Long-period, long-duration seismic events during hydraulic stimulation of shale and tight-gas reservoirs — Part 1: Waveform characteristics

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ABSTRACT

Long-period long-duration (LPLD) seismic events are relatively low-amplitude signals that have been observed during hydraulic fracturing in several shale-gas and tight-gas reservoirs. These events are similar in appearance to tectonic tremor sequences observed in subduction zones and transform fault boundaries. LPLD events are predominantly composed of S-waves, but weaker P-waves have also been identified. In some cases, microearthquakes are observed during the events. Based on the similarity with tectonic tremors and our observations of several impulsive S-wave arrivals within the LPLD events, we interpret the LPLD events as resulting from the superposition of slow shear-slip events on relatively large faults. Most large LPLD waveforms appear to start as a relatively slower, low-amplitude precursor, lacking clear impulsive arrivals. We estimate the energy carried by the larger LPLD events to be ~1000 times greater than a ~MW = 2 microseismic event that is typical of the events that occur during hydraulic stimulation. Over the course of the entire stimulation activity of five wells in the Barnett formation (each hydraulically fractured ten times), the LPLD events were found to cumulatively release over an order of magnitude higher energy than microearthquakes. The large size of these LPLD events, compared to microearthquakes, suggests that they represent slip on relatively large faults during stimulation of these extremely low-permeability reservoirs. Moreover, they imply that the accompanying slow slip on faults, probably mostly undetected, is a significant deformation process during multistage hydraulic fracturing.

INTRODUCTION

Although horizontal drilling and multistage hydraulic fracturing are successfully used to stimulate production from shale-gas, tight-gas, and tight-oil reservoirs, the principal deformation mechanisms responsible for the stimulation of production from these extremely low-permeability formations are still poorly understood. It has been well documented that, during multistage horizontal fracturing, a “cloud” of microseismic events surrounding the hydraulic fractures indicate the occurrence of shear slip on preexisting fractures and faults distributed around the induced hydraulic fractures (Rutledge and Phillips, 2003; Warpinski et al., 2004). However, the correlation between the cloud of microseismic event and the stimulated rock volume is still questioned (Sicking et al., 2013). In a case in which the microseismic cloud was fairly planar, Moos et al. (2011) showed that the number of microseismic events did not correlate with production from successive hydraulic fracturing stages.

The motivation of this paper is to investigate the other sources of deformation that occur in these unconventional reservoirs during hydraulic stimulation. In a preliminary study of seismic waveforms recorded at depth during multistage hydraulic fracturing in five parallel wells in the Barnett shale, Das and Zoback (2011) documented the occurrence of long-period long-duration (LPLD) seismic events. The observed LPLD events have similarities to seismic events known as tectonic tremors observed in subduction zones and transform boundaries (Southwest Japan, Cascadia, San Andreas Fault; Obara, 2002; Shelly et al., 2006; Nadeau and Guilhelm, 2009). Several authors (Rogers and Dragert, 2003; Obara et al., 2004; Hirose and Obara, 2010; Bartlow et al., 2011) have shown that tremor accompanies slow shear-slip events, in some cases independently seen geodetically, localized in the transition zone between the locked and sliding portion of major plate-boundary faults such as subduction thrusts and the San Andreas fault. Das and Zoback (2011) suggested that the LPLD signals observed provide evidence of similar
Deformation mechanisms happening during multistage hydraulic stimulation of gas fields. Zoback et al. (2012) expanded on this interpretation, and provides reasons why slow slip on preexisting faults appears to be an important deformation process during multistage hydraulic fracturing and could potentially contribute appreciably to reservoir production.

Geiser et al. (2012) observe seismic signals that differ from conventional microearthquakes during hydraulic stimulation. By stacking the energy from these signals over periods of minutes to hours, they have identified linear sources of low-frequency energy, which they have interpreted to be natural fracture pathways.

Eaton et al. (2013) undertook a study in the Canadian Montney gas reservoir looking at low-frequency signals generated during hydraulic fracturing and were able to observe signals which resembled LPLD events, although they were not so numerous or energetic as those Das and Zoback (2011) identified in the Barnett. LPLD events were also observed in a multistage hydraulic fracture stimulation project in the Cardium formation in west central Alberta (A. St-Onge, personal communication, 2013).

