Measurement of Q and cumulative attenuation from VSP data

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Summary

The measurement of attenuation in seismic data is described and analyzed. The measurement problem is defined as the estimation of the attenuation parameter Q or the estimation of the related quantity CA (cumulative attenuation) or both. Two very different estimation techniques are described: the spectral-ratio method (SRM), which is well-known, and the dominant-frequency method, which is mostly new here. The strengths and weaknesses of both methods are discussed and the extension to CA is given. It is demonstrated that CA estimates are more stable than Q estimates when attenuation is weak. The application of these techniques is demonstrated on a zero-offset VSP with a vibroseis source. Using the shallowest receiver (2185ft) as a reference, attenuation estimates were obtained for all receivers at depths equal to or greater than 5000ft. Consistent estimates were obtained from both the SRM and the DFM but it is demonstrated that any residual upcoming waves in the downgoing wave cause considerable error. The possibility of extending these measurements to the earth's surface by assuming the reference wave there is the Klauder wavelet is examined. The results are plausible and seem appropriate to apply to surface recordings for bandwidth enhancement.

Introduction

Frequency-dependent attenuation is arguably the single most important factor limiting the resolution of seismic images. Theory suggests that there are two principal causes of attenuation: anelastic loss and stratigraphic filtering. Anelastic loss is usually thought of as a conversion of wave energy to heat as the seismic motion causes internal friction in the rock. Stratigraphic filtering, first described by O'Doherty and Anstey (1971), is an interference effect resulting from the chaotic superposition of very short-path multiples arising from fine layering with nearly random layer properties. Both theory and experiment agree that these two attenuation mechanisms are very similar in their effects and nearly impossible to distinguish (Margrave, 2014) such that any practical measure of attenuation will always estimate the combined effect.

Theory and Method

Commonly, attenuation measurement has been taken to be equivalent to measuring Q, called the "quality factor" of an anelastic material. There is general agreement (e.g. Aki and Richards, 2000; Kjartansson, 1979) that the amplitude spectrum of a propagating seismic wave decays according to the attenuation law

$$|\hat{w}_t(f)| = |\hat{w}_0(f)|e^{-\pi f t/Q} \tag{1}$$

where $|\hat{w}_0(f)|$ is the source amplitude spectrum and $|\hat{w}_t(f)|$ is the amplitude spectrum after propagating for time t. There are a variety of methods of measurement of Q (e.g. Cheng, 2013; Tonn, 1991) and all suffer some degree of instability due largely to the fact that Q appears in the denominator of the exponent of equation 1. Attenuation becomes small when t/Q is small which can happen when t is small, or when Q is large, or some combination of both. The ratio t/Q has been called cumulative attenuation (Hauge, 1981) and here denoted CA. Thus, whenever CA is small, there will be little difference between the initial and attenuated spectra.

In a zero-offset VSP setting, the propagating wave is observed by downhole receivers at a variety of depths. For any two observations, label the shallower by '1' and the deeper by '2' and then it results that the log-spectral-ratio, or lsr, is

$$lsr_{1,2}(f) = \ln\left[\frac{|\hat{w}_2(f)|}{|\hat{w}_1(f)|}\right] = -\frac{\pi f(t_2 - t_1)}{Q_{1,2}} = -\pi f CA \qquad (2)$$

where $(t_2 - t_1)/Q_{1,2} = CA$. The spectral-ratio method, or SRM, uses equation 2 to estimate Q by fitting a straight line to the lsr. The slope of the estimated line is inversely proportional to Q but directly proportional to CA.

An alternative estimation method that avoids the spectral division needed to form the lsr is the dominant-frequency method. This is a simple variant of the technique introduced by Quan and Harris (1997) who called it the frequency-shift method. In this method, the spectra at levels 1 and 2 are measured and then the forward Q filter that best converts the observation at level 1 into the observation at level 2 is determined. This is done by matching the dominant frequencies where dominant frequency is defined by $f_d = \frac{\sum_f |\hat{w}(f)|^2 f}{\sum_f |\hat{w}(f)|^2}$. The power spectrum is used here while Quan and Harris used the amplitude spectrum. Testing suggests that the Amplitude spectrum gives too much weight to the higher frequencies. The method proceeds by direct search over all integer values of Q between 5 and 300 and the estimate of Q comes from the minimum of the objective function $O_{f,k} = Q_k\left(f_{d2} - f_{d2,k}\right)^2$ where Q_k are the integer Q values, f_{d2} is the dominant frequency observed at level 2 and $f_{d2,k}$ is the predicted dominant frequency from the observation at level 1.

