

Passive Seismic Tomography and Seismicity Hazard Analysis in Deep Underground Mines

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ABSTRACT

Seismic tomography is a promising tool to help understand and evaluate the stability of a rock mass in mining excavations. It is well known that closing of cracks under compressive pressures tends to increase the effective elastic moduli of rocks. Tomography can map stress transfer and redistribution and further forecast rock burst potential and other seismic hazards, which are influenced by mining. Recorded by seismic networks in multiple underground mines, the arrival times of seismic waves and locations of seismic events are used as sources of tomographic imaging surveys. An initial velocity model is established according to the properties of a rock mass, then velocity structure is reconstructed by velocity inversion to reflect the anomalies of the rock mass. Mining-induced seismicity and double-difference tomographic images of rock mass in mining areas are coupled to show how stress changes with microseismic activities. Especially, comparisons between velocity structures of different periods (before and after a rock burst) are performed to analyze effects of a rock burst on stress distribution. Tomographic results show that high velocity anomalies form in the vicinity of a rock burst before the occurrence, and velocity subsequently experiences a significant drop after the occurrence of a rock burst. In addition, regression analysis of travel time and distance indicates that the average velocity of all the monitored regions appears to increase before rock burst and reduce after them. A reasonable explanation is that rock bursts tend to be triggered in highly stressed rock masses. After the energy release of rock bursts, stress relief is expected to exhibit within rock mass. Average velocity significantly decreases because of stress relief and as a result of fractures in the rock mass that are generated by shaking-induced damage from nearby rock burst zones. The mining-induced microseismic rate is positively correlated with stress level. The fact that highly concentrated seismicity is more likely to be located in margins between high-velocity and low-velocity regions demonstrates that high seismic rates appear to be along with high stress in rock masses. Statistical analyses were performed on the aftershock sequence in order to generate an aftershock decay model to detect potential hazards and evaluate stability of aftershock activities.

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Contents

List of Figures	vi
List of Tables	viii
1 Introduction.....	1
2 Literature Review.....	3
2.1 Seismicity and Rock burst Potential	3
2.2 Seismic Energy	7
2.3 Aftershock.....	12
2.4 Tomography	16
2.5 Summary	23
Chapter 2 References	24
3 Seismic velocity change due to large magnitude events in deep underground mines	29
3.1 Abstract.....	29
3.2 Introduction.....	29
3.3 Velocity and seismicity investigation	30
3.4 Velocity fit using mining-induced seismicity	31
3.5 Case Study	32
3.6 Summary and discussion.....	39
Chapter 3 References	41
4 Imaging of Temporal Stress Redistribution due to Triggered Seismicity at a Deep Nickel Mine.....	43
4.1 Abstract.....	43
4.2 Introduction.....	43
4.3 Data and methods.....	44
4.4 Results.....	48
4.5 Summary and discussion.....	54
Chapter 4 References	57
5 Stress redistribution with the change of frequency of mining-induced seismicity	59
5.1 Abstract.....	59
5.2 Introduction.....	59
5.3 Data and methods.....	60
5.4 Results.....	64
5.5 Summary and discussion.....	73
Chapter 5 References	75
6 Passive tomographic study on velocity changes in underground mines	77

6.1	Abstract	77
6.2	Introduction.....	77
6.3	Large magnitude events	78
6.4	Average velocity analysis in tabulation areas	80
6.5	Summary and discussion.....	87
	Chapter 6 References	89
7	Statistical analyses of mining-induced seismicity from deep hard-rock mines	91
7.1	Abstract	91
7.2	Introduction.....	91
7.3	Data of mining-induced seismicity and analysis procedure.....	93
7.4	Results.....	97
7.5	Summary and discussion.....	102
	Chapter 7 References	104
8	Conclusions and Discussion.....	106

List of Figures

Figure 3.1. Linear fitting for single event of Creighton Mine	32
Figure 3.2. Linear fitting of travel time with travel distance	34
Figure 3.3. Residuals of robust linear fitting and ordinary linear fitting for Creighton Mine	35
Figure 3.4. Velocity change affected by large magnitude events in Creighton Mine.....	36
Figure 3.5. Residuals of robust linear fitting and ordinary linear fitting for Kidd Mine	37
Figure 3.6. Velocity change affected by the first large magnitude event in Kidd Mine	38
Figure 3.7. Velocity change affected by the second large magnitude event in Kidd Mine	39
Figure 4.1. The large velocity model.	47
Figure 4.2. The small velocity model	47
Figure 4.3. Cumulative number of events.....	48
Figure 4.4. The side view (top) and plan view (bottom) of the major events in the study area. The main drifts are plotted and noted with depth.Red balls: major events.	49
Figure 4.5. Tomography studies on large velocity model. Red trigangles: sensors in the seismic network.	51
Figure 4.6 Tomography studies on small velocity model. Red triangles: sensors in the seismic network.	54
Figure 5.1. Sensors location and raypaths coverage in Creighton Mine.....	61
Figure 5.2. Cumulative number of microseismic events in Creighton Mine	62
Figure 5.3. Cumulative number of microseismic events in Kidd Mine	63
Figure 5.4. Frequency-magnitude relationship for the Creighton Mine	64
Figure 5.5. Frequency-magnitude relationship for the Kidd Mine	64
Figure 5.6 Cross-Section of Creighton Mine	66
Figure 5.7. Velocity on level and seismicity distribution in July of Creighton Mine	67
Figure 5.8. Velocity on level and seismicity distribution in August of Creighton Mine	69
Figure 5.9. Cross section of Kidd Mine	70
Figure 5.10. Tomograms of Kidd Mine	73
Figure 6.1. Depth range of seismic events in Creighton Mine.....	78
Figure 6.2. Depth range of seismic events in Kidd Mine.....	80
Figure 6.3. Mesh grids of velocity model	81
Figure 6.4. Residuals of Creighton Mine (a) and Kidd Mine (b).....	81
Figure 6.5. Average velocity of cubes by velocity inversion of 2000 microseismic events of Creighton Mine	83
Figure 6.6. Average velocity of cubes by velocity inversion of Creighton Mine	84

Figure 6.7 Average velocity (Event 1) of 500 microseismic events of Kidd Mine	86
Figure 6.8. Average velocity (Event 2) of 500 microseismic events of Kidd Mine	87
Figure 7.1. Cumulative numbers of aftershocks are given as functions of magnitude for (a) the first mainshock in Creighton Mine and (b) the first mainshock in Kidd Mine. The solid straight line are the best fit of Gutenberg-Richter relation.	95
Figure 7.2. Change of b-values and number of aftershocks with the magnitude of completeness for (a) first mainshock of Creighton Mine (b) second mainshock in Creighton Mine. The cumulative magnitude distribution curve is approximately linear for $M > -1.5$ for both the aftershocks sequences. The b-values calculated by the least-square method and maximum likelihood method at magnitude = -1.5 are significantly closer than that with most other magnitudes.	96
Figure 7.3. Change of b-values and number of aftershocks with the magnitude of completeness for (a) first mainshock in Kidd Mine (b) second mainshock in Kidd Mine. The cumulative magnitude distribution curve is approximately linear for $M > -1.5$ for both aftershocks sequences. The b-values calculated by the least-square method and maximum likelihood method at magnitude = -1.5 are significantly closer than that with most other magnitudes.	96
Figure 7.4. Temporal change of the b-values for (a) one week scale (b) two weeks scale. Continuous line with open circles: b-values through all time periods; Asterisk: the specific time period with occurrence of mainshocks. The period noted by an asterisk includes the time of the mainshocks in July, 2011.	98
Figure 7.5. Creighton Mine cumulative number of aftershocks decay with time modeling for (a) the first mainshock (b) the second mainshock. Solid lines: practical cumulative number of aftershocks; Dash lines: fitted model using Generalized Omori's law.	100
Figure 7.6. Kidd Mine cumulative number of aftershocks decay with time modeling for (a) the first mainshock (b) the second mainshock. Solid lines: practical cumulative number of aftershocks; Dash lines: fitted model using Generalized Omori's law.	102

List of Tables

Table 3.1. Times, Locations, and Magnitude of Major Events in Creighton Mine	33
Table 3.2. Times, Locations, and Magnitude of Large Major Events in Kidd Mine	36
Table 4.1. Times, Locations, and Magnitude of Major Events	45
Table 4.2. Time frames with major events occurrence	49
Table 5.1. Seismicity rate of July in Creighton Mine	65
Table 5.2. Seismicity rate of August in Creighton Mine	68
Table 5.3. Seismicity rate in Kidd Mine	69
Table 6.1. Times, Locations, and Magnitude of Major Events in Creighton Mine	79
Table 6.2. Events Grouped for tomographic studies in Creighton Mine	79
Table 6.3. Times, Locations, and Magnitude of Major Events in Kidd Mine	80
Table 6.4. Events Grouped for tomographic studies in Kidd Mine	80
Table 7.1. Times, Locations, and Magnitude of Mainshocks in Creighton Mine.....	94
Table 7.2. Times, Locations, and Magnitude of Mainshocks in Kidd Mine.....	94

1 Introduction

To detect seismic hazards and improve safety in mines, a great number of seismic networks have been developed and used with an increasing depth of excavations. Considerable data sets are generated because of the good coverage of seismic monitoring systems in mines. It is an important issue to continue improving the extraction and analysis of the information from mining-induced seismicity and to comprehensively assess the seismic hazards. The professional evaluation of seismic risks of mines requires the incorporation of geophysical analysis methods and tools building on the fact that mining-induced seismic monitoring systems are designed based on the experiences of using seismic networks in natural earthquakes. Besides statistically-based studies on mining-induced seismicity, passive seismic tomography provides the engineer with an understanding of how stress changes temporally and spatially. Such information is fundamental for detecting potential rock seismic hazards and improving safety in mines.

The time-dependent analysis of the passive seismic tomography of the mines reveals changes in the relative state of stress in a rock mass. Passive seismic tomography is an imaging technique used to evaluate the velocity distribution of the P wave which crosses through the rock mass. P wave velocity varies with the different bulk moduli of a rock mass. For the same area, bulk modulus change is mainly impacted by stress state depending on both the geological structure and mining activities including excavating, blasting, and rock supports. Thus, velocity change in the rock mass reflects the stress distribution and redistribution.

Widespread application of microseismic monitoring systems guarantees the availability of tomography. Microseismic monitoring is a way of remote sensing to discriminate between the geological structures including shear zone and fault zone, as well as other discontinuities. Generally, the initial goals of microseismic monitoring are to identify the geological structure and potentially unstable rock mass by locating the seismic events. However, microseismic monitoring is less comprehensive since the seismic events oriented analysis is only restricted on the discrete events related area. Passive seismic tomography provides a way to further utilize the information of microseismic activities to yield velocity distribution around the mine area. According to the velocity distribution displayed on tomography, the relative state of stress would be evaluated based on how the tomographic images change. Associated with the in situ stress state measured by

geotechnical instrumentation, velocity can be scaled to indicate the corresponding quantitative value of stress.