In this paper, we discuss the characteristics of the LPLD events recorded during passive seismic monitoring of multistage hydraulic fracturing projects in four different case studies. We look in detail at LPLD events in two of these data sets from the Barnett Shale in Texas and illustrate their occurrence in a tight-gas sand reservoir in Canada and a shale-gas reservoir in Canada. In a companion paper (Das and Zoback, 2013), we explore in detail where, when, and why the LPLD events occur in a reservoir during hydraulic stimulation, and the physical processes responsible for their occurrence.

DATA SETS ANALYZED

The first data set analyzed in detail here was initially described by Das and Zoback (2011) and is henceforth referred to as Barnett Data 1. It consists of five subhorizontal wells, A, B, C, D, and E, spaced roughly 150-m apart from each other (Figure 1), all within a Barnett shale layer approximately 90-m thick. As described by Vermilyen and Zoback (2011), about 10 stages of slick-water hydraulic fracturing were carried out in each well. Wells A and B were fractured using a “simul-frac” method and recorded by a single array consisting of nine 3C sensors in well C as shown in Figure 1. Wells D and E were fractured using a “zipper-frac” method and recorded by a single array of 10 3C sensors, again in well C. Well C was fractured from toe to heel in a conventional manner and recorded by an array of 12 sensors deployed in the vertical section of well B, with the deepest sensor slightly above the Barnett shale. The total pumping volume (approximately 325,000 gallons), pumping rates (approximately 55 bpm), and amount of proppant (approximately 400,000 lbs) were nearly the same in every fracturing stage. Thus, twice as much pumping was done during the simul-fracs.

When the array was in the horizontal section of well C, one component of the three-axis geophones was oriented axially along the borehole (the axial component) while the other two were arranged orthogonal to each other and the well axis (the transverse 1, horizontal component, and the transverse 2, vertical component). When the array was in the vertical section of well B during hydraulic fracturing of well C, the horizontal sensors were rotated to north and east. The sensors in the horizontal arrays were separated by ∼15 m for the simul-frac and ∼12 m for the zipper-frac and the fracturing in well C.

The entire network was operational for about a month, although it was recording only during the fracturing jobs, which typically lasted ∼2.5 hours. The microseismic vendor located about 4500 microearthquakes during the monitoring period, with reported magnitudes between −1 and −4 (Figure 1). The actual number of microearthquakes that occurred is much larger, but most were not located due to their small size and the inability to identify clear P- and S-waves when the signal-to-noise ratio was too low.

The microseismic events are plotted with their size proportional to their magnitude and colored according to the fracturing stage they were recorded in. The variability in their locations and magnitudes is readily apparent in Figure 1, despite the relative uniformity of the hydraulic fracturing process. The area near stages 6–9 of wells A and B has higher density and larger events than near the toe. In fact, the stages near the toes of all the wells are practically devoid of microearthquakes. This suggests that different parts of the reservoir respond differently to the stimulation. In our companion paper (Das and Zoback, 2013), we suggest possible reasons for this.

The second data set studied in detail also comes from the Barnett Shale (Barnett Data 2). It consists of two horizontal wells fractured in 8 and 10 stages, respectively (Figure 2). The hydraulic fracturing was recorded by vertical arrays in two different vertical wells, each with 40 3C geophones. The geophones in both arrays were each separated by 15 m such that each array is roughly 600-m long. The deepest geophones in recording wells 1 and 2 are ∼450 m and 300 m, respectively, above the two horizontal treatment wells. Approximately 500,000 gallons of fluid was pumped at 100 bpm at every stage. There were roughly 7000 microearthquakes located by the microseismic vendor ranging in magnitude from −0.5 to −3. The reported microseismic events are plotted as in Figure 1, colored
LPLD events: Waveform characteristics

CHARACTERISTICS OF LPLD EVENTS

LPLD events were first identified in stacked spectrograms — sum of the nine geophones of each of the three components — the axial and two orthogonal transverse components (Das and Zoback, 2011). In Figure 3a, bursts of energy at frequencies between 10 and 80 Hz that persist for 10–100 s can be seen in stacked spectrograms for two contiguous stages of Barnett Data 1. This is shown more clearly in Figure 3b, with an expanded time scale. Figure 3c shows one set of LPLD events that corresponds to the spectrogram in Figure 3b. Black lines in Figure 3a show the pumping pressure records. Note that some LPLD events occur after pumping ends in stage 7 and before pumping begins in stage 8, including those shown in Figure 3b. This is not surprising considering that elevated pressure in the reservoir persists after a pumping cycle ends.

