A robust method for estimating attenuation in reflection seismograms

W. O. Raji[1,2] and R. Rietbrock[1]

ABSTRACT

A robust technique for determining attenuation (1/Q) in field seismogram is presented. The technique involves two key stages: (i) computation of centroid frequency for the individual signal using variable window length and fast Fourier transform, and (ii) estimation of the difference in the centroid frequency and travel time for the incident and transmitted signals. The robust features of the new method include the use of spectral shape factor and variable window length. The shape factor allows the user to represent the real seismic wavelet with several shapes without altering the mathematical formula. Variable window length helps to reduce noise interference during spectral computation. The new techniques is tested on synthetic data sets and compared with other methods. Further to this, the method is tested on a seismic data set from an oilfield. The traces were subdivided into four intervals: AB, BC, CD, and DE, using prominent reflectors and well logs signatures. The estimated mean attenuation, 1/Q_m over 500 traces are 0.0196, 0.0573, 0.0389, and 0.0220, for interval AB, BC, CD, and DE, respectively. To show an important application of accurate 1/Q estimate, the estimated 1/Q_m is used to perform image enhancement on the seismic traces. The compensation scheme is applied to the traces one after the other using a MATLAB programme specifically designed for the study.

The results show that accurate estimate of 1/Q can be used to improve seismic images of subsurface structures.

[1]Department of Geophysics, Stanford University, 397 Panama Mall, Stanford, CA 94305, California, U.S.
[2]Department of Geology, University of Ilorin, PMB 1515, Ilorin, Kwara State, Nigeria.

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INTRODUCTION

Attenuation analysis of subsurface geophysical data can provide crucial additional dimensions to the existing exploration techniques (Parra et al., 2006; Raji, 2012).

Lack of reliable methods for estimating seismic attenuation in field data is the major limitation to the further development of attenuation (Q^(-1)) technology for hydrocarbon exploration (SEG DPF, 2005). Methods for determining Q in seismic records include (SRM) spectral ratio method (Bath, 1974; Spencer et al., 1982), centroid frequency-shift method (Quan and Harris, 1997), and peak frequency-shift method (Zhang and Ulrych, 2002). While SRM provides robust Q estimates in VSP data, they are less successful for estimating Q in reflection seismic data because of the strong influence of reflections from closely spaced rock layers (Parra et al., 2006). Improvements to the classic spectral ratio method include the works of Dasgupta and Clerk (1998) and Rein et al. (2012a & b). Aside the sensitivity of the SRM to noise, the choice of a frequency bandwidth is still an issue.

To achieve reliable (Q) estimates in seismic data, we present a new mathematical model. The model is based on the concept of frequency down-shift. The inversion process consist of two key stages: computation of the centroid frequency for the individual signal using variable window length and fast Fourier transform; and estimation of the difference in the centroid frequency and travel time for a paired incident and transmitted signals. The inclusion of a shape factor in the mathematical model allows the users to represent the real seismic signal with any spectral shape that satisfies the basic characteristics of a seismic signal without altering the mathematical model. The spectral shapes considered in the new model include Gaussian, Ormsby (or boxcar), Ricker, and triangle wavelets.

THEORY AND METHOD

If a real seismic signal is represented by a Ricker wavelet, the source spectrum at initial time 0 in an Earth medium can be defined (Zang and Ulrych, 2002) as:

$$b_r(f,O) = \frac{2}{\sqrt{\pi}} \frac{f^2}{fm^2} \exp\left(\frac{-f^2}{fm^2}\right) \quad \text{.............(1)}$$

where f_m is the dominant or peak frequency and f is the sampling frequency. After travelling for time t in an absorbing medium, the frequency spectrum can be defined (Varela et al., 1993) as:

$$b(f,t) = b_o(f,0)h(f,t) \quad \text{.........(2)}$$

where h(f,t) is the earth absorption filter whose frequency response is:

$$H(f) = \exp^{(\ldots)}$$

The attenuation coefficient (α) is related to quality factor (Q) as:

$$\alpha(w) = \frac{1}{Q(w)} \frac{\pi f}{v(w)}$$

where f, is the frequency of the wave, and v is the velocity. Therefore, the spectrum of the seismic signal after a time t, can be given as:

$$b(f,t) = b_o(f,0)\exp\left(\frac{\pi f \Delta t}{Q}\right) \quad \text{.........(3)}$$

