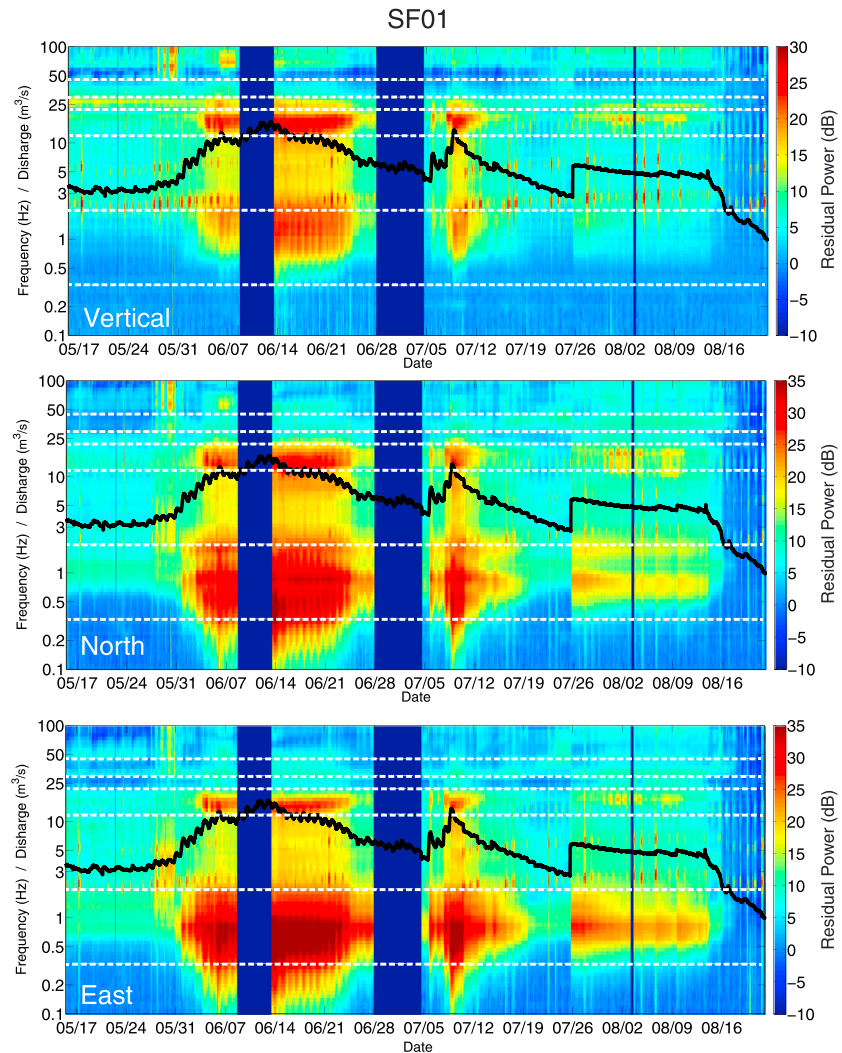


Figure 5. Residual acceleration spectrograms for the three components of SF01, obtained by normalizing with respect to median background power spectral density (background calculated after discharge drops below $2 \text{ m}^3/\text{s}$ in late August; Figure 3). Horizontal, white dashed lines indicate predominant power frequency bands identified as being especially sensitive to discharge (Figure 4). Dark blue areas are data gaps, as described in the text. Differential spectrograms for SF02 and SF03 show similar features of fluvial excitation and are available in the supporting information (Figures S6 and S7).

vertical components, we are unable to reliably characterize the reference background noise environment because the ambient seismic noise field was below the self-noise level of the 4.5 Hz geophones. The self-noise levels (i.e., electrical and mechanical noises internal to the sensor) of these instruments at low frequencies is readily identified in PSD probability density functions as a distinct linear feature during minimal discharge conditions (Figure 3). At higher frequencies and greater rates of discharge, seismic background noise (in the form of ground vibrations) exceeds the self-noise of the instrument and is recorded with high fidelity.

The three seismic stations display strong commonalities in their responses to discharge, but, despite close proximity and very similar installation conditions, also illustrate some notable differences. To characterize how changing discharge conditions influence seismic power at different frequencies at each station, we generate both differential PSDs between maximal discharge and background conditions (Figure 4) as well as corresponding differential spectrograms over the entire study period (Figures 5, S6, and S7). Although increases in seismic power with discharge are observed across the entire spectral range analyzed in this study (0.1–100 Hz), three distinct bands of response were especially notable (Figures 4 and 5).

Between 30 and 45 Hz, we find the highest levels of vertical acceleration power for all ranges of discharge (Figure 3a). This band encompasses frequencies that were reported to be excited by turbulent flow and bed load transport in previous fluvial studies where the sensor was located in close proximity (< 10 m) to the channel (Barrière et al., 2015; Roth et al., 2016). Between 12 and 22 Hz, we find highest levels of acceleration power on the horizontal components (Figures 3b and 3c) as well as distinct excitation of all components in response to increased discharge (Figure 4). Finally, at frequencies between 0.33 and 2 Hz, we also find signals that are strongly excited by discharge, particularly on the horizontal components. As noted above, PSD estimates in the lower-frequency part of this band were sometimes compromised by the limited long-period geophone response. Although this band is still fairly high-frequency by global seismology standards, this band is below frequencies typically studied in fluvial seismology and thus is here referred to as the “low-frequency” band.

Above ~ 25 Hz, robust relationships between increasing vertical component seismic power and discharge are solely observed on SF02, with the other stations observing no change or sometimes decreases in high frequency power with increasing discharge (Figures 3a, S3, and S4). When discharge rates exceed $10 \text{ m}^3/\text{s}$ (near or above bankfull conditions), vertical component power levels between 3 and 10 Hz increase by 15–20 dB over background conditions on all three stations (Figure 4a). However, the strongest increases in vertical component power (20–33 dB) occur within the 12–22 Hz band, observed as a distinct peak between 15 and 20 Hz. This response is more pronounced (up to 10 dB) on the horizontal components. An additional low-frequency peak also appears in the vertical component of the acceleration PSD at ~ 1 Hz during higher rates of discharge (Figures 3a and S5a).

