Body Wave Attenuation Structure in Southern Mexico

Ting Chen[1], Robbert W. Clayton[1]

1. California Institute of Technology, Seismological Laboratory 252-21, Caltech, Pasadena, CA 91125

Abstract

The 2-D model of attenuation structure is determined along a 550 km trench-perpendicular profile in southern Mexico. Velocity spectra from 14 moderate earthquakes recorded by the Meso American Subduction Experiment (MASE) array which consists of 100 broadband sensors from Acapulco to Tampico are used. By assuming a Brune-type source, a path-averaged frequency-independent Q is obtained for each seismogram in the frequency band 0.5 Hz to 7-30 Hz, depending on the signal quality. These measurements are then inverted for spatial variations in Q. The 1-D tomography result shows a pattern of Q qualitatively similar to other subduction zones, with low attenuation crust (Q ~ 1000), and high attenuation in the mantle wedge beneath the Trans-Mexico-Volcanic-Belt (Q < 250). The location of the low-Q region and the variation of the Q value also provides some constraints on the geometry of the subducting slab, or with the structure provided by other methods such as receiver functions, the Q estimates can be used to estimate variations in viscosity.

Method

Spectral Analysis

The Fourier velocity spectral amplitude of a body wave from event j, recorded at station i, can be written as [e.g. Garcia, 2004]

$$A_{ij}(f) = CS_j(f)I_i(f)\exp(-\pi f t^*_{ij}) \qquad (1)$$

where S(f) is the source spectra, I(f) is the instrument response, C is the frequency-independent amplitude term associated with geometric spreading, seismic moment, radiation pattern, and other static effects. The exponential term describes the attenuation effect. The term t* can be expressed as t* = t/Q, where t is the travel time, and Q is the quality factor.

Assuming a Brune-type source [Brune,1970], the source velocity spectrum of event j, can be written as

$$S_j(f) = \frac{fM_{0j}}{1+(f/f_c)^2} \qquad (2)$$

where M_{0j} is the signal moment, and fc is the corner frequency.

Since only data in the flat portion of the pass band of the recording system is used in this study, I(f) can be neglected. The corner frequency is estimated based on the result given by Daniel Garcia (2004) for central Mexico, which takes the form of

$$f_c = 1.956\times10^7 M_0^{-0.297} \qquad (3)$$

where M_0 is the seismic moment in dyne-centimeters. This result is actually for shear waves, so (3) needs to be modified in this study by taking the P - S wave corner frequency ratio to be 1.5 [Boatwright, 1985], which means the corner frequency for P wave has the form of

$$f_c = 2.934\times10^7 M_0^{-0.297} \qquad (4)$$

Now we can rewrite (1) as

$$A_{ij}(f) = \frac{Cf}{1+(f/f_c)^2}\exp\left(-\pi f\frac{t_{ij}}{Q}\right) \qquad (5)$$

where C is a frequency independent term.

Taking the logarithm of equation (5), we obtain

$$\log A_{ij} = \log C + \log f - \log(1+(f/f_c)^2) - \pi f t_{ij}/Q \qquad (6)$$

We then solve this equation by a least-squares method to obtain the average frequency-independent Q for each ray path.

Data and Analysis

We used moderate earthquakes with magnitudes ranging from 4.5 to 5.0. The dataset is shown in Figure 1, and consists of 14 local earthquakes recorded by the instruments of MASE from February 2005 to May 2006.