The observed difference in the frequency spectrum of the seismic pulse at time 0 and time t is shown in Figure 1. Taking the natural logarithm of (3) and rearranging the terms, we have:

$$\ln\{b(f,t)\} - \ln\{b_o(f,0)\} = \frac{\pi f \Delta t}{Q} \quad \text{.........(4)}$$

Equation 4 is the basis of the spectra ratio method (Bath, 1974; Spencer et al., 1982). If the amplitude spectral b(f,t) and b_o (f,0) are defined in terms of the spectral centroid frequency, cf_a and cf_b, and we removed the logarithm, the elements in equation 4 can be rearranged as:

$$Q = \frac{\pi f \Delta t}{cfa - cfb} \quad \text{.........(5)}$$

Equation 5 is the basis of the frequency-shift method for Q measurement, where Q is estimated from the difference in frequency and travel time ($cf_a - cf_b$ and Δt) between the incident and the transmitted signals (see, Quan and Harris, 1997; Zhang and Ulrych, 2002). The spectral parameter, cf, is known as the centroid frequency, and can be defined (Barnes, 1993) as:

$$cf_a = \frac{\int_0^\infty f I_a(f) df}{\int_0^\infty I_a(f) df} \quad \text{and} \quad cf_b = \frac{\int_0^\infty f I_b(f) df}{\int_0^\infty I_b(f) df} \quad \text{.........(6)}$$

For attenuation measurement in the exploration frequency band, it is common to use empirical equation to model attenuation (Gladwin and Stacy, 1974; Toverud and Ursin, 2005). Based on the physical observations shown in figure 1; the definition given in equation 5 and several synthetic experiments, a statistics-based empirical model for Q measurement is proposed as:

$$Q = \frac{t2cf_a cf_b}{\mu\sigma_{ab}(cf_a - cf_b)} \quad \text{.........(7)}$$

Where μ is the shape factor, a constant used to obtain the best fit to the data at the various spectral shapes. As pointed out by equations (12), (13) and (14) of Quan and Harris (1997), the constant μ depends on the actual spectral content of the wavelet, and μ varies with the spectral shapes. The average standard deviation of the incident and transmitted spectral is σ_{ab} and t is the travel time different. The average standard deviation,

$$\sigma_{ab} \text{ is: } \sigma_{ab} = \frac{\sigma_a + \sigma_b}{2};$$

$$\sigma_a^2 = \frac{(f - cf_a)^2 d(f) df}{\int_a(f) df} \quad \text{and} \quad \sigma_b^2 = \frac{(f - cf_b)^2 b(f) df}{\int_b(f) df} \quad \text{.........(8)}$$

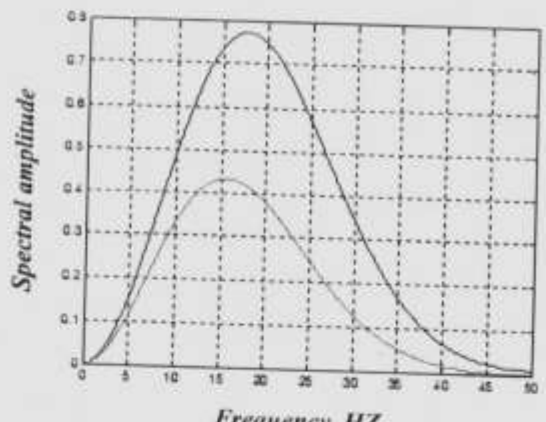

Fig. 1: The frequency spectrum of a Ricker wavelet at 1s (black) and 2.2s (blue) in an attenuating medium with Q value equals 30

For a source-receiver offset x, the travel time, t, can be expressed (Sheriff and Geldart, 1985; Burger et al. 1992) as:

$$t = t_o \left(\frac{1 - x^2}{v^2 rms\, t_o^2}\right)^{1/2} \quad \text{.........(9)}$$

where t_o is zero offset travel time and vrms is the root mean squared velocity. The time, t required for equation 7 can be approximated as:

$$t = t_o \; 1 + \frac{x^2}{2v^2 rms t_0^2} - \frac{x^4}{8v^4 rms t_0^4} \quad \text{.......................}(10)$$

