Case History

Spatiotemporal variations in seismic attenuation during hydraulic fracturing: A case study in a tight oil reservoir in the Ordos Basin, China

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ABSTRACT

During hydraulic fracturing (HF) stimulation for unconventional reservoir development, seismic attenuation has a significant influence on high-frequency microseismic data. Attenuation also provides important information for characterizing reservoir structure and changes to it due to HF injections. However, the attenuation effect is typically not considered in microseismic analysis. We have adopted the spectral ratio and centroid-frequency shift methods to estimate the subsurface attenuation (the factor $Q^{-1}$) in a tight oil reservoir in the Ordos Basin, China. The P- and S-wave attenuations are calculated using the 3C waveform data recorded by a single-well downhole geophone array during a 12-stage HF stimulation. Both methods provide similar results (with differences in $Q^{-1}$ of absolute values less than 0.010 for P- and S-waves). For individual events, their median $Q^{-1}$ values calculated from different geophones are selected to represent the average attenuation. Spatiotemporal variations in attenuation are obtained by investigating $Q^{-1}$ values along propagating rays linking different source–receiver pairs. The $Q^{-1}$ values derived at different HF stages reveal significant attenuation in the targeted tight sandstone layer (0.030–0.062 for $Q^{-1}_P$ and 0.026–0.058 for $Q^{-1}_S$), and the attenuation is apparently increased by fluid injection activities. We explain the sudden decrease in attenuation near the geophone array as a result of high shale content using log data from a horizontal treatment well. The consistency between the $Q^{-1}$ values and horizontal well-log data, as well as the HF process, indicates the reliability and robustness of the attenuation results. By studying spatiotemporal variations in attenuation, the changes in subsurface structures may be quantitatively characterized, thereby creating a reliable basis for microseismic modeling and data processing and providing additional information on monitoring the HF process.

INTRODUCTION

During unconventional oil and gas development, hydraulic fracturing (HF) stimulation is a key technique for enhancing resource production. Currently, most studies on microseismic monitoring of HF stimulation focus on characterizing hydraulic fractures with microseismic event locations and focal mechanisms (Baig and Urbancic, 2010; Maxwell, 2011; Li et al., 2021; Li and Chang, 2021). Compared to velocity or density parameters, attenuation is typically neglected in microseismic modeling and data processing. However, attenuation affects all seismic data, sometimes drastically (Kjartansson, 1979; Tonn, 1991; Abercrombie, 1998). Attenuation can be treated as a low-pass filter that changes the frequency content of seismic waves, thereby affecting their amplitudes and leading to phase shifts (Aki and Richards, 2002). Therefore, obtaining attenuation parameters is essential for reliable estimations of seismic source spectrum, mechanism, and magnitude (Abercrombie, 1995; Tomic et al., 2009; Chang et al., 2018). The impact of attenuation is more severe for microseismic waves than conventional reflection seismic data due to their higher frequencies (100–1000 Hz in downhole monitoring cases). In addition, because attenuation is sensitive to fracture density, fluid saturation, and other petrophysical properties, it can be a valuable parameter for reservoir characterization and indicating how HF
injection affects the subsurface rock formations between the seismic source and receiver (Aki et al., 1982; Klimentos, 1995; Vlastos et al., 2007; Blake et al., 2020).

There are numerous studies for estimating attenuation using data from laboratory HF experiments (Winkler et al., 1979; Lei and Xue, 2009; Subramanian et al., 2014; Zhai et al., 2017) or natural earthquakes (Aki and Chouet, 1975; Anderson and Hart, 1978; Zhao et al., 2013). However, only a limited number of studies have conducted attenuation measurements in microseismic field data processing, and even fewer have investigated attenuation variations during the HF process. Eaton (2011) estimates attenuation from high-quality microseismic records and demonstrates that attenuation potentially provides useful constraints for predicting magnitude-detection distance. Eisner et al. (2013) propose an attenuation inversion method based on the peak frequency of direct waves for microseismic events. Fieltz and Wegler (2015) study the high-frequency attenuation properties from fluid-induced microseismicity at the German Continental Deep Drilling site. Zhu et al. (2017) investigate the spatiotemporal changes in seismic attenuation during CO$_2$ injection and reveal that attenuation variation is an indicator of the movement and saturation of CO$_2$ plumes. Wcislo et al. (2018) characterize the variations in attenuation in wastewater injection activities in the High Agri Valley, Italy, and find anomalously high attenuation around the injection well. Zhang et al. (2019) estimate attenuation using perforation shot signals and microseismic waveform records and reveal that HF stimulation leads to a considerable increase in reservoir attenuation. Yu et al. (2020) find that seismic energy loss (attenuation) is higher near the HF well and interpret this as a result of higher fracture density and/or elevated pore pressure in the rock matrix due to HF stimulations. Ameri et al. (2020) investigate the source and attenuation properties of induced events in the Groningen gas field using spectral decomposition of borehole records.