At maximal discharge rates, low-frequency (< 2 Hz) power recorded on the horizontal components exceeds the L28 self-noise to below 0.1 Hz at all stations and even the Global Seismic Network High Noise Model on SF03 (Figures 3, S3, and S4; Peterson, 1993). During these times, 0.5 Hz horizontal power increases by at least 30 dB above ambient background levels for all stations and is overall the most sensitive band to discharge (Figures 4b and 4c). The strongest response to discharge on both horizontal components is observed at SF03, followed by SF01. Additionally, horizontal power exceeds vertical power by > 25 dB with the east component (perpendicular to the channel) showing a 5–10 dB increase relative to the north component. This east polarization on the horizontal components is again the most pronounced at SF03, followed by SF01. On the east component of all stations, a small, ~ 1 Hz peak can be seen even during times of minimal discharge. This is absent from the north component (Figures 3b, 3c, S3, and S4). As discharge increases to $> 3 \text{ m}^3/\text{s}$, this peak appears on both horizontal components of all three stations. During the rising and falling limb of snowmelt runoff, diurnal changes in power excited near 1 Hz are in phase with daily variations in discharge (Figure 5).

4.3. Hysteresis Observations During the 7–10 July Rainstorm Event

Clockwise hysteresis observations between discharge and seismic power in distinct frequency bands have been interpreted to arise from bed load transport in several fluvial seismology studies (e.g., Barrière et al., 2015; Burtin et al., 2008; Hsu et al., 2011; Roth et al., 2014; Schmandt et al., 2013). However, changes in bed morphology, such as created by migrating bed forms, could also produce similar hysteresis patterns (Gimbert et al., 2014). To assess the character and consistency of hysteresis patterns and to test for any signals that may be related to suspended sediment transport, we construct discharge-integrated seismic power relations over a 2 week time period centered on the distinct and short-lived July rainstorm, during which discharge varied from $\sim 3.5 \text{ m}^3/\text{s}$ to $13 \text{ m}^3/\text{s}$ (Figures 6, S8, and S9).

We observe virtually no hysteresis within the 30–45 Hz band during this event. Overall, velocity power within this band increases by at least 5 dB on each component of each station. However, none of the components of any station exhibit monotonically increases in seismic power with discharge in this frequency band.

The 12–22 Hz band shows significant but inconsistent hysteresis patterns between stations. Clear clockwise hysteresis is observed for all components of SF02, which could be interpreted as evidence of bed load transport, and overall power increases by ~ 3 –5 dB with increasing discharge. However, SF03 is more sensitive to discharge (increases of ~ 15 dB on all components) yet shows weak counterclockwise hysteresis. At SF01, seismic velocity power levels on all components rise in power function fashion with discharge during the rainstorm until flow reaches $\sim 8 \text{ m}^3/\text{s}$, after which seismic power levels remain relatively constant even as discharge continues to increase. No hysteresis is observed on this station.

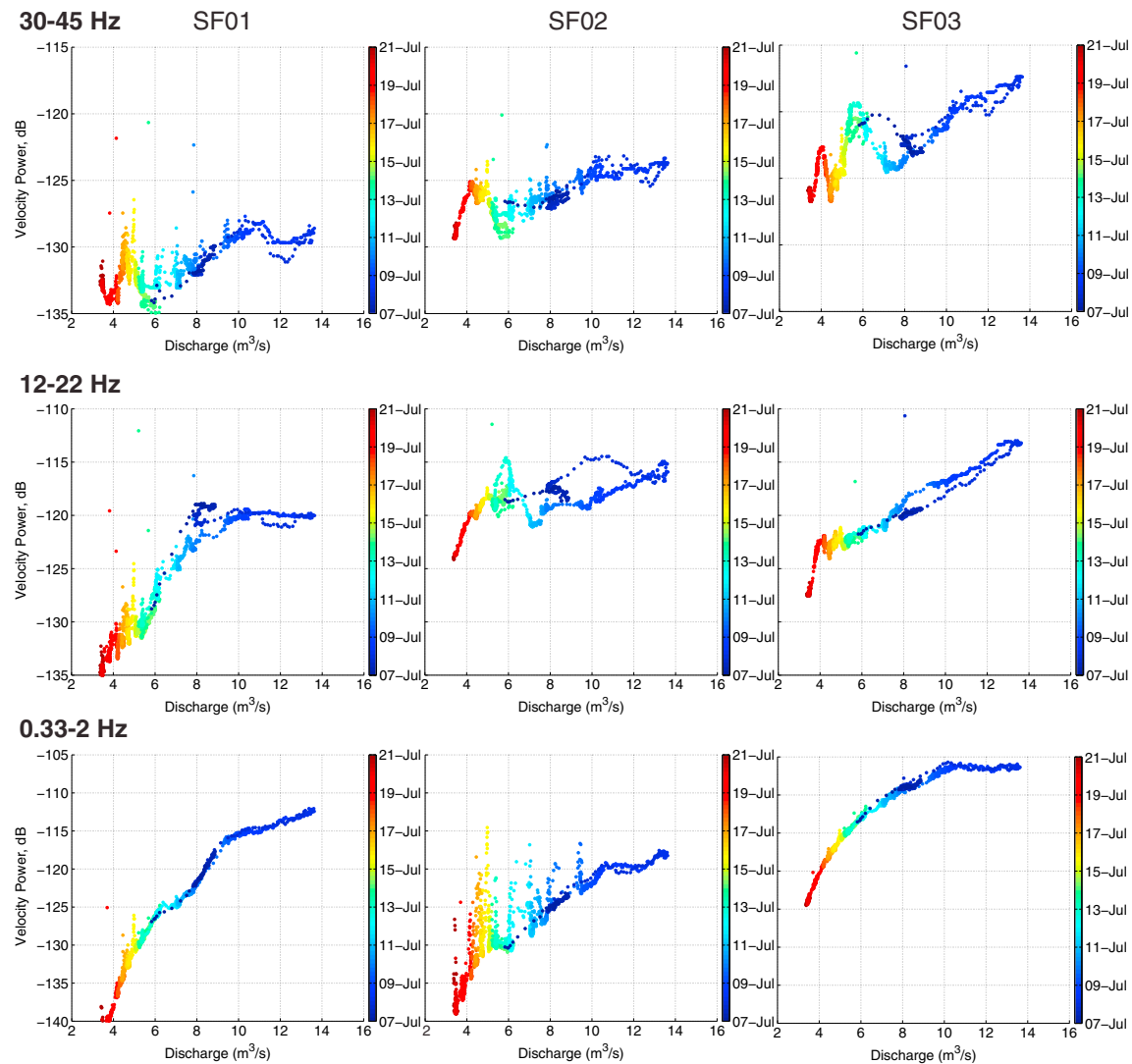


Figure 6. Ground velocity (dB rel. 1 (m/s)^2) versus discharge for the July rainstorm event recorded on the east component of all stations (given by column) for each of the three frequency bands (rows) identified to be particularly sensitive to discharge (Figure 4). Plots are color coded by date as indicated in the color bars. Y axis is constant for all stations for a given frequency band.