A double difference algorithm is applied in velocity inversion on computing tomography. Initial events location and absolute arrival time are available from the microseismic monitoring system. Travel times from the single event source to receivers, which captured the event, are computed. Background velocity is determined by the source–receiver distance and travel time for the velocity inversion. Differential catalog arrival times are subtracted from cross correlated waveforms. Both absolute arrival times and differential arrival times are used for velocity inversion. Velocity structure and seismic event locations are mutually solved to generate a velocity model that represents subsurface structure.

To detect potential seismic hazard, studies on crustal earthquakes have established frameworks for essential seismic characteristics and data analyses. Besides double-difference tomography, b-value has been proved the potential to be a predictor to forecast earthquakes. Applications of the b-value and aftershock decay models could be useful to improve safety in mines.

2 Literature Review

2.1 Seismicity and Rock burst Potential

An earthquake is a result of sudden shear planar failures in rock material (Aki and Richards 2002). Earthquake ruptures start at the hypocenter, the origin point propagating seismic waves on the fault-plane leading to the dislocation of two sides of the fault surface. The rupture of earthquakes excites seismic waves, including primary, compressional (P-waves) and secondary, shear waves (S-waves). Earthquakes are subdivided into shallow, intermediate, and deep earthquakes in terms of the depth of hypocenter. Evidence shows that shallow events exhibit a good grouping property following time (Kagan 1994).

Focal mechanism analysis is a standard method in seismology. It indicates the sense of first motion of P-waves, solid for compressional motion and open for dilatational (Kagan 1994). The mainshock is the large event that dominates the sequence of events. The sequences of small earthquakes following large earthquakes are called aftershocks. Also, earthquakes are preceded by foreshocks, which consist of weaker events. Foreshocks are usually much less frequent than aftershocks. Stress transfer is the main cause of aftershocks in the adjacent regions of a mainshock. Aftershocks relax the excess stresses, which are caused by a mainshock (Shcherbakov, Turcotte et al. 2005).

Theoretical understanding of the earthquake process and interrelations between earthquakes are analyzed by using statistical methods. Standard analytical methods are not applicable because of the intrinsic randomness of seismicity. Kagan (1994) proposed that statistical models are used to explore seismic properties, including stochastic, multidimensional tensor-valued, and point process. Statistical methods are necessary for seismicity research because of the randomness of seismicity. Case studies of earthquakes can only provide understandings for partial earthquake sequences. Seismicity can be observed and displayed with long-term and long-range correlations by using statistical analysis (Kagan 1994).

One of the methods widely used in seismicity study is detailed analysis of seismograms, which can provide a complex internal temporal and spatial structure of earthquake sources. Geometry of faults could be observed by geological investigations. Preseismic and postseismic deformations are usually involved with statistical analysis combined with seismic events if abundant and uniform coverage of seismic events are available (Kagan 1994). The most prominent feature of

earthquake catalogs is the spatial and temporal grouping of earthquakes. Earthquake groups are divided into foreshock, mainshock, and aftershock sequences. Utsu (2002) provides examples to demonstrate distribution of various earthquake sequences. Also, he mentioned that the aftershock epicenters scatter more widely than mainshock epicenters (Utsu and Seki 1954). Utsu and Seki (1955) introduce the relation between earthquake magnitude and the aftershock area as:

$$\log S = M_m - 3.9 \quad (2-1)$$

where S is the aftershock area in km^2 for a magnitude M_m .

Aftershocks are located near the margin of the fault areas which have experienced dislocation and displacement. However, as mentioned before, the surface rupture length and magnitude fail to keep the same consistency for the earthquake (Utsu 2002). Some studies obtain the empirical relationship between the magnitude and the length of fault rupture and show that the coefficient of magnitude ranges from 0.78 to 1.22 (Wells and Coppersmith 1994). For smaller earthquakes, the study is performed by Iio (1986) and shows:

$$\log L = 0.43M_m - 1.7 \quad (2-2)$$

where L is the length of aftershock area for a mainshock with magnitude M_m .

The reason aftershock occurs is that the mainshock increases regional stresses, which lead to the subsequent aftershocks (Shcherbakov, Turcotte et al. 2005). According to the dislocation model, regional stresses arise from the strain accumulation and release. It is widely known that aftershocks are the stress transfer to the vicinity of the hypocenter during the occurrence of mainshocks. Aftershocks are associated with the relaxation of stresses arising from mainshocks. Regions around the rupture of a mainshock are associated with stress increase that are greater than the yield stress. The aftershocks after the mainshock lead to stress relaxation associating with microcracking until achieving the yield stress. This transient process is interpreted by using damage mechanics combined with Omori's law (Shcherbakov, 2005).

Aftershocks satisfy the following laws (Shcherbakov, Turcotte et al. 2005):

1. It is found that Gutenberg-Richter frequency magnitude relation can be applied to the aftershocks sequence.

2. According to Båth's law, the magnitude difference between a mainshock and its greatest aftershock keeps roughly constant. The constancy corresponds to the scale-invariant aftershock along with mainshocks. That is:

$$\Delta m = m_{ms} - m_{as}^{max} \quad (2-3)$$

where m_{ms} is the magnitude of the mainshock, m_{as}^{max} is the magnitude of the largest aftershock, Δm is approximately equal to 1.2.

3. The temporal decay rate of occurrence of the aftershock sequence follows the modified Omori's law.

As the edge or face of the stope is advanced, the rock in the vicinity of the excavations is stressed approaching or even beyond its elastic limit, leading to inelastic deformation of the rock mass (McGarr 1971). It is well known that mine excavations influence the virgin state of stress and even lead to stress concentrations in the rocks. Seismic pulses are triggered when the rocks subject to stress approach or exceed the strength of the rock. The occurrences of seismicity are associated with the mine excavations and can be forecasted by using statistical analyses (Cook, 1976).

Rockbursts are mining-induced seismic events ($1.5 < M < 4.5$) that lead to violent and dynamic failure of the rock mass. Mining excavation on highly stressed rock mass can cause stress distribution and result in rockbursts (Young, Maxwell et al. 1992). It is shown that the occurrences of seismicity follow the progress of active mining. However, the seismicity cannot directly reflect the major faults in the mined area (Kusznir, Ashwin et al. 1980). The analysis of tomography and seismic events behavior with mining excavation could improve our knowledge of the influence of stress change on the velocity structure of a rock mass. Evidence supports the view that the seismicity and rock deformation are shown to be closely related in space and in time (McGarr and Green 1975). Tilt is the first derivative of the subsidence profile and can show the change in subsidence between two points. It is found that the rate of occurrence of tremors is closely correlated with the rate of tilting in a deep level gold mine (McGarr and Green 1975). As a result of compressional stresses with mining progresses, rock masses above and below the stopes converge (McGarr 1976).

Examination of the pre-burst and post-burst seismic data suggest that, a significant increase of microseismic events is followed by a dramatic decrease before the rock burst (Brady and Leighton 1977). To predict impending failure, seismicity has to be qualified on some conditions. A low-modulus inclusion zone is called primary inclusion zones. It is required that seismicity increases with the primary inclusion developing. Also, the primary inclusion zone should exhibit anomalously long ruptures along with seismic events (Brady and Leighton 1977). The seismic tomography study developed by Young and Maxwell implies that low-velocity regions are of low rock burst potential and high-velocity areas are of high rock burst potential. Tomographic imaging and induced seismicity were integrated to characterize highly stressed rock mass in rock burst prone mines. Locations of seismicity (foreshock and aftershock of rock burst) are superimposed on tomographic images to indicate the correlations between velocity anomalies and seismic events by analyzing the velocity structure and seismicity. It is indicated that a destressing as a result of rock burst of the sill pillar is represented by low velocity anomalies.

In contrast, mining-induced tremors and rock burst displayed a high velocity region (Young and Maxwell 1992). Considerable evidence from seismic data indicate that mine tremors are physically similar and part of them obey the same mechanisms to earthquakes. It is noted by McGarr that many mine tremors and earthquakes are the result of shear failure along a plane (McGarr 1971). Consequently, mine tremors and earthquakes obey the same Gutenberg-Richter magnitude-frequency relation (McGarr 1971). Stress drops are observed in natural earthquakes and the magnitude of stress change ranges from 0.3 to 50 MPa (Ishida, Kanagawa et al. 2010). The earthquake stress drop is regarded as a fraction of the regional differential stress because evident stress disturbance, such as pronounced rotation of principal stress axes, is associated with coseismic deformation (Hardebeck and Hauksson 2001). Underground observations indicate that an instantaneous convergence is associated with burst fractures. Convergence continues to be followed by a period of rapid convergence, which gradually reduces to the normal rate (Hodgson 1967, McGarr 1971). It is proposed that the rate of stope convergence decreases prior to a large seismic event, which promotes the dislocation of rock in stopes (Hodgson 1967).

The frequency of event occurrence and magnitude of events conform to the log-linear function based on power-distribution law. A drop of β value can indicate some major events and an abrupt increase of effective stress prior to the occurrence of them (Young, Maxwell et al. 1992).

Richardson and Jordan suggested that induced seismicity in mining are divided into two kinds of events (Richardson and Jordan 2002). The first one is high frequencies that usually swarm in small spatial and temporal ranges. Moreover, this kind of event is triggered by the development of fractures causing the rupture of rock mass when dynamic stresses change due to blasting and stress perturbations in excavations. The other kind of events spread out through the active mining areas. They mostly locate in shear zones including faults, and the analysis for them can be extrapolated from tectonic earthquakes. Also, this kind of event is caused by friction dominated ruptures due to the removal of ore over a long period of time rather than blasting activity. Radiated energy of this kind of events are lower than that of the first kind of events.

Seismic velocity increases the focal volume of the primary inclusion zone with accumulating strain energy. Because of the high seismicity rates and dense instrumentation in southern California, there are tens of thousands of well-recorded earthquakes which can be used to infer stress orientation, and most seismically active regions can be studied with a spatial resolution of 5-20 km.