Data Analysis

This work was done on a VSP recorded in the southern United States. This was a zero-offset VSP using a vibroseis source with an 8-96Hz linear sweep. The receivers were three component geophones spaced at 50 ft and extending from 2185ft to 13803ft. Processing of the VSP was done by a third party and a 300ms ribbon of the separated, flattened, downgoing wave is shown in Figure 1a) while Figure 1b) shows the same data after a depth averaging process using and averaging window of ± 400ft. Each trace of 1b) is the average of all the traces in an 800ft window centered on the corresponding trace of panel 1a). The slight differences are due to residual upgoing waves that have survived the wavefield separation process. As will be seen, much better attenuation estimates arise from the data of 1b) than from 1a).

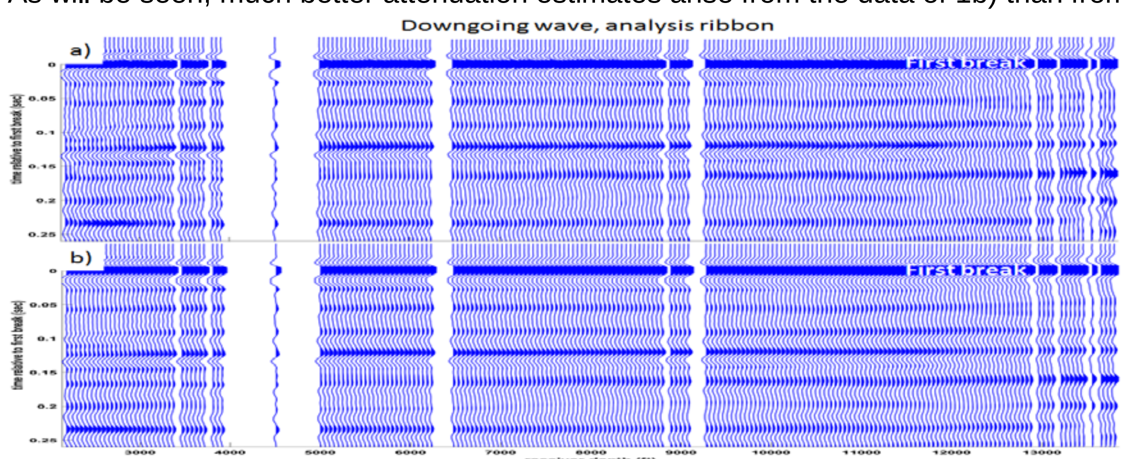

Figure 1: a) A ribbon of data from the VSP downgoing wave approximately 0.3 sec wide and beginning just before the first breaks. b) The result of spatial averaging of the ribbon in panel a) using an averaging half-width of 400ft. Each trace of panel b) is an average of the traces of panel a) over the depth range ± 400ft relative to the trace position.

Figure 2 shows the amplitude spectra of the traces in Figure 1. These are the data than will directly determine the attenuation estimates. Both panels of this figure show clear evidence of attenuation as the high frequencies are clearly decaying with depth. The shallowest receiver is at a depth of 2185ft and shows a bandwidth of roughly 10-80Hz while the deepest receiver, at a depth of 13803ft, shows a bandwidth of about 10-55Hz. Despite the high similarity between these panels, the data of Figure 2b) will be shown to give much better attenuation estimates than those of 2a).