Figure 4 shows an example of an LPLD event that was recorded during stimulation in Barnett Data 2. It is clearly identified on both recording arrays. They are similar, in temporal and spectral characteristics, to events observed in Barnett Data 1 (Figure 3). The LPLD events in Barnett Data 2 have a slightly lower maximum frequency (~60 Hz compared to ~80 Hz) than in Barnett Data 1. This might be partly due to the larger distance between the recording array and the events in Barnett Data 2 than in Barnett Data 1, which allows further attenuation of high frequencies. Typical recording distances in Barnett 1 are 100–300 m, whereas in Barnett Data 2, they are 600–900 m.

Figure 5a shows the presence of LPLD events during hydraulic fracturing in the Horn River area of Canada and Figure 5b shows an LPLD event in a Canadian tight gas sand. LPLD events in both of these data set are not as energetic or numerous as in the two Barnett data sets. As a consequence, they are typically hard to identify. They also exhibit much lower frequencies than the previous two data sets. Because distances between the recording arrays and events are not dramatically different between these two and the previous two data sets, it is possible that the formation properties have a role to play in the frequency content of LPLD events. Eaton et al. (2013) argues that the absence of a complex fracture network in their Canadian gas-shale study might be responsible for the smaller number and lower intensity of LPLD events. Nevertheless, the fact that LPLD events have been detected in a range of geologic environments during hydraulic stimulation suggests that they are a general phenomenon occurring during stimulation. Apparently, LPLD events have been overlooked during routine microseismic processing of hydraulic stimulation data sets due to their low amplitude, absence of clear impulsive arrivals, and noise-like appearance.

As pointed out by Das and Zoback (2011), the LPLD event waveforms (Figure 6d, 6e, and 6f) are similar in appearance to tectonic tremors (also known as nonvolcanic tremors) observed in subduction zones and strike-slip margins (Figure 6a, 6b, and 6c), although tectonic tremors are typically observed between 1 and 15 Hz, a frequency band that is not detectable with the 15 Hz geophones used for microseismic recording in the four data sets reported here. The similarity is striking despite the fact that tectonic tremors occur at depths ranging from 20 to 45 km (Obara, 2002; Shelly et al., 2006, 2007b; La Rocca et al., 2009; Shelly and Hardeback, 2010) and is...
recorded by surface seismometers spaced tens of kilometers apart, whereas the LPLD events reported here occur at depths of 2–3 km and are recorded by downhole arrays only a few hundred meters from the events. In the next section, we compare LPLD events, microseismic events, and tectonic tremors in the time and frequency domains.

TEMPORAL AND SPECTRAL ANALYSIS

Shelly et al. (2006) found that within tremor there are impulsive arrivals moving coherently across the stations with the velocity of S-waves (Figure 7a). P-wave arrivals, although too weak to be visible above the noise, were also detected using crosscorrelation. These low-frequency impulsive arrivals are referred to as low-frequency earthquakes or LFEs (Figure 7a). Several authors (Ide et al., 2007; Shelly et al., 2007a; Brown et al., 2009; La Rocca et al., 2009) have demonstrated that tremors can be explained as a swarm of such small LFEs, each of which occurs as shear faulting on the subduction-zone plate interface. Figure 7b is an example of an LPLD event recorded by three adjacent geophones in a single recording well from Barnett Data 2. Just like the LFEs observed within tremor, we can see clear coherent impulsive arrivals within LPLD events. These arrivals move across the array with velocities that are consistent with shear and compressional velocities in the reservoir.