2.2 Seismic Energy

The energy release in earthquakes is one of the most fundamental subjects to help understand earthquakes (Kanamori 1977). Energy released by an earthquake are principally in the form of strain energy, which includes radiated seismic energy, fracture energy and frictional heat released on the moving surfaces during the dislocation process. Fractures are generated when the local stress exceeds the local strength (Kranz, 1983). Fracture energy merely occupies less than 0.1% of the total energy released. The elastic energy and heat of friction are the principal forms of energy released (Krinitzsky 1993). The assumption, the rock mass surrounding a stope is elastic, is applied to estimate the energy released in the area of a stope. It is proposed that rock bursts are associated with high energy-release rates in strong, brittle rocks (Hodgson and Joughin 1966). McGarr suggests that elastic theory fails to provide accurate understanding of the mechanism of mine tremors because occurrences of seismicity are associated with inelastic phenomena (McGarr 1971). Evidences from underground observations show that seismicity in mining are the result of violent shear failures across planes. Fractures in planar zones are correlated with seismic events radiating seismic energy (McGarr 1971). By comparing densities of aftershock in an earthquake sequence, Watanabe proposed that a significant diminishing number of aftershocks were identified following a major event (WATANABE 1989). A quantitative assessment of energy release of a

rock structure under load is useful for understanding the behavior of a rock mass. Acoustic emission (AE) and ultrasonic wave propagation are fundamental forms of energy change. Acoustic emissions in a rock mass are mainly formed from the newly formatted cracks and sudden development of damaged surfaces in preexisting crack (Falls and Young 1998). Also, it is mentioned that microseismic events are triggered where the differential stress ($\sigma_1 - \sigma_3$), which is defined as the in situ crack ignition stress σ_{ci} , is highly concentrated (Falls and Young 1998). Damage mechanics is used to analyze the acoustic energy radiated with time (Shcherbakov, 2003). Also, Shcherbakov claimed good power-law scalings regarding energy radiated with time t and pressure were observed.

Xie proposed that the distribution of microseismic events can be characterized by fractals (Xie and Pariseau 1993). The fractal dimension decreases with the development of fractures in a rock mass. The occurrence of a rock burst in mines is more likely to be indicated by a low fractal dimension. Fractal dimension (D) of the group of fractures within the rock mass is exponentially correlated with the strain energy release (E),

$$D = C_1 e^{-C_2 E} \quad (2-4)$$

where C_1 and C_2 are constants.

The relations between magnitude of earthquakes and seismic wave energy were established by Gutenberg and Richter. Seismic energy is computed by magnitude of seismic events,

$$\text{Log} E = 9.9 + 1.9 M_L - 0.024 M_L^2 \quad (2-5)$$

where E is the seismic energy, M_L is the magnitude of a seismic event (Richter 1958).

The spatial rate of energy release (ERR) is the most suitable index to assess the relative difficulty of mining on a particular region due to mining excavation (Heunis and Msaimm 1980). An important characteristic of the spatial rate of energy release (ERR) is that rock bursts are strongly associated with it (Heunis and Msaimm 1980). However, rock falls fail to pose influence on it. It is clear from considerable research that most mining-induced seismic events are related to geological discontinuities in the surrounding rock masses. Planes of weakness caused by the discontinuities even if the mining-induced stress is lower than the threshold stress for fractures developing in rock mass. Also, production blasts appear to trigger and raise the risk of rock bursts in the surrounding rock mass.

Rock bursts in deep underground mines of hard rock include events triggered by high stress and those associated with large-scale slip on faults. Events induced by high stress are evaluated to be of lower magnitude and less damaging than large scale slip on faults (Swanson 1992). Coulomb failure criterion is employed to assess the potential fault slip in static stress field. The Brune model (Brune 1970) is used to estimate constant stress drop and the correlation between rupture size and magnitude of the seismic events. Seismic energy E is estimated by using Gutenberg and Richter's relation

$$\text{Log}E = 11.8 + 1.5M_L \quad (2-6)$$

where M_L is nearly the same as magnitude M (Gutenberg and Richter 2010). This relation is considered to present a reasonably accurate estimation of seismic wave energy for earthquakes in most cases. However, it is necessary to calibrate before applying it to great earthquakes, especially for the great earthquakes with rupture length over 100 km (Kanamori 1977). It is not accurate to estimate the energy based on magnitude for great earthquakes due to the little correlation between magnitudes and rupture length in the developing process of earthquakes. The main interpretation is that the magnitude M is determined at a short period, in which the rupture process of an earthquake fails to accomplish (Kanamori 1977). The change of strain energy is difficult to estimate because the amount of strain energy before and after an earthquake is unknown. Also, strain energy change cannot be calculated arising from the fact that the absolute stress level involved in faulting is unknown. Kanamori (1997) proposed that the minimum strain energy drop is able to be computed by estimating the energy of seismic waves. Based on the static source parameters of earthquakes registered by the seismic monitoring system, accurate estimation of energy for earthquakes are approached (Kanamori 1977). The seismic moment M_0 is an essential parameter that indexes the deformation at the hypocenter. It is observed that the seismic moment M_0 and the fault area S follow a linear relationship (Kanamori 1977). The relation is suggested as,

$$M_0 = 1.23 \times 10^{22} S^{3/2} \text{ dyn cm} \quad (2-7)$$

where the seismic moment M_0 is defined by μDS . μ is the rigidity and D is the average offset on the fault.

The correlation between the location of seismic activity and induced stresses is investigated. Combining microseismic monitoring with the numerical modeling is an efficient way to detect areas subject to rock burst hazards (Abdul-Wahed, Al Heib et al. 2006). Doublets are applied to

remove locating errors arising from the velocity model. The seismic energy of a seismic event is recorded by the receivers as flow energy, which is carried by both P and S waves and is computed from

$$F = \rho \alpha \int_{t_1}^{t_2} V^2(t) dt \quad (2-8)$$

where $V(t)$ is ground velocity (m/s), ρ is the density of the medium (KN/m^3), α is the velocity of P wave (m/s) or S wave (m/s), and $t_2 - t_1$ are signal duration. Based on the spherical wave hypothesis, the elastic energy release of the source i is

$$E = 4\pi R_i^2 F_i \quad (2-9)$$

where R is the distance (m) between a source and a receiver. The total seismic energy of a seismic event is

$$E_t = \frac{1}{n} \sum_{i=1}^n 4\pi R_i^2 F_i \quad (2-10)$$

where n is number of receivers recording the signal.

The most direct characteristic of stress change is the abrupt release of stress during earthquakes (Simpson 1986). The regions of low strength and geological discontinuities such as faults are more likely to experience the release of stress. Stress keeps accumulating until a threshold level is achieved to cause damage, which yields a drop in stress during an earthquake. In natural seismology, the drop of stress is followed by a recovery of stress until the next earthquake occurs. Some examples show that the repeat time between earthquakes in the same fault is from tens to hundreds of years. However, rock bursts occurred more often in mining since a rock mass loses the equilibrium due to mining excavation. Both the seismic moment and the stress drop can be interpreted in terms of the strain energy release in earthquakes. The seismic moment and the stress drop during earthquake could be used to estimate strain energy release. The elastic strain energy can be divided as

$$W = H + E \quad (2-11)$$

where $H = \sigma_f DS$ is the frictional loss and E is the wave energy. σ_f is the frictional stress during faulting. The difference of the elastic strain energy W before and after an earthquake based on the elastic stress relaxation model is

$$W = \sigma DS \quad (2-12)$$

where σ is the average stress during the form of fault, D is the average offset on the fault, and S is the area of the fault. According to the fact that stress drop $\Delta\sigma$ is approximately equal to 2σ if the stress drop is complete, there is

$$W = W_o = \frac{1}{2} \Delta\sigma DS = (\Delta\sigma/2\mu) M_o \quad (2-13)$$

where $\mu = (3 - 6) \times 10^{11}$ for crust-upper mantle conditions. W_o is the minimum strain energy drop in an earthquake.

It is demonstrated that there is a difference in energy released in earthquakes by using different methods (Kanamori 1977). The energy released proposed by Kanamori (1977) is higher than that calculated by the Gutenberg-Richter method for great earthquakes. However, the trend of energy change and the number of earthquakes displays a good correlation.

The degree of aftershock activity involved with the mainshock could be represented by

$$\sum_{i=1}^{\infty} E_i / E_m \quad (2-14)$$

where E_m and E_i are the energy of the mainshock and the i th largest aftershock, respectively. E_1 , the energy of the largest aftershock, is proportional to the total energy; E_1/E_m denotes the aftershock activity (Utsu 2002). Compared with aftershock, foreshocks are known as infrequent and various. Considerably large earthquakes are not preceded by foreshocks. All aftershocks contribute to decrease regional stress depending on the magnitudes of the aftershocks. It is believed that the energy radiated in aftershocks is possibly more than the elastic energy transferred to the higher stress region due to the mainshock. The main reason is that stress drop with aftershocks, triggered by the stress increase, is greater than the increasing amount of stress transferred (Shcherbakov, 2005). Damage mechanics are used to interpret the decay rate of aftershocks based on the hypothesis that stress transfer enhances the stress σ and strain ε exceeding the yield stress σ_y and yield strain ε_y in vicinities to the occurrences of mainshocks. The applied strain ε_o is defined as a constant and the stress σ decreases to the yield value σ_y due to stress relax during damage of aftershocks. The damage variable α is

$$\sigma - \sigma_y = E_0(1 - \alpha)(\varepsilon - \varepsilon_y) \quad (2-15)$$

with $\sigma = E_0 \varepsilon_y$, where E_0 is the Young's modulus of the material.

2.3 Aftershock

Mainshocks in an earthquake sequence are followed by aftershock sequences. The mainshock in an earthquake sequence has the largest magnitude. Aftershocks cannot have larger magnitude than the original mainshock, otherwise the original mainshock should be defined as a foreshock. The occurrences of aftershocks are triggered by the regional stresses increasing arising from the mainshock. Aftershocks play a role of relaxing the excess stress caused by a mainshock (Shcherbakov, 2005). In order to perform a reliable aftershock hazard assessment, it is essential to determine the pattern of aftershocks, analyze variations of seismicity parameters, and detect aftershock rate changes.

The b-value is involved in seismic hazard analysis to monitor the potential uncertainties of seismicity. It is known that the ratio of the number of large magnitude to small magnitude earthquakes follows

$$\log N = a + bM \quad (2-16)$$

where N is the cumulative number of earthquakes of magnitude M or greater, and a is an empirical constant (Gutenberg and Richter 1956). According to global and regional surveys of earthquakes, b-values of the Gutenberg-Richter relationship are generally limited to -1 ± 0.2 (Wesnousky 1999). Another method to estimate the b value is using magnitude of earthquakes,

$$b = \log e / (M - M_z) \quad (2-17)$$

where M is the mean magnitude of earthquakes of $M \geq M_z$, M_z is the threshold magnitude.