No hysteresis is observed within the 0.33–2 Hz band, with power varying smoothly by 20–25 dB on the horizontal components with increased discharge. A pronounced reduction in slope between seismic power and discharge is observed on the east component of all stations and on the north component of SF01 when discharge reaches $\sim 10 \text{ m}^3/\text{s}$. Notably, this discharge rate is roughly the bankfull capacity of this reach of the South Fork and higher rates of discharge cause out of bank flooding. All components of station SF02 have identical, discrete intervals where power increases abruptly. The differential spectrogram for this station (Figure S6) shows that these times coincide with quasi-diurnal, 0.1–10 Hz spikes which are prevalent throughout the study period and are correlated with high wind events identified in the infrasound record. We thus attribute these events to gusty local wind-generated noise at this site, possibly related to wind coupling at the nearby ($\sim 10 \text{ m}$) access bridge (Figure 1).

4.4. Synthetic Hydrograph From Seismic Observations

Power in the 0.33–2 Hz band recorded on horizontal components shows smooth, approximately exponential relations with flow rates (Figure 6), and discharge variations are clearly distinguished in acceleration spectra (Figure 3). We thus investigate the potential of mapping seismic power in this band to hydraulic discharge to establish a synthetic hydrograph relationship. To do this, we calibrate the strongly discharge dependent

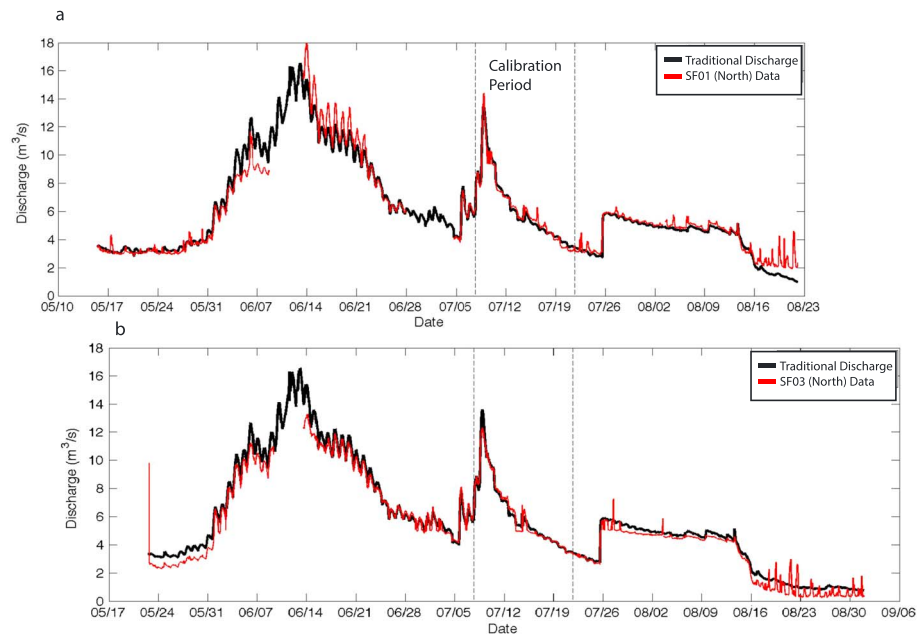


Figure 7. Synthetic hydrographs (red curves) calculated from 0.33 to 2 Hz horizontal component ground velocity power on the north components of (a) SF01 and (b) SF03. These hydrographs generated from seismic observations are compared with discharge estimates attained through traditional methods (i.e., pressure transducer and rating curves; black curves). The synthetic hydrographs match the pressure-transducer generated hydrograph to $1 \text{ m}^3/\text{s}$ at the 95% confidence interval. The dashed black horizontal lines indicate the 2 week period around the rainstorm event where seismic data was calibrated against discharge.

0.33–2 Hz power band, recorded on each channel of stations SF01 and SF03, to discharge during a 2 week period surrounding the July rainstorm (Figure 6). We then apply these relations to generate a synthetic hydrograph spanning the entire seasonal record. We do not use station SF02 in this analysis as power in the 0.33–2 Hz band at this station was strongly influenced by wind gust transients, as described previously.

Seismic velocity power in the 0.33–2 Hz band scales roughly as a piecewise power function with discharge. This becomes mapped to two linear relationships when seismic power is expressed in decibels. We use least squares linear regressions to fit linear relationships between 0.33 and 2 Hz seismic power (expressed in decibels) and discharge during the calibration period (e.g., Figure S10). A piecewise linear regression was determined to be the best approach, as both stations observe a drastic reduction in sensitivity to discharge once a critical threshold is reached as bankfull flow is approached ($\sim 9 \text{ m}^3/\text{s}$ at SF01, $\sim 10 \text{ m}^3/\text{s}$ at SF03 east, and $\sim 6 \text{ m}^3/\text{s}$ at SF03 north). At lower discharge rates, sensitivity between the four horizontal channels varied from 3.88 to 4.65 dB/(m^3/s) and dropped to 0.84–1.26 dB/(m^3/s) when flow exceeded these threshold levels.