If the spectral of a real seismic pulse is represented by a Ricker wavelet, the shape factor, μ is equal to $2\pi/3$. The spectrum of a real seismic signal is not exactly the same as that of a Ricker or a Gaussian wavelet (Ricker, 1953; Ryan, 1994). However, the shape can be approximated by any shape that satisfies the basic characteristics of the real seismic signal (i.e. that rises from zero to a peak and reduces to zero symmetrically). Approximating the spectrum of a real seismic signal to that of an Ormsby wavelet, a Gaussian wavelet, or a triangle wavelet, and following the procedures and definitions given in equations (7-10), the shape factor in equation 7 is modified to accommodate other spectral shapes without altering the mathematical model. The shape factor, μ is $\frac{6\pi}{7}$, $\frac{2\pi}{3}$ *and* $\frac{\pi}{2}$ for Ormsby, Gaussian, and triangular spectra, respectively. The use of shape factor in the mathematical model allows several wavelets to be used to represent the real seismic signal.

Application to Synthetic Seismograms

Seismic traces consisting of primary reflections, multiples, and noise are generated in a three-layer attenuating Earth model. The mechanism of attenuation in this model is fluid induced. The Earth model comprises shale–sandstone–shale layers. The sandstone layer represents reservoir interval, and therefore is assigned higher attenuation (due to the reservoir fluid), compared to the underlying and overlying shale layers. Q models in the sandstone layer are represented by 10, 20, 30, 40, 50, and 100. Q is measured in the seismic traces at the interval that corresponds to the sandstone layer using the proposed method with the various spectral shapes. The proposed method is here known as SFVQM. The Q estimates for the various input Q models are shown in figure 2. Further to this, SFVQM is compared with two published methods of attenuation measurement:

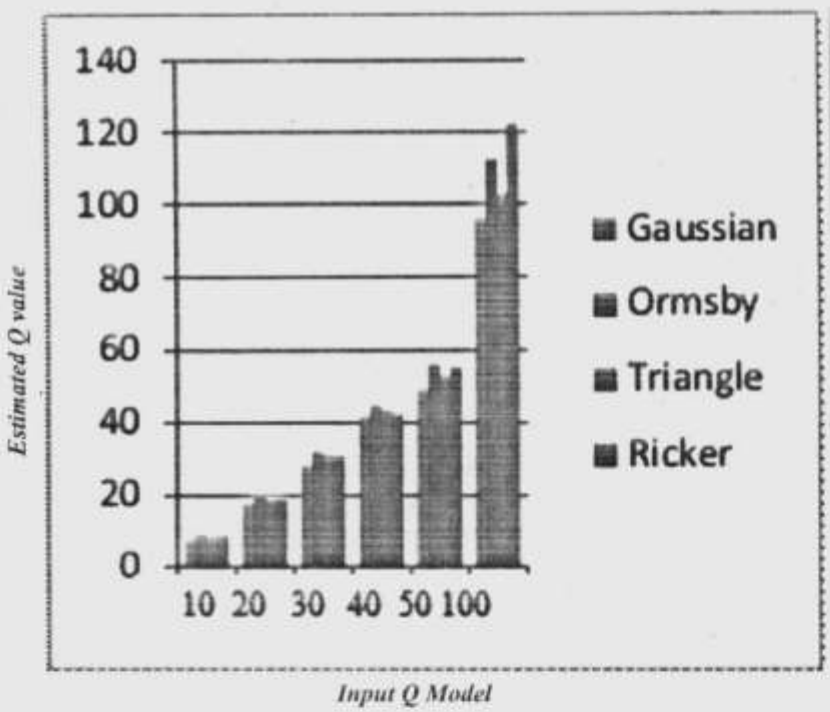

Fig. 2: Input Q model versus Q estimates. For Ormsby, Gaussian and Triangular spectral shapes, the frequency bandwidth is 10-40Hz. For the Ricker spectrum, the dominant frequency is 20Hz. See legend for the colour codes

i) The SFVQM (with Ricker shape spectrum) is compared with the spectral ratio method (SRM for short) for measuring Q in noisy seismograms. For each Q model, the noise level varies from 5%, 10%, 15%, to 20%. The mean Q estimates are plotted against the Q model in Fig 3.

ii) The Q model of Quan and Harris (1997) designed for a gaussian-shaped spectrum is compared with the SFVQM (using a gaussian shape spectrum). For each Q model, the frequency band of the source wavelet is varied as 10-40Hz, 10-50Hz, 20-50Hz and 20-60Hz. The mean Q estimates is again plotted with error bar in Fig 4.