Figure 8a shows the vertical component of 1.5 s of an LPLD event recorded in the Barnett Data 2. Two distinct phases can be seen clearly moving across the array with slightly different moveouts. The apparent velocity corresponding to arrival 1 is 4.9 km/s while that of arrival 2 is 2.9 km/s. From the equation

\[ \text{Angle of arrival} = \cos^{-1}\left( \frac{v_{\text{formation}}}{v_{\text{apparent}}} \right) \]  

(1)

it is clear that \( v_{\text{apparent}} \) can never be less than \( v_{\text{formation}} \). However, the lowest formation P-wave velocity from well logs is 3.6 km/s. Hence, arrival 2 with \( v_{\text{apparent}} \) of 2.9 km/s is an S-wave, whereas arrival 1 is a P-wave. This fact is illustrated clearly in Figure 8b, which shows the same 1.5 s of the LPLD event in all three components, but after correcting for the S-wave moveout. We find that the S-wave is strongly aligned in all three components and the P-wave is overcorrected. Using average P- and S-wave \( v_{\text{formation}} \), we get similar angles of arrival for arrival 1 and 2, respectively, demonstrating that they are coming from approximately the same area, which is reasonable.
considering that they occur in the same LPLD event.

Figure 8c is a wider view of Figure 8b, showing that after correcting for the S-wave moveout, we can see several similar strong S-wave arrivals and few weak P-wave arrivals. From this experiment, it appears that the LPLD signals are the result of superposition of numerous small shear events, all of them occurring in close proximity to each other. In most cases, the waveforms of individual events overlap and are hard to differentiate. The P-wave is visible above the background only in some very energetic LPLD events and when the LPLD event is far enough from the recording array to allow a clear separation between the P- and S-waves. In all the above aspects, LPLD events are very similar to tectonic tremors, which comprise superposed LFE waveforms.

A discerning characteristic of majority of large (40–80-s duration) LPLD event waveforms in both Barnett data sets is a precursory seismic event that occurs at the beginning of the LPLD event. It is evident from the seismograms in Figure 9, shown serially from left to right on an expanded time scale, that each LPLD event is comprised of a relatively low-amplitude precursor wave train of approximately 10–15 s duration, followed by several high amplitude impulsive S-wave arrivals lasting for 10–15 s, gradually diminishing in amplitude to the level of the precursor. This characteristic is also apparent from the corresponding spectrograms shown in the same figure. There is a distinct precursor of slightly lower frequency and low amplitude followed by the onset of a relatively higher-amplitude and higher-frequency signal, which finally dies down to the background level. One explanation for this typical waveform characteristic might be a relatively slow, low-amplitude rupture in the beginning, which abruptly accelerates to higher rupture speeds before decelerating and finally coming to a stop, assuming that the higher rupture velocity leads to higher frequency and greater radiated seismic energy.

The precursor waveforms do not have any distinct or coherent arrivals, making it difficult to identify phase velocities. Like the LPLD events themselves, the precursor events are probably a combination of weak P- and S-waves.

We also found evidence of microearthquakes occurring within LPLD events in addition to low-frequency impulsive arrivals. Figure 10a shows an LPLD event from Barnett Data 1, visible in the filtered data. Figure 10b and 10c shows 0.3 s of the same event on an expanded scale, filtered (10–80 Hz) and unfiltered, respectively. In Figure 10b, we can see the low-frequency impulsive arrivals discussed above, while in Figure 10c we can clearly see a microearthquake with distinct P- and S-waves. This microearthquake is not visible in Figure 10b because of the filtering. As shown by the dotted red lines, the moveout of the two different kinds of events are similar, suggesting that they might be coming from identical faults. This is shown to be the case in Das and Zoback (2013).

Figure 11a shows the velocity spectrum of an LPLD event from Barnett Data 2 compared with the spectra of the largest microearthquake ($M_w \sim -0.5$), a slightly smaller microearthquake...
(MW ∼−1) and background noise from the same data set. The instrument response of the recording geophones is also shown above the spectrum plots. Figure 11b shows the spectra of tectonic tremors and LFEs from Western Shikoku in Japan compared with two regular earthquakes and background noise (Shelly et al., 2007a). Although the range of frequencies in the two figures differs by almost an order of magnitude, the spectral characteristics of the LPLD event and tectonic tremors are similar and distinctly different from the microearthquakes (Figure 11a) and regular earthquakes (Figure 11b), respectively. Compared to them, the LPLD events and tectonic tremors are enriched in low-frequency energy and depleted in high-frequency energy.

It should be noted that tectonic tremors and LPLD events are fundamentally different from volcanic tremors (Julian, 1994), which are generally believed to be caused by oscillations of fluid-filled conduits transporting magmatic fluids or fluid-filled cracks. The spectra of volcanic tremors are usually dominated by one or more fundamental frequencies and their harmonics (Riuscetti et al., 1977; Konstantinou and Schlindwein, 2002) whereas tectonic tremor and LPLD event spectra, as shown above, are characteristically flat at low frequency and fall off in the manner of the spectra of regular earthquakes (Aki, 1967; Brune, 1970).