Using the piecewise linear mapping functions, we generate synthetic hydrographs over the full seismic record for each horizontal channel. To minimize the effect of transient spikes in the seismic record, discharge estimates were smoothed using a 1 h (six-sample) moving median filter. Visual inspection of the synthetic hydrographs shows that nearly all variations in discharge, including diurnal cycles, are well determined from 0.33 to 2 Hz horizontal component seismic data from SF01 and the north component of SF03 (Figure 7). The east component of SF03 is unable to resolve discharge above $10 \text{ m}^3/\text{s}$ because of seismic power saturation at high flow rates for this particular component (Figure 6). Correlation coefficients for the four channels between synthetic discharge and pressure transducer/rating curve estimates range from 0.96 to 0.99 and agree to within $\sim 1 \text{ m}^3/\text{s}$ at the 95% confidence interval. We note that pressure transducer and rating curve hydrographic estimates may also contain substantial uncertainty for streams of this size that can easily, for the full range of discharge, exceed the discrepancy between our seismically determined estimate and the available gauge estimate (e.g., Benson & Dalrymple, 1967). However, any such errors in the pressure gauge rating curve are propagated into our synthetic hydrograph during the calibration period.

The largest misfits between the two methods of estimating discharge occur at very low flow ($< 2 \text{ m}^3/\text{s}$), such as during late August when discharge levels become low enough that fluvial signals become overwhelmed by diurnal wind gusts (i.e., the spikes in the synthetic hydrograph in Figure 7; see also Figures 5, S6, and S7). Furthermore, static misfits (i.e., DC offsets) are observed on both SF01 and SF03 during these low discharge conditions. These misfits imply that we are unable to accurately extend our synthetic discharge estimates much beyond the range of what was observed during the calibration period. Ultimately, this underscores the importance of performing the calibration over a broad range of discharge conditions and suggests that the presented methodology may not be suitable for attaining accurate measurement of discharge during extreme events with discharge rates above those observed during the calibration period (e.g., strong flooding). Despite this difficulty in retrieving discharge during low-flow conditions, our results indicate that not only is the 0.33–2 Hz signal generated from fluid flow but that it can be used to determine discharge rates with a reasonably high degree of accuracy in moderate flow conditions using inexpensive and robust geophones.

5. Discussion

5.1. Sources of Seismic Noise Induced by Rivers

5.1.1. Sediment Transport

Bed load transport appears as high-frequency ($> 30 \text{ Hz}$) seismic energy in both mechanistic models (Gimbert et al., 2014; Tsai et al., 2012) and field observations of bed load transport performed in conjunction with seismic measurements taken in close ($< 10 \text{ m}$) proximity to the channel (Barrière et al., 2015; Roth et al., 2016). Thus, the only frequency band we considered that is likely to be sensitive to bed load transport at this site is the 30–45 Hz band. Field observations suggest that bed load transport during the study occurred in sporadic pulses primarily with high discharge during the snowmelt runoff. The seismic records do not match these field observations in which we note significant site-to-site variability in the 30–45 Hz band, inconsistent patterns of hysteresis during the July rainstorm, and no temporally correlated pulse signals in the differential spectrograms (Figures 5, 6, and S6–S9).

Modeling suggests that seismic energy generated by bed load transport should strongly scale with both grain size and impact rate (Tsai et al., 2012). Our lack of observations of candidate sediment transport signal in the seismic record likely arises because the grains transported during our study period (bed load $d_{50} = 2 \text{ mm}$ during peak runoff) are simply too small and/or sparse to generate appreciable seismic energy, even for streamside seismometer installations. In a smaller fluvial system similar to the South Fork, Barrière et al. (2015) were unable to detect seismic bed load signals on a sensor located 2 m from the channel until bed load impact rates reached 200 counts per minute with transported $d_{50} > 3 \text{ mm}$. We conclude that bed load transport is not a significant contributor of seismic energy in this particular fluvial system under the range of conditions measured in 2015 and suspect any ground motions generated by bed load transport occur at high frequencies ($> 25 \text{ Hz}$) where they are overwhelmed by signals generated by turbulent flow.

5.1.2. Turbulent Flow

High-frequency ($> 25 \text{ Hz}$) energy dominates the seismic spectrum throughout the entire study period, yet in contrast to the observations of Roth et al. (2016), is not consistently sensitive to increases in discharge from station to station or component to component (Figures 3–6 and S3–S9). Although we suspect this high-frequency energy is generated by turbulent flow and power in this band broadly increases with discharge, we are unable to consistently spectrally distinguish differing discharge conditions at high frequencies across the stations (Figures 3, S3, and S4). Additionally, high-frequency power shows $\sim 20 \text{ dB}$ of variability between the three stations at maximal discharge, suggestive of exceptional sensitivity to local sources and/or site effects. These observations indicate that mechanistic models and even calibrated field observations may be incapable of generating reliable estimates of discharge using high-frequency signals recorded in this manner.

Power in the 12–22 Hz band is overall more sensitive to discharge than any other frequencies $> 1 \text{ Hz}$ and shows a clear spectral response to discharge (Figures 3 and 4). Recent modeling suggests that $> 1 \text{ Hz}$ seismic power induced by turbulent flow, P_w , propagating as Rayleigh waves and generated by a turbulent flow/boundary roughness mechanism, should scale with discharge, Q , as a power function, with $P_w \propto Q^{7/5}$ proposed for surface streams (as opposed to sub-surface conduits; Gimbert et al., 2014, 2016). However, this theoretical relation is derived for far-field seismic observations at distances from the river that are much greater than the river or conduit width. In this near-field study, with sensors within or just above the water table and just $\sim 1 \text{ m}$ from the stream bank, we do observe clear power function scaling with discharge but find that the

