A Study of Mine-Related Seismicity in a Deep Longwall Coal Mine

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ABSTRACT

This study involves seismic monitoring of a deep coal mine. The purpose is to examine the processes responsible for induced seismicity. A seismic network consisting of five three-component short-period seismometers located above the mine recorded the seismic data. The events discussed here occurred from March 1, 2009 until April 7, 2011 during the mining of three longwall panels and the data was telemetered to Blacksburg, Virginia.

A correlation equation was developed to relate local magnitude estimated by automatic data processing software in near real-time to seismic moment for well-recorded events. Local magnitude is a relative measure of relative size for a suite of earthquakes, while seismic moment is an objective measure of the actual physical size. Using the calculated seismic moments, we calculated "moment magnitudes" (M_w) for all events, which allowed us to do further studies in terms of their absolute size as a function of both time and space.

The results indicate that there are two distinct classes of seismic events at the mine. The first class consists of small (M<=0) earthquakes recorded near the moving mine face. The second class of seismicity occurs in the mined-out “gob” area of the longwall panel at a greater distance behind the moving face. Their occurrence and relation to the mining history, depth of overburden and geology of the roof rocks is a significant interest.

Results show that thick overburden due to elevated topography has a positive correlation with the number of seismic events but is not the only controlling factor; other factors include gob size and geological variability. Another important observation is the high seismic attenuation of the rock mass above the mine. This appears to be the result of the fracturing and caving processes associated with the creation of the gob and the resulting subsidence of the ground surface.
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CHAPTER 1. INTRODUCTION

Seismic events associated with mining operations are common and have become an important topic. Seismologists have monitored mines instrumentally for many years in some areas, a few examples include, the deep gold mines in Australia and Africa and some coal mines in Germany and Poland. Coal mine monitoring began as early as the 1920’s, when seismologists installed a network of four stations at the Rozbark coal mine in Poland (Gibowicz & Kijko, 1994). In 1939, seismic monitoring in South African gold mines began with the installation of five seismographs (Gibowicz & Kijko, 1994). These networks inspired further research into the field of mining induced seismicity, but even today, monitored mines represent a small percentage of the world’s mines. Seismic monitoring of U.S. mines has also been very limited compared to that in several other countries, although recent examples of seismic monitoring have been reported in the news media. For example, the Crandall Canyon Mine collapse in 2007, which involved a deep coal mine in Utah.

Large seismic events near active mine workings can be hazardous to the miners and interrupt mining activities, while smaller events might be an indicator of stress buildup leading to the potential for a larger event. Real-time monitoring of this activity, using advanced instruments and data processing technology can be valuable in assessing the causes and potential for mining-related seismic events and to develop or validate strategies to reduce potential associated hazards. In addition to providing data for testing seismological hypotheses concerning the physics of the seismic source and wave propagation effects, the monitoring provides near-real-time hypocenter locations and estimates of event size (magnitude). This information can be used by my mining operators to assess conditions of changing stress that may affect mining operations. Currently, active seismic monitoring is being conducted at select mines located in Utah, Colorado and Virginia.

1.1. Background and Lithology Information

The study mine is a longwall operation located in the Appalachian Plateaus province, Virginia. The important geologic formations associated with the mine are the units around the
coal bed and the formations above it. The late Mississippian and early Pennsylvanian formations in the area, in ascending order, are the Bluestone, the Pocahontas, the Lee and the Wise.

The Bluestone formation ranges from 150 to 600 feet in thickness and is comprised of shale, siltstone and sandstone (Nolde, 1994). There are three recognized members within the formation: the Pride Shale Member, a middle sandstone member and an upper shale member (Nolde, 1994). The Pride Shale consists of grayish-black shale that has beds of siltstone, sandstone and limestone (Nolde, 1994). The quartzarenite of the Lee formation truncates the upper two formations before they reach the study area.

Directly above the Bluestone is the Pocahontas formation, which is about 700 feet thick and contains ten coal beds in a sequence of sandstone, siltstone, shale and underclay (Nolde, 1994). The study mine extracts coal from the Pocahontas Number 3 Coal Bed. In the mine area, the average thickness of the coal seam is 72 inches (personal communication, mine operators). The other coal seams in the Pocahontas typically have thicknesses of less than two feet and are not a target of extensive mining operations.

The Lee Formation, which is composed of three sandstone units overlays the Pocahontas formation. The Wise formation is above the Lee formation and has a maximum thickness of 890 to 1100 feet in the study area (Nolde, 1994). The formation is composed mainly of siltstone, shale, coal and sandstone that all contribute to the overburden of the mine.

1.2. Important Seismological Terms

The following sections will introduce and discuss various aspects of seismic source and wave propagation theory. Seismologists commonly describe the seismic source in terms of two quantities, the static seismic moment and a change in stress that occurs during faulting. For a shear dislocation source (earthquake), the seismic moment is equivalent of the product of the shear modulus in the rocks surrounding the fault zone, the area of rupture during the earthquake and the average displacement (Equation 2.1). This physical quantity can be estimated from the radiated seismic wave field. Brune (1970) developed a model for the Fourier amplitude spectrum of the far-field radiated seismic shear wave pulse. This model relates two source parameters, the static seismic moment and the stress drop. The amplitude spectrum of the Brune
model for the far-field shear wave displacement pulse is referred to as an $\omega^2$ spectrum because the high-frequency displacement amplitudes decay in proportion to frequency squared. The low-frequency displacement spectral amplitudes are constant and are proportional to the static seismic moment. The transition from constant spectral amplitudes, at low frequency, to rapidly decaying amplitudes at high frequencies is marked by the corner frequency. Brune (1970) derived a relationship for the corner frequency in terms of the stress drop and the seismic moment (Equation 1.1). Thus, measures of the spectral corner frequency and the low frequency displacement amplitude can be used to estimate seismic moment and stress drop under the assumption of an $\omega^2$ source spectrum.

<table>
<thead>
<tr>
<th>Equation 1.1</th>
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<tr>
<td>$\Delta \sigma = M_0 \left[ \frac{f_c}{4.91 \times 10^6 \beta} \right]^3$</td>
</tr>
</tbody>
</table>

- $\Delta \sigma$ is stress drop in Pascals
- $M_0$ is the static seismic moment in N-m
- $\beta$ is the S-wave velocity in km/s
- $f_c$ is the corner frequency

Wyss and Brune (1968) showed that two stress estimates for earthquakes, the apparent stress and the stress drop (described above), can be calculated from the seismic wavefield. The apparent stress is proportional to the product of the shear modulus and the radiated seismic energy divided by the seismic moment (Equation 1.2). The estimation of the apparent stress does not require the assumption of the $\omega^2$ model. Both apparent stress and stress drop in the context of the Brune model have been used in previous works to describe the seismic source in the case of natural earthquakes and mine-related seismic events.
Many magnitude scales developed over the years have been used to quantify the relative size of earthquakes. These include the original Local magnitude scale \( (M_L) \), the teleseismic surface wave and body wave scales \( (M_s, m_b) \) and regional scales such as \( m_b(Lg) \). Magnitude is a non-dimensional number comparing the source strength at a specific frequency of one event with that of a standard earthquake, and for that reason, the older magnitude scales have many limitations. Seismologists now routinely quantify earthquake “size” in terms of static seismic moment because it is a physical quantity directly related to the physical size of the fault its displacement and the strength of the rock (Equation 2.1). The concept of magnitude is still employed by seismologists because of tradition and convenience. However, in recent years, a magnitude based on the logarithm of the seismic moment rather than the logarithm of trace amplitude on a particular seismic instrument has come into universal use. This magnitude scale is referred to as the moment magnitude scale. However, the automatic data processing algorithms used for real time processing of the data stream from the study mine do not calculate seismic moment or moment magnitude; instead, the magnitude calculation uses the local magnitude described by Richter (1958). Magnitude 3.0 earthquakes are typically the smallest earthquakes that are felt.