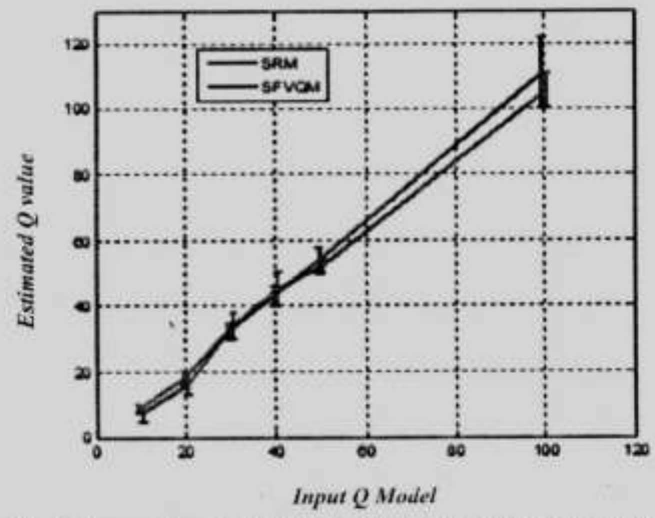

Fig. 3: Comparing the SFVQM with the Spectra Ratio Method - testing the effects of noise

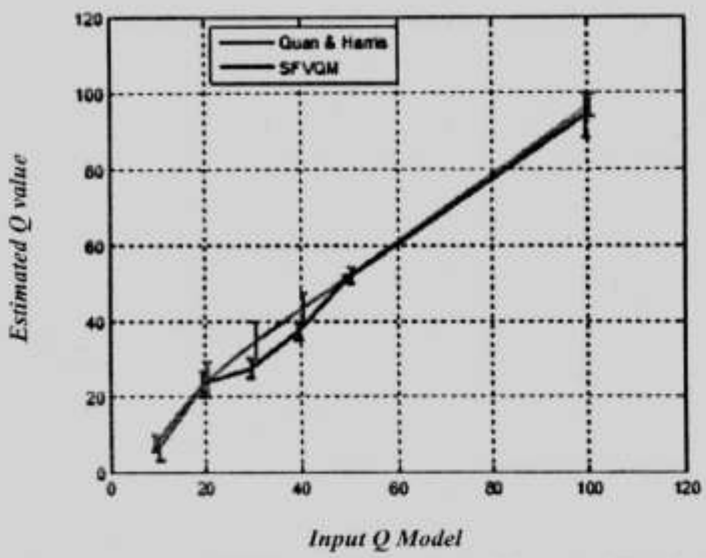

Fig. 4: Comparing the SFVQM with the method of Quan and Harris (1997) - testing the effects of the change in frequency bandwidth on the two methods

GEOLOGY OF STUDY AREA AND APPLICATION TO SEISMIC DATA

Gullfaks field is located in the Norwegian sector of the North Sea, Norway. It was formed during the Upper Jurassic to Lower Cretaceous as a sloping high, with a westerly structural dip, gradually decreasing towards the east (Peterson et al., 1990). The reservoir sands comprised: Middle Jurassic delta-deposited Brent Group; the Lower Jurassic shallow marine sandstones of the Cook Formation; and the Lower Jurassic fluvial channel and delta-plain deposits of the Statfjord Formation. Structurally, the field can be divided into three regions: Domino area with rotated fault blocks in the west, and a Horst area in the east; in between is a zone, characterized by folding structures. The north-south faults that divide the field, have a throw up to 300 meters. In the western part, the faults slopes downward to the east (Fossen and Hesthammer, 1998). The distribution of oilfields and structural cross-section of the field is shown in Fig. 5.