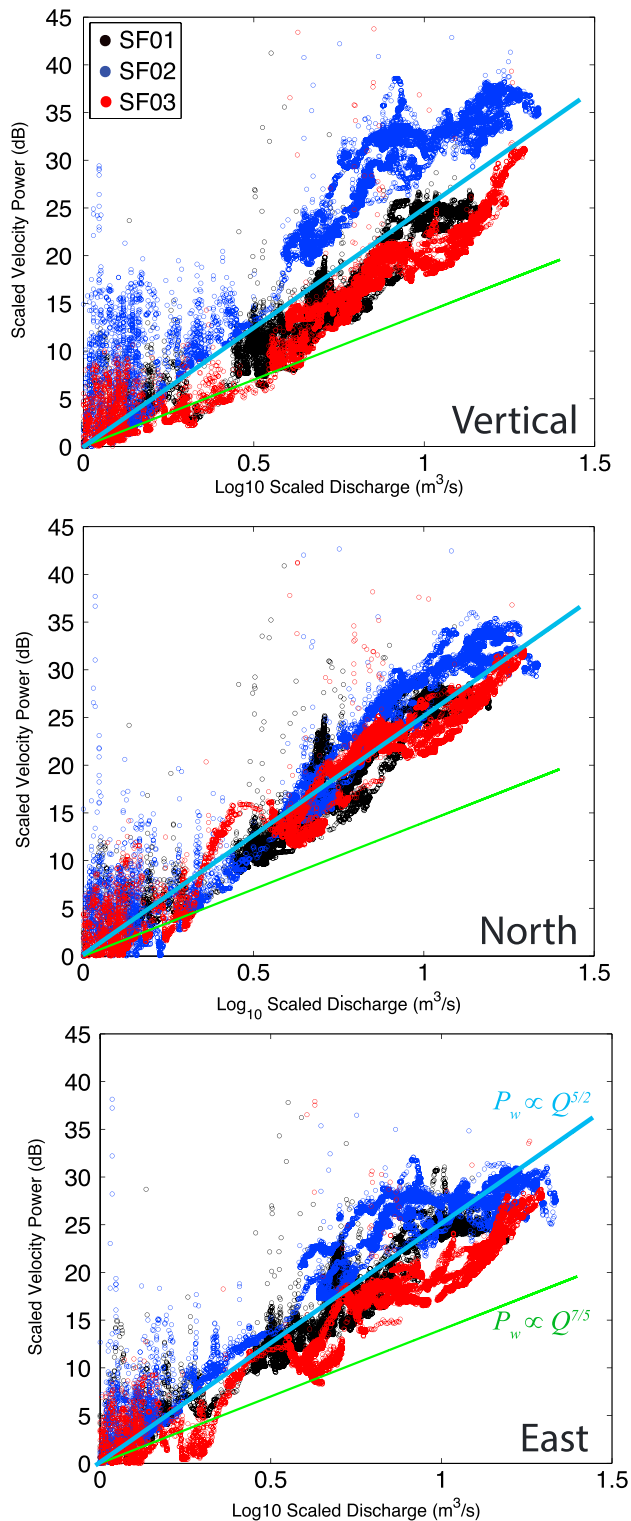


Figure 8. The 12–22 Hz velocity power versus discharge for all three components plotted on a log-log scale. Both velocity and discharge measurements have been normalized through subtraction by values attained during late August when discharge was minimal ($\sim 1 \text{ m}^3/\text{s}$). The green curve shows the Gimbert et al. (2016) relation in which seismic velocity power, P_w , generated by turbulent flow (no bed load transport) scales with discharge, Q , approximately as $Q^{7/5}$ (line curve). The cyan curve is for $P_w \propto Q^{5/2}$.

seismic power in this band is more sensitive to discharge ($P_w \propto Q^{5/2}$) than the scaling derived by Gimbert et al. (2016) (Figure 8). Using this relation and estimates of discharge from the pressure transducer, we are able to model horizontal power levels within ~ 3.5 dB and vertical power by 6.1 dB with 95% confidence across the entire range of discharge. However, our observation of differing and inconsistent hysteresis characteristics between stations in the 12–22 Hz band (Figure 6) suggests that the seismic response in this band is particularly sensitive to site characteristics, perhaps due to small changes in bed morphology or propagation effects.

Finally, our ability to create accurate synthetic hydrographs from the 0.33–2 Hz band on the horizontal components indicates that seismic signals at these frequencies are exceptionally sensitive to discharge (Figure 7). Given the spectrum of river processes, we attribute this to near-sensor forcing via turbulent flow. To our knowledge, these signals have not been previously reported in past fluvial seismology studies with more distant sensors, with the most spectrally similar signal (0.5–2 Hz) being observed during a controlled flood event (CFE; max discharge $\sim 1,200 \text{ m}^3/\text{s}$) at Hance Rapids in the Grand Canyon (Schmandt et al., 2013).

The low-frequency signal recorded at Hance Rapids was observed on a seismometer ~ 35 m from the channel and showed stronger variations in power on the vertical components than the horizontal components. The authors attributed the signal to turbulent flow, although Gimbert et al. (2014) subsequently suggested it may alternatively or additionally result from standing wave activity. We have processed the data from Hance Rapids identically to the South Fork dataset (Figures S11 and S12) and note that in contrast to our observations on the South Fork (Figures 3, S3, and S4), horizontal component power does not have a second low-frequency peak even during the CFE. In fact, despite discharge rates 70 times larger than observed on the South Fork, absolute horizontal component power at 1 Hz from Hance Rapids is at least 25 dB less than observed here. Finally, 0.33–2 Hz H/V ratios observed on the South Fork increase more than 25 dB during peak snowmelt runoff (Figure 9), whereas the low-frequency signals recorded at Hance Rapids during the CFE actually become slightly more vertically polarized (Figure S12). These large differences in both absolute amplitude and behavior under changing discharge conditions suggest that the 0.5–2 Hz signal observed at Hance Rapids results from a different process than we observe at the South Fork. We next investigate a possible mechanism for producing such strongly horizontally polarized seismic signals at low frequencies.

5.2. Sources of Strongly Horizontally Polarized Seismic Signals

Horizontal-to-vertical ratio (H/V) ratios (Figure 9) are attained through subtraction of the vertical component PSD from the horizontal component PSD since PSD power is expressed in dB (a logarithmic ratio). Both the 12–22 Hz and 0.33–2 Hz bands show systematically much higher levels of seismic power on the horizontal components than on the vertical. Such preferential horizontal excitation can arise from (1) local resonance caused by site conditions, (e.g., Bodin et al., 2001; Bodin & Horton, 1999; Nakamura, 1989); (2) source characteristics, such as predominant generation of horizontally polarized Love waves; and (3) instrument tilt (e.g., Peterson, 1993; Sorrells, 1971; Wilson et al., 2002). Understanding enhanced horizontal power in the 0.33–2 Hz and 12–22 Hz bands is particularly relevant for this study.