<table>
<thead>
<tr>
<th>Equation 1.2</th>
<th>[ \sigma_a = \mu \frac{E_s}{M_0} ]</th>
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<tr>
<td>• ( \sigma_a ) is apparent stress</td>
<td></td>
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<tr>
<td>• ( \mu ) is the shear modulus in Pascals</td>
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<tr>
<td>• ( E_s ) is radiated energy</td>
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<tr>
<td>• ( M_0 ) is the moment in N/m</td>
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1.3. Previous Work on Mine-Related Seismicity

One of the goals of this study is a better understanding of the physics of the seismic source involved with the mining induced events. Several previous studies have examined the relationship between these parameters using data from small earthquakes with magnitudes less than 3.0 (micro-earthquakes), similar in magnitude to those generated at the study mine during
normal mining operations. A major problem encountered in past studies is the effect of attenuation due to scattering and anelastic loss on the bandwidth of the observed seismic spectrum. The bandwidth of the observed seismic spectrum is used to calculate estimates of the stress drop and apparent stress. Attenuation (including both effects of anelastic absorption and scattering) alters the observed spectrum by reducing the high frequency amplitudes, which generally reduces the usable bandwidth and may complicate interpretation of the source spectrum. Previously published studies have dealt with this problem by attempting to independently estimate Q, the seismic quality factor, which quantifies a material’s ability to transmit elastic waves without loss of wave energy due to frictional heating. By definition, a low Q (<50) value represents an area with high attenuation. Frequency dependent loss by scattering mechanisms also contributes to actual field determination of Q. Given an accurate estimate of Q (containing both effects of anelastic absorption and scattering) the observed spectrum can be corrected, and stress drop can be estimated without bias.

Abercrombie (1995) examined source parameters for earthquakes in the Cajon pass scientific drillhole in southern California. She determined source parameters for over 100 nearby, tectonic earthquakes, from recordings at a depth of 2.5 km (in granite). She compared the recordings at depth from those at the surface and observed evidence of severe attenuation in the upper kilometers of the Earth's crust. Based on determinations of stress drop from spectral corner frequency, measured by inspection, she concluded that her data from the deep hole were consistent with constant stress drop source scaling. On the other hand, she observed a decrease in apparent stress with decreasing moment. Abercrombie (1995) concluded that the decrease in apparent stress with moment was consistent with constant stress drop scaling and could be attributed to the underestimation of radiated seismic energy for small, shallow events. The Abercrombie (1995) results are important for this study because they provide evidence for very high attenuation at shallow depths (less than 2 km) in the crust (Q values of approximately 30), and appear to support constant stress drop over the magnitude range from negative 1.0 to 5.5.

Richardson and Jordan (2002) studied mining-related seismicity in five very deep gold mines in the Far West Rand district in South Africa. They recorded data from in-mine arrays of three-component seismometers, with as many as 1000 seismic events occurring daily. The events recorded varied in size from moment magnitudes of negative two to three, and focal
depths were between two and four km. Richardson and Jordan (2002) noted two distinctive event types (their type-A and B). Type-A events were very numerous, small magnitude shocks that clustered in time and space near the active mining face. Type-A events were enriched in high frequencies and appeared to exhibit a maximum magnitude cutoff and significant isotropic (implosional) source mechanisms, with very high stress drops for their seismic moments. Richardson and Jordan attributed them to "fracture dominated" rupture of the competent rock mass due to blasting and closure of mine openings (stopes). Type-B events were temporally and spatially distributed throughout the active mining region, and with some large magnitudes (M>3), but showed a distinct lower magnitude cutoff at approximately M = 0. The type-B events were interpreted to represent "friction-dominated" slip in pre-existing shear zones, because they exhibited source-scaling properties similar to tectonic earthquakes and double-couple source mechanisms. The data Richardson and Jordan collected was near the source and the rocks were extremely hard, with large Q values: attenuation was not an issue for source parameter estimation. Richardson and Jordan observed that apparent stress decreased with moment for Type-A events, but increased with moment for Type-B events.

The Richardson and Jordan (2002) results are important for this study because they indicate that different types of source mechanism and source scaling relations may co-exist in the mining environment. Although the mining environment studied by Richardson and Jordan is very different from that in the coal mine studied here, analogous observations concerning two types of source mechanisms at the mine will be demonstrated in a later section.

Oye et al. (2005) examined small (-1.8 to 1.2 Mw) earthquakes in the Pyhasalmi ore mine in Finland. The mine is the deepest ore mine in Western Europe with depths up to 1.4 kilometers. Seismic data were recorded on 4 3-component geophones and 12 vertical-component geophones located within the mine with sampling rates of 3000 Hertz (sampling rates of 1000 to 500 Hz were used to save storage space). The b-value of the Gutenberg-Richter recurrence relation for the range of Mw -1 to 0 was one, the expected result for tectonic earthquakes (see section 4.2). Q values ranged from 200 to 500 within different earthquake clusters. The two methods used to estimate apparent stress, energy-to-moment ratios and multiple empirical Green’s function (MEGF), gave different results. The energy-to-moment ratio indicates an increase in apparent stress with magnitude while the MEGF results indicated
constant stress scaling. They concluded that it is not possible to choose between the two results because of the large scatter in the data.

Yamada et al. (2005) studied seismicity in a deep gold mine. Nine three-component accelerometers placed in a haulage tunnel about 2600 meters below the ground surface in close proximity to mining operations recorded very high frequency data. The study focused on five events with $M_w$ in the range 0.8 to 1.4 located within 150 meters of the recording stations. The data showed source corner frequencies of approximately 100 Hz, from the events occurring with normal faulting mechanisms in very hard gold-bearing quartzite at 2.6 km depth. They estimated radiated energy, and performed kinematic waveform inversion to estimate the rupture speeds of the events. They used $Q$ in the range 11 to 92 to correct for attenuation. They found static stress drops in the range of two to 20 MPa, similar values for apparent stress, and concluded that apparent stress does not decrease with decreasing seismic moment: i.e., the small events in this deep hard-rock mine exhibited stress drops and source scaling behavior similar to larger tectonic earthquakes. They concluded that their results are inconsistent with several previous studies that found decreasing apparent stress with decreasing seismic moment for small $M<3$ earthquakes (e.g., Abercrombie 1995; Prejean and Ellsworth, 2001; Gibowicz et al., 1991; Kanamori et al., 1993; Jost et al., 1998). They suggested (as was pointed out earlier by Ide and Beroza, 2001, and Ide et al., 2003), that the studies reporting violation of constant stress drop scaling at low magnitudes, based on measurements of apparent stress are biased due to underestimation of radiated energy caused by bandwidth limitations, or lack of correction for attenuation effects. Sonley and Abercrombie (2006) further examined the effect of attenuation on estimation of source corner frequency using data from the San Andreas Fault at Depth (SAFOD) High Resolution Seismic Network. They found that minor misrepresentation of $Q$ could result in major changes in the estimate of radiated seismic energy.

Grant (2006) conducted a study using datasets from the Moonee Colliery coal mine in Sydney, Australia and the Springfield Pike Quarry, an underground limestone mine, in Pennsylvania. That study showed that induced events in the coal mining environment feature very low stress drops and apparent stress as evidenced by source corner frequencies between 1 and 10 Hz for events in the magnitude -1 to 1.5 range. In addition, the study found evidence that apparent stress of the mining induced events increases with seismic moment. This finding
represents a divergence from constant stress drop scaling, regardless of effects due to attenuation. The results of Grant (2006) are important for this study because they show that in coal mines (at very shallow depths compared to the studies in South African gold mines), mining induced events appear to be different in terms of stress drop and source scaling from small tectonic earthquakes, and mining induced events in deep, hard-rock mines.

