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Attenuation ($1/Q$) estimation in reflection seismic records

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Abstract

Despite its numerous potential applications, the lack of a reliable method for determining attenuation ($1/Q$) in seismic data is an issue when utilizing attenuation for hydrocarbon exploration. In this paper, a new method for measuring attenuation in reflection seismic data is presented. The inversion process involves two key stages: computation of the centroid frequency for the individual signal using a variable window length and fast Fourier transform; and estimation of the difference in the centroid frequency and travel time for paired incident and transmitted signals. The new method introduces a shape factor and a constant which allows several spectral shapes to be used to represent a real seismic signal without altering the mathematical model. Application of the new method to synthetic data shows that it can provide reliable estimates of Q using any of the spectral shapes commonly assumed for real seismic signals. Tested against two published methods of Q measurement, the new method shows less sensitivity to interference from noise and change of frequency bandwidth. The method is also applied to a 3D data set from the Gullfaks field, North Sea, Norway. The trace length is divided into four intervals: AB, BC, CD, and DE. Results show that interval AB has the lowest $1/Q$ value, and that interval BC has the highest $1/Q$ value. The values of $1/Q$ measured in the CDP stack using the new method are consistent with those measured using the classical spectral ratio method.

Keywords: quality factor, attenuation measurement, seismic data, centroid frequency

(Some figures may appear in colour only in the online journal)

Introduction

Detailed knowledge of the attenuation structure of the Earth is essential for hydrocarbon exploration. Studies on the anelastic properties of rocks have linked attenuation ($1/Q$) to the presence of hydrocarbon and the saturation condition of rocks. Laboratory experiments and mathematical models (Gassmann 1951, White 1975, Toksoz *et al* 1979, Winkler and Nur 1979, Johnston and Toksoz 1980, Klimentos and McCann 1990, Klimentos 1995, Mavko *et al* 1998, Pride *et al* 2004) have shown that attenuation can indicate pressure, porosity, fluid type, and saturation level. Field studies also have shown that attenuation can be used for lithology discrimination (Stainsby and Worthington 1985, Dasgupta and Clark 1998, Singleton 2008) and anisotropy study of fractures (Maultzsch *et al*

2007, Zhu *et al* 2007, Behura and Tsvankin 2009, Clark *et al* 2009, Al-Harrasi *et al* 2011). In fact, the study by Winkler and Nur (1979) indicates that attenuation ($1/Q$) is more sensitive to changes in saturation than in velocity. Using Massillon Sandstone, they show that the effect of saturation on attenuation is greater on the order of magnitude than the effect of saturation on velocity. In addition to its use in seismic data processing (Wang 2006, Raji and Rietbrock 2012), knowledge of attenuation is useful in seismic acquisition design. It helps in the survey planning stage to know the portion of source energy that will reach the target layers.

Despite its numerous potential applications, the attenuation analysis of seismic reflection amplitudes is yet to become routine in the hydrocarbon industry. The lack of a reliable method for estimating seismic attenuation in field

data was cited at the 2005 SEG (Society of Exploration Geophysicists) Development and Production Forum as being the major limitation to the further development of $1/Q$ technology for hydrocarbon exploration. Another limitation in the use of $1/Q$ for hydrocarbon exploration is the question of how to determine the proportion of $1/Q$ that is solely due to the rock anelastic properties (e.g. saturation). This is because the $1/Q$ measured in seismic data is a combination of $1/Q_{\text{int}}$ (the intrinsic or inelastic attenuation) and $1/Q_{\text{ext}}$ (the extrinsic or elastic attenuation). $1/Q_{\text{int}}$ is the energy loss in waves from the conversion of a part of the wave's energy into heat. In $1/Q_{\text{ext}}$, there is no energy loss—energy is only redistributed or scattered into an unobserved wavefield. $1/Q_{\text{int}}$ depends on the presence of fluid(s) and the degree of saturation in the rock. On the other hand, $1/Q_{\text{ext}}$ is controlled by energy partitioning at lithology boundaries and thin-layering effects. Depending on the scale of heterogeneity, the mechanism of $1/Q_{\text{int}}$ can be classified into microscopic and mesoscopic (Biot 1956a, 1956b). The mesoscopic mechanism is believed to play a dominant role in the attenuation process taking place in reservoir rocks (Müller *et al* 2010). The attenuation ($1/Q$) measured in this paper is the effective or composite attenuation. Composite attenuation can be defined in terms of intrinsic and extrinsic attenuation (Dainty 1981, Richards and Menke 1983) as:

$$1/Q = 1/Q_{\text{int}} + 1/Q_{\text{ext}}. \quad (1)$$

Attenuation is commonly measured in terms of the rock quality factor (Q). The inverse of the rock quality factor ($1/Q$) is proportional to the attenuation coefficient (α),

$$\alpha(w) = \frac{1}{Q(w)} \frac{\pi f}{v(w)}, \quad (2)$$

where f is the frequency of the wave and v is the velocity. During propagation, the amplitude of a seismic wave decays, the peak frequency shifts towards the lower frequency band, and the wavelet width broadens. Methods proposed for Q measurement are based on the attenuation-induced changes in the spectral properties of seismic wavelets. Based on attenuation function and compensation function, Wang (2004) analysed Q in seismic traces using the Gabor transform spectrum to observe the time varying frequency content of the seismic trace. The Q analysis method avoids the issue of wavelet interference. Other methods for estimating Q in seismic records include the spectral ratio method (Bath 1974, Spencer *et al* 1982), the centroid frequency-shift method (Quan and Harris 1997), and the peak frequency-shift method (Zhang and Ulrych 2002). The spectral ratio and the centroid frequency-shift methods are designed for Q measurements in VSP data. While the methods provide 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). Further to this, the Q estimates obtained from VSP data does not give areal coverage for the attenuation properties in rocks. Improvements to the classic spectral ratio method include the works of Dasgupta and Clark (1998) and Reine *et al* (2012a, 2012b). Dasgupta and Clark (1998) improved the accuracy of Q estimates from the spectral ratio method by estimating Q

in the pre-stack domain—avoiding spectral distortion due to stacking and the normal move out stretch. Reine *et al* (2012a, 2012b) improved Q estimates from the spectral ratio method by accounting for the effects of source–receiver directivity, overburden anisotropy, and ray path differences. However, the choice of frequency bandwidth for the spectral ratio is still an issue in the spectral ratio method. Although the peak frequency-shift method (Zhang and Ulrych 2002) is designed for reflection seismic data, the technique largely depends on the peak frequency. The peak frequency of a seismic signal—the frequency that corresponds to the amplitude maxima—is easily influenced by a number of factors including the presence of noise, inter-bed multiples, and the wavelet extraction (windowing) process. These factors in turn affect the accuracy of Q estimates using the peak frequency-shift method.

In all the frequency-shift based attenuation methods, the shape of the seismic pulse is commonly model. This is to allow for the quantitative assessment of the attenuation-induced variation in the frequency content of the seismic spectrum. The commonly assumed shapes include Gaussian, boxcar, triangle, and Ricker. Different mathematical models are defined for the various methods due to the differences in the spectral shapes (see, Quan and Harris 1997, Zhang and Ulrych 2002). The new method presented in this paper uses a new mathematical relation and a shape factor to measure Q in seismic records. The method uses the centroid frequency difference between the incident and reflected signals. The centroid frequency is similar to a statistical average of the frequencies in the spectrum of a seismic signal (Barnes 1993). The procedure for Q measurement using the new method can be divided into two parts: 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 paired incident and transmitted signals. The use of variable window length reduces interference from seismic multiples and noise. The inclusion of a shape factor in the mathematical model allows the users to represent the real seismic signal with any wavelet shape that satisfies the basic characteristics of a seismic signal (as commonly done in attenuation measurement experiments) without altering the mathematical model. The wavelet shapes considered in the new attenuation model are Gaussian, Ormsby (or boxcar), Ricker, and triangle wavelets. Throughout this paper, we work with P-wave attenuation, but we simply refer to it as $1/Q$.

In subsequent sections, we describe the mathematical relation for the new method. Following this, the method is used to compute attenuation in synthetic seismic records, and in field seismic data recorded in an oil field in Norway. Finally, the method is compared with other published methods.

