

Introduction

Many subduction zones worldwide host large, devastating earthquakes, such as the 2011 M9 Tohoku-Oki earthquake. To increase resilience to such events, earthquake engineers utilize what are known as Ground Motion Prediction Equations (GMPEs) that quantify ground motion during an earthquake. GMPEs relate ground motion to a number of physical parameters such as earthquake magnitude, fault mechanism, local site characteristics, and namely, attenuation. Seismic-wave attenuation is the decay in amplitude of seismic waves as a function of distance from the earthquake source.

In addition to fast, seismic slip many subduction zones also host slow, largely aseismic slip. These "slow slip earthquakes" occur on timescales of weeks to months and are often accompanied by an episode of weak seismic signals known as "tectonic tremor," or simply "tremor." Tremor behaves differently than regular earthquakes in that it is comprised of many small earthquakes that radiate low-frequency seismic energy and originate at the plate interface down-dip of where large earthquakes typically occur (Beroza and Ide, 2011) (Figure 1). This behavior is thought to reflect variation in frictional properties, effective stress, or both in between the seismogenic and aseismic sections of the plate interface. Because tremor episodes have such a weak seismic signal and occur on long timescales relative to ordinary earthquakes, we typically do not feel them.

Our goal is to quantify seismic wave attenuation in Cascadia and determine its spatial variations using tectonic tremor. By inverting tremor ground motion data for an attenuation parameter, we can analyze if and how our results vary spatially in Cascadia and attempt to relate these variations to lithology and/or other physical properties.