In summary, source scaling of small mining induced earthquakes is a very difficult and controversial subject. There is abundant evidence that not all mines are the same, in terms of the seismic sources they generate. Soft-rock mines and coal mines at depths less than 1 km appear to generate events with very low stress drops, compared to most tectonic earthquakes and mining-related events in deep, hard-rock mines. In addition, evidence is slowly accumulating that different type of seismic sources may operate in different parts of a mine, be it shallow or deep.

1.4. Seismic Monitoring Network

The mine operators and the Virginia Tech Seismological Observatory (VTSO) installed a network of five surface seismometer stations at the mine in early 2009. The locations of stations was optimized to achieve accurate hypocenter locations, constrained by the requirement of direct line-of-sight radio links from the station locations to a centrally located 200 foot-high radio tower and equipment shelter, where there is a micro-wave radio link. The equipment at each station consists of a 3-component set of passive short-period (either 1Hz or 2Hz natural frequency) velocity transducers to record weak motions and a 3-component force-balance accelerometer to record strong motions. A Geotech Instruments, Inc. Smart-24D high-resolution digitizer with a GPS satellite receiver for location and timing digitizes the six channels. After the digitizer samples the seismic data at 100 samples per second, 900 MHz spread-spectrum radios telemeter the data to a central receive location, where it is further telemetered by microwave. The data are eventually routed by Internet to a second recording system at Virginia Tech, running the full suite of Geotech Instruments, Inc., SmartGeoHub and ancillary data acquisition and processing software.
Automatic real-time data processing involves event detection, phase arrival time picking, event location, and archiving of waveform data. The event detection uses the short-term/long-term signal level approach, with window lengths and station correlation times optimized to the spatial geometry and noise characteristics of the stations comprising the mine network. Once an event has triggered the detection algorithm, the software writes a log file and automatically attempts to pick phase arrival times and locate the event. If the event is successfully located, the computer writes an event file with the hypocenter and magnitude estimate. The processing uses the Geotech Instruments software package, which implements Geiger's method for event location and a half-space velocity model that is based on sonic logs from the mine area. The locations are subject to two main error types that affect hypocenter location accuracy. These error types are 1.) random location scatter caused by errors in arrival time measurements and 2.) systematic bias caused by incorrectness of the velocity model used to locate the event (Gibowicz & Kijko, 1994). One shortcoming of Geiger’s method is that it can be unstable with very sparse data, and in some circumstances, can converge to a local minimum error solution rather than to a global error minimum. The Geotech Instruments software package estimates a Local magnitude, using the Richter (1958) formula. This procedure involves correcting the recorded amplitudes to peak Wood-Anderson instrument amplitudes using the transfer functions of the instruments in the field.

Following the initial location of the event with Geiger's method, a second location is done automatically, using a more stable direct grid-search approach developed by M. Chapman (personal communication, 2011) that is more reliable with poorly constrained events. The grid-search algorithm is completely independent of the initial automatic location using Geiger’s method and only uses the phase arrival picks derived from the earlier stages of the automatic processing. Then the event data are archived (complete waveforms as well as parameter data), and the database developed by Virginia Tech is updated to include the new event. This database has extensive search and retrieval capability. The database software has graphical display and download capability that is available to the mine operators and VTSO staff via internet within a few minutes of the event origin time. This completes the automatic event processing sequence.

This study draws its data from the database that now consists of many thousands of events. The data requirements consist of well-recorded events at all 5 stations, with magnitudes
in the range from \( M_L \) -1.6 to 2.3, in the time interval from March 1, 2009 to April 7, 2011. These events satisfy the requirements of stable network operation (stable and consistent detection and location capability and calibrated instrument response).

Figure 1.1 shows the layout of the mine, with seismic station locations. The target coal bed is at an elevation of approximately 500 feet above mean sea level (msl), and is essentially horizontal. The thickness of overburden is variable because of rugged surface topography. Average surface elevation at the monitoring stations is 2,600 feet (msl). The mining was done in a sequence of “panels” progressing from left to right. Figure 1.1 shows the mined areas of the B, C, D, E and G panels (grey shading). The A and F areas are barriers between panels, and were not mined. Stations 1 through 5 monitored the seismicity during the mining of panels C, D and E. Station 4 was renamed and moved to become Station 6, after the completion of panel E. This was done to better monitor the mining of panel G. The coal in the active panel is extracted by cutting it from the longwall face using a mining machine (shearer). The shearer cuts the coal by traveling transversely, in the left-right direction, along the active longwall face that is 750-1,000 feet in length. The active longwall face advances to the north at a rate of approximately 50 feet per day in normal operations.

Shown in Figure 1.1 are the epicenters of seismic events with local magnitudes \( M_L > 0.0 \) that occurred during the period September 16, 2009 through April 7, 2011 during the mining of the D, E and G panels. Events associated with the mining of those two panels are the main subject of this study. Note that the seismic event epicenters clearly outline the active panels. As will be shown below, most events occur near the active face, and will be referred to below as "face events." However, a significant number of events occur behind the active face in the previously mined part of the panel known as the "gob." These gob events can occur weeks to months after mining. Examples of such late-stage gob events occurring in the previously mined C panel are evident in Figure 1.1.

Chapter 2 below describes the development of a local magnitude to seismic moment correlation equation that is used to convert automatically computed local magnitudes to seismic moments, for source parameter estimation. Chapter 3 discusses attenuation of ground motion at the mine as a function of frequency. Chapter 4 examines the spatial and temporal characteristics
of the mining-related seismicity, and Chapter 5 presents a discussion of the study that includes our conclusions and suggestions for future work.

Figure 1.1: Mine layout along with the network of seismic stations (numbered blue triangles). The network was installed in March 1, 2009 during the mining of the C panel. Red circles represent A and B quality seismic events with $M_L > 0.0$, during the period September 16, 2009 through April 7, 2011, when panels D, E and G were being actively mined. Dashed outlines represent planned mine panels. Grayed out boxes represent areas that have been mined by April 7, 2011.
CHAPTER 2. SEISMIC MOMENT

Estimates of seismic moment are needed for investigations of the physical nature of the source, and for comparison with other studies. An empirical relationship was developed using 52 of the best-recorded events to relate Local magnitude ($M_L$) to moment (Table 2.1). The ability to calibrate $M_L$ in terms of moment is needed because the automatic event processing software used to monitor the mine in real-time does not determine static seismic moment.

2.1. Seismic Moment Calculation

Equation 2.1 assumes a shear dislocation (double-couple) source, and knowledge of the fault dimension and average fault displacement to make calculations of the moment. Therefore, Equation 2.1 is impractical in cases where the fault does not break the ground surface, and cannot be applied to non-double-couple sources. A more general, purely seismological, estimate of the static seismic moment can be obtained by considering the seismic source to be a three dimensional system of time-varying force couples, acting at a point. The time integral of the couples is $M_0$, the static seismic moment. The amplitudes of the body wave components of the far-field P and S-waves are proportional to the source moment-rate function and share a common functional form. Given a recording of the far-field shear wave displacement, for example, $M_0$ can be estimated in either the time or the frequency domain if the source location and radiation pattern of the source (double-couple or otherwise) is known. Several approaches to accomplish this are available for large events recorded a sufficient distance from the source such that individual body wave phases can be distinguished and separated, but time domain estimation of the static moment for microearthquake events recorded at close distances in the mining environment is problematic. Shear waves arrive virtually simultaneously with the P and surface waves, and it is usually impossible to separate them. The problem is complicated by the fact that, with sparse networks of stations, the source mechanism may be difficult if not impossible to determine, and in general may be non-double-couple.
The waveforms were visually inspected with the aid of Seismic Analysis Code (SAC) (Goldstein & Snoke, 2005) to determine signal and noise windows. Signal windows varied from four to five seconds depending on the duration of the event. Signal windows were chosen half a second prior to the event and half a second after the event so that the application of a 10 percent trapezoid taper to the windowed signal would not alter the signal waveform (Figure 2.1). Signal and noise windows were consistently chosen to be equal time lengths (same number of samples), with noise windows following the termination of the event (signal windows). The windowed signal and noise time series for each component (Figure 2.2) at all available stations were transformed to the frequency domain and converted to ground displacement using the transfer function for each component.