Theory and methods

The effect of the absorption property of the Earth on seismic records is usually modelled as the convolution of a seismic impulse and the Earth reflectivity series with an Earth filter. The filter is minimum phase, causal, and depends on the quality factor Q (Aki and Richard 1980). The real seismic impulse is

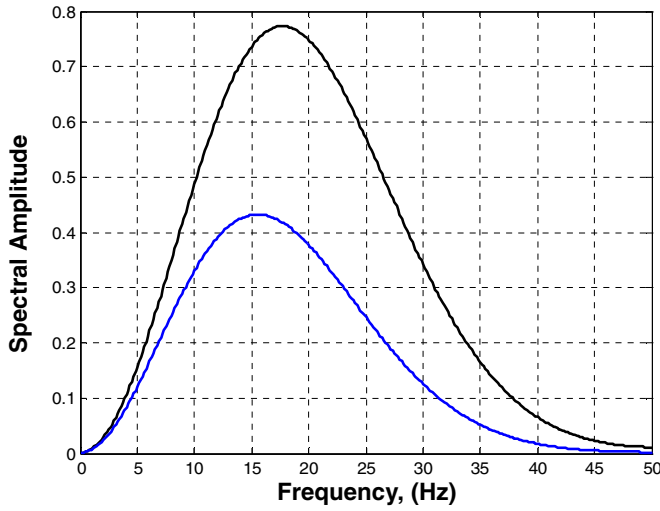


Figure 1. The frequency spectrum of a Ricker wavelet at time 0 (black) and time $t = 2.2$ s (blue) in an attenuating medium with the Q value equal to 30—showing a loss in frequency due to the effect of attenuation.

commonly modelled by a Ricker or a Gaussian wavelet. If a real seismic signal is represented by a Ricker wavelet, the source spectrum signal at initial time 0 in an Earth medium is given (Zhang and Ulrych 2002) as:

$$b_0(f, 0) = \frac{2}{\sqrt{\pi}} \frac{f^2}{f_m^2} \exp\left(-\frac{f^2}{f_m^2}\right), \quad (3)$$

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:

$$b(f, t) = b_0(f, 0)h(f, t), \quad (4)$$

where $h(f, t)$ is the Earth absorption filter whose frequency response is,

$$H(f) = \exp(-f \alpha(w, l) dl). \quad (5)$$

The attenuation coefficient α is related to Q as given in equation (2). Therefore, the spectrum of the seismic signal after a time t can be given as:

$$b(f, t) = b_0(f, 0) \exp\left(\frac{\pi f \Delta t}{Q}\right), \quad (6)$$

where $\Delta t = t - 0$. The observed difference in the frequency spectrum of the seismic pulse at 0 and time t are shown in figure 1, and their corresponding amplitudes are shown in figure 2.

Taking the natural logarithm of equation (6) and rearranging the terms, we have:

$$\ln\{b(f, t)\} - \ln\{b_0(f, 0)\} = \frac{\pi f \Delta t}{Q}. \quad (7)$$

Equation (7) is the basis of the spectra ratio method (Bath 1974, Spencer *et al* 1982). If the amplitude spectral $b(f, t)$ and $b_0(f, 0)$ are defined in terms of the spectral centroid frequency, cf_a and cf_b , and we remove the logarithm, the elements in equation (7) can be rearranged as:

$$Q = \frac{\pi f \Delta t}{cf_a - cf_b}. \quad (8)$$

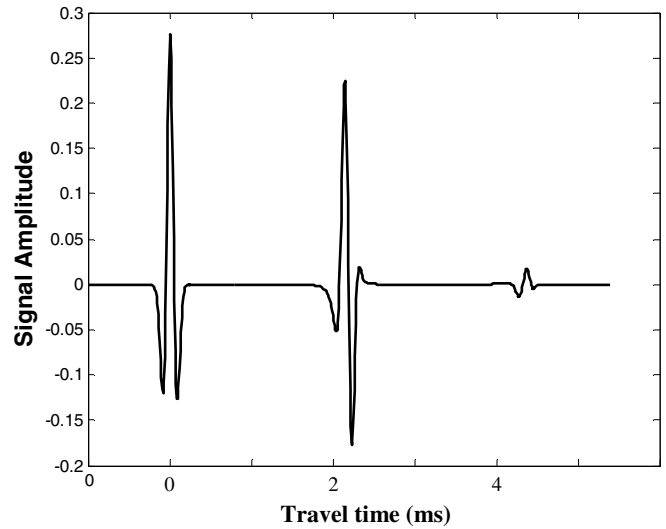


Figure 2. The amplitude spectrum of a Ricker wavelet at time 0 and 2.2 s in an attenuating Earth model with the Q value equal to 30—showing amplitude reduction and waveform distortion due to the effect of attenuation. The signal at 4.2 ms is a multiple reflection.