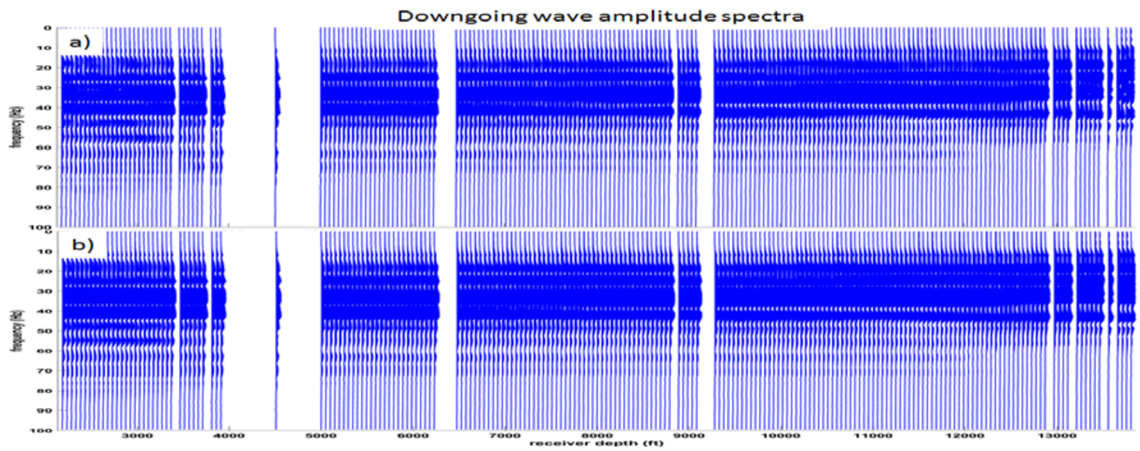

Figure 2: The amplitude spectra of the traces in Figure 1 are shown. The decay of high frequencies with increasing depth is evident in both panels.

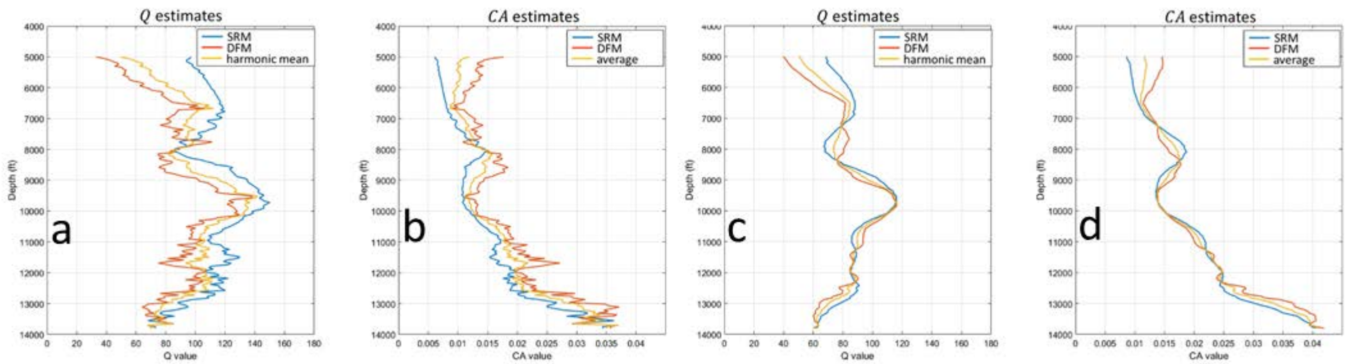

Figure 3: a and b are Q and CA estimates using the data of Figure 1a by both the SRM and the DFM. c and d are estimates using the data of Figure 1b.

Figure 3 a-b shows the results of Q and CA for the data of Figure 1a. The shallowest receiver (2185ft) was taken as the reference w_1 and the w_2 traces were selected starting at 5000ft and continuing to the deepest receiver. The Q estimates are therefore all average values representing the average effect from 2185ft to the depth noted on the vertical axis. The results from both SRM and DFM track one another quite well but show a lot of fluctuation and the SRM estimates are mostly higher than those from DFM. Shown between the two estimates is the harmonic average (e.g. $Q_{hm}^{-1} = Q_{SRM}^{-1} + Q_{DFM}^{-1}$) which may be a reasonable compromise[1]. The CA estimates of Figure 3 a-b were computed from the estimated Q values and the first break traveltimes, while the intermediate curve is the arithmetic average of the two estimates. The rapid fluctuations in these estimates are caused by residual upgoing waves that have survived the wavefield separation process. The data of Figure 1b result in the estimates in Figure 3 c-d and there is a dramatic reduction of the oscillations. This is clearly a judgement call since the true values are unknown but the preference is for consistency between the two methods and smoothness of the result