<table>
<thead>
<tr>
<th>Equation 2.1</th>
<th>$M_o = \mu AD$</th>
</tr>
</thead>
<tbody>
<tr>
<td>• $\mu$ is shear modulus in Pa or N/m$^2$</td>
<td></td>
</tr>
<tr>
<td>• $A$ is area of the shear rupture in m$^2$</td>
<td></td>
</tr>
<tr>
<td>• $D$ is the average displacement on the fault in m</td>
<td></td>
</tr>
</tbody>
</table>
Figure 2.1: Seismograms recorded on vertical components at all five stations for event on June 13, 2010. Blue arrows indicate the arrival of the P-waves. Vertical black lines indicate the signal window.
Figure 2.2: Seismograms recorded on all three components on Station 3 for June 13, 2010 event. Blue arrows indicate the arrival of the P-waves. Vertical black lines indicate the signal window.

Visual inspection of the waveforms shows that the majority of signal energy is associated with the shear and surface waves. The S-wave cannot be separated from the surface waves due to the small propagation distances and data sample rate (100 samples per second). The source radiation pattern is unknown, although it is clear that it is neither pure double-couple nor pure implosion, but possibly a combination of both. These unknowns and complications contribute to uncertainty in the estimates of the seismic moment. The basic measurement is the zero frequency spectral amplitude that is determined from the entire waveform. An estimate of the hypocentral distance is also required (Equation 2.4). The seismic moment is directly proportional to the zero frequency spectral amplitude. However, in this study the value of the constant of proportionality is uncertain for the previously mentioned reasons. Equation 2.2 is used to estimate the static seismic moment. It assumes an isotropic source mechanism with a
uniform radiation pattern value of unity, body wave geometrical spreading and a free surface
effect equal to 2.0. The shear wave velocity, $\beta$, appears in Equation 2.2 because most of the
energy is associated with the surface and S-wave instead of the P-wave. The choices of body
wave spreading, free surface effect and radiation pattern are judgmental.

\[
M_0 = \frac{4\pi \rho \beta^3 \Omega_0 \Delta}{2}
\]

- Where $M_0$ is the static seismic moment in N-m
- $\rho$ is density in kg/m$^3$
- $\beta$ is velocity in km/s
- $\Omega_0$ is the zero frequency displacement amplitude in km/Hz
- $\Delta$ is hypocentral distance in km

The zero frequency displacement amplitude in Equation 2.2 was estimated with the use of
all three short-period components at each recording station. The spectral displacements were
summed and averaged over the frequency range from one to seven Hertz. Event selection was
such that within this frequency band, the signal to pre-signal noise ratio exceeded two (Figure
2.3). The mean amplitudes for each component were combined according to Equation 2.3. This
procedure was repeated for all recording stations, giving, in most cases, five estimates of $M_0$ for
each event. Equation 2.4 determines the distance from the hypocenter to the receiver with the
assumption of a constant-velocity medium.

\[
\Omega_0 = \sqrt{\bar{z}^2 + \bar{n}^2 + \bar{e}^2}
\]

- $\bar{z}^2$ is the average vertical component displacement
- $\bar{n}^2$ is the average north-south component displacement
- $\bar{e}^2$ is the average east-west component displacement
\[ \Delta = \sqrt{Z^2 + R^2} \]

- \Delta is the hypocentral distance from the hypocenter of the event to the station
- \( Z \) is depth to the hypocenter
- \( R \) is epicentral distance

**Figure 2.3**: Displacement amplitude spectrum for one event recorded on June 13, 2010. Station 2 recorded this seismogram on the East-West component. As is demonstrated by the red bracket in the figure, amplitudes in the range from one to seven Hz are nearly constant. Therefore, the mean of those spectral values in this range are used as an estimate of the zero amplitude displacement.

### 2.2. Seismic Moment Results

An empirical relationship between seismic moment and \( M_L \) for the study area was determined using linear regression. The mean estimates of \( M_0 \) for each event were regressed against the mean estimates of \( M_L \). Figure 2.4 shows the results. The resulting model is
Equation 2.5

<p>| | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Log(M₀) = 1.04 * Mₜ + 9.61</td>
<td></td>
</tr>
</tbody>
</table>

- M₀ is the seismic moment in N-m
- Mₜ is the Local Magnitude

The data are well-modeled by the linear function. The residuals show no significant trends, the R² value is 0.98, and the standard error of estimate for both the intercept and the slope is 0.003.

Figure 2.4: Relationship between local magnitude and log of seismic moment in N-m.
Table 2.1: Data from 52 seismic events with Local magnitude calculations and the log of seismic moment.

<table>
<thead>
<tr>
<th>Date</th>
<th>(M_L)</th>
<th>Log Seismic Moment</th>
<th>Date</th>
<th>(M_L)</th>
<th>Log Seismic Moment</th>
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<td>0.16</td>
<td>16.79</td>
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</tr>
</tbody>
</table>
CHAPTER 3. WAVE PROPAGATION IN THE STUDY AREA

The loss of seismic wave amplitude with increasing distance is a result of two kinds of attenuation: elastic and anelastic. Elastic effects include geometrical spreading due to the expansion of wave fronts with time as the wave energy propagates through the medium, scattering, and dispersion (Yamada, 2005b). These elastic effects result in the loss of wave amplitude and cause an increase in signal duration with time. Anelastic attenuation (absorption) (see section 1.2) is due to internal friction that results in the transfer of kinetic wave energy to heat. It is difficult to isolate anelastic and elastic wave propagation effects. Furthermore, source and path effects can also be difficult to isolate.

Estimation of stress drop and/or apparent stress, in addition to the static moment described in Chapters 1 and 2, is very problematic for this study due to complications caused by path effects. The effect of anelastic attenuation reduces the amplitudes of the spectrum at high frequencies. If anelastic attenuation is sufficiently large, this will affect the estimate of the source spectrum bandwidth. If the spectral shape has been affected by attenuation, it may be impossible to identify a true source-related corner frequency. Radiated energy is very sensitive to the amplitude of the spectrum near the corner frequency. The energy can be significantly underestimated if anelastic attenuation affects that part of the spectrum. In fact, to estimate stress drop or apparent stress, it is necessary to quantify and account for the effects of attenuation.

Aside from estimation of source parameters, the material properties of the rock above the mine is a topic of interest. In particular, the area immediately above the mined out panels is subsided and is likely to be fractured to an extent that may affect the attenuation of seismic wave energy. If it is possible to quantify attenuation over the mine gob, it may be possible to study the extent of fracturing in the gob. The following sections address these issues.