Equation (8) is the basis of the frequency-shift method for attenuation measurement—where Q is estimated from the frequency and travel time difference ($cf_a - cf_b$ and Δt) between the incident and the transmitted signals (see, Quan and Harris 1997, Zhang and Ulrych 2002). cf is a spectra parameter known as the centroid frequency. The centroid frequency is an average spectra parameter (Barnes 1993) defined as:

$$cf_a = \frac{\int_0^\infty f|a(f)|df}{\int_0^\infty |a(f)|df}, \quad \text{and} \quad cf_b = \frac{\int_0^\infty f|b(f)|df}{\int_0^\infty |b(f)|df}. \quad (9)$$

For attenuation measurement in the exploration frequency band, it is common to use an empirical equation to model attenuation (Gladwin and Stacy 1974, Toverud and Ursin 2005).

Based on the physical observations shown in figures 1 and 2; the definition given in equation (8); and several synthetic experiments, we propose a statistics-based empirical model for Q measurement as follows:

$$Q = \frac{t2cf_acfb}{\mu\sigma_{ab}(cf_a - cf_b)}, \quad (10)$$

where μ is the shape factor, a constant used to obtain the best fit to the data for the various spectral shapes. As pointed out by equations (12)–(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. σ_{ab} is the average standard deviation of the incident and transmitted spectra, and t is the travel time difference. The average standard deviation, σ_{ab} is:

$$\sigma_{ab} = \frac{\sigma_a + \sigma_b}{2} \quad (11)$$

$$\text{where } \sigma_a = \sqrt{\sigma_a^2}, \quad \text{and} \quad \sigma_b = \sqrt{\sigma_b^2}. \quad (12)$$

$$\sigma_a^2 = \frac{\int (f - cf_a)^2 a(f) df}{\int a(f) df}, \quad \text{and} \quad \sigma_b^2 = \frac{\int (f - cf_b)^2 b(f) df}{\int b(f) df}. \quad (13)$$

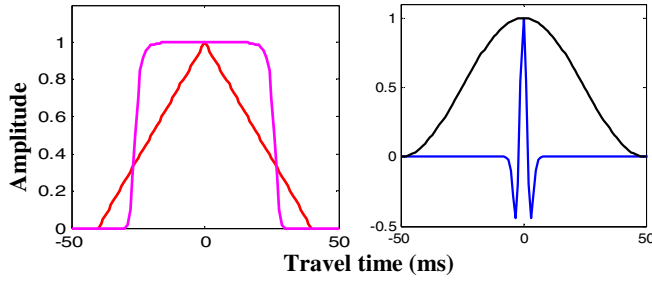


Figure 3. The various wavelet shapes used in the attenuation model to represent the real seismic signal. Red is the triangular spectrum, pink is the Ormsby spectrum (or a boxcar), black is the Gaussian spectrum, and blue is the Ricker spectrum.

For a source–receiver offset x , the travel time t can be expressed (Sherrif and Geldart 1985, Burger *et al* 1992) as:

$$t = t_0 \left(1 - \frac{x^2}{v_{rms}^2 t_0^2} \right)^{1/2}, \quad (14)$$

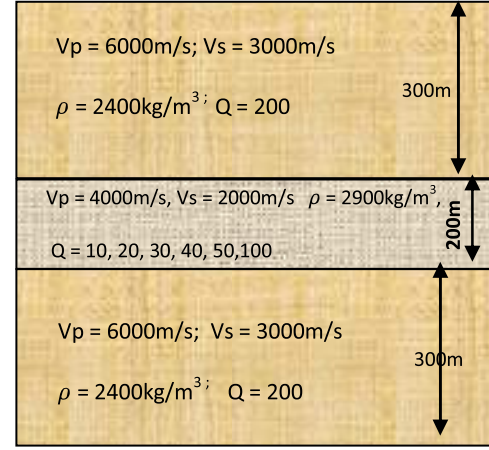
where t_0 is zero offset travel time and v_{rms} is the root mean squared velocity. The time, t , required for equation (10) can be approximated as:

$$t = t_0 \left(1 + \frac{x^2}{2v_{rms}^2 t_0^2} - \frac{x^4}{8v_{rms}^4 t_0^4} \right). \quad (15)$$