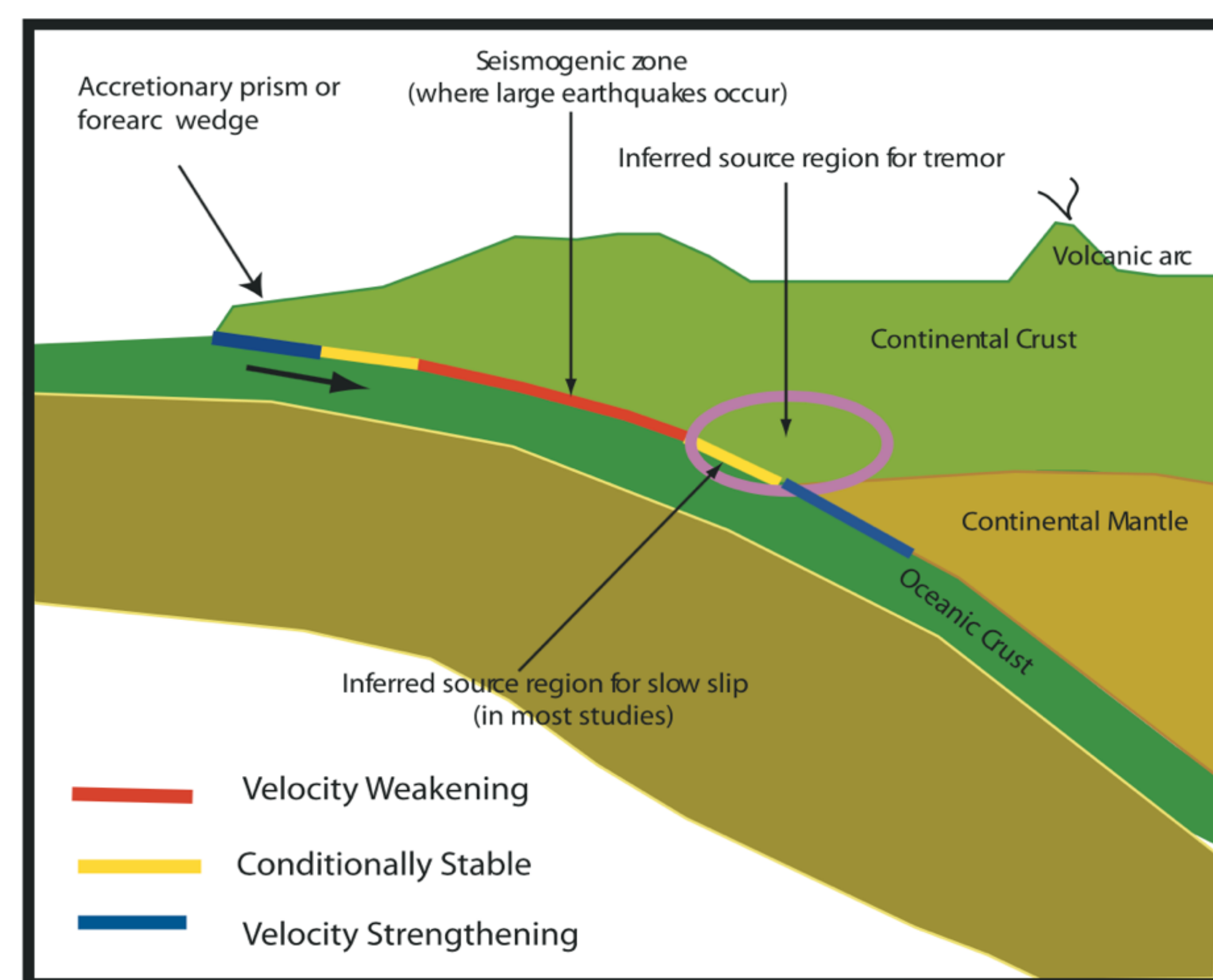
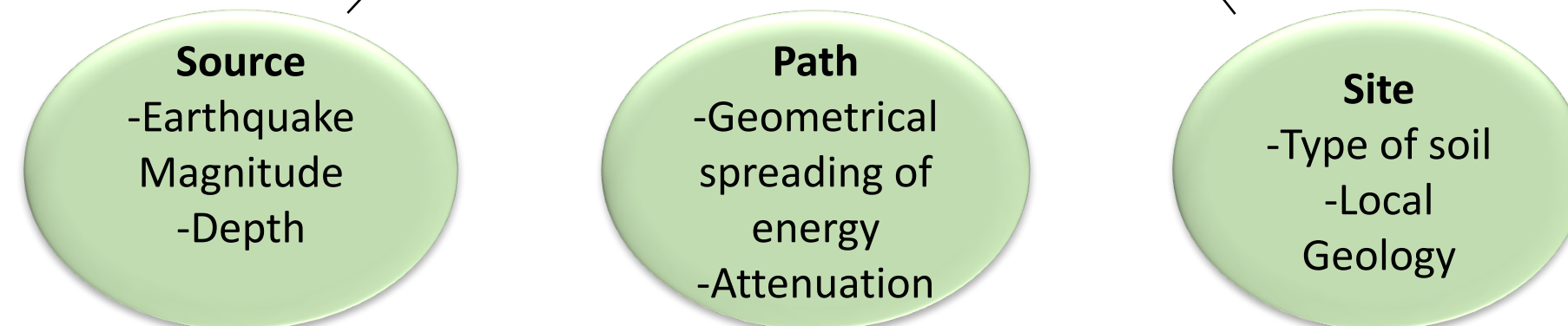


Figure 1. Cross-section of a subduction zone. The pink circle indicates where slow slip earthquakes and associated tremor are generated, down-dip of the seismogenic zone where larger earthquakes occur. Schwartz and Rokosky, 2007.

Ground Motion Prediction Equations (GMPEs)

$$\text{Ground Motion} = (\text{Source}) \times (\text{Path}) \times (\text{Site})$$



Here, we're looking at Path Effects:

$$A_{ij} = A_{io} \exp\left(\frac{-\pi R_{ij} f}{Q\beta}\right) \frac{1}{R_{ij}} S_j$$

- Describes the amplitude of the seismic wave at a certain station as combination of the initial source amplitude, geometrical spreading (following a 1/R relationship) and attenuation.
- Amplitude follows a 1/R relationship with distance, where R is the hypocentral distance from the earthquake to the station. However, attenuation causes the energy of the wave to dissipate more quickly along its path.
- We define the **anelastic attenuation parameter** as $c2 = \frac{\pi f}{Q\beta}$, according to Baltay and Beroza (2013).