3.1. Attenuation Correction

The data exhibit the effects of very strong attenuation. Figures 3.1 through 3.5 show the recorded displacement amplitude spectra for a 2.18 M_L event at all five recording stations. The
recorded spectra show apparent corner frequencies of approximately 5 Hz. At lower frequencies, the spectral amplitudes are approximately constant, and are interpreted to indicate the zero frequency displacement amplitude. Amplitudes at frequencies greater than the corner frequency decay rapidly. Spectral amplitude decay in proportion to the inverse two-and-a-half power of frequency at the closest station, ranging to the inverse fourth power of frequency, for the most distant station are shown by the figures. Also shown in Figures 3.1 through 3.5 are “corrected” displacement amplitude spectra. The corrected spectra assume a source spectrum with high frequency behavior according to the Brune (1970) \( \omega^{-2} \) spectrum model, and a frequency-independent Q. The following discussion describes the development of the corrected amplitude spectra.
Figure 3.1: Corrected (red), uncorrected (blue) and noise (green) displacement spectra for a 2.18 M_L event with a hypocentral distance of 1.53 km. Notice the difference in amplitudes at high frequencies between the corrected and uncorrected spectra.
Figure 3.2: Corrected (red), uncorrected (blue) and noise (green) displacement spectra for a 2.18 M$_L$ event with a hypocentral distance of 1.73 km. The corrected spectrum shows a decay of one power similar to the Figure 3.1, that has a similar hypocentral distance.
Figure 3.3: Corrected (red), uncorrected (blue) and noise (green) displacement spectra for a 2.18 $M_L$ event with a hypocentral distance of 1.06 km. Notice the smaller difference in decay compared to the other figures that have larger hypocentral distance and as a result greater attenuation in the uncorrected spectra. Pay particular attention to the following two figures that have the largest hypocentral distances and significantly more attenuation.
Figure 3.4: Corrected (red), uncorrected (blue) and noise (green) displacement spectra for a 2.18 M<sub>L</sub> event with a hypocentral distance of 2.08 km. Notice the difference in amplitudes at high frequencies between the corrected and uncorrected spectra. The corrected spectrum shows that it fits an $\omega^{-2}$ decay compared to the uncorrected spectrum that decays much quicker as $\omega^{-3.5}$. 
Figure 3.5: Corrected (red), uncorrected (blue) and noise (green) displacement amplitude spectrum for a 2.18 M<sub>L</sub> event with a hypocentral distance of 3.51 km. The corrected spectrum shows that it fits an $\omega^{-2}$ decay compared to the uncorrected spectrum that decays much quicker as $\omega^{-4}$.

3.2. Evidence for Path Dependent Attenuation and Q Estimation

Examination of the Fourier amplitude spectra of 42 well-recorded events reveals a systematic change of spectral shape. Spectra from all stations exhibit apparent source corner frequencies in the range 1 to 10 Hz. However, the spectral slopes at frequencies greater than 10 Hz are different at the different recording stations for a given earthquake. For example, Station 3, centrally located over the mining activity, shows an essentially flat spectrum at high frequency for most of the events examined, whereas the other stations exhibit variable, but significant degrees of amplitude decay with increasing frequency.
The issue to be examined is whether the observed behavior is a source effect, a path effect or perhaps a combination of both. The obvious corner frequency in the range of 1 to 10 Hz is common to all the stations and appears to be consistent for a given earthquake. This is expected if that feature is a source property. The small values (less than 10 Hz) of the observed corner frequencies suggest low stress drops for the events examined here, compared to natural earthquakes of the same moment.

If the observed corner frequencies are associated with the common Brune (1970) $\omega^2$ type source spectrum shape expected for a simple seismic source, the Fourier amplitude spectra of ground acceleration due to shear waves can be modeled using the following expression:

**Equation 3.1**

$$ S(\omega) = \frac{RM_0}{4\pi \beta^3 r} \frac{\omega^2}{1 + \left(\frac{\omega}{\omega_c}\right)^2} \exp\left(\frac{-\omega r}{2Q\beta}\right) $$

- $M_0$ is the seismic moment
- $\omega_c$ is the source corner frequency
- $R$ is the S-wave radiation pattern
- $Q$ is the (constant) quality factor of S-waves
- $r$ is hypocentral distance
- $\beta$ is S-wave velocity

For $\omega >> \omega_c$, we have

**Equation 3.2**

$$ S(\omega) = \frac{RM_0}{4\pi \beta^3 r} \omega_c^2 \exp\left(\frac{-\omega r}{2Q\beta}\right) $$

- $M_0$ is the seismic moment
- $\omega_c$ is the source corner frequency
- $R$ is the S-wave radiation pattern
- $Q$ is the (constant) quality factor of S-waves
- $r$ is hypocentral distance
- $\beta$ is S-wave velocity
Substitution of $\omega = 2\pi f$ and $\omega_c = 2\pi f_c$ and taking logarithms results in

$$
\text{Equation 3.3} \quad \ln S(f) = \ln \left[ \frac{R M_0}{4 \pi \rho \beta^3 r} (2\pi f_c)^2 \right] - \frac{\pi r}{Q \beta} f
$$

- $M_0$ is the seismic moment
- $\omega_c$ is the source corner frequency
- $R$ is the S-wave radiation pattern
- $Q$ is the (constant) quality factor of S-waves
- $r$ is hypocentral distance
- $\beta$ is S-wave velocity

The above expression is of the form $\ln(S) = a + b f$ and an estimate of the slope, $b$, provides an estimate of $Q$ if both $r$ and $\beta$ are known. Thus, it is possible, in principle, to determine $Q$ from a single recording of the shear wave.

Note the slope $\left( \frac{-\omega r}{2Q\beta} \right)$ of Equation 3.2 is independent of the source, because of the high-frequency assumption ($f >> f_c$). The high-frequency slope depends on $Q$, $\beta$ and hypocentral distance, $r$.

Under these assumptions, if multiple observations are available from $n$ different earthquakes at the 5 stations over a range of distances and if the quantity $Q^*\beta$ is constant, a plot of the slope $b_i$ versus distance $r_i$ from $i = 1, 2, 3 ... (n \times 5)$ observations should exhibit a linear trend: i.e.,

$$
\text{Equation 3.4} \quad b_i = \left( \frac{\pi}{Q \beta} \right) r_i \quad i = 1, 2, 3, ..., n \times 5
$$

- $\omega_c$ is the source corner frequency
- $R$ is the S-wave radiation pattern
- $Q$ is the (constant) quality factor of S-waves
- $\beta$ is S-wave velocity
Figure 3.6 shows, as individual symbols, least-squares estimates of the slopes of Equation 3.3, derived from high frequency (15-35 Hz) acceleration spectra from 42 seismic events recorded at the five stations. The $S(f)$ values used in Equation 3.3 are the geometric means of the two horizontal components. Different colors indicate recordings from different stations. Station 3 provides observations at the smallest hypocentral distances, whereas station 5 is consistently the most distant station from the source.

Figure 3.6 shows that the values of spectral slope decrease (become more negative) as hypocentral distance increases. This systematic behavior with distance is evidence for extremely strong attenuation along the path from source to receiver. The observation was unexpected because anelastic absorption effects were expected to be negligible given the relatively large $P$ and $S$-wave velocities determined for the mine area, combined with the small hypocentral distances involved in this study. The slope of a line with zero intercept fit by least squares to these data (e.g., the form of Equation 3.4) is $-0.033 \pm 0.005 \text{ sec/km}$. Assuming a value of 2.54 km/s for $\beta$, from sonic logs, this indicates $Q = 37 \pm 5$. This is a remarkably small value for the quality factor of Paleozoic sedimentary rocks with $\beta$ in excess of 2.5 km/s. These observations may point to an unusual attenuation mechanism. A highly fractured mass above the mine may be responsible. The ground surface may subside as much as three feet following the mining of a longwall panel. If this subsidence reflects inelastic brittle deformation, e.g., fracture, faulting and caving of the rock mass, rather than elastic strain, then the fracture density of the rock would dramatically increase, possibly leading to increased wave scattering at high frequencies and the low observed $Q$ value.
Figure 3.6: Slope of the log acceleration amplitude spectra versus frequency as a function of hypocentral distance.