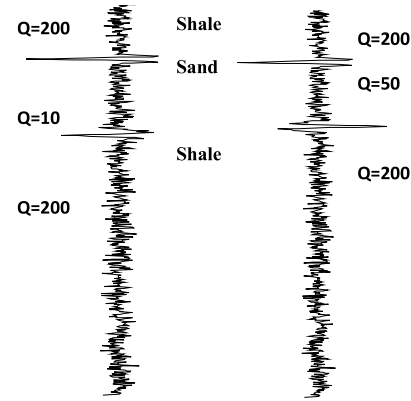
The use of a shape factor in the mathematical model allows several wavelet shapes to be used to represent the real seismic pulse. If the real seismic pulse is represented by a Ricker wavelet, the shape factor, μ , is equal to $\frac{2\pi}{3}$. The exact shape of a real seismic signal is generally unknown; however knowledge about the source can be obtained from full waveform inversion of the seismic data (e.g., Pratt 1999). 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, it 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 spectrum, a Gaussian spectrum, or a triangular spectrum, and following the procedures and definitions given in equations (9)–(13), the shape factor in equation (10) 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 corresponding spectral shapes are shown in figure 3. It is also possible to assume that a real seismic wavelet is a combination of all the wavelets shapes considered above. In this case, the shape factor will be the mathematical mean of the shape factors for the four wavelets. But we chose to use the various wavelet shapes in our computation in order to compare our results with other methods that used similar wavelets.

Synthetic tests and model comparison

Futterman's causal absorption model of 1962 (Futterman (1962)) is incorporated into the plane wave reflectivity algorithm of Müller (1985) to generate attenuated P-wave seismograms using a Ricker source. The synthetic wave



(a)



(b)

Figure 4. (a) Three-layer Earth model comprising shale–sand–shale, and the layer parameters. (b) Some of the synthetic P-wave seismograms generated (with multiples and noise) from the Earth model shown in (a): left—seismogram generated with Q of 200, 10, 200. Please note that the effect of high attenuation ($Q = 10$) caused polarity reversal at the bottom of sand layer. Right—seismogram generated with $Q = 200, 50, 200$. No polarity reversal in this case.

generation code is written in MATLAB. The Earth model consists of shale–sand–shale. The layer parameters are shown in figure 4(a). The P-wave synthetics are generated with multiples and noise. The P-wave quality factor (Q) values for the sand are 10, 20, 30, 40, 50, and 100, while the Q for the top and bottom shale are fixed at 200. Examples of the synthetic records generated are shown in figure 4(b).

Q is measured in the synthetics using the model presented in the above section (equation (10)), by comparing the seismic events corresponding to the top and base of the sand. The Q estimates are plotted against the input model in figure 5.

The proposed Q measuring algorithm is hereafter known as the SFVQM (spectral frequency variation Q method). The SFVQM is compared to two published methods of Q measurement. The two methods are the frequency-shift method (Quan and Harris 1997), and the spectral ratio method (Spencer *et al* 1982). The basis for the choice of these methods is that they are highly referenced in the literature on attenuation studies.

- (1) The spectral ratio method (SRM) assumes a Ricker source, and is noted for high performance in noisy seismic data.

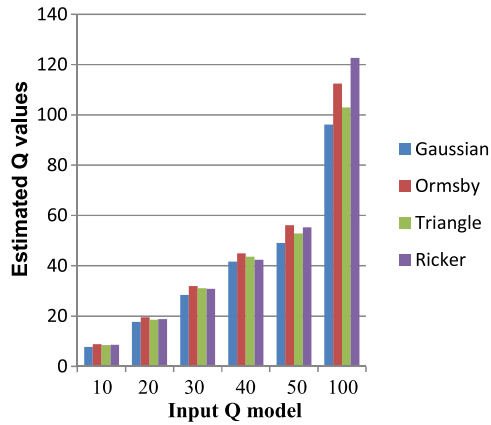


Figure 5. Input Q model versus Q estimates. The colour code represents the various spectral shapes. For Ormsby, Gaussian and triangular spectral shapes, the frequency bandwidth is 10–40 Hz. For the Ricker spectrum, the dominant frequency is 20 Hz. See the legend for the colours.