Some studies have validated the Gutenberg-Richter relationship. The Gutenberg-Richter law is used to ascertain the relationship between the cumulative number of aftershocks with magnitudes greater than m. Examples refer to Landers earthquake (m = 7.3, June 28, 1992), Northridge earthquake (m = 6.7, January 17, 1994), Hector Mine earthquake (m = 7.1, October 16, 1999), and San Simeon earthquake (m = 6.5, December 22, 2003). Although the b-value changed slightly for different earthquakes, the data presented a good linearity between the cumulative number of aftershocks and magnitudes (Shcherbakov, Turcotte et al. 2005).

Isacks and Oliver (1964) claimed that it is reasonable that the hypothesis of the constant b value is used to predict the earthquakes in the future by extrapolating to higher magnitudes based on frequency magnitude relations. Extrapolation of the relation to magnitude supports that an earthquake of magnitude 6 is likely to occur in 600 years (Isacks and Oliver 1964). It is known that b -value variations could be used as an indicator of failure both in the rock samples and crustal earthquakes (Lockner 1993). The variability of b -lines can alter significantly prior to and after an earthquake. Considerable evidences indicate that earthquakes were preceded by periods of high b -values (Smith 1981). The variation of parameter b with earthquakes were investigated in five regions of New Zealand from 1955 to 1979. There was an earthquake in the drop of b -value after achieving the peak of 1.9 in 1794. Similarly, another earthquake of magnitude 5.7 in 1973 was preceded by high values of b around 1.6. Also, an earthquake with the magnitude 5.9 in 1971 followed high values of b (Smith 1981). Besides, b -value before a large event tends to decrease due to the occurrence of foreshocks. A routine procedure to detect the anomaly in b -values with aseismic region with forthcoming shocks could help predict earthquakes. It is noted that temporal changes of b -value is more important for evaluating than the absolute peak b -value (Smith 1981).

The seismic potential in the seismic source zones is determined from the Gutenberg-Richter relationship between the magnitude of earthquakes and the frequency of occurrence because the curve can be projected to investigate the larger and less frequent earthquakes that have not occurred yet (Krinitzsky 1993). It has been observed from the frequency-magnitude relations that b -value changes prior to failure of rock in laboratory studies (Mogi 1963). From laboratory experiments, it was found to be that Gutenberg-Richter relationships can be applied for microfracturing events in rock as that for earthquakes (Mogi 1962). Lab tests with acoustic emissions (AE) monitoring shows that frequency-magnitude distribution of acoustic emission events satisfy power-law relationships (Guarino, 1998). Along with the occurrence of events, the radiated energy and the remaining time untill failure follow an inverse power-law relationship (Shcherbakov, 2003). Shcherbakov and Turcotte (2003) used damage mechanics to analyze the power-law distribution in radiated energy and the time to failure with increasing pressure arising from the good agreement between the time to failure and the applied pressure. The b -value is inversely proportional to the regional stress level (Scholz 1968, Huang and Turcotte 1988).

Motivated by efforts to decrease uncertainties in seismic hazard analysis, significant changes prior to an earthquake are concerned. Some studies show that b -values decrease with the increasing

stress. In addition, the velocity change of wave propagation in a rock mass is considered to be indicative of failure of the rock mass. It is well known that a decrease in velocity is generally associated with an attenuation of rock mass and an increased number of fractures (Lockner, Walsh et al. 1977). However, a few measurements in laboratory studies show that there are still some exceptions. It is found that attenuation fails to increase at low stress levels, and even decreases when stresses are able to close the preexisting fractures in a rock mass. Uniaxial tests indicate that P and S waves attenuate at high loading levels close to the ultimate strength because of increasing shear stress and vertical cracks are normal to the direction of wave propagation (Lockner, Walsh et al. 1977). Although there are some studies regarding velocity and attenuation, more studies in the field are needed to be performed.

Seismic attenuation could be used to predict rock bursts and earthquakes if the same characteristics with lab tests could be observed (Lockner, Walsh et al. 1977). Uniaxial tests in laboratories show that P wave velocity increases along with the closing of preexisting cracks and pores due to the stress in the early loading regime. Then, velocity continues to reduce with the opening of new fractures as a result of increasing stresses (Masuda, Nishizawa et al. 1990). Major earthquakes appear to exert influences on stress orientation. Two assumptions are established in the inversion of earthquake focal mechanisms. First, stress is relatively homogeneous over the spatial and temporal extent of the events. Second, focal mechanisms are significantly diverse and can be investigated by displaying P and T axis distributions (Hardebeck and Hauksson 2001). It is proposed that b-lines follow pronounced power-law progressions when there are considerably smaller earthquakes (Krinitzsky 1993). According to the relation of Gutenberg and Richter magnitude and recurrence, probabilistic seismic hazard analysis is conducted based on some assumptions that are as follows (Krinitzsky 1993):

1. The b-values of large regions can indicate special geological structures including faults and zones.
2. Earthquakes occur uniformly and randomly in certain spatial and temporal scale.
3. There is no influence between sources.
4. Projected b-lines can be employed to predict potential occurrences of earthquakes through time.

It is concluded that both small earthquakes and large earthquakes contain self-similarities. Earthquakes of different sizes have similar magnitude-frequency relations and yield similar b-values. The b-value of aftershocks of Fukui earthquake sequence is about 0.9 and the b-value calculated from the microearthquakes for ten years is about 1.1. The difference between them is not significant in the statistical examination (Watanabe 1989). The linearity of occurrences of earthquakes and magnitude is especially interrupted by large earthquakes (Krinitzsky 1993). In addition, the distribution of earthquake magnitudes in the Gutenberg-Richter relationship b-line indicates combinations of individual fault dimensions (Krinitzsky 1993). Seismic hazard analysis by using b-lines fulfills compelling needs on rock bursts forecasting. Combined with practical experience in structural engineering, seismic hazard maps are established for seismic hazard analysis and to alert potential dangers and mitigate hazards (Wesnousky, Scholz et al. 1984, Wesnousky 1986). Underground observations indicate that seismicity in mining (mine tremors) obey the same magnitude-frequency relationship as earthquakes. Magnitude-frequency data for events over 1 year at Harmony Gold Mine has been found to follow the Gutenberg-Richter relationship (McGarr 1971). Moreover, some work has been done concerning the implications of b-value change with regard to the stress change of the rock mass in mining. The probabilistic methodology is applied to estimating the occurrence of earthquakes, despite the shortcomings, which are:

1. The insufficiency of seismic data.
2. The unreliability of forecasting large earthquakes by using b-lines.

The temporal decay of aftershock activities are found to agree with Omori's law from considerable evidences. Omori's law quantifies the decay of seismic activity with elapsed time after a mainshock triggered at the origin of time (Ouillon, 2005). Omori's law is generally described as

$$r = \frac{dN}{dt} = \frac{K}{(c+t)^p} \quad (2-18)$$

where $r = dN/dt$ is the rate of occurrence of aftershocks with magnitudes greater than m and t is the time that has elapsed starting from the mainshock. K , p , and c are involved parameters. Omori (1894) originally defined the value of p as 1.

There is an important scaling law concerning aftershocks in terms of magnitude (Shcherbakov, Turcotte et al. 2005). Båth (1965) presented that the difference in magnitude between the mainshock and its largest aftershock keeps constant independent of the magnitude of the mainshock. Båth's law is interpreted as

$$\Delta m = m_{ms} - m_{as}^{max} \quad (2-19)$$

where m_{ms} is the magnitude of the mainshock, m_{as}^{max} is the magnitude of the largest aftershock, the value of Δm is approximately 1.2. Other studies show that the difference between mainshock and the largest aftershock is impacted by both magnitude scaling and the aftershock productivity (Marsan, Prono et al. 2013). Mainshock and the largest aftershock can be statistically estimated by using Båth's law and extrapolated Gutenberg-Richter scaling (Shcherbakov, Turcotte et al. 2005). Båth's law is further interpreted in the perspective of energy partitioning. It is found that the average ratio of the total energy radiated in an aftershock sequence to the energy radiated by the preceded mainshock is constant. The ratio of the drop in stored elastic energy due to the aftershock to the drop in stored elastic energy due to the mainshock is the same with the previous ratio, which is the radiated energy to the total drop in stored elastic energy (Shcherbakov, Turcotte et al. 2005). The ratio of total radiated energy of aftershocks to the radiated energy of the mainshock is interpreted as

$$\frac{E_{as}}{E_{ms}} = \frac{b}{\frac{3}{2}-b} 10^{-\frac{3}{2}\Delta m^*} \quad (2-20)$$

where E_{as} is the total radiated energy of aftershocks, E_{ms} is the total radiated energy of the mainshock, and Δm^* is the magnitude difference between the largest aftershock and the mainshock.

2.4 Tomography

Tomography was proven to be a greatly useful tool to examine stress distribution in a rock mass by generating images of its interior using mining-induced seismicity (Young and Maxwell 1992; Westman, 2004). It allows for an examination of stress distribution remotely and noninvasively. By using underlying framework of earthquake analyses, numerous studies related to the use of seismic arrays were carried out to improve mining safety (Friedel et al., 1995; Friedel et al., 1996). Mining-induced seismicity provide sources for imaging the structures of rock masses in mines. Techniques used in earthquake tomographic studies provide insight into improving safety

in underground mines. Tomographic studies are of great importance for describing the earth's structure. Earthquake location and tomography are widely used tools to infer active and passive structures of the earth's interior. The features of the inside structure of Earth are analyzed by imaging seismic velocities and earthquake locations (Monteiller, 2005). It is known that high resolution imaging of earthquake tomography needs both available seismological data and appropriate inversion scheme. First, a simple smooth model is established to interpret weighted average of the observations. Then, this initial velocity model keeps being modified until a reasonable range of error estimation between observed and predicted values is achieved (Kissling, Ellsworth et al. 1994). A minimum 1-D velocity model as an initial reference model for the 3-D local earthquake tomography is established. The 1-D velocity model needs to be tested for assessing the quality of the model. In the tests of velocity inversion, initial models are assigned with average velocities, which are greatly higher or lower than the minimum 1-D model. Also, event locations are perturbed randomly on all three spatial directions (shifted in x, y, and z) before using as input in the 1-D model. Velocity change in a 1-D model is mainly determined by lateral variations and local geology in the shallow subsurface. The tests conducted by Haslinger (1999) show that the velocity model is not well constrained in the shallow layer (0 km - 0.5 km) and below 30 km. The ratio of V_p/V_s is approximately 1.85 from 0.5 km to 20 km. The P, S velocity, and the ratio of V_p/V_s are arbitrary on the top layer (0 – 0.5 km) and deep depth (> 30 km).