A "corrected" ground acceleration spectrum, which accounts for geometrical spreading, and the combined effects of scattering and absorption is computed by multiplying the observed Fourier amplitude spectrum of instrument corrected ground motion by the hypocentral distance and a frequency dependent factor, as expressed in Equation 3.5. It is important to recognize that the value of Q determined here assumes an $\omega^2$ (or Brune, 1970) source spectrum.
3.3. Stress Drop

Figures 3.1 through 3.5 show the Fourier amplitude spectra of ground displacement for one of the best-recorded events in the study. From inspection of the figures, it appears that the source corner frequency lies in the range five to 10 Hz. This range for the corner frequency is similar to that observed for many other events of similar magnitude from the mine. The Brune (1970) static stress drop would be in the range 1 to 7 bars (0.1 to 0.7 MPa). This is very low compared to tectonic earthquakes, which typically are in the range 20-200 bars (2-20 MPa), e.g., Yamada et al. (2005). Additional study would be needed to evaluate this discrepancy.

\[
A_c(\omega) = A(\omega)\Delta e^{\omega\Delta/2Q}\beta
\]

- \(A(\omega)\) is the recorded acceleration amplitude
- \(\Delta\) is the hypocentral distance from the hypocenter of the event to the station
- \(\beta\) is shear wave velocity in km/s (2.54 km/s)
- \(Q\) is the quality factor (37 +/- 5)
- \(\omega\) is the angular frequency
CHAPTER 4. MOMENT RELEASE IN TIME AND SPACE

Three complete panels have been mined since the installation of the network: these panels are about 700 feet wide and ~11300 feet long. The three-mined panels are designated D, E, and G. Seismic events within those three panels were studied in terms of magnitude, temporal and spatial distribution, for insight into the relation between mining process, overburden thickness and the occurrence of the seismic events. The work described previously to develop a correlation relationship between \( M_L \) and seismic moment allows us to use the moment magnitude (Kanamori (1977); Hanks and Kanamori (1979)) as the basis for statistical descriptions in the mine.

\[
M_w = \frac{2}{3} \log M_0 - 10.7
\]

- \( M_w \) is the moment magnitude Hanks and Kanamori (1979)
- \( M_0 \) is the seismic moment is in N-m

This analysis indicates the presence of two distinct event classes similar in some respects to the event classes recognized by Richardson and Jordan (2002): events that occur near the active mining face (Class 1) and events that occur in the gob well after mining (Class 2). The events near the mining face are typically much more frequent and exhibit a large proportion of small events relative to large events, compared to the seismicity in the gob after mining. Class 1 events are apparently caused by stress concentrations near the active mine face. Class 1 events may be associated with fresh cracking of rock and coal and likely result in relatively small amounts of slip (Andrews, 1976; Ida, 1972) resulting in their smaller magnitudes. Class 2 events occur in the gob, the area of collapsing rock at greater distances behind the active face, and during closure of the void created by mining of the coal. The locations of Class 2 events are likely controlled by the geometry of the panel, and the nature of the roof rock. The Class 2 events may result from slip on pre-existing zones of weakness, such as fractures, bedding planes and faults (Richardson & Jordan, 2002), in response to large deviatoric stresses induced by the subsidence of the overburden over the panel void.
4.1. Location Quality of Event Hypocenters and Panel Events

An assessment of the quality of hypocenter locations is essential for spatial analysis. Each event is assigned a location quality, based on the number of phases used to locate the events, the epicentral distance from the closest recording station to the epicenter, the arrival time variance (root-mean-square value of the travel-time residuals), and the maximum azimuthal gap between the recording stations viewed from the epicenter. Table 4.1 shows the criteria for the quality assignment.

<table>
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<th>Condition</th>
<th>Description</th>
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<tr>
<td>C</td>
<td>Gap &gt; 180 deg. Dmin &lt; 3000 ft, N &gt; 4</td>
</tr>
<tr>
<td>D</td>
<td>Gap &gt; 180 deg. Dmin &lt; 3000 ft, N = 4</td>
</tr>
<tr>
<td>D</td>
<td>Gap &gt; 180 deg. Dmin &gt; 3000 ft</td>
</tr>
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**Dmin**: Distance of closest station to epicenter of event.

**Gap**: Maximum azimuthal gap between stations, viewed from epicenter.

**N**: Number of non-zero weighted arrival times after 3rd iteration of the location algorithm. Residuals greater than one second are given zero weight after the first iteration. Residuals greater than 0.1 seconds are given zero weight after 2nd iteration.

During mining, defined as the period when the continuous miner is within a given panel, the seismicity within that panel may change due to changes in stress. It is of interest to compare seismicity with the changing panel-gob geometry. For further analysis, the hypocenter locations with quality A and B were used to determine the spatial statistics. The catalog of events was filtered to determine which events occurred within the three panels of interest.

4.2. b-values

The relationship between the frequency of occurrence and the magnitude of seismic events has been a subject of interest to seismologists for decades. Gutenberg and Richter (1941) noted that the frequency of earthquakes, conditional on their magnitude, shows a power-law relationship. Various magnitude scales have been used in the past in connection with Equation...
4.2. Current practice is to employ the moment magnitude scale when estimating the Gutenberg-Richter recurrence relationship for a given region.

<table>
<thead>
<tr>
<th>Equation 4.2</th>
<th>( \log_{10} N = a - bM_w )</th>
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<tbody>
<tr>
<td>N is the cumulative number of earthquakes</td>
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<tr>
<td>( M_w ) is the moment magnitude, held constant</td>
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<tr>
<td>a is the y intercept and is constant</td>
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<tr>
<td>b is the slope of the distribution</td>
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On the global scale, the slope of the Gutenberg-Richter recurrence relationship (b-value in Equation 4.2) is approximately 1.0. Observations of variation in the b-value over small regions, and in different tectonic environments, have spawned many studies. For example, some studies report variations in the “b-value” for induced earthquakes in areas undergoing fluid injection. Most seismologists assume that the b-value is related to the size distribution of potential failure surfaces, loading rates and fault friction.

In this study, the spatial and temporal behavior of the b-value was examined at different stages of the mining process for three of the panels in the mine. The motivation was the possibility that the systematic variation of the b-value might reflect systematic differences in the source processes of the seismic events. The following figure shows the log N versus M relationship determined from earthquakes that occurred while there was active mining within the D, E and G panels, in comparison with the distributions derived after active mining in the D and E panels.
Figure 4.1: b-values for panels D, E and G during mining (A, B and C) and, for D and E, after mining (D and E). During mining, there are 2673 events in panel D with a b-value of -2.4, 2864 events in panel E with a b-value of -1.5 and 305 events in panel G with a b-value of -2.1. After mining, there are 812 events in panel D with a b-value of -1.1 and 87 events in panel E with a b-value slope of -0.9. The black lines represent b-values of -1.
Figure 4.1 indicates that the b-values of seismicity within a given panel changes dramatically after the panel has been mined. This change is indicative of two event classes. During active mining of a panel, there are many Class 1 events in the immediate vicinity of the active face resulting in b-values that are more negative. After mining, the b-values for panels D and E are similar to the expected b-value of one for tectonic earthquakes. Thus, in terms of the frequency-magnitude distribution, Class 2 events (i.e., gob events that occur after mining) appear to be similar to tectonic earthquakes, whereas Class 1 events are different.