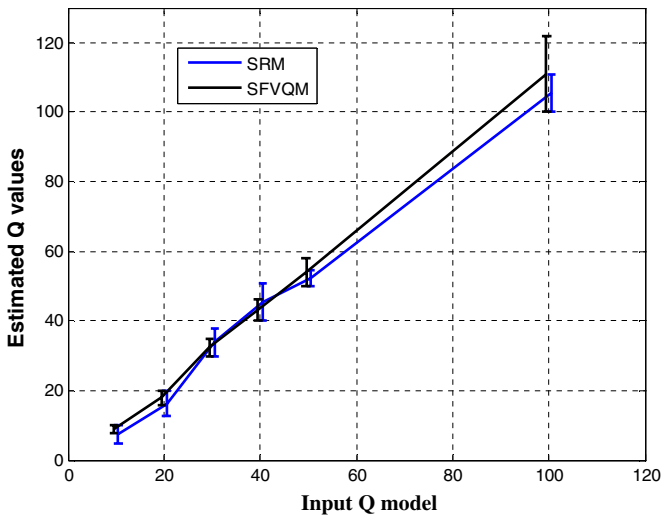


Figure 6. Comparing the SFVQM with the spectra ratio method (SRM)—testing the effects of noise on Q measurements. For each Q model, the noise level in the seismogram is increased from 5 to 10, 15 and 20% of the maximum signal amplitude. The average Q estimates are plotted on the vertical axis (with an error bar). See the legend for a colour code.

In order to comment on the sensitivity of the SFVQM to noise, we compare the SFVQM (with a Ricker shape wavelet) to the spectral ratio method, for measuring Q in noisy seismograms. For each Q model, the noise level is varied from 5%, 10%, 15%, to 20%. The mean Q estimates are plotted against the Q model in figure 6.

- (2) The Q model of Quan and Harris (1997) designed for a Gaussian-shaped wavelet is also noted for its insensitivity to changes in the frequency bandwidth. To comment on the sensitivity of the SFVQM to changes in frequency bandwidth, we compare the SFVQM (using a Gaussian shape spectrum) to the model of Quan and Harris (1997). For each Q model, the frequency band of the source wavelet is varied as 10–40, 10–50, 20–50 and 20–60 Hz. The mean Q estimates are again plotted with an error bar in figure 7.

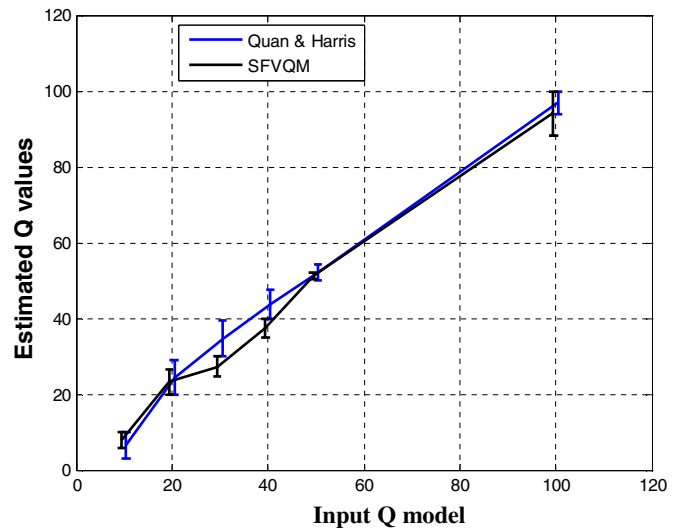


Figure 7. Comparing the SFVQM with the method of Quan and Harris (1997)—testing the effects of the change in frequency bandwidth. For each Q model, the frequency band of the source wavelet is varied as 10–40, 10–50, 20–50, and 20–60 Hz. The average Q estimates are plotted on the vertical axis (with an error bar). See the legend for a colour code.

The motivation for comparing our method (SFVQM) to the other two methods is to comment on the performance of the SFVQM with respect to the special features for which the other two are known. One advantage of measuring Q in seismic data using different methods is that it provides a basis for the assessment of the reliability of the new method. The Q measuring algorithm described in equation (9)–(15) is written as a code in Matlab, and all the computations took place in a Matlab environment. The results plotted in figure 5 show that the SFVQM can provide reliable estimates of Q with synthetic data and realistic seismograms using any of the spectral shapes. Generally, the SFVQM is less sensitive to noise interference compared to the SRM (figure 6) and thus, can give more reliable estimates of Q compared to the SRM.