Terms:
 A_{io} = Initial amplitude of earthquake i
 A_{ij} = Amplitude of earthquake i recorded at station j
 R_{ij} = Hypocentral distance
 $\frac{1}{R_{ij}}$ = Geometrical spreading of energy
 S_j = Site term, unique for each particular station
 Q = Quality factor
 β = Shear wave velocity

The Data

- For the tremor episodes of 2009, 2012 and 2015-2016:
- We obtain data from the Pacific Northwest Seismic Network (PNSN) and Incorporated Research Institutions for Seismology (IRIS).
- Ground motion data are waveforms containing the amplitude versus time of individual earthquake tremor events that comprise the tremor episode (Figure 3). Data are corrected for instrument response, filtered in the 1-10 Hz frequency range and converted to velocity, acceleration and displacement.
- We utilize the Peak Ground Acceleration (PGA) and Peak Ground Velocity (PGV) as amplitude measures in the inversion. These are taken as the peak amplitude in each one-minute time window for acceleration and velocity, respectively.
- Baltay and Beroza (2013) found that amplitudes are consistently higher in the daytime due to anthropogenic noise, as shown in Figure 4, so we limit our data to those occurring at night. In addition, we further constrain our dataset to the largest 10% of events in each tremor episode.

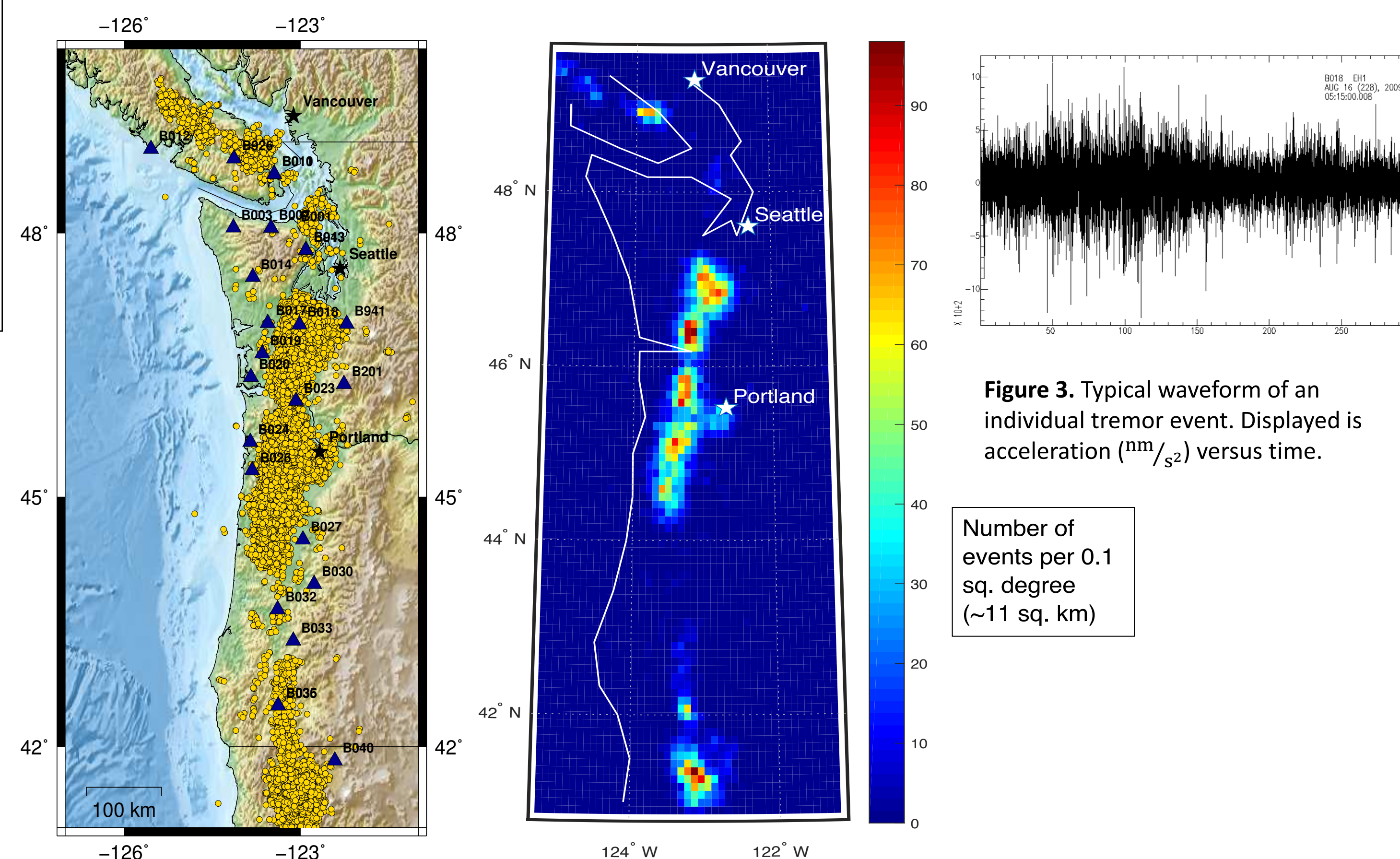