4.3. Minipanel Events

To further examine the time-space characteristics of the seismicity, panels D, E and G were each subdivided into 10 equal areas or “minipanels.” Events were filtered by minipanel location using the same procedure described above (Figure 4.2).
The number of events with A and B quality and moment magnitudes greater than zero that occurred during the period from March 1, 2009 to April 7, 2011 in D, E and G panels are plotted in Figures 4.3 through 4.5. The minipanels are numbered sequentially from south to north (see Figure 4.2 above). Dots in Figures 4.3 through 4.5 indicate the number of events in each minipanel. It is apparent that the frequency of seismicity is highly variable along the individual panels. The elevation of the ground surface is plotted along with the frequency of seismic events in each minipanel in Figures 4.3 through 4.5. The mine level is approximately 500 feet above sea level, with relatively small elevation changes along the length of the panels. Figures 4.3 and 4.4 show that the frequency distribution of events along the longitudinal axis of each panel correlates with the thickness of overburden. Overburden thickness range from a
minimum of about 1,300 feet up to a maximum of just over 2,200 feet. Lithostatic pressure varies by 40 percent in panel D, 35 percent in E, and 43 percent in G due to changes in overburden thickness.

Figures 4.3 and 4.4 appear to show that panel E is approximately three times more active than panel D. The network detection capabilities changed during the mining of these two panels. Telemetry problems affected data recovery during approximately one-half of the period when panel D was being mined. Under a worst-case interpretation (zero events detected), one-half of the events that would potentially have been detected and recorded in panel D were actually cataloged. No significant telemetry dropouts occurred during the mining of panel E. Therefore, we would expect panel E based on network performance alone to be appear approximately twice as active as panel D. Despite this difficulty, it appears that panel E actually was significantly more active than panel D. The increase in seismicity associated with the mining of panel E may be due to a larger mined area (gob) adjacent to the E panel, than was the case during mining of D or G. Figure 1.1 shows that during the mining of panel D, the gob to the left only consists of the mined B and C panels. However, by the time E was mined this gob area had increased substantially to include the panel D overburden as well. There is a possibility that (seismicity) increased in relation to gob size. However, this is not firmly established because of the above mentioned temporal changes in the detection capabilities of the network.

In contrast, panel G does not exhibit a correlation of seismicity with overburden thickness (compare Figure 4.5 to Figures 4.3 and 4.4). This may be related to the introduction of a coal barrier, panel F, which may have reduced stresses in panel G resulting in significantly fewer events than in either the D or E panels. Seismicity appears to be a very complex phenomenon that may involve the geometry of the gob adjacent to an actively mined panel, as well as overburden thickness and likely additional complications due to geological variability.
Figure 4.3: All events located in panel D that are A and B quality and have $M_w > 0$. Notice the maxima are near the center of the panel (minipanels 4-6). The change in frequency of events is minimally affected by the reduction of the acceptable location quality range. The blue line is the topographic profile over the panel.
Figure 4.4: All events located in panel E that are A and B quality and have $M_w > 0$. Notice the maxima are near the center of the panel (minipanels 5-7). Also note, this panel is much more active than panel D (see Figure 4.3 above). The change in frequency of events is minimally affected by the reduction of the acceptable location quality range. The blue line is the topographic profile over the panel.
Figure 4.5: All events located in panel G that are A and B quality and have $M_w > 0$. Compared to Figures 4.3 and 4.4, the number of events does not follow the topography trend as well. There is also an anomalous position at minipanel 8 but in this case, the number of events is lower than expected. The blue line is the topographic profile over the panel.

4.4. Time-Space Behavior of Seismicity in the Three Panels

Figure 4.6 shows the number of events located in each D minipanel on a weekly basis, during the period when the active face was within the D panel. Plot A in Figure 4.6 shows all events with $M_w > 0$ in the D panel regardless of face position. Plot B shows only events occurring in a given minipanel when the face was located in that minipanel (Class 1). Plot C shows only events in a given minipanel subsequent to the advance of the face outside that minipanel (Class 2). Mining commenced in mid-September 2009. However, the first events began to occur in early October. These events occurred in the minipanel containing the active face (face events) and in adjacent minipanels to the south (gob events). Seismic activity reached a maximum during December 2009. Seismicity began to decrease in January 2010, and face activity essentially ceased during early January 2010 until early March 2010. Note that, in contrast, gob
activity was largely continuous subsequent to early October 2009, primarily in minipanels two through six. Minipanels 4, 5 and 6 exhibited the largest numbers of both face and gob events and were beneath the thickest overburden. In the case of panel D, it appears that overburden thickness may have played an important role in the frequency of face events, as well as the persistence of gob events. Areas of thick overburden appear to result in high numbers of face events (minipanels 4 through 6, middle plot in Figure 4.6) and define areas where gob events persist long after the face has advanced away from those areas (bottom plot of Figure 4.6).
Figure 4.6: Weekly distribution of events per minipanel during mining of the panel. Minipanels show the time variance of the mining. Mining in panel D began on September 16, 2009 and ended April 10, 2010. Plot A shows all events that occurred during mining, plot B shows only face events and plot C shows only gob events. The black column shows the mining face position.
Figure 4.7 is analogous to Figure 4.6 but for panel E events. Similar to the situation in panel D, seismicity in E began about one month after mining of the panel began. There was a rapid increase in seismicity beginning in July 2010 that lasted into September 2010, involving both face and gob events. The gob events were most active in minipanels 5, 6 and 7. Here the correlation with topography is not as clear as was the case with the D panel. Minipanel 7 is to the north of the area with thickest overburden. It appears that the majority of events in the E panel are shifted slightly (one minipanel) to the north of the area of highest surface elevation. This suggests that overburden thickness may not be the sole controlling factor in regards to the frequency and location of seismic events. Figure 4.3 and Figure 4.4 show that, to a first approximation, the frequency of events along the axis of the panels has a crude triangular shape, with the maximum number of events near the center of the panel. It turns out that the thickest overburden is also near the centers of the panels. Thus, the geometry of the underground excavation and the variable thickness of overburden may act in combination to create the stresses responsible for the seismicity. Of course, lateral variability of material strength of the roof strata is also expected to play a role in this complex problem.
Figure 4.7: Weekly distribution of events per minipanel during mining of the panel. The events are divided into minipanels to show the time variance of the mining. Mining in panel E began on April 21, 2010 and ended October 11, 2010. Plot A shows all events that occurred during mining, plot B shows only face events and plot C shows only gob events. The black column shows the mining face position.
Figure 4.8 shows the event distribution for panel G. Seismicity began about one month after initiation of mining, as was the case with panels D and E. However, G behaves differently concerning gob events. In panels D and E, the gob events decay with distance from the active mining face gradually, whereas in G, they decay quickly with respect to distance and there are fewer late-stage events in the G gob. Gob events in G are most frequent in minipanels 5, 7 and 9 while face events are most frequent in minipanels 5, 6, and 7. There is poor correlation between overburden thickness and the number of events in the G panel (Figure 4.5).

The different behavior observed in panel G is likely related to the coal barrier (unmined panel F) separating the G gob from the larger gob formed by the mining of panels B, C, D and E. This barrier may have reduced deformation in the G gob roof area, as evidenced by the lack of gob events and the reduction in the number of face events in comparison to the previously mined panels. The presence of the barrier seems to have reduced the effects caused by variable overburden thickness (changing topography).
Figure 4.8: Weekly distribution of events per minipanel during mining of the panel. The events are divided into minipanels to show the time variance of the mining. Mining in panel E began on October 17, 2010 and ended April 7, 2011. Plot A shows all events that occurred during mining, plot B shows only face events and plot C shows only gob events. The black column shows the mining face position.
The spatial distribution of seismicity in the D, E and G panels, during the period when each panel was mined is a combination of face events and gob events. Figure 4.9 shows the numbers of events as a function of distance from the active face in the panels, during the period when each panel was being actively mined. There is a general exponential decay in the number of events with distance from the face. Fifty percent of the events occurred within 1200 feet (D), 800 feet (E) and 400 feet (G) of the active mining face. The ninety percentile distances are 3,800 feet, 3,400 feet and 1,200 feet for D, E and G respectively. Thus, the spatial frequency of events as mining progresses along the panel appears stable: i.e., the numbers of events decay rapidly with distance from the face, in a manner similar for both panels. This observation is noteworthy because of the large overall difference in seismicity rates: panel E was much more active than D or G during mining.