The error analysis of the Q estimates presented in figure 6 shows that the mean error from the SFVQM is 9.5%, while the mean error from the SRM is 14.7%. The mean error is estimated as the average of the errors from the six measurements plotted in figure 6. The result plotted in figure 7 shows that the SFVQM compares favourably with the method proposed by Quan and Harris (1997). The mean errors from the SFVQM and the method proposed by Quan and Harris (1997) are 9.8% and 13.4%, respectively. However, it is observed that the SFVQM gave less robust estimates, compared to the other two methods, when the input Q is 100. The reason is that the SFVQM is designed for high attenuation (low Q) cases, as commonly observed in oilfield data.

Q measurement in seismic field data

A set of 3D data from the Gullfaks field is used to test the suitability of the SFVQM for Q measurement in field data. The Gullfaks field is located in the Norwegian sector of the northern North Sea. The data was acquired by STATOIL in

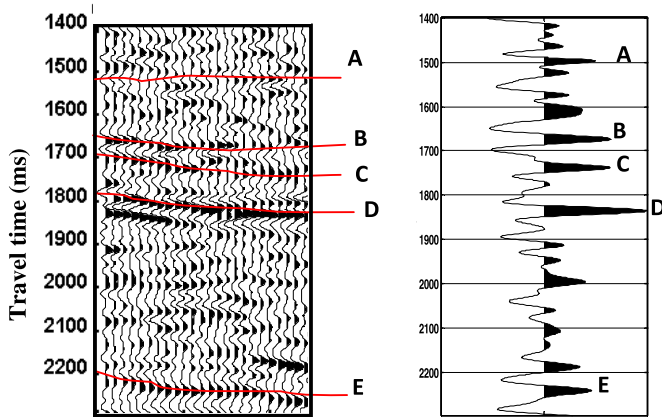


Figure 8. Left—seismic traces in the CDP 602. Right—stacked trace along the CDP 602 line. The trace length is subdivided into four intervals; AB, BC, CD and DE. The subdivision is for the purposes of Q measurement only.

1985. The available seismic records represent a travel time interval of 905 ms, starting from 1397 to 2302 ms, and it cuts across two series of sandstones separated by a shale layer. The well data accompanying the seismic cube consist of time, porosity, permeability and gamma logs covering a depth range of 1500 to 2416 m. The information synthesized from the well logs is used to generate a lithology log and an acoustic impedance log, to provide guidance in discriminating the reflections that correspond to lithological boundaries from those that are due to artefacts, noise, and multiples. Based on this approach, the entire time-depth section covered by the seismic traces is partitioned into four intervals, as shown in figure 8. The selected traces were edited to remove traces with an unacceptable level of noise.

For Q measurement in field data, we use the SFVQM with a Ricker spectrum (where μ is $2\pi/3$). Because Matlab cannot read seismic data in Segy format, a Seislab interface on Matlab was used to import and manipulate the seismic data within the Matlab command window. Seismic signals that correspond to the top and base of the interval to be analysed for attenuation were isolated using a Hanning window. To minimize interference from noise, variable window lengths are used (see also Reine *et al* 2009). The minimum and maximum window lengths used are 76 ms and 98 ms, respectively. Fourier transform is computed within the window containing each desired event and the spectral parameters required for Q measurement are computed as defined in equations (9)–(13). Whenever possible, the same window length is used for a signal pair being compared in the measurement. For the purpose of Q estimation, the total travel time represented by the traces is divided into four intervals: AB, BC, CD, and DE, as shown in figure 8. Following this, the traces corresponding to CDP 602 are stacked. Attenuation (or $1/Q$) is measured in the stacked traces using the SFVQM and the SRM. The results of the measurements are presented in table 1. Attenuation profiles plotted from the $1/Q$ measured in the stacked trace, using the SFVQM and the SRM are shown figure 9. The aim of the new method is to show that the special features that other methods are known for are combined in the SFVQM. The other aim is

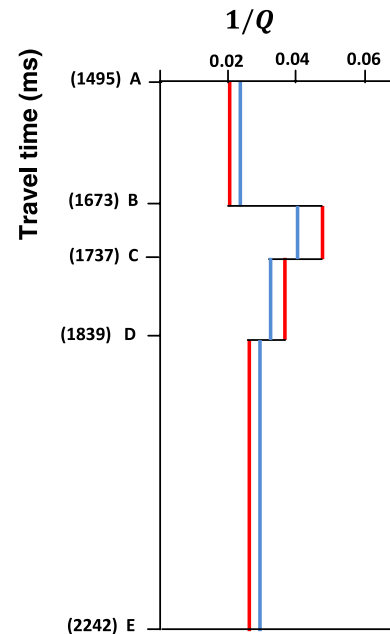


Figure 9. Attenuation profile plotted from the $1/Q$ estimated from the stacked trace. Red is from the SFVQM, blue is from the SRM.