SIZE OF LPLD EVENTS AND THE ENERGY BUDGET

In this section, we compare the energy released by the microseismic and the LPLD events with the stored elastic energy in the volume of rock affected by stimulation. The total elastic strain energy in an earthquake (or slow slip event) is partitioned in several different ways. Figure 12, adapted from Kanamori and Rivera (2006), shows three simplified models of energy partitioning. Here, EF is the interface frictional energy and can be equated to the energy dissipated as heat, and EG is called the rupture energy. It is essentially the sum of all kinds of energies associated with faulting, other than that due to interface friction. It can include the fracture energy, energy dissipated beyond the crack tip due to anelastic processes, energy used to create off-fault cracks, and thermal energy involved in melting and fluid pressurization. The radiated seismic energy is denoted by ER. The sum of EG and ER is the energy expended during rupture propagation in addition to frictional dissipation. LPLD events probably lie somewhere in between regular earthquakes and slowly slipping faults. This implies that measuring the radiated seismic energy is arguably not a good method for estimating the elastic strain energy released by microearthquakes and LPLD events.

Seismic moment MO (defined as the product of the average displacement, the fault area, and the shear modulus) is a static or absolute measurement of the size of the earthquake, irrespective of how fast or slow the earthquake occurred. Seismic moment is also the only parameter that can be reliably measured from the low-frequency amplitude level of the displacement spectrum. According to Kanamori and Anderson (1975), there is a relation between
the change in elastic strain energy and the seismic moment given by
the equation
\[ \Delta W = \frac{M_0 \bar{\sigma}}{\mu}, \]
where \( \Delta W \) is the elastic strain energy released during the event, \( M_0 \) is the seismic moment, \( \bar{\sigma} \) is the mean stress, and \( \mu \) is the rigidity. As we are concerned with a particular reservoir, \( \bar{\sigma} \) and \( \mu \) will have the same range of values for the events considered. For the two Barnett reservoirs, \( \bar{\sigma} \) and \( \mu \) are in the order of tens of MPa and tens of GPa, respectively. This indicates that the elastic strain energy is roughly a thousandth of the seismic moment.

Zhang et al. (2011) used the difference between the low-frequency amplitudes of tremors and of regular earthquakes of known moment to estimate the tremor moment. We follow the same methodology, but instead of the difference, we use the low-frequency spectral ratio of LPLD events to microearthquakes of known magnitude to estimate the moment of LPLD events. This is somewhat similar to the empirical Green’s function (EGF) method (Frankel and Kanamori, 1983; Mori and Frankel, 1990; Venkataraman et al., 2002). In this method, a small earthquake located close to a larger earthquake is used as an EGF. The EGF is deconvolved from the larger earthquake to remove the path and site effects and generate a clear source spectrum for the large earthquake. In the frequency domain, this is just a simple division of the two amplitude spectra, and hence we call it the spectral ratio. The deconvolution assumes that the larger earthquake and smaller earthquake (LPLD event and microearthquake, in our case) are located close to each other such that the path effects are almost the same for both events, which is a reasonable assumption for this data set.

Figure 9. Top panels show the filtered seismograms of an LPLD event with progressively increasing time scales. Arrows delineate the low-amplitude precursor and the high-amplitude main event within a single LPLD event. Bottom panel shows the corresponding spectrograms with low-amplitude, low-frequency, precursor followed by the high-amplitude, relatively high-frequency main event.
than the LPLD event; hence, it is an ideal candidate for the EGF. In addition, as explained in Das and Zoback (2013), this microearthquake is in the general area in the reservoir where the LPLD event is located, again satisfying the criteria of an EGF.