In this study, the spectral ratio (SR) method (Kjartansson, 1979) and the centroid-frequency shift (CFS) method (Quan and Harris, 1997) are used to investigate subsurface attenuation during an HF stimulation process. First, we briefly introduce the HF stimulation project and these two methods. Then, we apply these two methods to a multicomponent microseismic data set recorded during the HF process and calculate the compressional wave (P-wave) and shear wave (S-wave) attenuation parameters. Finally, we analyze the spatiotemporal variations in attenuation during the entire HF process and link the results to the fluid injection activities and log data from a horizontal treatment well.

**BACKGROUND AND METHODOLOGY**

**Overview of the HF project**

We study a single-well downhole microseismic monitoring data set from a tight oil reservoir in China. The study area is located in the south of the Ordos Basin, a large sedimentary basin near the western margin of the North China Craton in central China (Wang et al., 2015). The Ordos Basin consists of six substructures: the Yimeng uplift in the north, the Weibei uplift in the south, the Tianhuain depression and western thrust belt in the west, the Jinxin fold belt in the east, and the Yishan slope in the central part (see Figure 1; Tang et al., 2014). The HF project is conducted through a horizontal well in the Triassic Yanchang Formation, which is characterized by complicated pore-throat structures, abundant fractures, low permeability, and good crude properties (Zeng and Li, 2009; Yao et al., 2013). The targeted reservoir is located roughly from depth 1550 to 1650 m and primarily occurs in the Chang 7 oil-bearing member, which is largely composed of sandstones and shales with high interstitial content and rich natural fractures (approximately two to ten fractures every 10 m, e.g., Li et al., 2015; Fan et al., 2016).

There are 12 treatment stages in this HF project (map and side views are shown in Figure 2a and 2b). The HF process is monitored with a 12-level 3C geophone array in a nearby vertical monitoring well, with its sensors located from approximately 1317 to 1631 m in depth with a depth interval of approximately 30 m. The time-sampling interval of the waveform records is 0.5 ms. In total, 4811 microseismic events are detected and processed by a contractor using standard industrial practices. One hundred and nine events with cross-seismic events are detected and processed by a contractor using standard industrial practices. One hundred and nine events with cross-seismic events are selected to conduct the attenuation measurements. The numbers of total and selected microseismic events within each HF stage are listed in Table 1. In the contractor’s processing, a homogeneous velocity model (with P-wave velocity $V_P = 4.36$ km/s and S-wave velocity $V_S = 2.40$ km/s) is built based on sonic logs from the monitoring well (Figure 2c) and microseismic events are located.

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Figure 1. Map showing tectonic divisions in the Ordos Basin, China. The star denotes the location of the HF site and the microseismic monitoring well.
by minimizing the misfit between the picked and predicted arrival times. The distributions of selected microseismic events grouped by individual HF stages are shown in Figure 2.

\(Q^{-1}\) estimation using the SR method

In this study, seismic attenuation is described by the inverse quality factor \(Q^{-1}\), which is assumed to be independent of frequency (Kjartansson, 1979). For a moment-tensor seismic source at \(s\), the \(n\)-th component (\(n = 1, 2, 3\) denote north, east, and down, respectively) amplitude spectrum at the geophone \(i\) (marked as \([A_n(f)]_{is}\)) is given by (Vera Rodriguez and Stanchits, 2017):

\[
[A_n(f)]_{is} = M_0 \cdot S(f) \cdot [G_{npq} \cdot M_{pq}]_{is} \cdot \exp \left( -\frac{\pi f r_{is}}{V} Q^{-1} \right), \tag{1}
\]

where \(M_0\) is the scalar seismic moment, \(S(f)\) is the source spectrum, \(M_{pq}\) (\(p, q = 1, 2, 3\)) is the normalized moment-tensor component, \(G_{npq}\) is the Green’s tensor denoting the \(n\)-th spectrum component at geophone \(i\) caused by \(M_{pq}\) at source \(s\) in an elastic medium (Aki and Richards, 2002). \(f\) is the frequency, \(r_{is}\) is the source-receiver distance, and \(V\) is the wave velocity (with \(Q_{p}^{-1}\) and \(V_p\) for P-wave and \(Q_{s}^{-1}\) and \(V_s\) for S-wave, respectively). From equation 1, higher \(Q^{-1}\) corresponds to stronger attenuation. For a homogeneous isotropic medium, the logarithmic SR (LSR) of a geophone pair \(i-j\) can be derived as (Tomn, 1991; Vera Rodriguez and Stanchits, 2017):