3-D inversions of the travel time data set with the same layer velocities as the minimum 1-D velocity model are performed and show fewer artifacts than a priori 1-D model (Kissling, Ellsworth et al. 1994). A 1-D velocity model with corresponding stations and travel times is used to relocate events by using a trial and error procedure at the first step. Selected data including high-quality events are used in 3-D tomography inversion (Haslinger, 1999). In tomography inversion, a 3-D grid is established for providing a good coverage on raypaths. The velocity along a raypath and the velocity partial derivatives are computed by linear interpolation on the surrounding grid points (Eberhart-Phillips, 1986). It is shown that the velocity value at each point is usually calculated from the velocity values at the surrounding eight nodes by using tri-linear interpolation (Thurber, 1999). To ensure enough information for calculating the velocity of the grid point, the spacing between the grids is arranged to include abundant raypaths around each grid point. It is not necessary that all the spacing are uniform (Eberhart-Phillips, 1986). For example, both a

horizontal grid and a vertical grid with 10×10 km node spacing cover an area of 100×100 km, and the depth ranges from the surface to 40 km depth (Haslinger, 1999).

The damped least-squares method is a commonly used velocity inversion technique. Parameters including damping have to be set up before inversions. Damping is added to the diagonal elements of the separated medium matrix by trying to suppress significant model changes, which usually occur for singular values near zero. Optimal damping value can be determined by the trade-off between model variance and data variance on multiple tests (Eberhart-Phillips, 1986). It is a traditional approach that the ratio of data variance to model variance is picked as the damping value. However, the value is likely to be unreasonably small. The oscillation of velocity from one grid point to another is large if the damping is small. Consequently, the damping value is empirically picked for the optimal iteration. Multiple iterations with different damping values are performed with decreasing both the data variance and solution variance. As mentioned by Haslinger (1999), an appropriate damping value contributes to a pronounced weakening in data variance and a moderate growth in model roughness. The damping and smoothing are two regularization parameters. The “L-curve” trade-off analysis of data variance versus model variance are used to select the optimal parameters when damping is applied. Zhang (2007) provided an example that how the trade-off curves of data variance and model variances were used for different damping and smoothing weight values.

In Eberhart-Phillips’s (1986) study, the modeled area was designed not to cover all the earthquakes and stations. The events and stations outside the modeled area contribute to improve the raypaths distribution in the modeled area. It is known that the density of raypath coverage changes significantly over the modeled area. High density of raypath coverage provides necessary information for more detailed velocity model. The principle to arrange the locations of grid points is that the reasonable resolution can be generated at most grid points from distribution of stations and hypocenters. It is known that the high and low velocity regions keep a very similar distribution and slightly changed amplitudes of the velocity anomalies even if the arrangement of grid points is different. According to this feature, the range of velocity anomalies can be estimated by using different arrangements of grid points. It is noted that high resolution regions of tomography needs the smaller grid spacing for velocity inversions (Eberhart-Phillips, 1986). A starting model with initial velocities is established for the initial inversion. Tests using different initial velocities can be performed to select the optimal velocity values. Some indexes are used to assess how the

velocity is constrained and solved at each grid point. The hit count is a simple and direct estimate of the effectiveness of the model. It is the total number of raypaths that are involved at one grid point in the solving process. The relative amount of raypaths traveling around the grid points is another important index, which can be indicated by derivative weight sum (DWS). The DWS is expressed as

$$DWS_j = \sum_{i=1}^N \sum_{l=1}^L \frac{\partial T_{ij}}{\partial m_l} \quad (2-21)$$

where N is the number of events; Travel time T is from an earthquake i to a seismic station j ; m_l are the L parameters of the velocity model. The model partial derivatives $\frac{\partial T_{ij}}{\partial m_l}$ are line integrals along the raypath. The influence of model parameters m_l on the travel time T_{ij} is indicated by $\frac{\partial T_{ij}}{\partial m_l}$.

The DWS quantifies the relative raypath density in the volume of a grid point and weights the significance of each raypath by its distance to the grid point (Haslinger, 1999). Only the inversion grid nodes whose derivative weight sum (DWS) values are greater than five are used in the tomographic analysis. DWS can be used to assess the model resolution of inversion, especially for large inverse problems (Zhang, 2007). Another important index for measuring the inversion quality is the resolution matrix, which assesses the dependency of the solution for one model parameter on all the other model parameters. The density of raypath or the values of the resolution of the grid space can be enlarged by decreasing the damping value for iteration. However, the velocities of grid points are likely to be unreasonable due to the large oscillations in velocity arising from the small damping value. The general approach to choosing optimal damping value is to select the damping value generating both moderate data variance and solution variance. Moreover, accuracy fails to be improved if large values of resolution are generated in the practical grid space having low density raypath (Eberhart-Phillips, 1986). Tomographic studies can merely demonstrate the features of the regions having a larger area than the grid spacing. Despite the fact that velocity variations are generated in the grid spacing, velocities at gridpoints reflect the average velocity distribution within the surrounding volume. Reasonable assessments on velocity anomalies can be depicted by the size and shape of tomographic distribution. However, it is not a rigorous and exact estimate of boundaries of velocity features based on the size and shape of velocity anomalies in tomographic studies. The reliability of tomographic inversion methods and the quality of the images can be assessed by using a synthetic velocity model, such as the checker-

board test (Lévêque, 1993). Numerical tests prove that it is necessary to arrange source-receiver geometries for building a checkerboard test (Monteiller, 2005). A checkerboard test is usually applied with perturbation in the velocity model to compute theoretical travel time with the real source-receivers geometry. Studies of checkerboard tests indicate that the spatial extent of valid tomographic reconstruction is mainly controlled by ray distribution. An example of a checkerboard test for double difference tomography is presented by Monteiller (2005). A 200m/s sinusoidal perturbation is deployed in the velocity model, which is iterated in grids of 1 km. The tests show that the model is estimated in a reasonable volume with a correct resolution of the tomographic inversion (Monteiller, 2005).

It is believed that the velocity of waveforms propagation is uniform along the straight line raypaths from sources to receivers due to the assumption of the uniform medium. Various algorithms are applied in ray tracing techniques to estimate the raypath. A fast two-point ray tracing algorithm is developed for ray approximation by minimizing the travel time (Um, 1987). It tries to achieve a balance on the computation time and the accuracy of travel times. Iterations on perturbations are performed to find the raypath with the minimal travel time based on an initial guess of the raypath. Another enhancement using the simplex search method is performed on the fast two-point ray tracing algorithm by distorting the raypath from a starting path in a systematic way until obtaining the minimized travel time. Simplex method applied in raypath search can decrease the required amount of travel time (Prothero, 1988). The inversion of tomographic studies is usually linearized even for non-linear seismic tomography problems. The final iteration provides the model resolution matrix to summarize the quality of inversion. Synthetic tests can be used to estimate the model resolution and uncertainty by performing velocity inversion on a synthetic data set, which has the same distributions with the real data (Zhang, 2007). Inversion of synthetic data can indicate the quality of an inversion scheme. Synthetic models are usually combined with velocity anomalies, which are generated from the real data. At the same time, synthetic travel times are computed based on the real distribution of sources and receivers. The synthetic is computed with the same parameters used in real data.

In seismic tomography research, it is believed that local earthquake tomography (LET) can provide a higher-resolution imaging of velocity structures than teleseismic tomography (Thurber, 1999). Meanwhile, the shortcomings of local earthquake tomography is the high variability of model sampling arising from the non-uniform source distribution of seismic events. Double –

difference tomography algorithms are the basis of some form of the linearization of the travel time equation in a first order Taylor series that relates the difference between the observed and predicted travel time to unknown adjustments in the hypocentral coordinates through the partial derivatives of travel time with respect to the unknowns (Waldhauser 2001). The DD technique is based on the fact that if the hypocentral separation between two seismic events is minute compared to the event-station distance and the scale length of velocity heterogeneity, then the ray paths between the source region and a common station are similar along almost the entire ray path. In this case, the difference in travel times for two events observed at one station can be attributed to the spatial offset between the events with high accuracy.

Double-difference equations are built by differencing Geiger's equation for earthquake location. In this way, the residual between observed and calculated travel-time difference (or double-difference) between two events at a common station are related to adjustments in the relative position of the hypocenters and origin times through the partial derivatives of the travel times for each event with respect to the unknown. HypoDD calculates travel times in a layered velocity model (where velocity is assumed depending only on depth) for the current hypocenters at the station where the phase was recorded.

Double – Difference tomography is first presented by Waldhauser and Ellsworth (Waldhauser and Ellsworth 2000). It is an efficient approach to determine hypocenter locations over large distances because it can incorporate ordinary absolute travel – time measurements and cross – correlation P and S wave differential travel – time measurements. The double-difference tomography arrival time T from an earthquake source i to a receiver k is written using ray theory as a path integral,

$$T_k^i = \tau_i + \int_k^i u ds \quad (2-22)$$

where τ_i is the origin time of event i , u is the slowness field, and ds is an element of path length.

The source is defined as the centroid of the hypocenters. Residual between observed and calculated differential travel time between the two events is defined as

$$dr_k^{ij} = (t_k^i - t_k^j)^{obs} - (t_k^i - t_k^j)^{cal} \quad (2-23)$$

On the basis of previous approach using waveform cross-correlation, Haijiang Zhang and Clifford H. Thurber developed a double-difference seismic tomography method (tomoDD)

employing both absolute and relative arrival time (Zhang and Thurber 2003). The relative arrival times are directly used to determine relative event locations and absolute arrival time picks are employed to minimize differences among relative arrival times.

Due to the fact that event locations and the velocity model affect each other, simultaneously solving on event locations and velocity structure is suggested by Haijiang Zhang (2007). The tomoDD (double difference tomography code) is designed to jointly solve the velocity structure and seismic locations. Although jointly solving increase parameters for velocity inversion and iteration, optimal parameters can be selected by test and trade-off analysis. It is an important goal of seismic tomography to improve the estimates of the model parameters including velocity structure and locations of seismic events. In order to minimize the root mean square (RMS) misfit, perturbations on model parameters are solved by iterations (Thurber, 1999). Zhang and Thurber (2007) compared the velocity inversion with and without smoothing constraints to the slowness perturbations. It is shown that the iteration system with smoothing constraint is better conditioned and more stable. In the same manner, the trade-off analysis on damping value is carried out. The event locations and velocity model can be solved using the LSQR or SVD method. The LSQR method is better than the SVD method for applying large sensitivity matrices since it is usually much faster than the SVD method.