The data set used for the study is a marine 3D seismic acquired in 1985 by an unknown company for Statoil. Details of acquisition is not available to the author. The data was pre-processed by Statoil.

Following the synthetic data test, a set of 3D data from the Gullfaks Field, North Sea, Norway was used to test the suitability of the SFVQM method for Q measurement in field data. The seismic records represent a travel time interval of 905 ms, starting from 1397ms to 2302ms. A total of 500 traces were selected

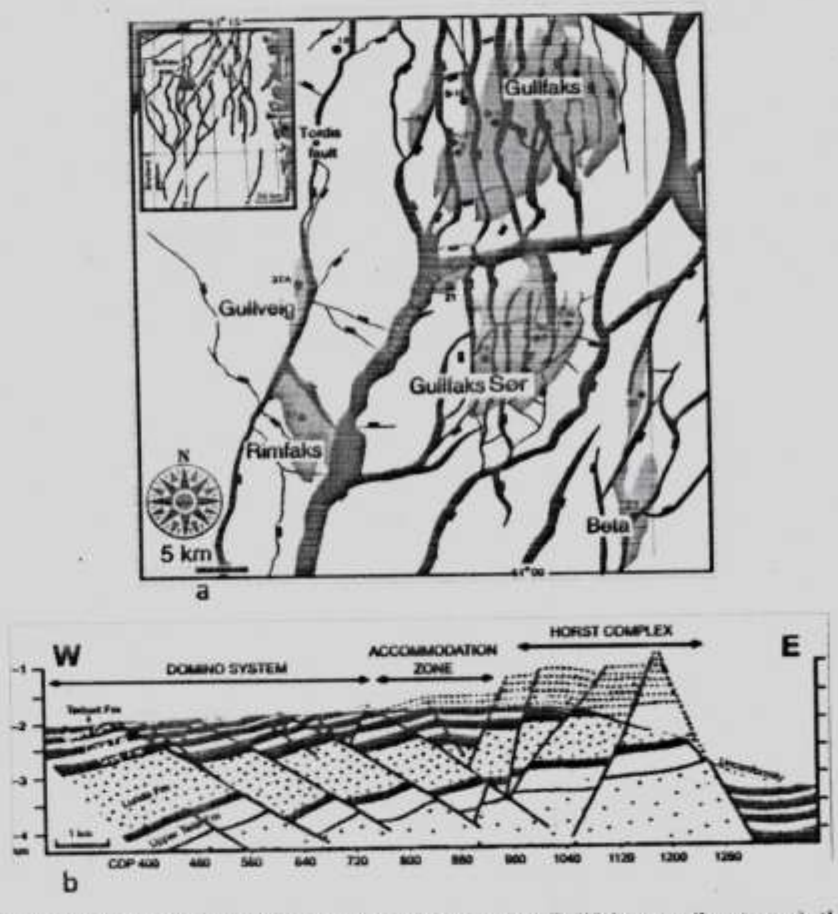

Figure 5: (a) Regional map of the northern North Sea showing Gullfaks area (inset map), the Gullfaks-Statfjord area showing oilfields (b), and a structural profile across the Gullfaks Field (c). Oil fields are shown in green in figure b. (Hesthammer et al., 2002)

for the test. For the purpose of Q estimation, the total travel time represented by the traces was subdivided into four intervals: AB, BC, CD, and DE, as shown in Figure 6. For Q measurement in the Gullfaks data, the SFVQM with a Ricker spectrum was used to minimise interference from noise, variable window length was used. The minimum and maximum window lengths used are 76ms and 98ms, respectively. Whenever possible, the same window length was used to comp the two signals being compared in the measurem Fourier transform was computed within the win containing each desired signal and the spec parameters required for Q measurement was comp as defined in equations (7) - (10). The distributions estimates at the four intervals are shown in histograms in figure 7.

