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Estimation of seismic attenuation from zero-offset VSP acquired in hard rock environments

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SUMMARY

Understanding of seismic attenuation plays an important role in successful application of seismic imaging and subsurface characterisation techniques based on amplitude analysis. Zero-offset vertical seismic profiling (VSP) is one of the principal tools which can be used to study seismic attenuation.

Apparent attenuation estimated from seismic data analyses comprises of scattering and intrinsic components. Scattering mechanism can play significant role in hard rock environments in areas associated with fracture zones or other complex structures. As such seismic attenuation can be an important seismic exploration attribute.

Meanwhile attenuation analyses from VSP data are almost routinely done in oil and gas industry they are still uncommon in mineral exploration.

In this study we analyse zero-offset VSP data acquired in Western Australia on one of DET CRC test sites using both hydrophones and 3C geophones as receivers. We compare several methods for apparent attenuation estimation and evaluate their applicability to VSP data acquired in crystalline rocks. Extensive wire line log coverage for the well allows us to investigate relative contribution of different attenuation mechanisms.

Key words: Vertical seismic profiling, Attenuation, Mining geophysics.

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Vertical seismic profiling and cross-hole seismic surveys are the most reliable data sources for getting apparent attenuation estimates due to the relatively high signal to noise ratio, especially for the direct wave arrivals. A large set of different methods were developed for that. One of the most basic ones, the amplitude decay method was published by McDonal et.al (1958). Several methods are based on evaluation of the changes in the shape of the wavelet, caused by attenuation, in the time domain. These include pulse broadening (Ricker 1953), pulse rise (Gladwin and Stacey 1974), pulse power (Stainsby and Worthington 1985) and some other methods. The obvious alternative to these is a set of spectral domain methods, such as the spectral ratio (Hauge 1981), the spectral amplitude matching technique (Blias 2011) and the centroid frequency shift method (Quan and Harris 1997).

An excellent study of reliability of different attenuation estimation techniques was published by Tonn (1991). The main outcome of the study was that not a single method is superior and estimation of apparent attenuation requires very high quality input data. We recently compared several techniques applied to the VSP data acquired in siliciclastic sedimentary section (Pevzner et al. 2012) and established a workflow based on the modified centroid frequency method, which seems to be reasonably robust. However it is not quite applicable to hard rock environments due to the very high seismic velocities typical for crystalline rocks.

In this study we analyse zero-offset VSP data acquired in Western Australia in one of the DET CRC test sites using both