The velocity in the shallow layers of the crust is significantly affected by physical factors, including composition, fluids, saturation, temperature and ambient pressure. The analysis of seismic wave behavior can reveal the influence of physical conditions. Crack density and saturation ratio of P-wave and S-wave can be computed from inversion of P-wave and S-wave arrival times. The high resolution seismic tomography provides the information of crack density and saturation ratio in specific regions. The ratio between P-wave and S-wave identifies crack density and saturation ratio. The S-wave is useful for inferring fluid accumulation in the rock mass because the velocity of S-wave depends on density and rigidity of the rock and cannot pass through liquids. Crack density and saturation ratio of rocks can be discriminated according to the same P-wave velocity and the different S-wave velocity (Serrano, 2013). High porosity (Ψ) areas usually have the same distribution as high crack density (ϵ) areas. The porosity Ψ has been defined as the product of P-wave and S-wave,

$$\Psi = V_P \times V_S \quad (2-24)$$

$$\ln \Psi = \ln V_P / \ln V_S \quad (2-25)$$

$$d \Psi / \Psi = d V_P / V_P + d V_S / V_S \quad (2-26)$$

where $d \Psi / \Psi$ is the change in porosity; $d V_P / V_P$ and $d V_S / V_S$ are perturbed values of V_P and V_S . Also, the relation between ratio of V_P / V_S and Poisson's ratio is

$$\left(\frac{V_P}{V_S}\right)^2 = 2(1 - \sigma)(1 - 2\sigma) \quad (2-27)$$

According to these relations, it is inferred that areas with active aftershock activities and low-Poisson's ratios are more likely to be brittle part of the fault zones. Similarly, the hypocenters region of the mainshock are found to be with high-Poisson's ratio and low velocities of P-wave and S-wave (Zhao, 1998).

2.5 Summary

By showing the usefulness of earthquake data on assessing the seismic risk in the earth's interior, the studies provide insight into the evaluation of the seismic risk from mining-induced seismicity in deep underground mines. Such information is important to engineering applications in underground mines. In addition to the applications in natural earthquakes, some studies indicate that tomography is useful to detect rock burst prone zones by examining the stress distribution. However, these studies claim the potential seismic risk merely on location of groups of microseismicity and the stress distribution. In order to explore the application of double-difference tomography for predicting the potential seismic risk in mines, historical seismicity sequences including major events and microseismicity from rock bursts need to be analyzed using double-difference tomography.

The stress distribution response to the occurrence of major events gives mining engineers an understanding of conditions of potential rock bursts. Owing to an apparent lack of tomographic studies on changes of the stress distribution associated with mining-induced major events in underground mines, it is necessary to accurately examine the stress distribution surrounding the major events. Empirical verifications by double-difference tomographic studies on seismic sequences in underground mines are useful to provide professional judgment on the stress distribution and the potential seismic risks.

Chapter 2 References

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3 Seismic velocity change due to large magnitude events in deep underground mines

3.1 Abstract

Passive seismic monitoring was used to measure wave propagation in deep rock masses. About 30,000 mining – induced seismic events in two hard rock mines were recorded and investigated. The two mines were the Creighton Mine and the Kidd Mine. The accurate travel times of source-receiver pairs allowed us to demonstrate temporal change in velocity. It was observed that the average P wave velocity was about 6000 m/s in the mining regions of Creighton Mine and Kidd Mine. The coseismic and postseismic velocity changes were examined in crustal earthquakes. Velocity investigations on three months of data around large events in deep hard rock mines show a significant decrease in the average seismic wave velocity after the large events. The sudden drop in average velocity change after large events implies the creation of new fractures in a rock mass due to large events. Additionally, it was observed that the average velocity returned to a normal level after several days. A reasonable explanation for the postseismic velocity increase is that opened cracks began to heal due to crack close.

3.2 Introduction

Large magnitude events are defined as mining-induced seismic events with a moment magnitude greater than 0. It is also suggested that the moment magnitude of rockbursts usually ranges from 1.5 to 4.5, and low magnitude microseismic events are with a moment magnitude less than 0 (Young, Maxwell et al. 1992). Large magnitude events studies in underground mining were driven by a desire to predict rock bursts and mine failures (Ouillon and Sornette 2000). By investigating the average velocity of P-wave traveling through rock mass during and after large magnitude events occurred, it is possible to evaluate the stability of rock masses. Velocity changes in natural earthquake studies have provided an important framework for how the large seismic events affect the velocities. Previous studies attempted to establish a functional relationship between velocity and stress changes in shallow crust (Nun 1971).

The effects of stress on seismic velocities in dry rock at temperatures and pressures have been investigated extensively and the mechanism for attenuation is interpreted in different ways (Lockner, Walsh et al. 1977). Significant progress has been made in the study of velocity change caused by large magnitude events in seismology (Schaff and Beroza 2004). However, not much

work has been performed regarding the measurements of velocity change influenced by large magnitude events caused by underground mining. More studies of how mining-induced large magnitude events influence velocities around the rock mass of mining are needed. A considerably stratified rock mass containing abundant layers is regarded as a homogeneous and transversely isotropic continuum if the dimension of mining excavation is sufficiently large (Salamon 1968). It is feasible to regard layers of rock mass as homogeneous and transversely isotropic due to the stratum structure. The highly stratified rock mass can be regarded as an equivalent medium exposed to homogeneous stress distribution under large excavations (Salamon 1968). It is found that large magnitude events could likely be responsible for velocity change with stress redistribution of the rock mass.

The purpose of this paper is to evaluate the effects of large magnitude events on the variation of P-wave velocities. In addition, the importance of the observation of velocity change caused by large magnitude events in mining is discussed. In order to interpret the effects of large magnitude events in terms of seismic P-wave velocities, we present two cases studying velocity change before and after large magnitude events in two hard rock underground mines.

3.3 Velocity and seismicity investigation

In crustal earthquakes, microearthquakes are used to study the effects of large earthquakes (Schaff and Beroza 2004). The microearthquakes are detected and registered by a seismic network. Coseismic and postseismic velocity changes were found in the work by Schaff and Beroza (2004). It was found that a large approximately 2.5% coseismic velocity decrease was recorded after the 1989 Loma Prieta earthquakes. The velocity subsequently returned to its original value.

Also, it was claimed that velocity increases prior to a crustal earthquake, which reduces the seismic wave speeds and subsequently recovers to original level (Vidale and Li 2003). The velocity change of the P-wave and S-wave before and after the crustal earthquake were investigated. It is inferred that the rupture arising from the earthquake impedes wave propagations, and seismic wave speed's recovery is consistent with healing caused by closure of cracks (Vidale and Li 2003).

Mining-induced seismicity arises in the vicinity of excavations in rock masses, which are originally in a state of static balance. Mining activities, including extractions, blasts and injections, are the main triggers of seismicity. A rock mass generates seismic waves as the original state of stress is disturbed by human activities such as excavation and production blasts (Cook 1976).

Mining-induced seismicity is the source of wave propagation, which is governed by the characteristics of media through which waves travel. Increasing mining rate is a signature of fractures developing and coalescing. Mining-induced seismicity is triggered more frequently as more ruptures develop through the rock mass (Iannacchione, Esterhuizen et al. 2005). In addition, geological structures can be inferred by seismicity, which tend to occur along shear faults due to high shear wave energy (Christensen and Mooney 1995).

It is known that an increase in the level of seismicity occurs prior to a large magnitude event (Nun 1971). The noticeable pattern of microseismicity before, during and following the large magnitude events reflects the typical seismic energy accumulation. In underground mining, a direct observation of seismic events is performed by the seismic stations network. Seismic waveforms are recorded by microseismic monitoring systems. In addition to locating seismic events, the travel times from hypocenters of seismic events to seismic stations are estimated and analyzed. Velocity fitting is then performed on travel distance and travel times to get the average velocity of all the wave raypaths. It is postulated that, after the occurrence of large magnitude events, the seismic velocity of the region around the earthquakes will decrease.

3.4 Velocity fit using mining-induced seismicity

The P-wave raypaths vary due to the lateral heterogeneity of rock mass (Panza, Romanelli et al. 2001). It has been proven that raypaths are approximate circular paths of varying radii of curvature in crustal structure (Thurber 1983). Factors including sedimentary geologic structures and fluid mediums affect the propagation of raypaths. Due to the absence of fluid within the zone of underground mining, the raypaths can be assumed to be straight lines within the scale of mine sites. Figure 3.1 illustrates the arrival times of one single event recorded by multiple stations. According to the arrival times of the waveforms from seismic sources and the distance between sources and stations, the linear fitting line in Figure 3.1 represents the arrangement of arrival time–distance pairs of the same event very well. It appears that the records of seismic events are reliable because the arrival time changes linearly and positively in correlation with the distance between the same event and different stations.

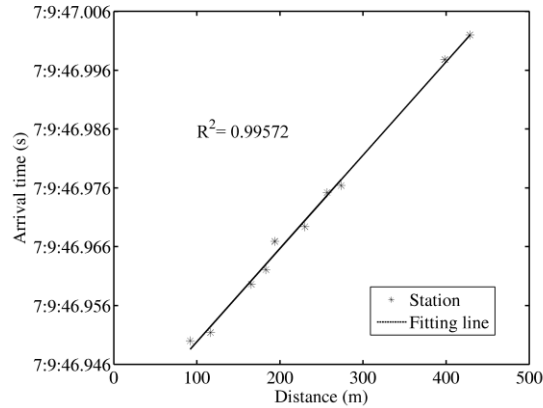


Figure 3.1. Linear fitting for single event of Creighton Mine; showing arrival time of seismic energy versus the distance between the event and sensor locations.

The microseismic events were divided into groups of equal time duration (two weeks). Average velocity of each group was examined by velocity linear fit. Velocity change provides us with key information on the stability of rock masses in underground mining. The propagation velocity of seismic waves is governed by density, fractures, and geologic structures of the rock mass. In addition, tremors during large magnitude events would cause changes in seismic velocity (Wegler, Nakahara et al. 2009). As a consequence of stress variation, seismic velocity change reflects the stress change in space and time.

Temporal change in seismic velocity in underground mines showed that the sudden drop of velocity agreed with the irregular seismic activities, including large magnitude events and highly active microseismicity. Fracture development triggered during and after large magnitude events could weaken the subsequent wave propagation. There is thus a coseismic decrease in velocity with a large magnitude event. The velocity generally stays at a lower level for a duration of time after the large magnitude event due to the fractures around the rupture zone. In natural earthquake studies, some measurements suggested that a velocity drop of 5% preceded $M = 4$ earthquakes (Lukk and Nersesov 1978). The velocity finally recovered to normal levels due to crack healing effects (final closure of the cracks normal to the stress) after a time duration following the large magnitude event.