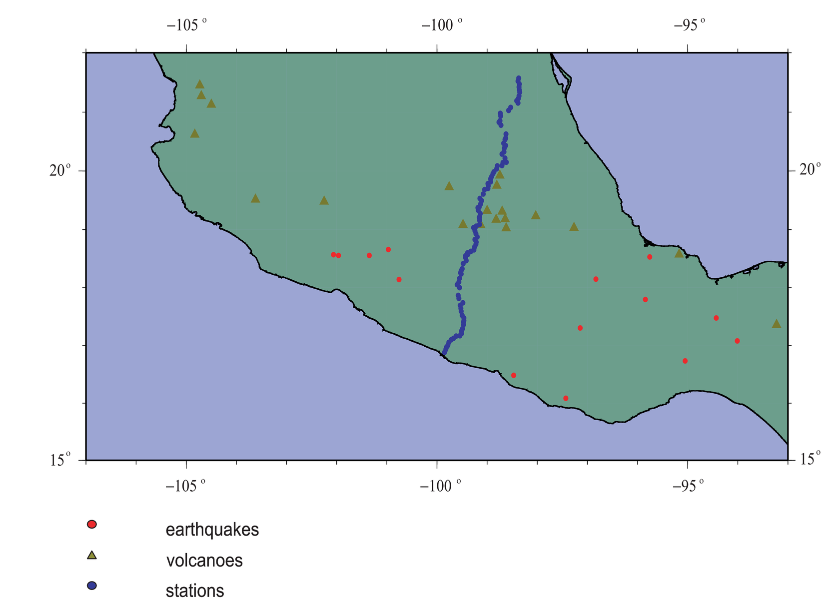

Figure 1

The smoothed spectra of the vertical velocity-component for P wave is calculated from a 3.5 s time window, beginning 0.5 s before the arrival pick, after 5% cosine taper is applied. A smoothed spectra of noise is also calculated from a 3.5 s time window immediately preceding each signal window in the same way. Tests show that changing the window length from 2 - 5 s does not produce significantly different spectra in the frequency band used, so a constant 3.5 s window length was used for all the events. The signal is kept for further analysis if the signal-to-noise ratio is greater than 2, in a frequency band 0.5 to 7-30 Hz.

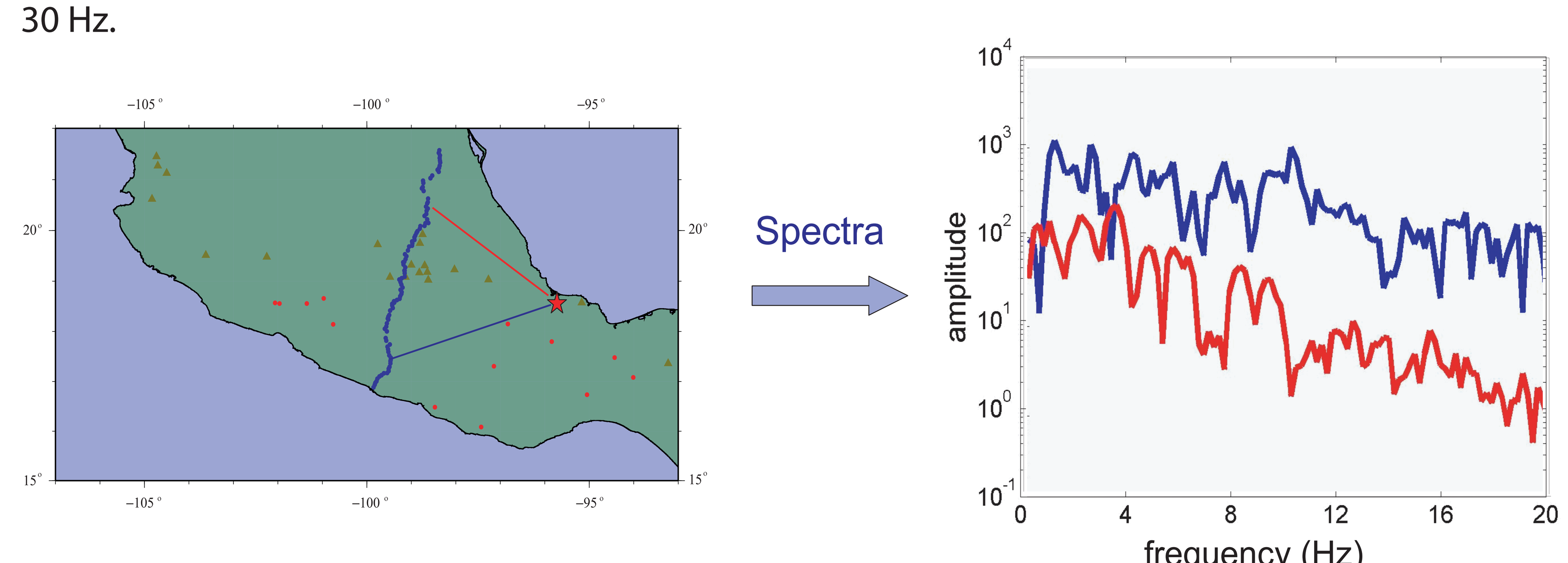

Figure 2

In Figure 2, two velocity spectra are calculated for two paths with the same distance from an event. The blue one is mainly through the crust, and the red one is mainly through the mantle wedge. From figure 2, we can see that the wave passing through the mantle wedge attenuates more than the wave passing through the crust.