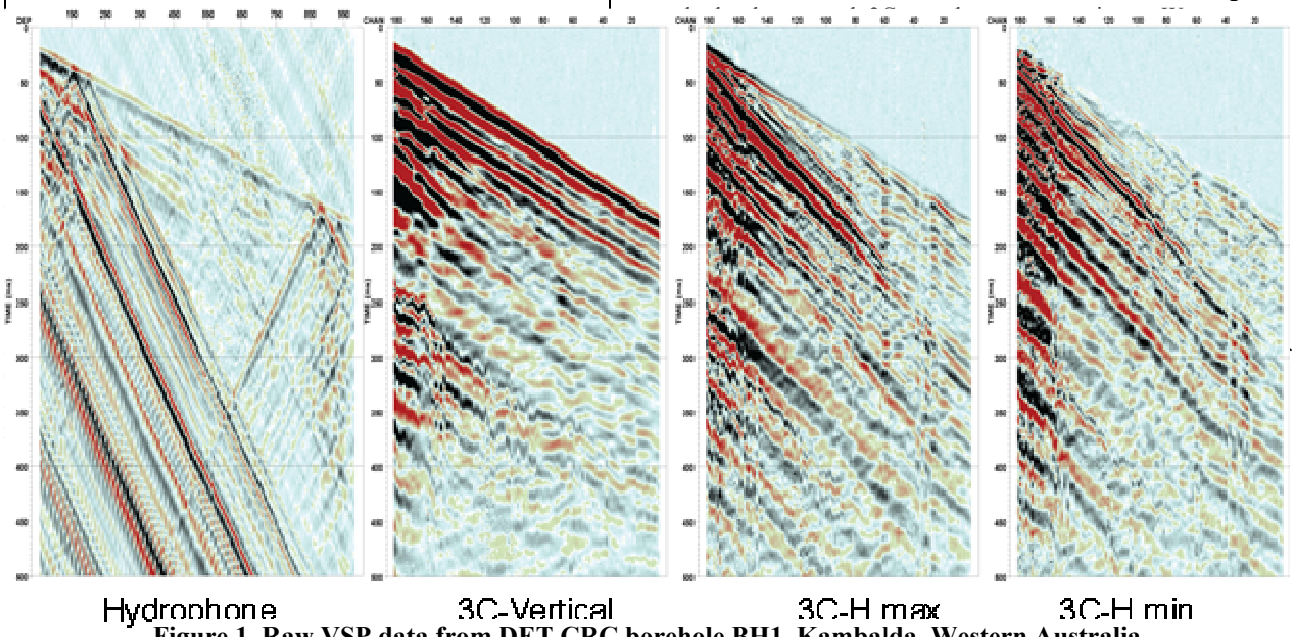

Figure 1. Raw VSP data from DET CRC borehole BH1, Kambalda, Western Australia.

Both data sets were acquired using equivalent acquisition geometry of 5 m station spacing from 100 to 1000 m depth. A near offset shot point of 28 m and an 800 kg weight drop hammer source was used. The 3C data were collected with two 5 m spaced AMC-VSP-3-48M shuttles supplied and operated by ASTO Geophysical Pty Ltd. A reference geophone was placed near the borehole collar to correct trigger timing variances. The hydrophone data were collected using a 24 channel string manufactured by V-Cable and a 24 channel DAQ-Link III seismograph. A reference geophone was not possible with the hydrophone configuration. Borehole BH1 was drilled to a depth of 1061 m, its path is deviated with an approximate 85 degree dip to the South.

Comparison of hydrophone and 3C geophone data from the well is presented in Figure 1. Signal-to-noise ratio on the 3C geophone data is substantially higher compared to the hydrophone data. For this reason we use only the vertical component of the 3C geophone data for further analysis.

ESTIMATION OF APPARENT ATTENUATION FROM VSP DATA ACQUIRED IN HARD ROCK ENVINRONMENTS

The spectrum $S_R(f)$ of the direct wave recorded by a receiver can be expressed using linear systems theory (Quan and Harris 1997) as:

$$S_R(f) = G(f)D \cdot H(f)S_0(f)\,, \qquad (1)$$

where $S_0(f)$ is the spectrum of the emitted wave, $G(f)$ is the factor responsible for source and receiver properties (i.e. instrument response, coupling, directivity pattern), D is the divergence of the wavefront and $H(f)$ is frequency dependent attenuation. Seismic attenuation along the ray path can be expressed using the following equation:

$$H(f) = \exp\left(-f\int_{ray} \pi\left[Q(l)v(l)\right]^{-1} v dl\right), \qquad (2)$$

where Q is so called quality factor and v is velocity. Here we assume that both quality factor and velocity are independent of frequency.

All methods of Q estimation from VSP data rely in our ability to capture changes in the overall shape of the wavelet (or its spectrum) or some sort of characteristic parameter, such as amplitude, due to the seismic attenuation.

Presence of steeply dipping interfaces in the area, resulting in the presence of reflected waves with similar travel time curves to the direct wave, large number of mode-converted waves observed in all three components and overall high velocity of the seismic waves (Vp is varying within 5000-7000 m/s along the well) substantially complicate wave field separation. This makes attempts to use any method which requires estimation of the whole downgoing wavelet shape/spectrum difficult.

For this reason here we focus on those methods which operate with the very part of the wavelet, namely pulse rise, amplitude decay and centroid frequency decay method. The last one will be applied using the first half period of the downgoing wavelet.

We always assume that the wave propagates along the well, source signature remains stable throughout the acquisition, coupling conditions for the receiver remains constant for all receiver positions in the well.