In Figure 13d, we divided the low-frequency displacement amplitude spectrum of the LPLD event with that of displacement spectrum of the microearthquake shown in Figure 13c to obtain the moment of the LPLD event relative to the moment of the microearthquake, which is known. Zhang et al. (2011) demonstrated that the displacement spectra of the Cascadia tremors have corner frequencies around 3–8 Hz and fall off at $f^{-2} - f^{-3}$ at higher frequencies essentially agreeing with the observations of regular earthquakes. Therefore, they suggest that tremors could be analyzed using standard spectral models. Although we cannot be sure at this point whether tectonic tremors and LPLD events are generated by similar mechanisms, due to the apparent similarity in waveforms (Figure 6) and in spectra (Figure 11), we consider it reasonable to use the standard models in our calculations, too. Therefore, we fit the resultant spectrum after spectral division with a Brune (1970) source model to determine the exact value of the relative moment and also the corner frequency. For this particular 1-s time window shown in Figure 13a and 13b, we estimate a spectral ratio of $\sim 8$ which corresponds to a moment release rate of $2.5 \times 10^8 \text{N} \cdot \text{m/s}$ for the LPLD event (Figure 13d). To get the cumulative moment of the entire LPLD event, either we can take the summation of the spectral ratios from each 1-s windows for the entire duration of the LPLD event or we can take the area under the displacement seismogram filtered from 10-100 Hz (98% of the LPLD event energy). Both methods give very similar results. From either method, we get a cumulative moment of about $3 \times 10^9 \text{N} \cdot \text{m}$. Using Hanks and Kanamori (1979),

$$M_W = \frac{2}{3} (\log_{10} M_O - 9.1),$$

we calculate that this moment is equivalent to a $M_W \sim 0.3$ earthquake. It implies that a large LPLD event has more than 1000 times the energy of a typical $M_W \sim 2$ microearthquake.

As mentioned earlier, approximately 4500 microearthquakes were reported in Barnett Data 1 ranging from $M_W \sim -1$ to $-4$. In Barnett Data 2, there are about 7000 microearthquakes in the magnitude range from $-0.5$ to $-3$. This indicates that just a few large LPLD events in each data set have more energy than all microearthquakes recorded during the entire stimulation. Knowing the moments (approximately) of LPLD events and microearthquakes, we can do a quantitative analysis of the energy budget during hydraulic stimulation.

Maxwell et al. (2008), Maxwell (2010), Cioppa et al. (2011), and Warpinski et al. (2012) have already pointed out that the cumulative energy of microearthquakes is a small fraction of the total injection energy and that slow, tensile opening and expansion of the hydraulic fracture accounts for roughly 15%–20% of the injection energy.

**Figure 10.** (a) LPLD event observed during stage 7 of the simulation in Barnett Data 1. (b) A 0.3-s window of the filtered LPLD waveform in (a) shown on an expanded time scale. Clear coherent low-frequency impulsive arrival can be seen moving across the array shown by red dotted line. (c) Same time window as in (b), but without any filter applied. A microearthquake with clear P and S phases, distinctly different from the low-frequency arrivals in Figure 9b, but with similar moveout, is found to occur within the LPLD event.

**Figure 11.** (a) Velocity spectrum of a large LPLD event of duration 50 s from Barnett Data 2 (red solid line) compared with the spectrum of a $M_W = -0.5$ microearthquake (blue dotted line) and a $M_W = -1.0$ microearthquake (green dotted line), both $\sim 450 \text{ m}$ away from the recording geophones used for the calculation. The brown dotted line shows the background noise level. A multitaper spectral method was used to calculate the spectrum and a 1-s window was used in all cases. The instrument response of the 15-Hz recording geophones is shown by the black dotted line on top. (b) Velocity spectrum of tectonic tremor (black solid line), LFEs (gray solid line) and neighboring regular earthquakes greater than $M_W > 2.5$ (black dotted line) or $M_W < 2.5$ (gray dotted line), all from western Shikoku in Japan (Shelly et al., 2007a). The thin gray dotted line shows the noise spectrum. Here, 2.5-s windows were used for all the spectrum calculations.
It is also important to consider the microseismic and LPLD events in terms of the relative amounts of elastic energy stored in the rock that they release. The amount of elastic strain energy in a volume of rock affected by stimulation that is available for shear slip on faults can be estimated from the deviatoric stresses in the rock from the following equation (eq. –5.151 in Jaeger et al., 1969):