\[
\ln \left( \frac{[A_n(f)]_{is}}{[A_n(f)]_{js}} \right) = -\pi \Delta t_{ij} Q^{-1} f + \ln M. \tag{2}
\]

where \([A_n(f)]_{is}/[A_n(f)]_{js}\) is the SR from the same source \(s\) for component \(n\) at two different geophones \(i\) and \(j\) (here assuming \(r_{is} > r_{js}\)). \(M = [G_{npq}M_{pq}]_{is}/[G_{npq}M_{pq}]_{js}\) is a factor which is composed of geometric spreading and source radiation pattern but should be frequency-independent, and \(\Delta t_{ij} = (r_{is} - r_{js})/V\) is the traveltime difference between geophones \(i\) and \(j\). From equation 2, we see that \(\ln M\) is a constant for a given geophone pair; thus, the relationship between \(\ln ([A_n(f)]_{is}/[A_n(f)]_{js})\) and \(f\) is linear, with the slope \(k = -\pi \Delta t_{ij} Q^{-1}\). Therefore, the \(Q^{-1}\) can be obtained by fitting a straight line to the observed LSR.

\(Q^{-1}\) estimation using the CFS method

The CFS method (Quan and Harris, 1997; Zhu et al., 2017) is also used to estimate \(Q^{-1}\) in our study. From Quan and Harris (1997), the centroid frequency (CF) (or dominant frequency) of an observed amplitude spectrum \(A(f)\) is defined as

\[
\text{Table 1. Number of total and selected microseismic events within each HF stage.}
\]

<table>
<thead>
<tr>
<th>Stage Number of selected events</th>
<th>Number of total events</th>
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<tbody>
<tr>
<td>1</td>
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<td>2</td>
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<td>11</td>
<td>8</td>
</tr>
<tr>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Total</td>
<td>109</td>
</tr>
</tbody>
</table>
\[ \text{CF} = \frac{\int fA(f)df}{\int A(f)df}, \]

and its variance \( \sigma^2 \) is given by

\[ \sigma^2 = \frac{\int (f - \text{CF})^2A(f)df}{\int A(f)df}. \]

For a geophone pair \( ij \) \((r_{ij} > r_{jk})\) with the \( n \)-th component amplitudes \( [A_n(f)]_{ij} \) and \( [A_n(f)]_{jk} \), their relationship can be derived from equation 1 as

\[ [A_n(f)]_{ij} = [A_n(f)]_{jk} \cdot M \cdot \exp(-\pi f \Delta t_{ij} Q^{-1}). \]

where \( [A_n(f)]_{ij}^{\text{syn}} \) is the synthetic \( n \)-th component amplitude spectrum of geophone \( i \), predicted from the observed spectrum \( [A_n(f)]_{ij} \) of geophone \( j \), and \( M = \Delta t_{ij} \) are the same as those defined in equation 2. Therefore, we can estimate \( Q^{-1} \) by finding the best fit between \( [A_n(f)]_{ij} \) and \( [A_n(f)]_{ij}^{\text{syn}} \) when \( Q^{-1} \in [\text{min}, \text{max}] \). In this study, the following objective function based on the \( L_2 \)-norm is adopted for \( Q^{-1} \) estimation:

\[ \text{Misfit} = \frac{\|\text{CF}_i - \text{CF}_{ij}^{\text{syn}}(Q^{-1})\|_2}{\max(\|\text{CF}_i - \text{CF}_{ij}^{\text{syn}}\|_2 Q^{-1} \in [\text{min}, \text{max}] \} + \frac{\|\sigma^2 - \sigma_{ij}^{\text{syn}}(Q^{-1})\|_2}{\max(||\sigma_i^2 - \sigma_{ij}^{\text{syn}}||_2 Q^{-1} \in [\text{min}, \text{max}] \)}, \]