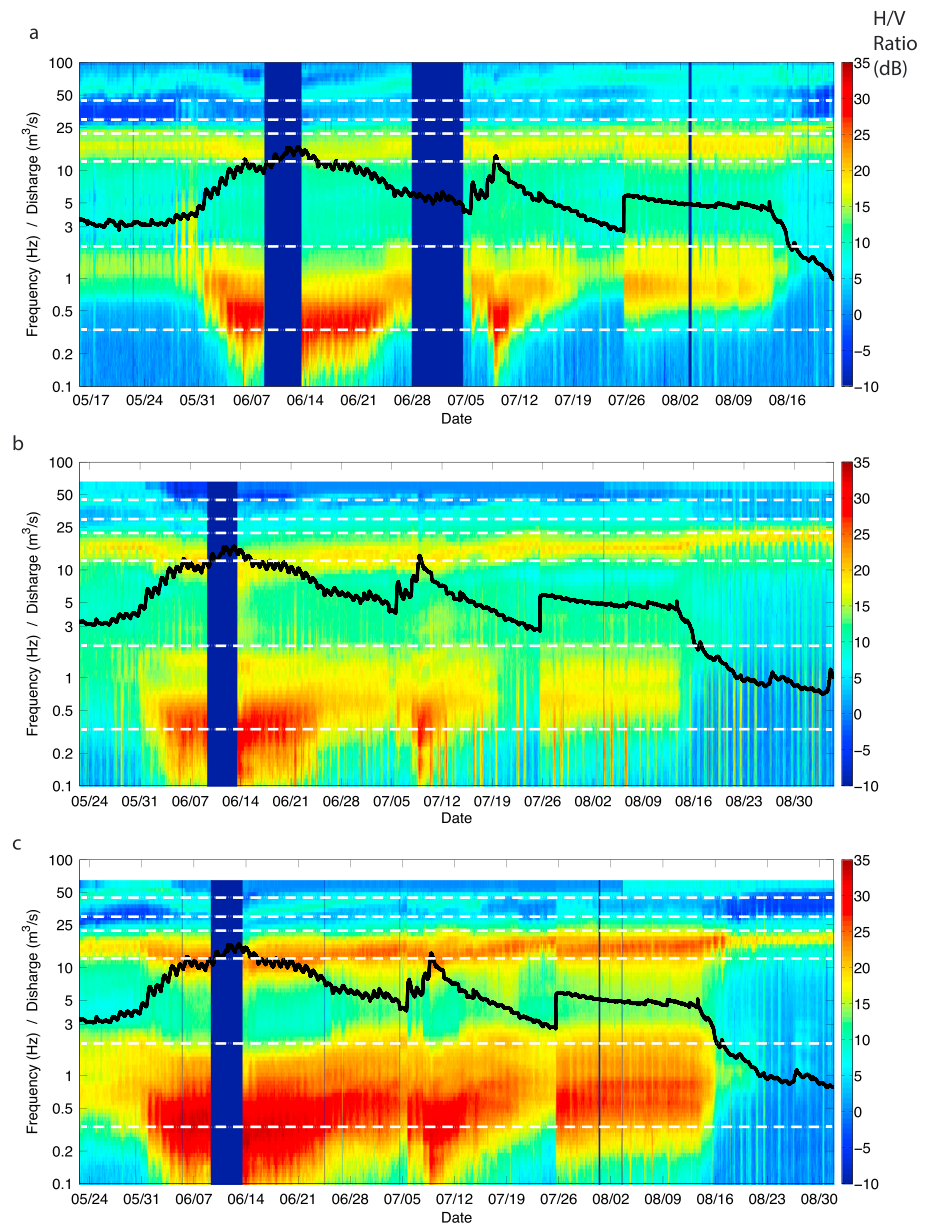


Figure 9. Horizontal-to-vertical (H/V) ground acceleration power ratio spectrograms at (a–c) SF01–SF03 with peak H/V ratio bands occurring between 12–22 Hz and 0.1–2 Hz. Discharge is shown as a black curve, and white horizontal dashed lines delineate the three period bands excited by discharge as described earlier in the text.

5.2.1. Resonance From Local Site Conditions

The 0.33–2 Hz and 12–22 Hz signals are identifiable as peaks in H/V spectrograms (Figure 9) but behave differently under changing discharge. Both the amplitude (Figure 9) and the frequency of H/V acceleration power ratios within the 12–22 Hz band remain relatively constant over a wide range of discharge conditions and persist even during the minimal flow period at the end of August. The only observable changes in the peak frequency coincide with flooding at all stations when the frequency of the H/V peak shifts slightly from 16 Hz to 11 Hz (13.5 Hz on SF01). Thus, we propose that elevated H/V levels within the 12–22 Hz band are not related to fluvial process and instead represent the site response of seismically low-velocity alluvium overlying higher velocity bedrock. We attribute the shifting frequency during flooding events to further decreases in velocity of the overlying sediment as it becomes water saturated.

In a simple two-layer system, where a higher seismic impedance substrate is overlain by low impedance sediments, the fundamental seismic resonance frequency, f_r , is related to the sediment layer thickness, Z , by

$$f_r = \frac{V_s}{4Z}, \quad (1)$$

where V_s is the average-shear wave velocity in the sediment layer (e.g., Ibs-von Seht & Wohlenberg, 1999). The fundamental frequency can be determined from the frequency of peaks in the H/V ratio of ambient seismic noise (e.g., Nakamura, 1989). Assuming an average near-surface alluvium shear wave velocity of 137 m/s (consistent with near-surface, silty-clay such as in which the sensors were emplaced in; e.g., Cuny & Fry, 1973), the 16 Hz H/V peak corresponds to a reasonable low velocity layer thicknesses of ~ 2 m in our study area. Assuming a constant sediment thickness, the average velocity would have to drop to 94 m/s during flooding to solely enable the shift in H/V peak to 11 Hz. Considering water content is known to have a strong effect on shear wave velocities in silts and clays (e.g., Hamilton, 1976) and that the near-surface shear wave velocity of saturated silty-sand at the surface has been measured to be ~ 88 m/s (e.g., Cuny & Fry, 1973), this explanation seems quite plausible. We note that the above calculations are rough estimates and that the actual low-velocity layer thickness and velocities adjacent to fluvial systems vary spatiotemporally due to changes in the saturated thickness and heterogeneous deposits from the migrating channel. The ~ 2 m low-velocity thickness we calculate is consistent with sediment depths obtained near the headwaters of South Fork of the Cache la Poudre (Rubin et al., 2012) as well as with limited field observations from our study site.