Figure 2. (Left) Distribution of epicenters of tremor events for the tremor episode in 2009, indicated with yellow dots, and the stations used in the analysis, indicated with triangles. (Right) Density of tremor events for the same 2009 tremor episode.

Figure 3. Typical waveform of an individual tremor event. Displayed is acceleration (mm/s^2) versus time.

Inversion Scheme to Obtain the Attenuation Term

Starting from our original equation,

$$A_{ij} = A_{io} \exp\left(\frac{-\pi R_{ij} f}{Q\beta}\right) \frac{1}{R_{ij}} S_j$$

we take the natural log and leave out the site term to solve for a preliminary $c2$ value, and designate a $c1$ term, describing the initial amplitude for each event:

$$\ln A_{ij} = c1_i - c2 * R_{ij} - \ln R_{ij}$$

We use differential ground motion amplitudes by comparing the amplitude of each event at two different stations. Subtracting the amplitude of event i at station 1 and the amplitude of event i at station 2, for all possible station combinations:

$$c2 = \frac{(\ln A_{i1} - \ln A_{i2}) + (\ln R_{i1} - \ln R_{i2})}{(R_{i2} - R_{i1})}$$

We compute a $c2$ term for each event and each station combination, then substitute our preliminary $c2$'s to determine the $c1$ terms for each event:

$$c1_i = \frac{1}{n} \sum_{j=1}^n \ln A_{ij} + \ln R_{ij} + c2 * R_{ij}$$

Plugging back in to recover site terms for each station:

$$\ln S_j = \frac{1}{m} \sum_{i=1}^m \ln A_{ij} - c1_i + \ln R_{ij} + c2 * R_{ij}$$

And there we have it! We use all of our previously determined terms to find our final attenuation parameter:

$$c2 = \frac{\ln A_{ij} - \ln S_j - c1_i + \ln R_{ij}}{-R_{ij}}$$

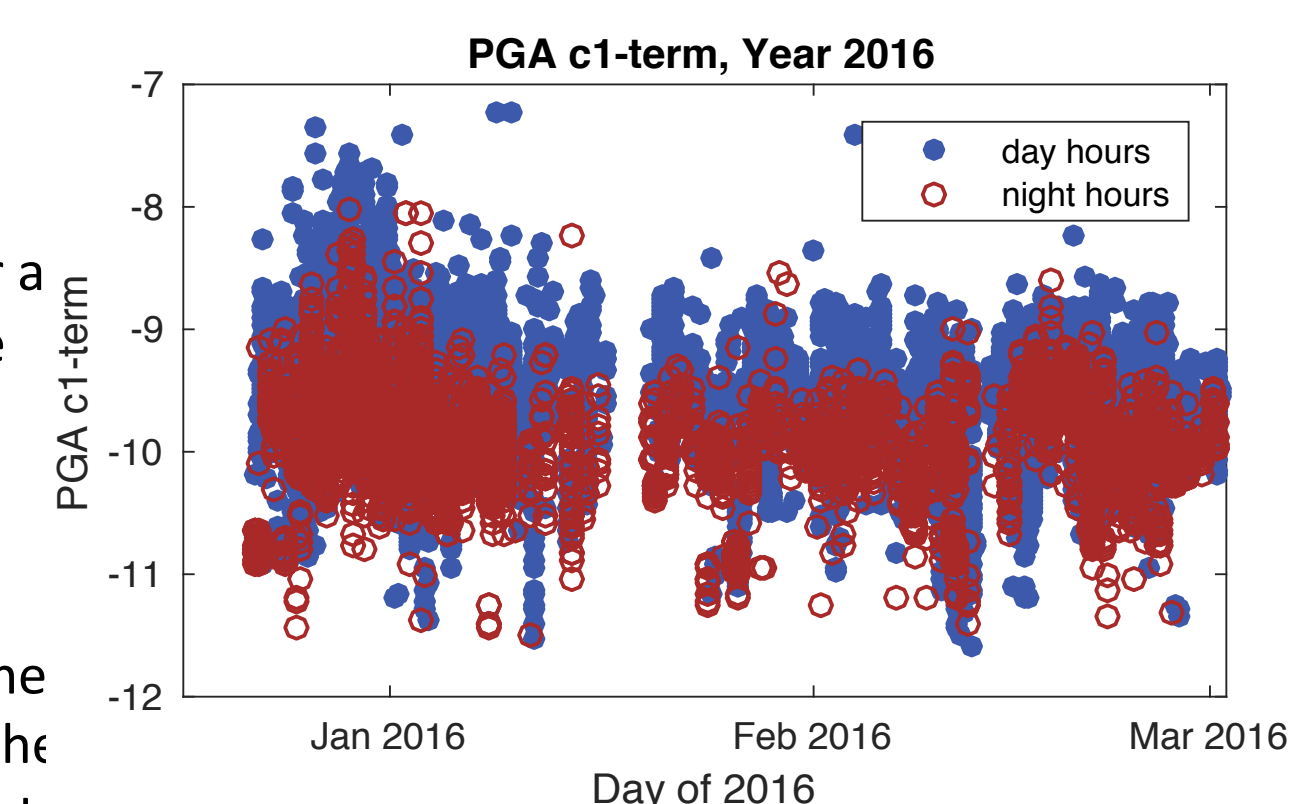
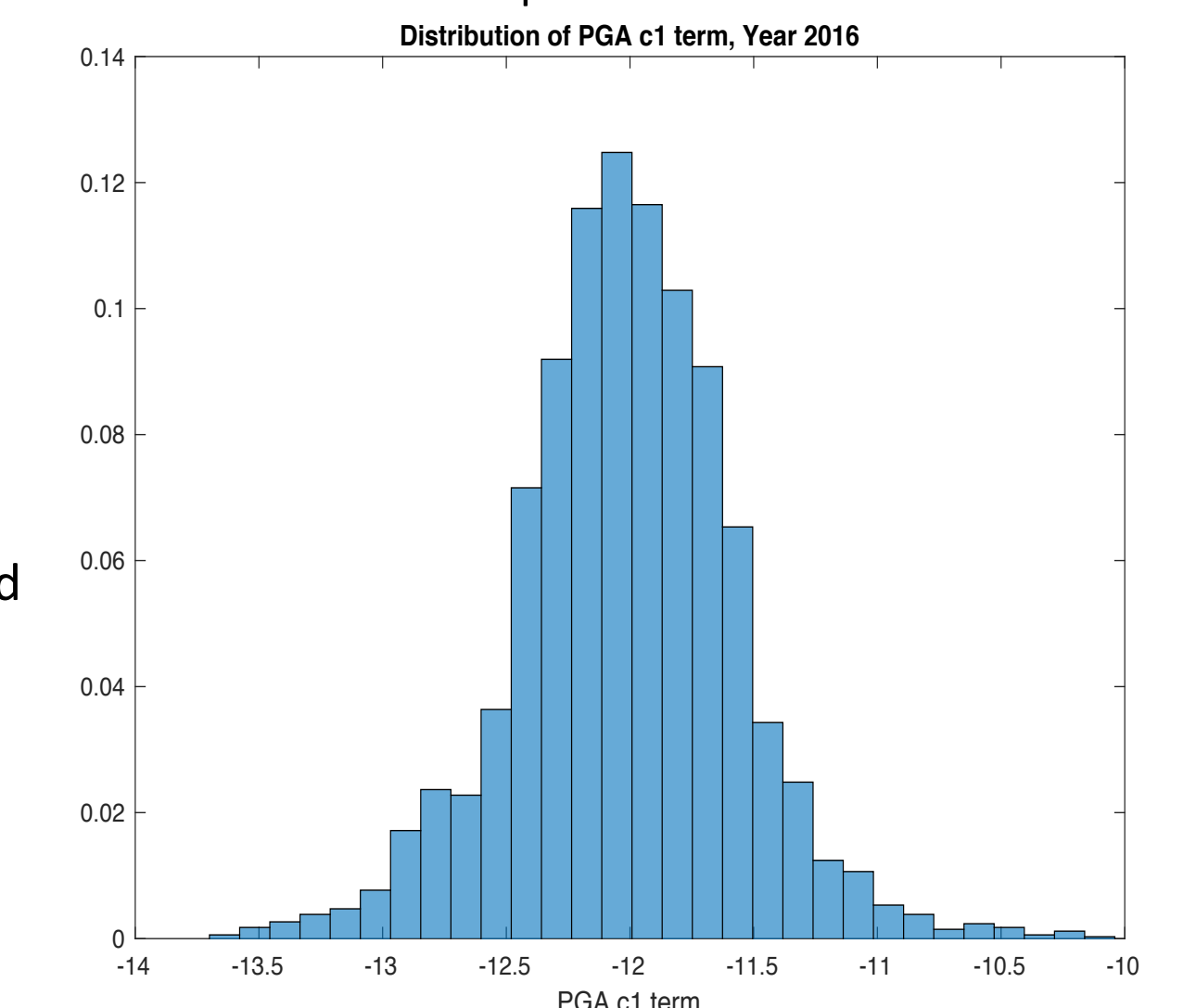