where \( \text{CF}_i \) is the centroid frequency of the observed spectrum \( [A_n(f)]_{ij} \), \( \sigma^2 \) is its variance, and \( \text{CF}_{ij}^{\text{syn}}(Q^{-1}) \) and \( \sigma_{ij}^{\text{syn}}(Q^{-1}) \) are the centroid frequency and variance of the synthetic spectrum \( [A_n(f)]_{ij}^{\text{syn}} \) for a given \( Q^{-1} \), respectively. From equations 3–5, \( \text{CF}_{ij}^{\text{syn}} \) and \( \sigma_{ij}^{\text{syn}} \) are independent of the term \( M \) because \( M \) is frequency-independent. Therefore, the objective function (equation 6) is only related to \( Q^{-1} \). In data processing, we calculate all misfit values with \( Q \) ranging over all integers from 5 to 500 (i.e., \( Q^{-1} \in [0.002, 0.200] \)) and use \( Q^{-1} \) at the minimum misfit as the estimated attenuation. The variations of \( Q^{-1} \) (or \( Q \)) corresponding to 50% rise of the misfit function from its minimum value are used for the uncertainty measurements of the CFS method.

Comparing the SR and CFS methods with each other, although the SR method is more commonly used, it requires simultaneous processing with the field data

\[ \text{Processing with the field data} \]

Figure 3 illustrates the waveforms and spectrograms of two sample microseismic events, E1 and E2, with their locations shown in Figure 2a. From Figure 3a and 3b, we can see apparent changes in time-domain P- and S-waves from the top channel (geophone no. 1) to the bottom channel (geophone no. 12). A visual examination can clearly reveal that the dominant frequencies at deeper geophones are higher than those at shallower geophones because the waves received by the shallower geophones propagate through longer distances. We attribute this change in frequency content to the attenuation effects along the raypaths. Moreover, by comparing the z-component spectrograms of geophones no. 2 and 10 (Figure 3c–3f), we find the relative loss of high frequencies with increasing source–receiver distance. This phenomenon is more prominent for the S-wave, with the frequency at its maximum energy declining from approximately 150 Hz at G10 to approximately 100 Hz at G02. This can be explained by the fact that S-waves encounter stronger scattering attenuation than P-waves, and high-frequency signals have greater attenuation than low-frequency signals along the same raypath (Tonn, 1991).

We use SR and CFS methods to measure subsurface \( Q^{-1} \) values. These methods calculate \( Q^{-1} \)
values experienced by seismic signals along their travel path from the sources to the receivers (Vera Rodriguez and Stanchits, 2017; Zhu et al., 2017), that is, the average attenuation effect over the entire seismic raypath. Owing to the limited coverage of a single-well downhole microseismic monitoring system, we assume a constant attenuation model (i.e., $Q^{-1}$ is homogeneous within the whole model) for each microseismic event. We use a Tukey window to sample the P- and S-wave signals (Zhang et al., 2019). The time durations of the sampling windows are approximately 1.5 dominant periods of the observed P- and S-wave traces for all microseismic events. Before the Fourier transform, we add zeros to the extracted P- and S-waveforms after the time series, making the total time length 0.1 s to increase the frequency-domain sampling rate. According to equations 2 and 6, we can obtain an attenuation measurement ($Q_{ij}^{o}$) for each pair of observed spectra $[A_i(f)]_{ij}$ and $[A_j(f)]_{ij}$ from the same source s. To determine the frequency band for attenuation measurement, we calculate the CF from $[A_i(f)]_{ij}$ and $[A_j(f)]_{ij}$ (marked as $CF_i$ and $CF_j$, respectively). Assuming that $CF_i < CF_j$, we use $[CF_i - 30 \text{Hz}, CF_i + 30 \text{Hz}]$ as the frequency band to calculate $Q_{ij}^{o}$. In other words, the frequency band will be adapted for specific source, geophone pair, component, and wave type (i.e., P- or S-wave) combinations. By testing different methods, we find that this frequency band selection makes the results from the SR and CFS methods more consistent. As examples, Figures 4 and 5 illustrate the attenuation measurement procedure with the SR and CFS methods using the waveforms in Figure 3c–3f. From Figures 4 and 5, the estimated $Q_{ij}^{o}$ and $Q_{ij}^{s}$ using the SR method are (0.033, 0.030) and (0.040, 0.031) from E1 and E2, respectively, and those using the CFS method are (0.038, 0.032) and (0.037, 0.031), respectively. The results from these two methods are similar and indicate strong subsurface attenuations for P- and S-waves. In Figures 4 and 5, the CF values are always larger than F (the frequency at the peak spectral amplitude). This is because the CF is defined as the average frequency weighted by amplitudes (equation 3) and the factor $fA(f)$ at higher frequency contributes more to the CF than that at lower frequency. Figures 4d, 4h, 5d, and 5h illustrate the misfit functions versus the $Q$ values, where the magnified figures show the variations of $Q$ values corresponding to 50% rise of the misfit functions from their minimum values. They provide uncertainty evaluations of a single measurement using the CFS method.