$$E_d = \frac{1}{4G}[(s_1^2 + s_2^2 + s_3^2)]_s,$$  

where $E_d$ is the deviatoric strain energy density, $G$ is the shear wave modulus of the rock, and $s_1$, $s_2$, and $s_3$ are the deviatoric stresses. Note that we have used the effective stresses to calculate the deviatoric stresses. For Barnett Data 1, we estimate the deviatoric strain energy density to be $4.65 \times 10^3 \text{ J/m}^3$ based on stress magnitudes and a value of 30 GPa for $G$ based on observations of Sone (2012). From Figure 1, a conservative estimate of the volume of the rock affected by stimulation for Barnett Data 1 is $1000 \text{ m} \times 700 \text{ m} \times 100 \text{ m}$. This gives us a total deviatoric stored energy of 335 GJ. This amount of energy could potentially be released in the form of shear slip events during stimulation.

Summing up the moments of all the reported microearthquakes yields $3.5 \times 10^9 \text{ N} \cdot \text{m}$. Using equation 2 (Kanamori and Anderson, 1975), we estimate the elastic strain energy released by all microearthquakes together to be in the range of $3.5 \times 10^9 \text{ J}$. Microearthquakes that were not reported due to poor signal-to-noise ratio are not expected to contribute significantly to the total moment because of their small size, as discussed previously. The sum of all LPLD moments is about $150 \times 10^9 \text{ N} \cdot \text{m}$, which corresponds to energy in the range of $150 \times 10^9 \text{ J}$. This is a conservative estimate because we likely weren’t able to detect all the LPLD events because of the instrument response of the 15-Hz sensors and overall low signal-to-noise ratio of the events. In addition, we only estimated the moment from the portion of LPLD events that had distinct body wave phases. Nonetheless, the calculations tell us that for the entire Barnett Data 1, LPLD events released about 40 times as much energy as the microearthquakes. Yet the LPLD events released only about 0.05% of the total deviatoric strain energy stored in the reservoir prior to stimulation, compared with 0.001% released by the microearthquakes.

In Barnett Data 2, the total deviatoric strain energy in the volume of the reservoir affected by the stimulation is estimated to be $285 \times 10^9 \text{ J}$ using the same calculations as above. Nearly 50 LPLD events were identified in the entire operation, which is fewer than the LPLD events identified in the Barnett Data 1. Reported microearthquakes, on the other hand, were almost double the number (and larger in size) than in Barnett Data 1. From the moment calculations for this data set, it was found that the microearthquakes released about 0.006% of the total deviatoric strain energy whereas LPLD events released about 0.03%.

Figure 13d also shows that the estimated corner frequency of the LPLD event is 15 Hz. The range of corner frequency for all LPLD events observed in the two data sets is from 10 to 20 Hz. Using Madariaga (1976),

$$r = \frac{0.21\beta}{f_c},$$

where $\beta$ is the shear-wave velocity and $f_c$ is the corner frequency, we get the radius of the fault patch to be in the range of 25–50 m for a corner-frequency range of 20–10 Hz. Using the Brune (1970) model would give us 1.76 times higher patch radius. However, the calculation is based on the assumption that rupture velocity is 0.9, which is most likely not true for LPLD events. Hence, this estimated patch size should be treated as the upper limit. The microearthquakes with the range of magnitudes observed in the two data sets (–0.5 to –4) will rarely have corner frequencies below 100 Hz even with stress drop as low as 0.1 MPa (Allman and Shearer, 2009). Using the same equation 5, we get an upper limit of the patch size radius of 4 m with the majority having radii less than a meter. Thus, compared to microearthquakes, LPLD events are associated with fault patches that are an order of magnitude larger.

The estimates of corner frequencies of LPLD events (10–20 Hz) are much lower than would be seen for regular earthquakes of comparable seismic moment (Allman and Shearer, 2009). This observation is consistent with those of Zhang et al. (2011), who found that tremors and LFEs also have much lower corner frequencies for comparable moment. Two factors that might contribute to such low corner frequencies are very low stress drops and very low rupture velocities. Using the stress-drop relation for a circular crack according to Eshelby (1957),

$$\Delta \sigma = \frac{7}{16} \times M_0 \times \left( \frac{2 \times \pi \times f_c}{2.34 \times \beta} \right)^3,$$

where $\Delta \sigma$ is the stress drop, we get values ranging from 0.2 to 5 kPa for the LPLD events, much smaller than the 0.1–10 MPa of conventional earthquakes (Abercrombie and Leary, 1993; Allman and Shearer, 2009). Low stress drop is most likely indicative of low effective stress during rupture (Shelly et al., 2006; Audet et al., 2009), which is consistent with the fact that LPLD events occur in regions of high fluid pressure during pumping, another topic addressed in Das and Zoback (2013).