3.5 Case Study

Data sets of mining-induced seismicity recorded in two hard-rock underground mines, Creighton Mine and Kidd Mine, were analyzed. The greatest depth of excavation is 2440 m in Creighton Mine and 1500 m in Kidd Mine, respectively. Mining-induced seismicity were well

located by their seismic networks. A total of about 30000 seismic events, nearly 25000 in Creighton Mine and about 5000 in Kidd mine, were investigated in the velocity change study.

Creighton Mine

Seismicity data from Creighton Mine, an underground nickel mine in Sudbury, Canada, were recorded from May 25, 2011 to September 25, 2011. The mining methods used have varied significantly over the course of its mine life. Since 2008, the large-diameter blast holes method associated with vertical retreat mining has been adopted at the Creighton Mine. The depths of main production area ranged between 1829 m (6000 ft) to 2438 m (8000 ft) (Malek, Espley et al. 2008). The microseismic monitoring system consists of 62 stations, including 10 triaxial and 52 uniaxial accelerometers. The depth of stations were arranged from 1508 m (4947 ft) to 2392 m (7847 ft) to provide a good coverage on the rock masses adjacent to main production areas.

The microseismic monitoring system provided real-time seismic monitoring and recorded the location and trigger time of each event. As shown in Table 3.1, Creighton Mine experienced four large magnitude events, which are seismic events with a moment magnitude larger than 1.0, at a depth of approximately 2500 m (8200 ft) in July, 2011.

Table 3.1. Times, Locations, and Magnitude of Major Events in Creighton Mine

Major Events	Date	Time	North (m)	East (m)	Depth (m)	Magnitude
1	July, 6 th 2011	8:41 AM	1927	1399	2332	3.1
2	July, 6 th 2011	8:46 AM	1914	1478	2333	1.2
3	July, 6 th 2011	8:47 AM	1879	1456	2284	1.3
4	July, 10 th 2011	2:44 AM	1853	1385	2392	1.4

According to the seismic records, microseismicity was remarkably active between the large magnitude events and post large magnitude events period. Additionally, the common characteristic of these large magnitude events was that they all followed shortly after a production blast. An 1134 kg production blast preceded the initial three large magnitude events. Similarly, the last large magnitude event was triggered by a production blast performed one minute prior to it.

Based on the premise of the linear relationship between arrival time and source-receiver distances, the velocity of wave propagation on each raypath could be calculated by the travel distance and travel time on the raypath (Figure 3.1). It was observed that the distribution of travel time-distance pairs agrees with a linear relationship (Figure 3.2). The average velocity of the region was computed by velocity fitting using all the raypaths traveling through the target area.

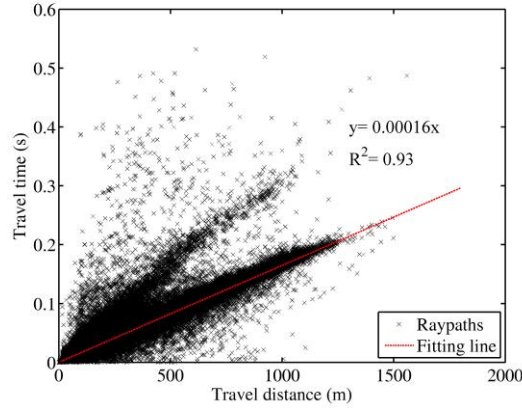


Figure 3.2. Linear fitting of travel time with travel distance

Using all of the raypaths propagating through the rock mass, the arrangement of travel time-distance pairs from sources to stations follows a line (Figure 3.2), the slope of which is the average velocity of all raypaths propagating through the rock mass. Note that the fitting line crosses the origin as a consequence of there being no travel distance if the travel time is zero. As can be seen in Figure 3.2, there is a linearly distributed swarm of scatter besides the linear dominating swarm fitted by the regression line. It implies that waveforms including S-waves (scatters with low velocity) are combined with P-wave in the data set. The robust fit method was used to perform the velocity linear fitting. According to the velocity fit line, the travel times corresponding to velocities larger than 9000 m/s or smaller than 3000 m/s are removed in the velocity fit analysis. Robust fit method is used to lessen the effects of outliers (Dumouchel and O'Brien 1991; Cleveland 1979).

Error analyses on different regression methods were performed, indicating that residuals of the robust fitting were lower than that of the ordinary fitting (Figure 3.3). Besides, the fluctuations of the robust residual were much smaller. According to the linear regression analysis, the robust fitting provided an accurate velocity fit for the seismic waves.

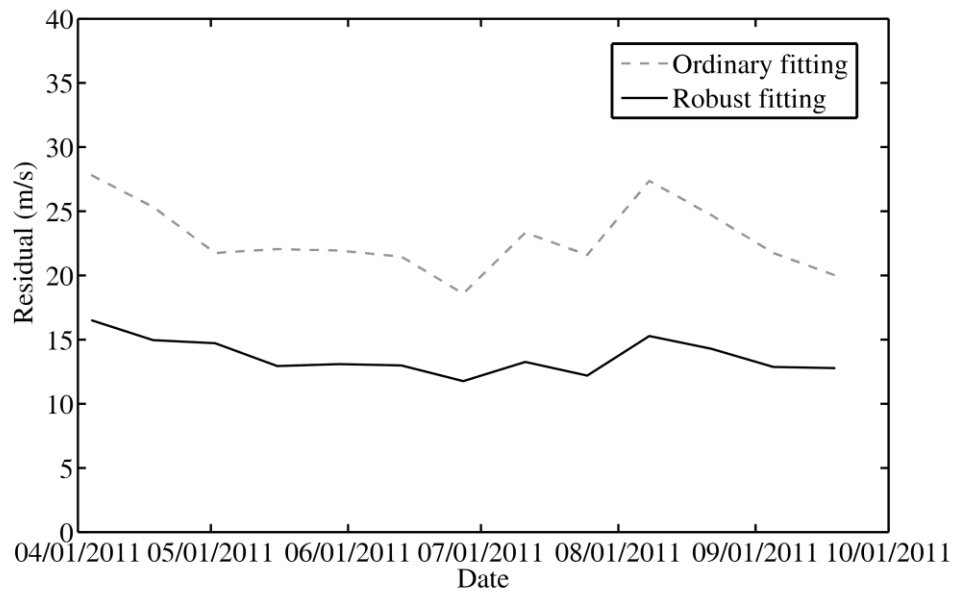


Figure 3.3. Residuals of robust linear fitting and ordinary linear fitting for Creighton Mine

The average velocity fit results indicate that the velocity within the near-mining rock mass was influenced by large magnitude events. The average velocity in the period during the large magnitude events was the lowest of all the time periods. A steady increase in the average velocity was found from data sets before the occurrence of large magnitude events (Figure 3.4). As a consequence of large magnitude events, the velocity dropped from 6049 m/s during June, 13th - June, 26th to approximately 6017 m/s during the period from June, 27th to July, 10th. Because large magnitude events caused fractures to develop and collapse, wave transmissions in the surrounding rock mass attenuated due to the developed fractures. A few days later, the velocity increased to the initial level of 6050 m/s. This was a likely result of the fractures in the rock mass closing due to the crack healing effect. The increased damage to the rock mass likely results in the velocity reduction after the large magnitude events (Bieniawski 1970).

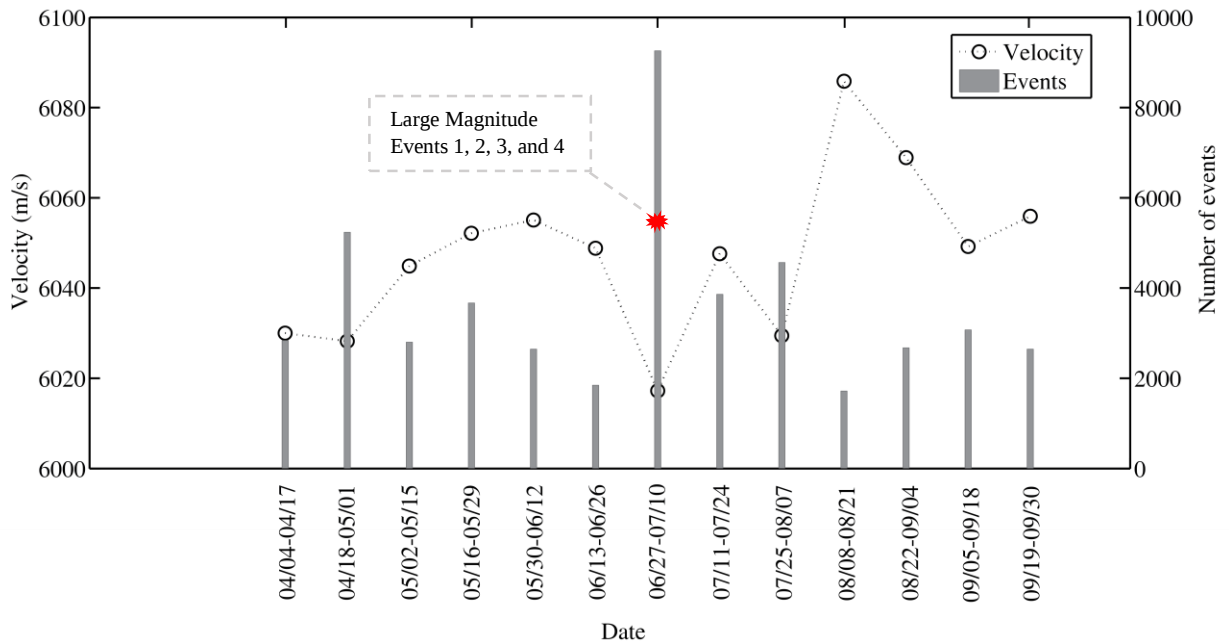


Figure 3.4. Velocity change affected by large magnitude events in Creighton Mine

Kidd Mine

Kidd Mine, situated in near Sudbury, Ontario, Canada, was investigated to see how large magnitude events affected seismic velocities. Kidd Mine is the deepest copper and zinc mine in the world. There was a high level of seismicity due to the stresses experienced at such a depth (Guha 2000). The three-dimensional seismic network, at Kidd Mine, contains 23 sensors.

Two large magnitude events occurred in 2009. The earlier one occurred on January 6th and the other occurred on June 15th. The same method used for Creighton Mine was applied in analyzing the velocity change of wave propagation affected by large magnitude events in the rock mass. Since the origin time of the large magnitude events are five months apart, the velocity changes around large magnitude events were analyzed individually.