A complete set of path-average Q determined for one event is provided in Figure 3 for example.

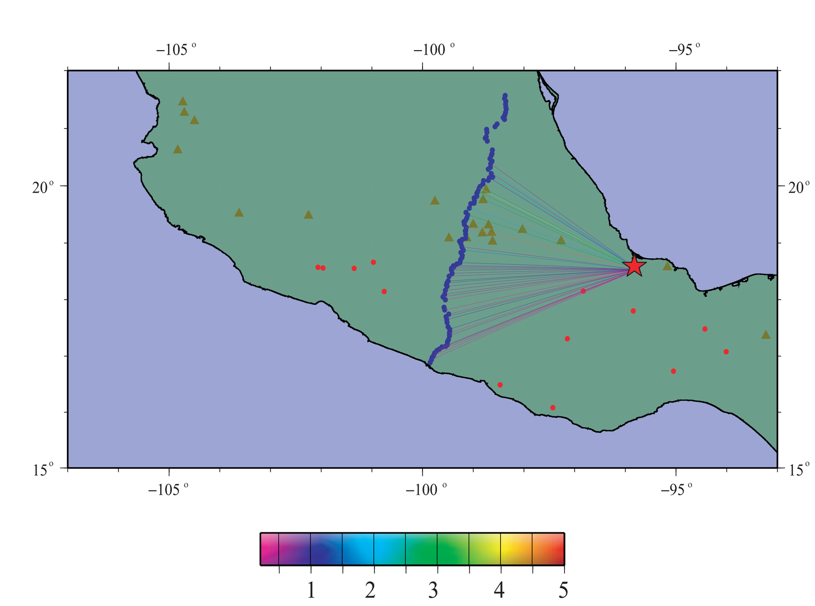

Figure 3

We can see that the average Q for the wave path through the mantle wedge is smaller than that for the path through the crust. This indicates that the mantle wedge is characterized by higher attenuation property than the crust.

Tomographic Inversion

One-dimensional tomographic inversion is used to determine the attenuation structure. Based on the data coverage, the study region is divided into eight blocks parallel to the trench, and each block is assumed to have constant Q. The observed t^* for the ith ray path is

$$t^*_i = \sum_j t_{ij}(Q^{-1})_j \qquad (7)$$

where t is the travel time in block j for the ith ray, and Q is the quality factor of block j. To compute the travel time, we currently assume that the ray path is a straight line connecting the source and station.

The inversion problem can be written in matrix form as

$$[t]_{NM}[Q^{-1}]_M = [t^*]_N \qquad (8)$$

where M is the number of blocks and N is the number of data.

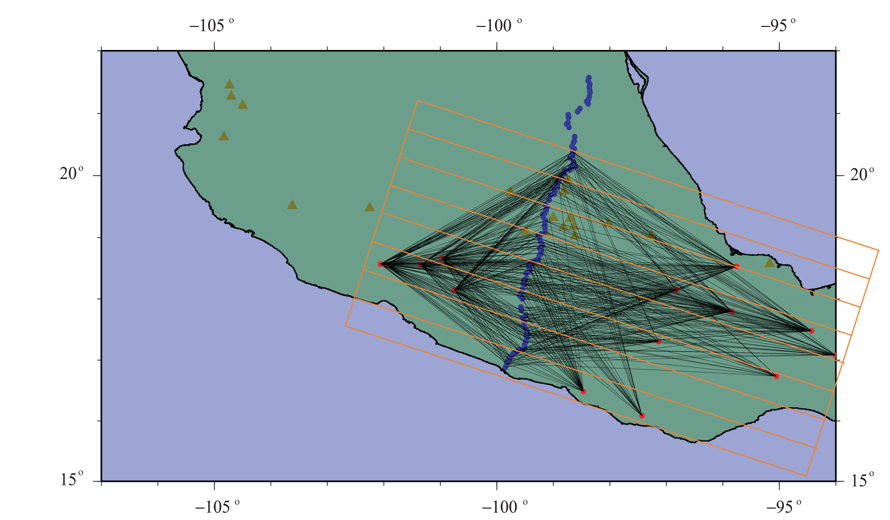

Figure 4

Inversions show high Q in the first four blocks, low Q in the next three blocks, and a slightly higher Q again in the last block (Figure 5). The low Q blocks correspond to the Trans-Mexico-Volcanic-Belt on the map. We interpret the low-Q region as the mantle wedge, as has been shown in attenuation studies of other subduction zones.