Ambient noise modeling of H/V ratios indicates that when Rayleigh waves are generated at the surface and recorded at source-receiver distances within 4–50 times the low-velocity sediment thickness, the resulting H/V peak is primarily an indicator of horizontal ellipticity of the fundamental mode Rayleigh waves (Bonnefoy-Claudet et al., 2006). Thus, our observations are consistent with Rayleigh waves locally (< 200 m) generated by turbulent flow dominating the wavefield in the 12–22 Hz band as considered in Gimbert et al. (2014) model. We suggest that elevated ~ 15 –22 Hz horizontal noise observed over a wide range of discharge conditions by Barrière et al. (2015) may be the result of a similar resonance effect of locally generated Rayleigh waves propagating in low-velocity soils and not necessarily evidence of Love wave generation from turbulent flow.

Complete amplitude characterization of H/V energy below 0.33–1 Hz is not possible due the vertical noise levels being dictated by the self-noise of the instrumentation (Figure 3). Thus, our calculated H/V ratios below these frequencies represent minimum amplitudes and the strong apparent decrease in the frequency of the H/V peak with increasing discharge is primarily driven by horizontal power levels beginning to exceed the self-noise levels of the sensor while the verticals do not (Figures 3 and 9). However, despite these limitations, there is still enough observational evidence to rule out local geological structure as the source of elevated horizontal noise between 0.33 and 2 Hz. This is most easily demonstrated by showing (below) that the low-frequency signals observed in this study are not propagating as elastic seismic waves and thus are not sensitive to the deeper geological structure required to produce elevated H/V ratios at low frequencies (Bodin & Horton, 1999).

5.2.2. Elastically Propagating Seismic Waves

Although the generation of Love waves from both bed load and turbulent flow has been proposed (e.g., Barrière et al., 2015; Gimbert et al., 2014; Schmandt et al., 2013), to our knowledge no observations have yet confirmed strong fluvial excitation of these surface waves. Consistent with Love waves, we observe substantial horizontal polarization in the 0.33–2 Hz band. However, as explained below, our amplitude observations exclude any type of elastically propagating seismic wave from creating this signal.

During maximal discharge, ambient horizontal component power observed on the east component of SF03 exceeds the Global High Noise model between 0.1 and 0.2 Hz and is 12 dB larger than signals observed on SF01 and SF02 at 0.2 Hz (Figures 3 and S5). Ambient seismic noise within the 0.1–0.2 Hz frequency band is well known to be typically dominated by surface waves generated by ocean wave interactions known as microseisms (Aster et al., 2008; e.g., Peterson, 1993). Due to the low-frequency and attenuation of surface waves, microseism signals propagate from coastal regions deep into the continental interior and are ubiquitously observed on seismic stations across the globe. However, the 12 dB amplitude difference that we observe at 0.2 Hz between seismic stations separated by much less than one wavelength is inconsistent with observations of elastically propagating seismic waves in this period band.

5.2.3. Instrument Tilt

We next consider local, nonseismic signals recorded by seismometers. In a gravity field characterized by acceleration g , a tilt of the sensor of angle θ appears as an apparent horizontal acceleration of amplitude $g \sin \theta$ (e.g., Wielandt & Forbriger, 1999). Thus, small tilts can generate large seismic signals on the horizontal components. The accelerations observed on the horizontal components in this study could be produced through tilts of a few microradians (e.g., $\sim 10 \mu\text{rad}$ at SF03). This phenomenon commonly dominates long-period ($> \sim 20$ s), horizontal component noise on near-surface seismometers and is caused by local ground deformation in response to atmospheric pressure (e.g., Peterson, 1993; Sorrells, 1971; Wilson et al., 2002). In contrast to long-period, elastically propagating seismic waves, tilt signals induced in this manner can vary substantially between seismometers placed just a few meters apart from each other (Aderhold et al., 2015; Rohde et al., 2017).

Tilt-induced signals can occur at any frequency at which the sensor is oscillating about its vertical axis. For instance, elevated horizontal noise levels and an H/V peak between 0.33 and 2 Hz observed at exposed rock-sited seismic stations in the Transantarctic Mountains were interpreted to arise through this phenomenon as the sensor oscillated during strong wind events (Anthony et al., 2015). We thus attribute the long-period horizontal signals observed here to instrument tilt caused by local, viscoelastic deformation within the hyporheic zone. We explore the hypothesis that large, turbulent eddies create temporally varying pressure on the banks and induce near-field tilt oscillations at frequencies of ~ 0.1 –2 Hz at these stations.

5.3. Large-Scale Eddies as a Potential Forcing of Near-Field Signals in the Hyporheic Zone

The hyporheic zone is defined as the region below streambeds and adjacent to stream banks where channel water and groundwater mix (e.g., White, 1993). During snowmelt runoff, strong precipitation events, and upstream dam releases, water moves from the stream to be taken up by the stream banks (e.g., Winter et al., 1998). By placing the geophone in exceptionally close proximity to the channel (~ 1 m) and at or below the water table, the local, saturated soil may experience changes in pore pressure caused by incoming bank water storage or flow during high discharge events (Chen & Chen, 2003). Recent modeling work has suggested that advective pumping forces related to large-scale eddy formation likely induce the deepest flow paths of hyporheic exchange (Boano et al., 2011). Here we perform simple calculations to investigate if the low-frequency, horizontal signals we observe are spectrally consistent with local pressure variations on the banks induced by turbulent eddies in a flat (e.g., no bedforms) stream. These changes in hydraulic head will induce a time-varying force applied to a given point not only on the channel bed (as considered in Boano et al., 2011) but also on the channel banks. Specifically, the total force applied to the banks of a rectangular channel is proportional to the square of the flow depth (e.g., Turcotte & Schubert, 2005). We note that pressure variations on the channel bed should result in the generation of seismic waves (perhaps observed as low-frequency energy during the Grand Canyon controlled flood studied by Schmandt et al., 2013); however, the high-self noise of the sensors used in this study coupled with modest discharge rates prohibit characterizing these weaker, nontilt signals.