Pulse rise method

Gladwin and Stacey (1974) found a simple empirical dependence of pulse rise time τ (see Figure 2) on the time of propagation of the pulse t:

$$\tau = \tau_0 + C\int_0^t Q^{-1}dt\,, \qquad (3)$$

where τ_0 is the initial pulse rise time at $t = 0$, and C is a constant. Gladwin and Stacey (1974) obtained $C = 0.53 \pm 0.4$ from ultrasonic measurements in tunnels constructed in hard rocks.

The pulse rise time estimates obtained from Z-component of field VSP record is shown in Figure 3. We plot it as a function of transit time to match the equation 3.

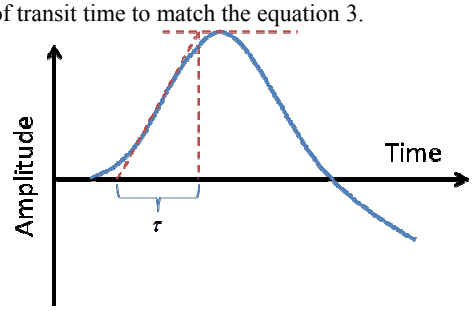

Figure 2. Rise time r of a pulse is here defined in terms of the tangent at the point of maximum slope, extrapolated to zero and to the peak pulse amplitude (modified from Gladwin and Stacey, 1974)

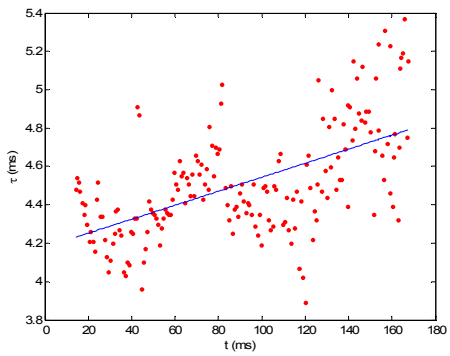

Figure 3. Pulse rise time as a function of the transit time.

Steady increase of the pulse rise time observed for the whole well corresponds to Q^{-1} = 0.007 (for C=0.53). It is important to note that we can observe significant (measurable) seismic attenuation in this well. However the C constant was derived empirically for ultrasound measurements; it will not necessarily remain the same for the data acquired with different set of sources, receivers and frequency range.

Centroid frequency shift

Hauge (1981) noticed that propagation of the seismic wave in the medium with constant Q will result in linear decrease of

the centroid frequency with the propagation time. Quan and Harris (1997) derived equations to compute actual Q values for certain assumptions regarding the initial shape of the amplitude spectra (e.g. Gaussian, boxcar or triangle). Recently we proposed to slightly modify the method (Pevzner et al., 2012) to accommodate changes in the shape of the amplitude spectrum with depth/time (eq. 4).

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$$2\frac{(f_{R1}-f_{R2})}{\sigma_{R1}^2+\sigma_{R2}^2} \approx \int_{ray} \frac{\pi}{Qv} dl, \qquad (4)$$

where $f_{R1,2}$ is the centroid frequency at a pair of receivers (R1 and R2), $\sigma_{R1,2}^2$ is the variance of the spectrum.

Due to the reasons explained above robust estimate of the centroid frequency is complicated. However we can use the time width of the first half-period(cycle) of the wavelet to get a rough estimate of the visible central frequency decay (Figure 4).

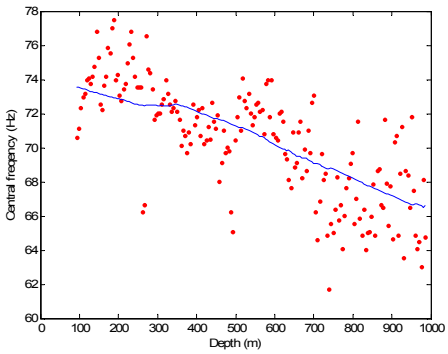

Figure 4. Central frequency decay as a function of depth

Problems with the seismic interference in the analysis window prevent us from accurate estimation of the variance of the spectra, so to get a very rough indicative Q estimate we assume that $\sigma_{R1}^2 \approx f_{R1}^2/4$, we also treat central visible frequency as a proxy for centroid frequency. These assumptions will give us Q^{-1} = 0.014 computed for the whole well.