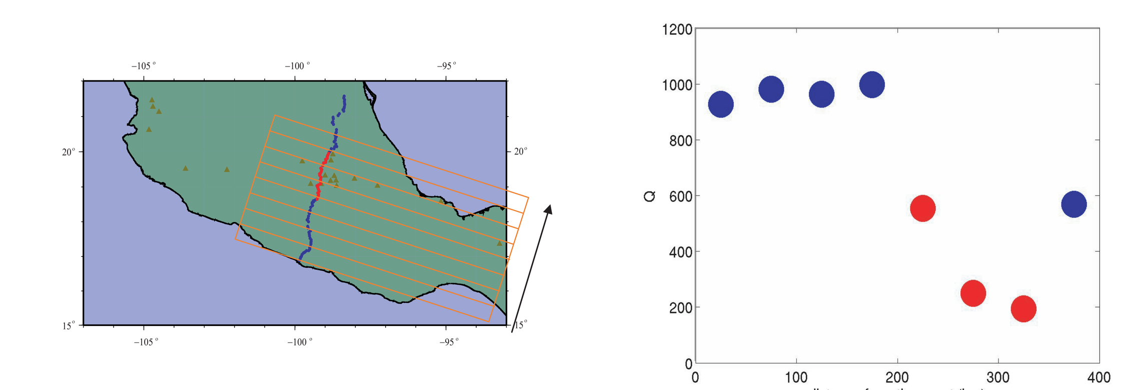

Figure 5

Our results indicate that the mantle wedge begins about 200 km, and ends about 350 km from the coast, and gradually transitions to normal upper mantle beyond that (Figure 6).

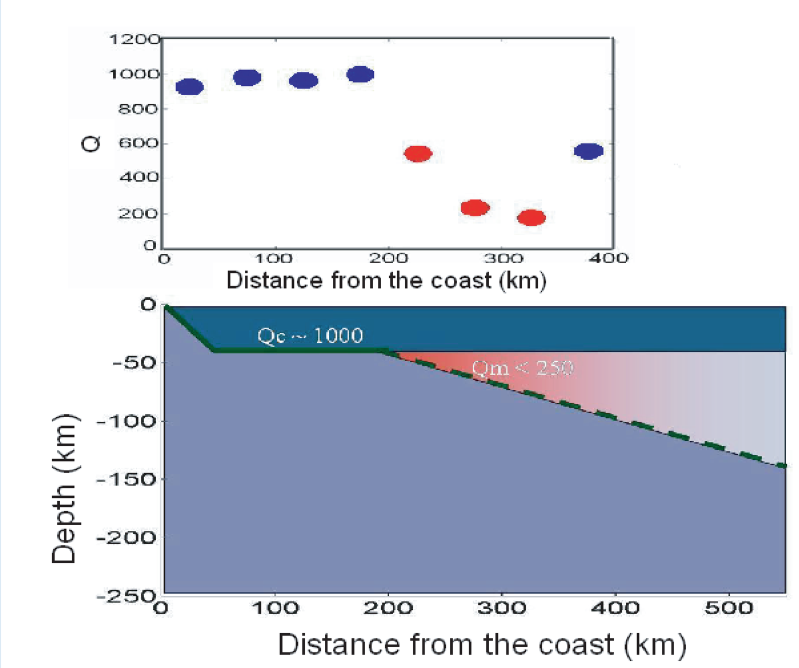

Figure 6

Summary

We have studied the attenuation structure in southern Mexico using the spectral decay method. The results show a low-Q zone beneath the Trans-Mexico-Volcanic-Belt, which we interpret to correspond to the mantle wedge. We suggest that the mantle wedge lies between about 200 km and 350 km from the coast, and has a low Q value less than about 250. The attenuation structure obtained in this study is similar to that of many subduction zones.

Our future work is to obtain a more detailed 3-D tomographic attenuation model, which takes Q variation in depth and varying subducted slab geometry into consideration. This work may involve earthquake relocation and ray tracing. Then, we can convert the attenuation model into viscosity model by using the approximation $\eta/\eta_0 = (Q/Q_0)^{(1/\alpha)}$ [Billen and Gurnis, 2001]. The viscosity value obtained will help us to conduct a more realistic geodynamic modeling in southern Mexico.