Figure 4.10 shows the result. The cumulative moment release versus distance from the active face, using the filtered catalog, is very similar to the results shown in Figure 4.9: both frequency of events, and cumulative moment versus distance from the active face show very similar trends. Fifty percent of the moment release contained in the filtered catalog occurred within 1,600 feet, 800 feet, and 400 feet of the active face during mining of panels D, E and G respectively. The ninety percent distances for cumulative moment release (filtered) are 4,200 feet, 3,600 feet, and 1,200 feet for D, E and G, respectively. This similarity between the frequency of events and cumulative moment when plotted versus distance from the active face may imply a spatially constant b-value, for a given panel during mining of that panel.
Figure 4.9: Plots A, B and C display the number of events as a function of distance from the active mining face for panels D, E and G. Plots D, E and F are the cumulative number of events percentage as a function of distance from the active mining face in the three panels.
Figure 4.10: Plots A, B and C display the seismic moment for events shown in Figure 4.9. Notice the increased seismic moment in panel E compared to D and G. The moment data shows a different trend for D; this may be due to, at least in part, to two relatively large events. If the largest 10% of events are removed (plots D, E and F) from the data set, a similar trend is shown for all panels. This trend shows a dependence on distance from the active mining face.
The depth dimension of the hypocenter location has much more uncertainty than the horizontal dimension. Accurate focal depth determination with a small network is a difficult problem particularly in situations where most stations lie at epicentral distances greater than the focal depth. Figure 4.11 shows the depths of events based on their distance from the active face. The events shown in Figure 4.11 are A and B quality representing the best-located events. In some cases, the uncertainty in the focal depth can exceed the focal depth and some of the scatter in Figure 4.11 is due to random noise in the depth determination. However, a trend in the depths of the events is apparent. The depths shown in Figure 4.11 for both the D and E panels exhibit a trend that suggests that events become more shallow with distance from the face, consistent with increasing time from the mining of that section of the panel. When active mining occurs, the events are located near the mine level (500 ft msl) but with time the events become more shallow up to depths 500 to 1000 feet above the mine over a period of about two months (approximately 4000 ft of mining). However, panel G did not follow this trend; the majority (90%) of events remained within 1200 feet of the active face and did not show an upward shallowing trend.
Figure 4.11: Elevations, with respect to sea level, of events based on their distance from the active mining face. Mining occurs approximately 500 feet above sea level. The average ground surface elevation over the mine is approximately 2,200 feet. The ground surface elevation datum assumed by the location program is 2,600 feet msl.
CHAPTER 5. DISCUSSION

Due to the rapid advance of active faces, underground monitoring networks are difficult to operate in coal mines. This study shows that a deep underground, longwall coal mine can be successfully monitored with a relatively sparse network of surface stations. This study demonstrates that epicenter locations using as few as five stations with good azimuthal coverage can be determined with sufficient accuracy to identify, in the horizontal plane, the location of the active longwall face position to within approximately 300 feet. The hypocenter depth accuracy is much less, but is sufficient to resolve major trends in the spatial distribution of events as mining progresses.

The seismicity at the study mine is spatially associated with the geometry of the mining panels and can be characterized as consisting of two classes of events. The volume of rock in the immediate vicinity of the advancing, active mining face produces the most events. The frequency of these Class 1 or "face events" decreases rapidly with distance into the gob area behind the active face. The Class 1 events exhibit "b-values" smaller (more negative) than the value of negative one typical of tectonic earthquakes, indicating a high frequency of smaller magnitude events. The events in the gob, well behind the advancing face, represent Class 2 events. These events are relatively infrequent compared to the intense Class 1 seismicity near the active face, but represent the larger events observed to date at the mine. The gob events can persist for months following the active mining at a given location. The frequency-magnitude distribution of Class 2 events, derived from the long-term gob events in two different panels subsequent to the mining of those panels, is similar to that of tectonic earthquakes (b-value near -1). Also, whereas the Class 1 (face events) appear to be concentrated near the active mine workings (longwall face), the Class 2 (gob events) occur along or parallel to the longitudinal axis of the mined panel and focal depths may decrease with time. The focal depth accuracy of the monitoring network is not sufficient to unequivocally resolve the maximum elevations of these late-stage gob events. Based on data gained from monitoring two adjacent panels, it appears that the late-stage gob events mostly occur at elevations between 1,000 and 2,000 feet, msl. The average elevation of the ground surface is 2,600 ft msl, and the mine is at 500 feet msl. The apparent location of the late-stage gob events well above mine level may reflect the upward migration of the deformation process that ultimately leads to surface subsidence.
The frequency of both Class 1 and Class 2 events depends upon location along the longitudinal axis of the D and E panels. Frequency of events appears highest near the centers of the panels, and decreases toward the ends. From a geometrical perspective, one might expect an increase of stress at the active face as it advances to the vicinity of the center of the adjacent panel gob, followed by a gradual decrease as it advances past it. However, there is also a correlation between the frequency of events and overburden thickness, in that areas of thickest overburden correlate with positions along the panel with the highest seismicity (both Class 1 and Class 2). The complication is that the maximum overburden thickness is near the centers of the D and E panels. Therefore, it is unclear whether panel-gob geometry, overburden thickness, or sandstone thickness influences seismicity. It is likely that all three of those factors contribute.

It appears that the increasing width of the adjacent gob area in deep cover could cause an increase in seismicity. The E panel appears to have been more active than either panel D or G. Panels D and E represent conditions involving a large gob area adjacent to an actively mined panel. The character of the seismicity is changed by the introduction of a wide barrier between active and previously mined panels. The behavior observed in the G panel indicates that the effect of the barrier reduces seismicity in the gob area distant from the active mining face and causes a general reduction in the number of face events. The barrier may also mask the effect of overburden thickness, since stresses were observed to be reduced in the gob area of the active panel. In such cases, the seismicity may be controlled by variability in the strength, proximity, and thickness of the overlying strata.

The seismic signals recorded at close distances from the shallow mine-induced seismicity are very poor in high-frequency energy, compared to tectonic earthquakes of similar magnitude. This appears to be the result of two factors. The first is the very low value of $Q = 37$ estimated for the overburden rock from the slope of the high-frequency amplitude spectra of the horizontal component acceleration records. Values of $Q$ less than 100 are typically found in areas where seismic waves propagate through poorly consolidated sediments, such as the Cretaceous and Cenozoic sediments in the Atlantic and Gulf coastal plains (Chapman et al., 2008). These materials have P-wave velocities in the range 1.5 to 2.0 km/s and S-wave velocities less than 1.0 km/s. At the mine, the P and S-wave velocities are ~4.4 and 2.5 km/s, respectively, measured as an average over the entire ray path. Those values seem to imply higher $Q$ than observed.
Assuming that the measurements of $Q$ are accurate, I speculate that fracturing of the rock mass in the gob contributes to scattering that may be responsible for the low $Q$ measurements. In addition, the source corner frequencies of the mine-induced events appear to be in the range five to 10 Hz for the larger mining events, which imply static stress drops of 10 bars (0.1 MPa) or less for these events. This estimate of stress drop is low compared to values reported for tectonic earthquakes, and for some studies of mining induced events in deep, hard-rock mines. In addition, the P-wave first-motions of the events exhibit an almost ubiquitous dilatational first motion, indicating a significant implosional component in the seismic source. Thus, the mining related events at this coal mine are probably occurring in a highly heterogeneous, crack filled medium, ultimately as the result of gravitational potential energy being converted to strain energy during closure of the mine excavation.
REFERENCES


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