Figure 4. (Above) PGA $c1$ -terms are indicative of the relative magnitude of each individual tremor event. Figure 4 displays the $c1$ terms over the duration of the 2016 tremor episode, illuminating the dependence on the time of day. Day hours exhibit consistently larger amplitudes, amplified during the daytime due to noise from humans! (Below) Normal distribution of PGA $c1$ -terms for the 2016 tremor episode.



Preliminary Results

PGA

Year / Value	Median	Mean
2009	0.0069	0.0064
2012	0.0060	0.0061
2016	0.0061	0.0055

PGV

Year / Value	Median	Mean
2009	0.0052	0.0051
2012	0.0050	0.0056
2016	0.0047	0.0049

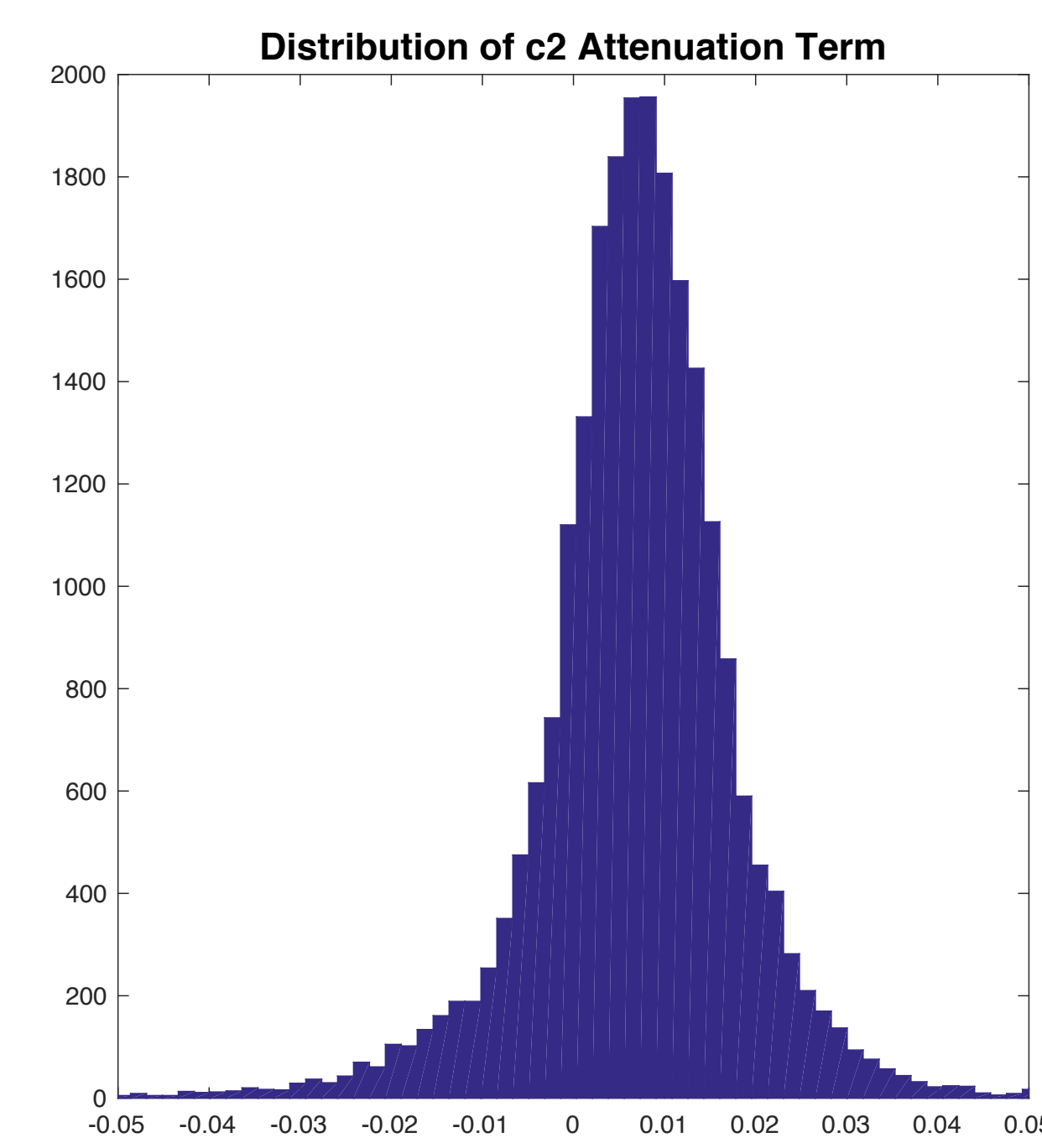
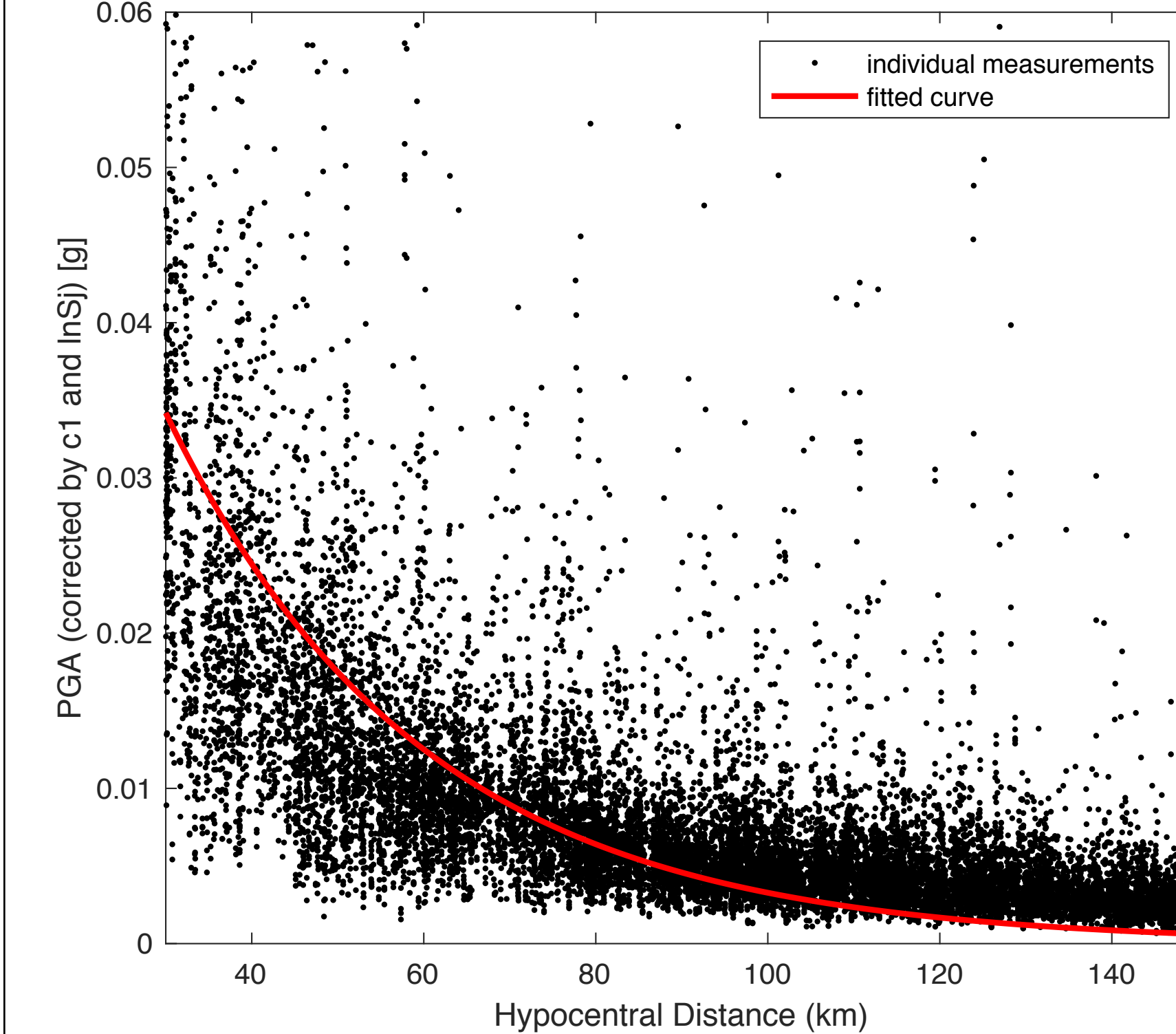


Figure 5. (Above) Final $c2$ values for each year, computed using both PGA and PGV. The similarity between the mean and median $c2$ values is what we expect for the normal distribution as shown on the right, and indicates that these are relatively stable results.

PGA versus Distance for Corrected Dataset *



Our attenuation terms for PGA are similar to those of Baltay and Beroza (2013), which found a median $c2=0.00647$ using data from the years 2010, 2011 and 2012, where the term varied from 0.00644 and 0.00666. Although our PGV $c2$ terms are slightly larger than those of Baltay and Beroza (2013), 0.00467-0.00480, our results are very similar.

Figure 6. PGA amplitudes versus distance, corrected with $c1$ and site terms obtained from the inversion.

What's Next?

Quantifying seismic-wave attenuation allows a straightforward way of determining its spatial variations in Cascadia. Next, we will repeat our analysis regionally to determine if the attenuation parameter experiences significant changes. Changes in seismic-wave attenuation could result in significantly different ground motions in the event of a very large earthquake, hence quantifying attenuation may help to better estimate the severity of shaking in densely populated metropolitan areas such as Vancouver, Seattle and Portland.

References

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