Given a microseismic event, there are 12 3C geophones, which provide 36 waveforms. For the SR and CFS methods, we select two geophones to form a geophone pair ($i$ and $j$) provided that they have different arrival times. Waveforms of the same component (north–south, east–west, or z) in a geophone pair are used in a single estimation. In some instances, calculations yield unrealistic $Q^{-1}$ values (e.g., negative or near zero), which may be caused by the noise in the data, or the fact that the traveltimes difference of the selected geophone pair ($\Delta t_{ij}$) is too small. Therefore, we perform quality control (QC) on the results by accepting only those values with $Q^{-1} \in [0.01, 0.10]$. This criterion is set according to previous HF-induced attenuation studies (Vera Rodriguez and Stanchits, 2017; Wcislo et al., 2018; Yuan et al., 2019; Zhang et al., 2019; Xing and Zhu, 2021).

Figure 6 shows the P- and S-wave attenuation results for E1 and E2 (Figure 3a and 3b) using the SR and CFS methods. From Figure 6, both of these methods provide similar mean, standard deviation (STD), median, and median absolute deviation (MAD) values of $Q^{-1}$. Moreover, the valid number of $Q_i^{-1}$ (96, 96, 60, and 59) is greater than that of $Q_P^{-1}$ (45, 43, 44, and 41). This is probably because in microseismic field data (e.g., Figure 3) the P-waves are relatively weak and can be easily affected by the background noise. Therefore, the number of $Q_i^{-1}$ results rejected by the QC is greater than that of $Q_P^{-1}$ due to data noise and numerical errors. Furthermore, there are two different ways to calculate the average subsurface attenuation from measurements of individual rays, the arithmetic mean or the median value. From the histograms of frequencies for the calculated $Q_i^{-1}$ values in Figure 6, we find that most of the results are closer to the medians than the means. This...
likely stems from the fact that the mean value is more affected by the maximum and minimum values of a sample than the median value, and $Q^{-1}$ with large values (e.g., 0.080–0.100 in Figure 6) only occupy a small part of the total results. This is consistent with the previous work of Vera Rodriguez and Stanchits (2017), who reveal that the arithmetic mean is always susceptible to a larger approximation error, whereas the median value is relatively stable and can be used to represent the average attenuation. Therefore, we choose the median values to represent the average $Q_{P}^{-1}$ and $Q_{S}^{-1}$ for individual events. From Figure 6, the median $Q_{P}^{-1}$ and $Q_{S}^{-1}$ values for events E1 and E2 are similar (approximately 0.032
0.037). This corresponds with the results in Figures 4 and 5. In addition, the MAD values are not small (0.015–0.022) for P- and S-waves, likely due to the heterogeneous attenuation distribution in the study area, errors in the arrival-time picking, influence of background noise, or contamination by other events or phases. Despite this, we consider that the median $Q^{-1}$ values are relatively reliable and can be used to represent the average attenuation properties because we use as many geophone pairs as possible to compute $Q^{-1}$, and we have a large enough set of valid values of $Q^{-1}$ (defined as $Q^{-1} \in [0.01,0.10]$, with numbers greater than 40 in Figure 6).

**Calculated $Q^{-1}$ results**

The $Q_{P}^{-1}$ and $Q_{S}^{-1}$ values are measured using the SR and CFS methods with data from 109 microseismic events in 12 HF stages. The results shown in Figure 7 reveal that the $Q^{-1}$ values (indicated by black dots in the figure) obtained with the SR and CFS methods are similar, with their differences in absolute $Q^{-1}$ values less than 0.010 for P- and S-waves. This demonstrates the reliability of the measured attenuation. Moreover, the $Q_{P}^{-1}$ and $Q_{S}^{-1}$ values obtained in stage 6 are smaller than those from the other stages, indicating a sudden decrease in attenuation in areas covered by source-receiver rays from events in HF stage 6. Furthermore, the mean, median, median ± MAD, maximum, and minimum $Q^{-1}$ values increase from HF stages 4 to 5, decrease from stages 5 to 6, and increase again through stages 6 to 9 in Figure 7, whereas those from HF stages 1 to 4 and 9 to 12 exhibit no distinct or consistent variance trends. We also observe that mean and median $Q^{-1}$ values from stages 8 to 12 are higher than those from stages 1 to 4.