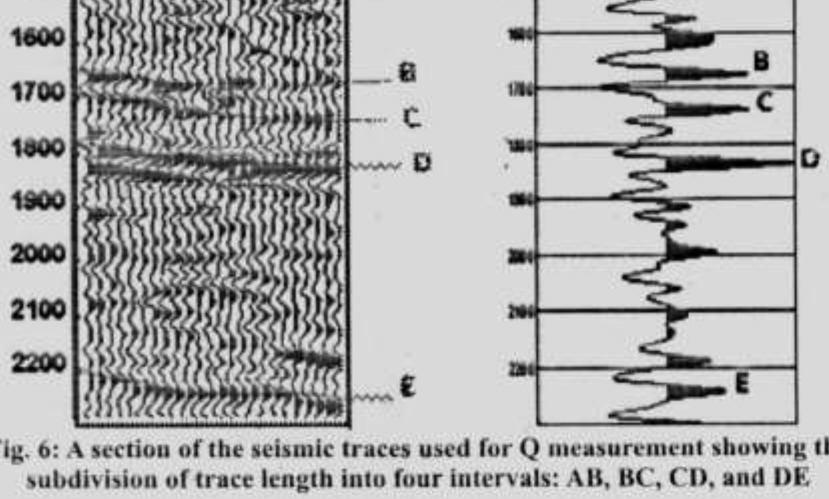

Fig. 6: A section of the seismic traces used for Q measurement showing the subdivision of trace length into four intervals: AB, BC, CD, and DE

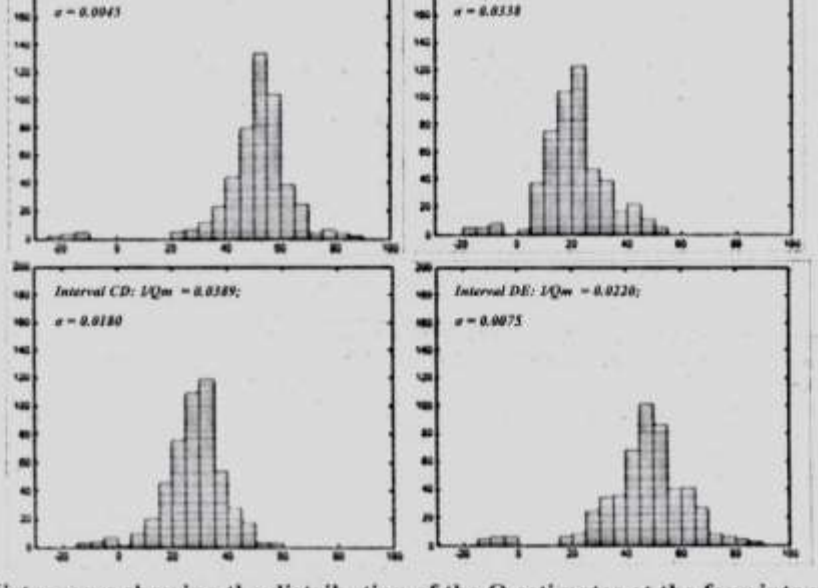

Fig. 7: Histograms showing the distribution of the Q estimates at the four intervals. The negative values of Q are not considered in the computation of the mean attenuation (1/Qm)

Application to Seismic Image Improvement

The amplitude and waveform of seismic waves are strongly impacted by wave attenuation. This usually results in loss, in parts of the desired information in the recorded seismic records. The loss is often expressed in the form of amplitude diminution, wavelet distortion and delayed sign signal arrivals, which in turn degrade the quality of seismic data. Attenuation causes poor seismic interpretability and contributes to seismic-well data mis-tie .In order to enhance the interpretability of seismic records, seismic data processing usually involves a stage where the effects of attenuation on seismic records are corrected. In this section, we show the application of attenuation estimate (1/Q) to seismic image processing, using inverse Q filtering.

Detailed processes of seismic image improvement using Q can be found in Wang, (2006) and Raji and Rietbrock, (2012).

A basic equation describing inverse Q filtering is:

$$(z+\Delta z,w) = A_e\ (z,w)\ \exp\ (jw/\beta(w)\Delta z)\ \exp\ (w/2\beta(w)Q\,\Delta z) \ldots\ldots(10)$$

The two exponential operators in equation 10 correct the effects of velocity dispersion and amplitude diminution, respectively. Figures 8 and 9 show the improvement in seismic image by the use of attenuation factor (Q) in seismic image processing.