From in situ measurements of stream velocity fluctuations in a gravel bed river, large-scale turbulent eddy structures were found to have lengths, L , of 3–5 times the flow depth, h , by Roy et al. (2004). Using the assumptions of Boano et al. (2011), we consider a single sinusoidal pressure pulse of length L propagating downstream at the average stream velocity, U . The dominant frequency of pressure fluctuations, f_E , generated by large-scale eddies can then be approximated as

$$f_E \sim \frac{U}{Ch}, \quad (2)$$

where the constant C varies from 3 to 5 based on the findings of Roy et al. (2004). Measurements of flow velocity and depth were attained from the access bridge (Figure 1) over the study period to constrain the relation between stage height and discharge. Using these measurements, we calculated f_E for a variety of discharge conditions (using U varying between 0.3 and 1.9 m/s and h varying between 0.4 and 1.0 m) and find that f_E varies between 0.1 and 0.7 Hz, which is consistent with the spectrum of elevated H/V ratios in the seismic record. We argue that pressure variations on the banks of the South Fork generated from large scale turbulent eddies represent a viable forcing mechanism to generate low-frequency tilt oscillations observed by our near-channel sensors. This mechanism is similar to that described by Aderhold et al. (2015) regarding localized tilt observed during noise analysis of broadband sensors deployed in soft alluvium near the Rio Grande.

In this model, we envision changing pressure on the banks of the river directly generating viscoelastic deformation, including sensor tilt, within the water-saturated hyporheic zone. As the turbulent eddies considered here induce changing pressure over the entire depth of flow (Roy et al., 2004), the magnitude of the force generated on the channel banks is proportional to both the amplitude of the pressure variation and the area of bank submerged below water level. Under these conditions, the total force generated by these pressure variations scales strongly with stage height until out-of-bank flooding occurs. During these times, additional increases in stage do not increase the area on the banks that pressure variations may act on and changes in the total force become primarily driven by the channel-perpendicular hydraulic head.

This conceptual model has additional features that are consistent with our low-frequency seismic observations besides similar frequency content. First, because of the bank geometry of this straight reach, we would expect tilts induced by this process to occur predominantly in a direction perpendicular to the channel. The east components (aligned perpendicular to the channel; Figure 1) indeed show both larger amplitude horizontal accelerations (Figure 3) as well as the strongest response to changing discharge conditions (Figure 4). Next, we would expect the amplitude of tilt oscillations to scale strongly with discharge until flooding occurs, after which the signal should become less responsive to flow rate. This is reflected in a large reduction in east component sensitivity to discharge once it exceeds the bankfull capacity of the South Fork of $\sim 10 \text{ m}^3/\text{s}$ (Figure 6). Finally, measurements of macroscale turbulence behind flow obstructions, such as boulders, have been modeled and observed to significantly enhance turbulence and generate pressure pulses orthogonal to streamflow through vortex shedding (Tritico & Hotchkiss, 2005). This may explain why we observe the largest horizontal signals and highest sensitivity to discharge at SF03 (Figures 4 and S5), which is located immediately downstream of several large boulders (Figure 1).

Fundamentally, the low-frequency ($< 2 \text{ Hz}$) horizontal component signals observed in this study may thus arise from pressure variations on the banks generated from large-scale turbulent eddies. These findings suggest that the forces associated with macroscale eddies may be able to be remotely and continuously monitored from outside the channel using seismic observations. As large-scale turbulent structures in fluvial systems are thought to have a significant impact on hyporheic exchange at depth, monitoring the frequency and accelerations induced by hydraulic pumping forces may have use in modeling hyporheic flow (e.g., Boano et al., 2011).

As with other efforts to model seismic signals from fluvial systems, quantitative numerical modeling of this process is made difficult by the highly nonlinear dynamics of turbulent flows and the influences of bed obstructions whose geometries vary with flow depth (e.g., Boano et al., 2011; Tritico & Hotchkiss, 2005). Instead, we suggest that future targeted measurements of turbulence using acoustic Doppler velocimeters in conjunction with near-channel seismic observations can validate and constrain the influence of large-scale eddies on the frequency and magnitude of hydraulic forces on the channel banks. We suggest that such observations be conducted with low-profile, waterproof broadband sensors directly buried at varying distances from the channel to more fully characterize signal behavior, including long-period tilt and its attenuation with distance from the channel through H/V analysis.

6. Conclusions

Analyzing seismic records from three-component geophones deployed in the hyporheic zone ($\sim 1 \text{ m}$ from the channel) of a small mountain river during spring and summer runoff periods in conjunction with measurements of discharge, we find a wide frequency range (0.1–45 Hz) of signals to be excited by discharge. However, we do not observe seismic signals consistent with periodic measurements of bed load transport or continuous measurements of suspended sediment that were also taken during the course of the study and conclude that the grain size of both bed load and suspended sediment in this case are insufficient to generate detectable seismic signals, even on sensors located in close proximity to the channel.

Although the ambient noise field is dominated by high-frequency ($> 25 \text{ Hz}$) acceleration energy over the course of the entire study period, we are unable to generate consistent relations between high-frequency seismic energy and discharge. In contrast, power between 12 and 22 Hz approximately scales with discharge, Q , as $Q^{5/2}$ and has consistently higher power levels on the horizontal components. We show that the horizontal polarization in this frequency band is likely a resonance effect due to locally generated Rayleigh waves propagating within a shallow ($\sim 2 \text{ m}$) low-velocity sediment layer.

Finally, low-frequency (< 2 Hz) signals observed overwhelmingly on the horizontal components also show a power function sensitivity to discharge. The amplitude and attenuation properties of these signals from station to station are not consistent with elastically propagating seismic waves, and we instead suggest that they arise through sensor tilt in response to local viscoelastic deformation within the hyporheic zone. We present a simple conceptual model in which these tilt signals are generated by large-scale eddies inducing temporally varying pressure fluctuations on the banks of the stream. The model is consistent with several observations of the low-frequency signals on all the sensors including spectral content, polarization, and relationships between amplitude and discharge. Although further field measurements are needed to validate this interpretation, these findings suggest that continuous temporal monitoring of the frequency and forces associated with large scale eddies may be feasibly measured using seismographs emplaced in close proximity to the channel.

Calibrating horizontal seismic power with flow rates over the course of a several-day rainstorm event for individual sensors, 0.33–2 Hz signals can be readily used to invert for stream discharge to within a $1 \text{ m}^3/\text{s}$ (95% confidence) discrepancy relative to estimates derived from a local pressure-based gauging system. As these signals are large and occur at lower frequencies than anthropogenic noise and bed load transport processes, they may have utility for continuously monitoring discharge, even in noisy urban environments and/or where appreciable bed load transport is occurring.

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