**Spatial and temporal variations of $Q^{-1}$**

Next, we want to learn more details about the spatiotemporal variations in $Q_{P}^{-1}$ and $Q_{S}^{-1}$ during the entire HF process. As the 109 selected microseismic events occur at different times and places, calculated $Q_{P}^{-1}$ and $Q_{S}^{-1}$ values from these events represent subsurface attenuation at different times and places. By comparing these $Q^{-1}$ values, we can analyze the spatiotemporal variations in subsurface attenuation. Figure 8 illustrates the source–receiver ray coverage of all selected microseismic events. Figure 8a and 8c shows that many seismic rays pass through the study area, with their stages coded by different colors. Therefore, we calculate the average subsurface attenuation using all of the $Q^{-1}$ results obtained from these rays. For HF stages 1 to 12, we discretize the entire space with $10 \times 10 \times 10$ m$^3$ cubes and calculate the average attenuation property of each cube using the following equation:

$$Q^{-1}(x, y, z) = \text{median}[Q^{-1}_i(x, y, z)],$$

where $i$ is the index of a ray and $N$ refers to the total number of rays passing through the cube $(x, y, z)$ (where $x$, $y$, and $z$ denote north–south, west–east, and up–down, respectively), $Q^{-1}_i$ is the calculated $Q^{-1}$ value related to the $i$th ray, and $\text{median}[Q^{-1}_i(x, y, z)]$ denotes the median value of all $Q^{-1}_i$ values. From Figure 2, the 12 HF stages are mainly distributed along the $x$ (north–south) direction. Therefore, for the cube $(x, y, z)$, we combine the $Q^{-1}$ values of its neighboring cubes $(x-1, y, z)$ and $(x+1, y, z)$ to calculate the median value $Q^{-1}(x, y, z)$. This procedure is equivalent to adopting an interpolation along the $x$-direction or assuming that the ray is fat, which increases the $N$ value and improves the calculation of $Q^{-1}$ distributions. During calculation, $N$ is required to be no less than 10. Figure 8b and 8d illustrates the ray number (with values of $\geq 10$) in each cube, that is, the ray density. We see that the maximum ray density is located in areas approximately 800 to 900 m along the north–south direction (the location of the geophone array).

Figure 9 shows spatial distributions of the $Q_{P}^{-1}$ and $Q_{S}^{-1}$ values using the SR (left) and CFS

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**Figure 5.** Similar to Figure 4, except for event E2 (shown in Figure 3e–3f).
(right) methods. The depth and vertical slices are arranged in the same way as those in Figure 8, except the colors here refer to the average $Q^{-1}$ values in individual cubes calculated using equation 7. Compared to Figure 7, Figure 9 illustrates the subsurface P- and S-wave attenuations in different areas more clearly and characterizes the spatial variations in attenuation along depth and vertical slices. From Figure 9, the minimum $Q_P^{-1}$ and $Q_S^{-1}$ values are located near the event locations of HF stage 6 (approximately 800 to 900 m along the north–south direction) and the $Q^{-1}$ values from HF stages 8 to 12 (approximately 1000 to 1400 m along the north–south direction) are relatively larger than those from stages 1 to 4 (approximately 350 to 700 m along the north–south direction). These phenomena are consistent with the observations from Figure 7.

To illustrate the temporal variations in attenuation during HF, we calculate the average attenuation using only the results of HF stages 3 to 4 and 6 to 8. Figure 10 shows the temporal changes in $Q_P^{-1}$ and $Q_S^{-1}$ during these HF stages. The left column in Figure 10 shows that for a given method (either SR or CFS), the $Q_P^{-1}$ and $Q_S^{-1}$ of stages 3 to 4 have similar values, which means that the attenuation of the ray-covered areas may be at the same level. In contrast, the middle and right columns in Figure 10 indicate a clear increase in $Q_P^{-1}$ and $Q_S^{-1}$ values from HF stages 6 to 8. Figure 10 also evidently shows that the attenuations of HF stage 6 are weaker than those of the HF stages 3, 4, 7, and 8, corresponding with the observations from Figures 7 and 9.

**DISCUSSION**

Attenuation is typically not considered in microseismic analysis. Our study provides a sample case study of estimating attenuation during the HF process based on a single-well microseismic monitoring system. Currently, single-well monitoring schemes provide the bulk of downhole microseismic data. Nevertheless, inverting attenuation using single-well data is quite difficult, mainly due to its very limited ray coverage (particularly a lack of crossover rays). Another difficulty is that the attenuation distribution linked to the HF process is intrinsically 4D (space plus time), which further dilutes the data constraints. Other impacts are from the data noise, due to the low microseismic magnitudes, the huge amounts of fractures that continuously happen during the HF process, and noise from the engineering equipment.