Attenuation is estimated in the stacked trace (Figure 6, left) using the method described in the previous section. The measured attenuation is used to build a seismic image enhancement scheme, as summarised in equation 10. The scheme is then applied to the dataset to show that attenuation estimates can be used for seismic image improvement. Figure 8 shows the original seismic trace (black) and the improved seismic trace (blue). Figure 9 shows Time amplitude data plotted in wiggle format: the top plot is the original field data, while the bottom plot is the improved image. Figure 9 is the difference-plot between the improved and the original data

DISCUSSIONS AND CONCLUSIONS

A new method for measuring attenuation in reflection seismic records is discussed. The appropriateness of the method for measuring attenuation is tested using

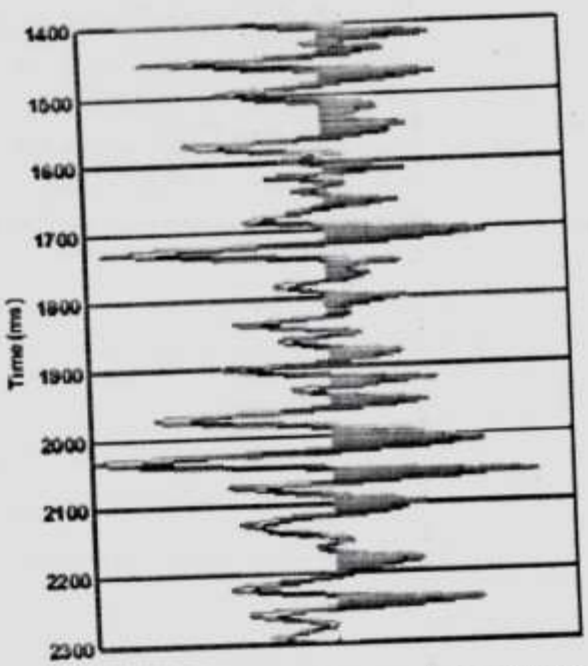

Fig. 8: Attenuated and compensated seismic traces-showing the effects of seismic Q compensation. Black is original (attenuated) traces, blue is after Q compensation

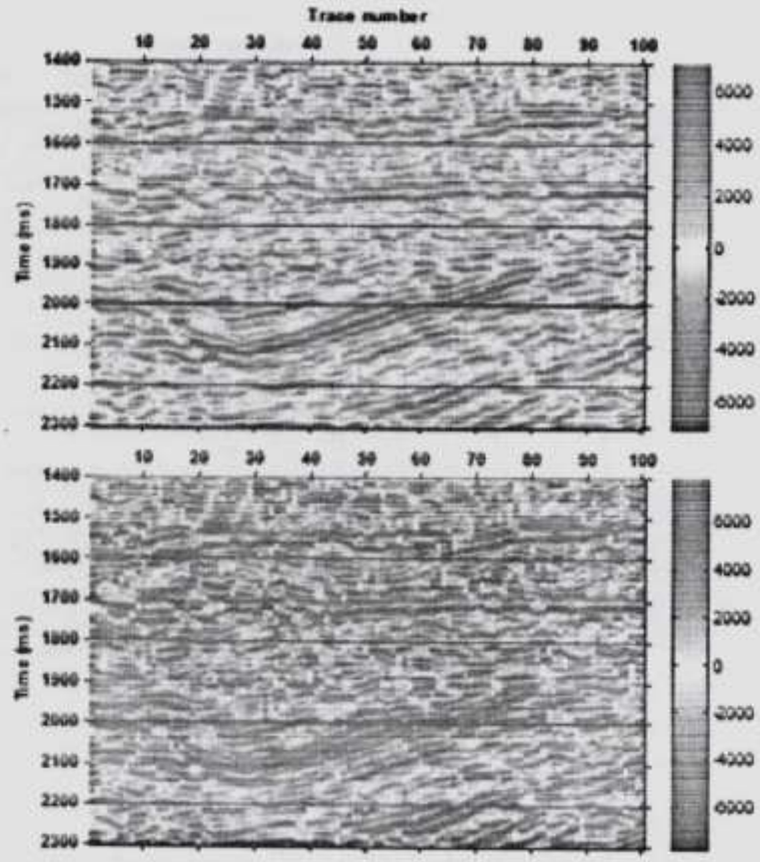

Fig. 9: Time-amplitude data plotted in pixel format. (Top) the original traces before Q-compensation, (bottom) traces after Q-compensation.

synthetic and field data. Results show that the new method (SFVQM) can give robust estimates of attenuation in synthetic and real data using any of the four wavelet shapes. The new method is less sensitive to interference from noise and the change of frequency bandwidth.