![Figure 12. Simplistic models of energy partitioning during earthquakes; $E_R$ is seismic radiation energy, $E_U$ is rupture energy, and $E_F$ is interface frictional energy. (a) Constant friction model, in which fault stress drops to final value instantly and all the available energy is radiated as seismic energy. (b) Slip weakening model in which a portion of the available energy is used for rupture propagation, the rest is radiated as seismic energy. (c) Quasistatic model in which rupture is so slow that there is no radiated seismic energy.](image-url)
DISCUSSION

Based on the frequency content of microearthquakes and assuming a kinematic rupture model, Maxwell and Cipolla (2011) estimated that the total microseismic slip area is about 0.4% of the total area of one modeled planer hydraulic fracture, whose dimensions were assessed from the microseismic locations. Thus, microseismic events that are routinely observed and located during multistage hydraulic fracturing operations clearly do not represent all the deformation modes of hydraulic fracturing. We have shown that in some cases, LPLD events, by themselves, account for almost two orders of magnitude higher energy than the cumulative energy of all microearthquakes. More importantly, an order of magnitude larger patch radius associated with LPLD compared to microearthquakes implies two orders of magnitude larger area of failure. This implies there is possibly more permeability enhancement during stimulation due to LPLD events than due to microearthquakes. 

Knowing that the microseismic and LPLD events release only a small fraction of the stored elastic strain energy, it might be possible to refracture the reservoir multiple times to activate the faults responsible for slow slip to enhance the productivity.

We have also shown that LPLD events are similar, in temporal and spectral characteristics, to tectonic tremor, which is interpreted to result from the superposition of slip on numerous asperities driven by aseismic slip on the surrounding larger fault (Rogers and Dragert, 2003; Ito et al., 2007; Bartlow et al., 2011). Presumably, a large part of remaining energy deficit is accounted for by slow slip events associated with LPLD events that are undetected due to the geophone instrument response. Baig et al. (2012) and Viegas et al. (2012) have shown how the low-frequency response of 15-Hz geophones can truncate and even completely mask larger seismic events. They also point out how the signals in the time domain are distorted relative to clear discrete arrivals observed for the same event when using lower frequency geophones. Thus, with the current recording instruments typically used in the types of project studied here, we are probably seeing an incomplete picture of the LPLD events and the entire suite of larger deformation processes associated with hydraulic stimulation.

CONCLUSIONS

We discuss LPLD seismic events observed in microseismic data sets from hydraulic fracturing operations in horizontal wells in several geologic environments. These LPLD events typically last for tens of seconds and are most conspicuous in the 10–80-Hz band, in which they are similar to tectonic tremors observed routinely in subduction zones and strike-slip margins. LPLD events are comprised of low-amplitude but coherent impulsive arrivals with finite moveouts. Apparent velocity calculations establish that they are predominantly comprised of S-waves, although weaker P-waves have also been identified. Sometimes microearthquakes are also found within the LPLD events. It appears that LPLD signals are a result of the superposition of numerous slow slip events in close succession of each other. The spectra of LPLD events are enriched in low frequencies and depleted in high frequencies. Spectral ratio analysis indicates that the moment carried by a large LPLD event is more than 1000 times that of a typical $\text{MW} - 1$ earthquake. Energy balance calculations shows that, for both data sets, the cumulative LPLD event energy is one to two orders of magnitude higher.

Figure 13. (a) Filtered velocity seismograms of a large LPLD event and (b) unfiltered velocity seismogram of a $\text{MW} - 1$ earthquake. Black dotted lines show 1-s windows selected for the generating displacement spectra in (c) The result of dividing the displacement spectrum of the LPLD event with that of the microearthquake is shown in (d). The microearthquake acts as the empirical Green’s function. The resultant spectral ratio, corner frequency, and spectral fall off are $\sim 8$, 15 Hz, and $-3.7$, respectively.
than the cumulative energy of microearthquakes. Further, slow slip on large-scale faults, which is inferred to accompany LPLD events, appears to be a significant process during hydraulic stimulation.

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