We adopt the SR and CFS methods to calculate the $Q_P^{-1}$ and $Q_S^{-1}$ values using 3C waveform records from 109 microseismic events separated in 12 HF stages. We select geophone pairs as long as they have different arrival times to maximize the availability of valid data with $Q^{-1} \in [0.01, 0.10]$. Both methods give comparable results (Figures 4–7 and 9–10). According to the uncertainty analysis of the CFS method (Figures 4 and 5), the variations of $Q$ values corresponding to 50% growth of the misfit function from its minimum value are rather small. Therefore, the obtained subsurface attenuation results during HF stimulation are relatively reliable.

HF is usually conducted in a space-time manner, which increases the difficulty in investigations but also naturally decomposes the data in space and time. Therefore, even with a limited acquisition system, by properly processing the data set, we can still analyze the spatiotemporal variations during the HF process.

According to the results, the P- and S-wave attenuations in HF stage 6 are weaker than those in other stages (see Figures 7, 9, and 10). To check the origin of this observation, we compare the attenuation in the targeted tight sandstone layer (1550–1650 m in depth) with the log data from the horizontal treatment well (Figure 11). According to the results in Figure 7, the $Q_P^{-1}$ and $Q_S^{-1}$ vary within 0.030–0.062 and 0.026–0.055, respectively, indicating that the P- and S-wave attenuations in the targeted tight sandstone layer are significant. This corresponds with previous laboratory attenuation studies on tight sandstones (e.g., Tisato and Quintal, 2013; Chapman et al., 2016). In Figure 11a, a prominent increase in the shale contents can be observed near HF stage 6 (also where the geophone array is). The area where shale content increases is more or less overlapped with the locations of the minimum $Q_P^{-1}$ and $Q_S^{-1}$, indicating that the former may be responsible for the latter. Previous studies (e.g., Xing and Zhu, 2021) also found that the shale has lower attenuation than that of sandstone. However, from Figure 11a, a sudden increase in shale content can also be observed in the areas between HF stages 5 and 6, and between HF stages 6 and 7, but the P- and S-attenuations of HF stages 5 and 7 are higher than those of HF stage 6.

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**Figure 6.** Calculated P- and S-wave attenuations from individual microseismic events E1 and E2 (Figure 3a and 3b). The valid number of geophone pairs with $Q^{-1} \in [0.01, 0.10]$, the arithmetic mean $Q^{-1}$ (the blue lines) and their STDs, and the median $Q^{-1}$ (the red lines) and their MADs are listed at the upper right in each panel, respectively. (a–d) Histograms of frequencies of the P- and S-wave attenuation measurements from events E1 and E2 using the SR method. (e–h) Similar to (a–d) but using the CFS method.
(Figures 7, 9, and 10). From the ray coverage in Figure 8a and 8c, the source-receiver rays of events in HF stages 7 to 12 propagate through areas where HF stimulations have been conducted, whereas those in HF stages 1 to 5 primarily propagate through areas where HF stimulations have not yet been performed. Studies have found that HF stimulation can significantly increase subsurface attenuation (e.g., Wcislo et al., 2018; Zhang et al., 2019). Therefore, the change in the structure and impact of HF stimulations should be simultaneously considered to interpret the attenuation variance.

In Figures 7 and 9–11, the average P-and S-wave attenuations in HF stages 8 to 12 are apparently higher than those in HF stages 1 to 4. Given that the structures are similar (Figure 11a), this clearly reflects different attenuations in fractured and unfractured media. The increase in attenuation from HF stages 6 to 7 (Figure 10) is also probably due to the HF stimulation (because their ray-covered structures are similar, both with high shale content), whereas the sudden decrease in attenuation in HF stage 6 compared to the other stages is probably due to the high shale content (because it is difficult to determine whether the source-receiver rays of events in HF stage 6 pass through fractured areas or not), and the increase in attenuation from HF stages 7 to 8 (Figure 10) is likely due to the joint impact of the structural change (increase in sandstone content) and the HF stimulation.

By comparing the obtained results with the sonic log data and HF activities, reasonable explanations can be provided for most of the observed spatiotemporal attenuation variations. There are some phenomena that remain unexplained, such as the sudden attenuation increase in HF stage 5 (Figure 7), which may be caused by lack of data (only five events in this stage) or other factors such as a large natural fluid flow channel or increasing fracture density. Despite this, the consistency between the $Q^{-1}$ results and sonic log data, as well as the HF stimulation, indicates the reliability and robustness of the subsurface attenuation results from our study. This also demonstrates the importance of attenuation in characterizing the subsurface structure and monitoring the HF process. In addition, the attenuation property provides basic information for microseismic modeling and estimations for microseismic source spectra and magnitudes.