Table 1. Attenuation ($1/Q$) estimates in the stacked traces using the SFVQM and the SRM.

Interval (m)	AB 1495–1673	BC 1673–1737	CD 1737–1839	DE 1839–2242
SFVQM	0.0202	0.0489	0.0372	0.0260
SRM	0.0226	0.0383	0.0329	0.0291

to show that the SFVQM can compute reliable estimates of Q in real data.

Discussion

The use of different wavelet shapes to represent a real seismic signal is a common practice in seismic attenuation experiments. Different models have been proposed for attenuation measurements based on the shape of the wavelet used to represent the real seismic signal. The method proposed for attenuation measurement in this paper uses a new mathematical relation between the difference in the centroid frequency of a paired seismic event, the travel time difference between the two events, and a shape factor.

The shape factor allows the use of any shape (that satisfies the basic characteristic of a real seismic pulse) to represent the real seismic signal without altering the mathematical relation. The wavelet shapes considered in the new model include Ricker, Ormsby (or boxcar), triangular, and Gaussian. A more realistic approximation for the wavelet can be obtained by computing a mixture of all the wavelet shapes, where the shape factor will be the average of all the shape factors. However, we chose to use the various shapes in our computation to provide a basis for comparing our method with other methods that used similar wavelet shapes.

The proposed method (SFVQM) is used to measure Q in synthetic seismograms using various wavelet shapes. Results

show that the SFVQM can provide robust estimates of Q using any of the wavelet shapes. The maximum difference in the Q estimates using any of the shapes is about 5% for an input Q model of 10, 20, 30, 40, and 50. The difference peaks to 12% when a Q model of 100 (i.e. low attenuation) is considered. Compared to the SRM, the SFVQM is less sensitive to noise interference. This is because the SFVQM relies on the centroid frequency—an average of the entire frequency content of the signal; as it averages the spectral frequency, it decimates the effect of noise.

The 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, the SFVQM shows less sensitivity compared to the method proposed by Quan and Harris (1997) and, thus, gave better estimates of Q . The errors are 9.8% and 14.7% for the SFVQM and the method of Quan and Harris (1997), respectively. The $1/Q$ values measured in the field data using the SFVQM are 0.0202, 0.0489, 0.0372, and 0.0260, for intervals AB, BC, CD, and DE, respectively. These results are consistent with those measured using the SRM. Because the SRM is generally acceptable for computing $1/Q$ in field data, the consistencies in the results therefore show that the SFVQM is suitable for measuring $1/Q$ in field data. Interval BC, which recorded the highest attenuation value, approximately coincides with the shale break that separates the top and base sandstones. The factor responsible for the high attenuation in interval BC compared to the overlying and underlying sandstone cannot be ascertained from this study. However, results from attenuation measurement by Stainsby and Worthington (1985) in VSP data recorded in a well from the North Sea, Norway, show that shale in the area has a Q value of 25 ± 3 or $1/Q$ attenuation ($0.0357 \leq 1/Q \leq 0.0455$).

Conclusion

A new method, the SFVQM, for measuring attenuation ($1/Q$) in seismic records is presented. The SFVQM combines the special features that the SRM and the model of Quan and Harris (1997) are known for: it is less sensitive to interference from noise and the change of frequency bandwidth compared to the SRM and the method of Quan and Harris (1997), respectively. Further to this, the SFVQM can compute reliable estimates of Q using different wavelet shapes without altering the mathematical model—a feature that is not obtainable in other methods. The consistencies in the attenuation profiles plotted from the $1/Q$ measured in field seismic data using the SFVQM and the SRM show that the SFVQM can provide robust estimates of attenuation in real and synthetic seismic data. Because $1/Q$ is sensitive to pore fluids, attenuation measured in field seismic records can provide useful guides during hydrocarbon exploration and reservoir characterization.

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