Amplitude decay method

The ratio of these amplitude spectra observed on two receivers in the well can be expressed by

$$\ln\frac{S_{R1}(f)}{S_{R2}(f)} = \pi f \cdot \frac{\Delta l}{v} \cdot \frac{1}{Q} + M = \pi f \cdot \Delta t \cdot \frac{1}{Q} + M, \qquad (5)$$

where Δl is the distance between the receivers, Δt is the difference in the arrival times and M is a constant responsible for frequency-independent factors such as the divergence of the wavefront and transmission losses.

To get a Q estimate in between these two receivers using the amplitude decay method, one can simply re-arrange equation (5):

$$\frac{1}{Q} = \frac{v}{\pi f \Delta l}\left(\ln\frac{S_{R1}(f)}{S_{R2}(f)} - M\right) = \frac{1}{\pi f \Delta t}\left(\ln\frac{S_{R1}(f)}{S_{R2}(f)} - M\right). \qquad (6)$$

The traditional way of excluding M from equation (6) is to correct the data for wavefront divergence as well as to estimate (or simply neglect) transmission losses. Another problem with this method is that it requires amplitude to be measured at the same frequency for both receivers.

Figure 5 shows the logarithm of the amplitude of the direct wave (measured along the first extremum of the wavelet, corrected for spherical divergence of the wavefront).

Assuming that we can use central frequency to get estimates from this curve, we obtain average value of Q^{-1} = 0.014 (an average value for whole well).

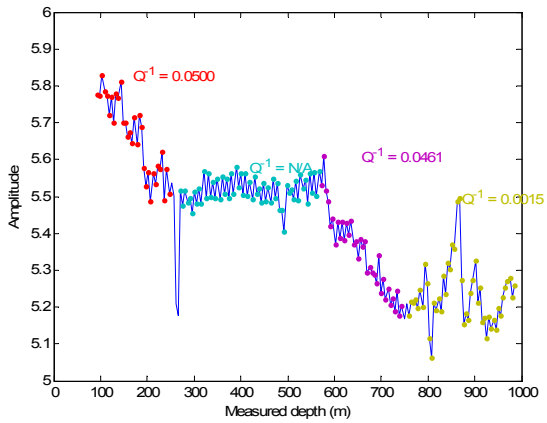

Figure 5. Amplitude of the direct wave (corrected for divergence of the wavefront)

The amplitude decay curve exhibits 4 distinct intervals with different behaviour (Figure 5). In fact, the same intervals can be identified on pulse rise time and central frequency decay curves (Figure 3 and 4). The top part of the curve (> 250 m) corresponds to the interval with rapid alternating mafic and ultramafic rocks, Q^{-1}=0.05. Depth range 250-800 m corresponds to a granitite body, at the top part (250-570 m) we do not see significant amplitude decay caused by attenuation, lower part (570 m – 800 m) exhibits significant attenuation (Q^{-1}=0.0046). We speculate that it is caused by a large shear and fractured zone there (Vp velocity is also decresing in the interval). The bottom part of the well (>800 m) passes through rapidly changing intervals of granites and mafic basalts, the amplitude of the direct wave is unstable there (likely caused by rapid change in the impedance and/or coupling conditions).

At the final stage of the analysis we estimate relative contribution of scattering into apparent attenuation. We make an attempt to estimate input of 1D scattering using reflectivity modelling using the velocity model obtained from the log data. We found that 1D scattering alone cannot be responsible for low Q values observed in 0-250 and 570-800 m depth intervals.

CONCLUSIONS

Here we analyse seismic attenuation in hard rock environments using vertical seismic profiling data acquired in

one of the wells in Kambalda, WA. We found that the amplitude decay method, the centroid frequency shift method and the pulse rise technique give us results which broadly agree with each other. Pulse rise method technique may require recalibration of the empiric constant used there due to the source/receiver parameters. We observe significant seismic attenuation in two intervals ($Q^{-1} \sim 0.05$) which corresponds to either rapid alterations of the material or possible presence of a fractured zone. Finally we found that 1D scattering cannot explain high attenuation in this interval, further 2D/3D modelling study is required.

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