There are still some related issues to be solved. In this study, $Q^{-1}$ is assumed frequency-independent and constant along the raypath. Although current results are useful for characterizing the spatiotemporal variations of attenuation during HF stimulations, a frequency-dependent and spatially variable $Q^{-1}$ model would be physically more realistic. Moreover, a formal tomographic inversion may be more appropriate to investigate the spatial variations in $Q^{-1}$. To constrain a more complex attenuation model using single-well monitoring data, a more sophisticated inversion method, for example, the twofold SR method (Matsuzawa et al., 1989; Yamada and Oda, 2019) may be considered. In addition, the attenuation could be composed of intrinsic and scattering attenuations. Further research is needed to investigate which mechanisms actually contribute to the observed attenuation variations. Because the multicomponent elastic waveform data provide P-
Figure 9. Spatial distributions of P- and S-wave attenuations. (a and b) Map views of the $Q_{P}^{-1}$ spatial distributions using the estimation results of the SR and CFS methods, respectively. (c and d) Side views of the $Q_{P}^{-1}$ spatial distributions using these two methods. (e-h) Similar to (a-d) but for $Q_{S}^{-1}$.

Figure 10. Temporal variations of P- and S-wave attenuations in several HF stages. Panels (a-d) are attenuation distributions using the results of the SR and CFS methods for the P- and S-waves, respectively. Shown in different panels are vertical slices along north–south depth directions. Left, central, and right columns indicate the results from stages 3 to 4 ($S3\rightarrow S4$), 6 to 7 ($S6\rightarrow S7$), and 6 to 8 ($S6\rightarrow S7\rightarrow S8$), respectively.
Figure 11. Log data of horizontal treatment well and the $Q^{-1}$ estimation results of targeted tight sandstone layer. (a) Horizontal well-log data. The green, orange, and gray colors indicate the relative contents of shale, sandstone, and porosity, respectively. The red and blue lines refer to the variation curves of acoustic (AC) and natural gamma-ray (GR) data, respectively. (b–e) North–south depth slices of estimated $Q^{-1}_P$ and $Q^{-1}_S$ values in the targeted sandstone layer using the SR and CFS methods.

and S-wave observations, investigating relations between the P- and S-wave attenuations and their links to other material properties are also possible.

CONCLUSION

In this study, we adopt the SR and CFS methods to estimate subsurface seismic attenuation during an HF stimulation case in a tight sandstone reservoir in the Ordos Basin, China. The P- and S-wave attenuations are measured using a 3C microseismic waveform data set obtained by a vertical downhole monitoring array. To maximize the available data, we select geophone pairs with different arrival times and conduct QC on the calculated results. During attenuation measurement using records from individual events, we find that the median values of $Q^{-1}_P$ and $Q^{-1}_S$ are more appropriate for representing the average attenuation than arithmetic means. Using records from 109 microseismic events in 12 HF stages, we analyze the subsurface attenuations and their spatiotemporal variations. The $Q^{-1}_P$ and $Q^{-1}_S$ values obtained using the SR and CFS methods exhibit good consistency with each other. According to the results, the attenuation effect of the targeted tight sandstone layer is relatively strong (approximately 0.030–0.062 for $Q^{-1}_P$ and 0.026–0.058 for $Q^{-1}_S$), and the HF stimulations apparently increase the P- and S-wave attenuations. These results are consistent with previous studies on HF-induced attenuation changes. Moreover, to investigate detailed attenuation distributions, we calculate median $Q^{-1}$ values along the seismic rays. This enables us to characterize the spatial distributions and temporal variations of subsurface attenuations during the entire HF process. By linking the $Q^{-1}_P$ and $Q^{-1}_S$ with the horizontal well-log data, we explain the prominent decrease in $Q^{-1}$ values in some areas as a result of higher shale content. The consistency between the attenuation and the horizontal well-log data, as well as the HF process, indicates the reliability of the results. Due to the limitations of the single-well acquisition system, high background noise, and complicated spatiotemporal distribution of the microseismic events, the current investigation is preliminary. Despite this, our study provides a useful example of microseismic attenuation measurements and how to apply them in characterizing the variations of underground structures during HF activities.

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DATA AND MATERIALS AVAILABILITY

Data associated with this research are confidential and cannot be released.

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Biographies and photographs of the authors are not available.