While the estimates of Figure 3 c-d seem self-consistent and reasonable, they are all referenced to the shallowest receiver at 2185ft while data processing needs estimates relative to z=0. Since the source was a vibrator it might be assumed that the wavelet at z=0 is the Klauder wavelet, or autocorrelation, of the sweep. This ignores the vibrator-earth interaction. Baseplate accelerometer recordings or ground-force estimates were not available in this case. Figure 4 shows the resulting attenuation analysis when the reference trace, w_1, is taken to be the Klauder wavelet at $z = 0$. The Q values are now lower and the range of estimation has been extended upwards from 5000ft to 2185ft. These values are now referenced to $z = 0$ and, in theory, could be suitable input for an inverse Q filter to be applied to surface seismic data. The

[1] The harmonic average was chosen simply because this is the wave that interval Q values "add" to give and average Q.

maximum Q has decreased from about 115 to about 55. These values may be too low because there are many potential mechanisms that could cause the actual radiated spectrum to lose strength at high frequencies.

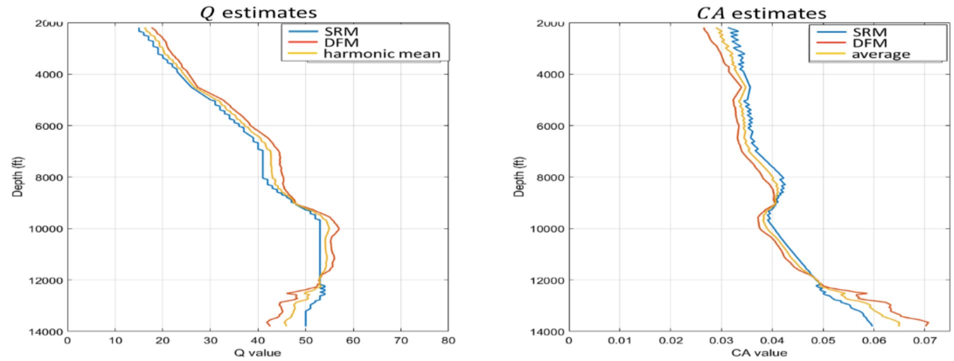

Figure 4: The result of an attenuation analysis where the reference signal, w_1, is taken to be the Klauder wavelet at depth $z = 0$. The input data were those of Figure 1b so this result should be compared to Figure 3 c-d. Note the change in the depth axis which now extends to 2185ft. The stair-stepping on the Q curve from the DFM happens because that method estimates Q to the nearest integer only. The resulting CA estimate shows saw teeth.

Conclusions

The theory and application of seismic attenuation analysis have been discussed as manifest in the estimation of both Q and CA (cumulative attenuation). Two computational methods were described: the spectral-ratio method (SRM) and the dominant-frequency method (DFM). The SRM is well established and frequency used while the DFM is introduced here. Once a Q estimate is obtained, a CA estimate is then found as the differential traveltime divided by Q. It was demonstrated, and explained theoretically, that, when attenuation is low, CA estimates are much more stable and accurate than Q estimates. These techniques were then demonstrated by application to a zero-offset VSP. Data processing was done to isolate the downgoing wave and 300ms data ribbon beginning at the first breaks was used. Using the shallowest receiver at 2185f as a reference, attenuation estimates were obtained for all receivers at 5000ft and deeper. It was demonstrated that consistent results were obtained from both estimation methods and, it was argued, this lends credibility to the results. It was also demonstrated that even very small residual upgoing waves in the downgoing field cause considerable instability in the estimates. In a final analysis, the possibility of extending the reference depth to the earth's surface by using the Klauder wavelet was examined. This extension resulted in lower Q values as expected but may be too low because the vibrator-ground interaction has been neglected. In future work, it would be beneficial to repeat this analysis on other wells and to apply the estimated attenuation values to surface seismic data.

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