Compared to the spectral ratio method (SRM), SFVQM is less sensitive to noise interference. Error in Q estimates due to the effects of noise interference is 9.5% for the SFVQM and 13.4% for the SRM. When tested for the effects of the change of frequency bandwidth, SFVQM shows less sensitivity compared to the method proposed by Quan and Harris (1997). The errors are 9.8% and 14.7%, for SFVQM and the method of Quan and Harris (1997), respectively.

The mean attenuation ($1/Qm$) computed for interval A BC, CD, and DE are 0.0196, 0.0573, 0.0389, 0.02 respectively; while the standard deviation (σ) a 0.0045, 0.0338, 0.0180, 0.0075, respectively. Interv BC which recorded the highest attenuation val approximately coincides with the shale break th separates the top and base sandstones. The fact responsible for high attenuation in interval E compared to the overlying and underlying sandsto cannot be ascertained from this study. Further to this, v show that attenuation estimates can be used to impro the quality of seismic image. The improved ima showed amplitude restoration and higher seismic ima resolution.

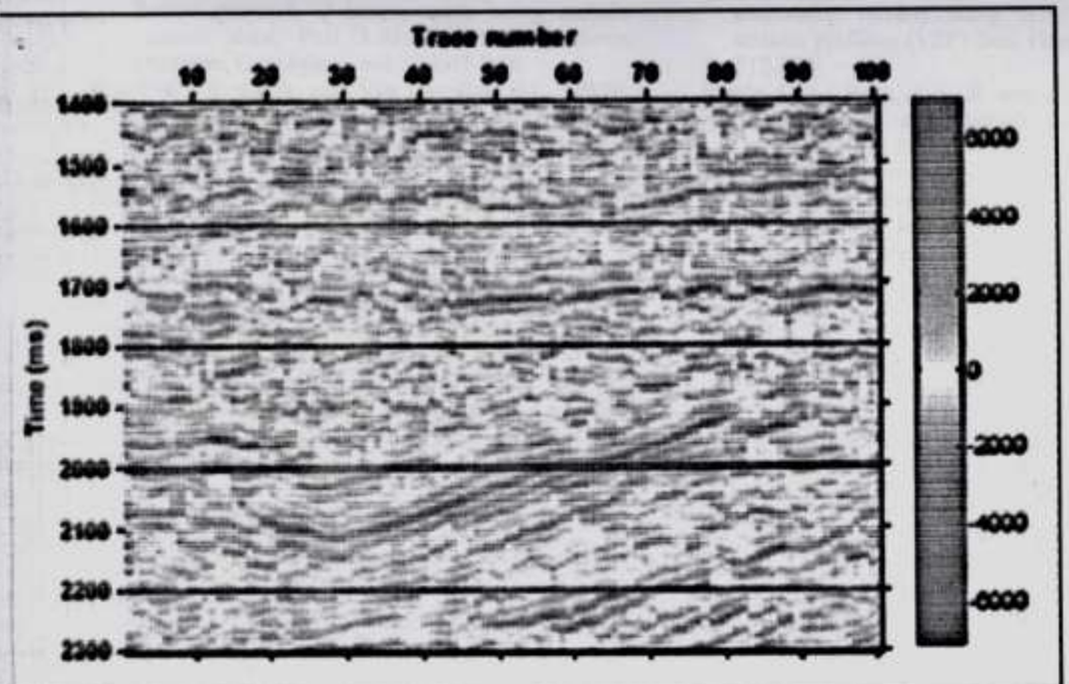

Fig. 10: The difference plot between the compensated and original data- showing the improvement in the seismic image achieved by the Q compensation process

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