Table 3.2. Times, Locations, and Magnitude of Large Major Events in Kidd Mine

Large Magnitude Events	Date	Time	North (m)	East (m)	Depth (m)	Magnitude
1	January, 6 th 2009	4:40 AM	65733	65686	1150	3.8
2	June, 15 th 2009	7:01 PM	65861	65737	1035	3.1

Properties of the two large magnitude events are listed in Table 3.2. The microseismicity data were divided into two week groups. The microseismicity within each two week group was analyzed and compared with the other time periods.

For the velocity fitting of the microseismicity from Kidd Mine, it was observed that the robust fitting was superior to the ordinary linear fitting for these data sets. As shown in Figure 3.5, the residual of the robust fitting is smaller and fluctuates less than that of the ordinary fitting.

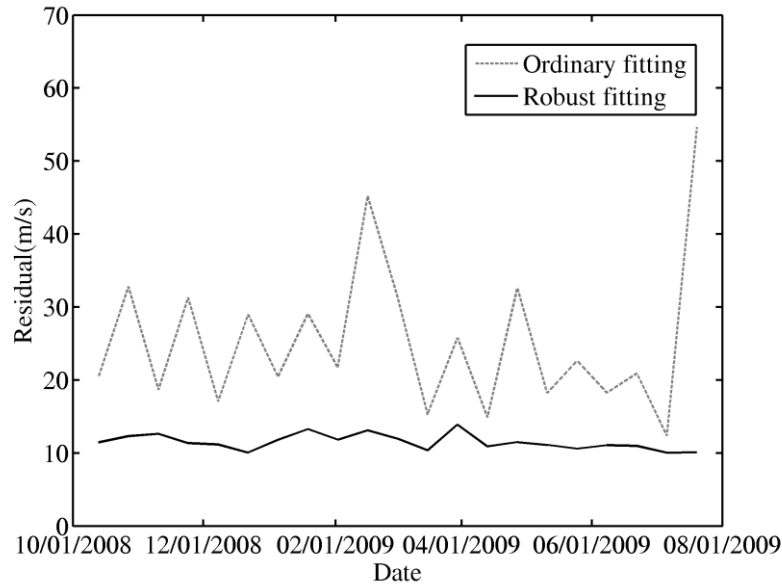


Figure 3.5. Residuals of robust linear fitting and ordinary linear fitting for Kidd Mine

Regarding the large magnitude event of January 6th, 2009, the average velocity from January 5th to January 18th, 2009 experienced a significant drop in comparison to the subsequent two weeks after this drop (Figure 3.6). As expected, the velocity decreased significantly after the large magnitude event and then returned to the initial level. It was postulated that the fractures actively developed around the large magnitude event and mining-induced microseismicity weakened the wave propagations. There was a noticeable phenomenon involving velocity that was shown at Creighton Mine as well. The average velocity decreased again after first recovering due to the loss of structural stability in fracturing of the rock mass even though it was still higher than the lowest velocity historically experienced.

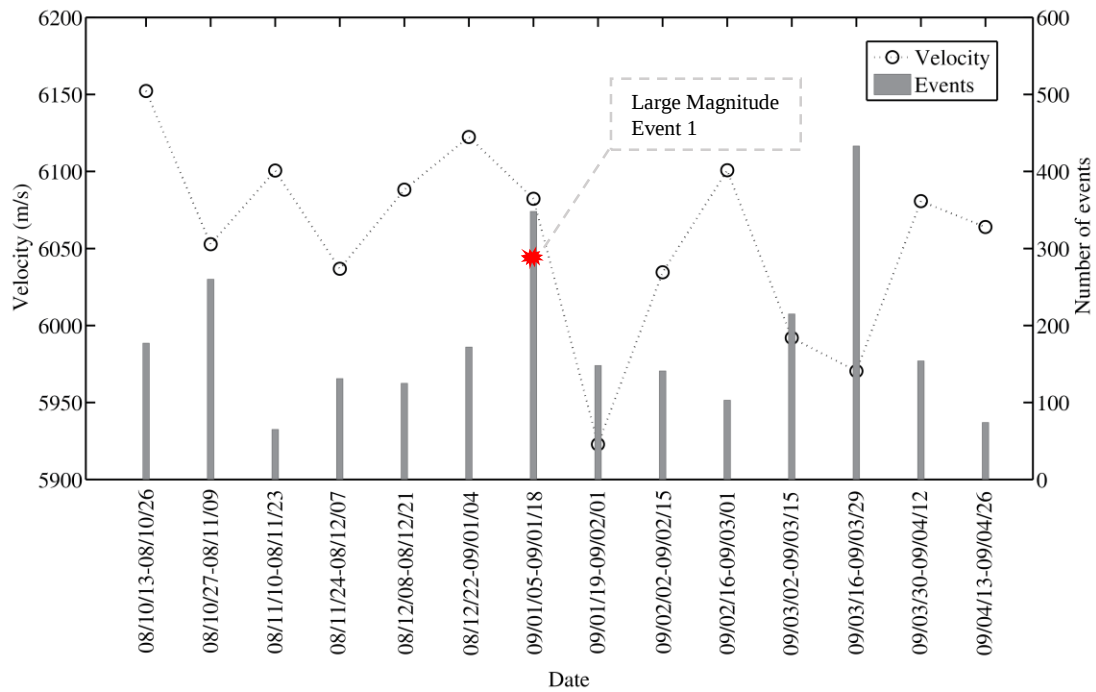


Figure 3.6. Velocity change affected by the first large magnitude event in Kidd Mine

The same analytical procedures were used to investigate the velocity variations affected by the second large magnitude event. The velocity experienced a significant drop as a consequence of the second large magnitude event (Figure 3.7). The trend of average velocity suggests that the large magnitude event reduced the average velocity due to fractures developing. In addition, the average velocity grew back to the normal level after the postseismic activity, as can be seen in Figure 3.7.

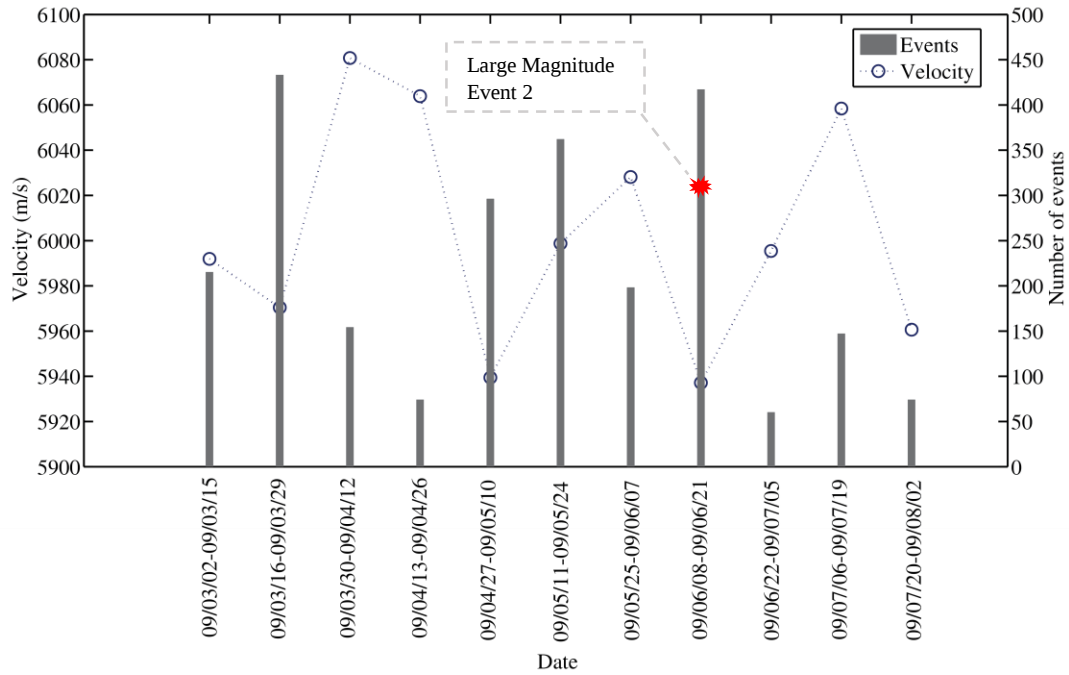


Figure 3.7. Velocity change affected by the second large magnitude event in Kidd Mine

Some differences were manifested between the velocity changes from these two large magnitude events in Kidd Mine. For the first large magnitude event, the velocity decreased to 5923 m/s after the first large magnitude event period and then returned to background level at 6100 m/s. The magnitude of velocity change of the first large magnitude event was smaller than that of the second large magnitude event. The velocity change before the first large magnitude event was within 100m/s. However, the magnitude of velocity fluctuations after the first large magnitude event were about 130 m/s. The possible explanation is that the structure of rock mass is less stable after the fractures induced by large magnitude events.

3.6 Summary and discussion

Velocity changes have been examined to show the response to large magnitude events in underground mines. The robust linear fitting method is effective for computing the average velocity from all raypaths because coefficients of every observation are estimated by a weighting function to reduce residuals. During the periods including large magnitude events, the magnitude of average velocity reduction for P waves is 0.6% for Creighton Mine and 1.8% for Kidd Mine. Similar results from crustal earthquakes are claimed that the path-average velocity changes are 1.5% for P waves. Analyses on S waves of mining-induced seismicity are needed in future studies to compare with a coseismic velocity decrease as much as 3.5% for S waves in crustal earthquakes.

A significant decrease for S-waves indicates shear failures in the rock masses, providing insight into potential fault-slip risks.

The primary advantage of continuous velocity measurements is that the mining-induced seismicity can be used to consistently assess the potential seismic risk in rock mass covered by seismic networks in underground mines. Once general velocity anomalies are observed, some portions of the regions experience significant stress change including stress concentration and stress relaxation. One limitation of this analysis is the use of seismic events which are distant from objective regions in underground mines. The raypaths from these events give more information with respect to the stress distribution along the raypaths. However, part of the information cannot be used for the stress distribution analysis because some portions of the raypaths are beyond objective regions. Thus, trade-offs on using events beyond objective regions could be made for giving rise to optimum results.

The velocity tends to increase to the original level from a duration of recovery after the occurrences of large magnitude events. The velocity changes suggest that velocity reduces because the fractures, caused by large magnitude events and microseismicity, slow the wave propagation in the rock mass. The fact that the velocity eventually returns to the previous state implies a recovering effect due to the closure of the cracks (crack healing effect). An increase in velocity can be attributed to the fact that bulk modulus increases, possibly arising from the closure of cracks. It is noted that the velocity fluctuates by a significant amount after large magnitude events because the induced fractures reduce the stability in a rock mass.

Chapter 3 References

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4 Imaging of Temporal Stress Redistribution due to Triggered Seismicity at a Deep Nickel Mine

4.1 Abstract

Imaging results from passive seismic tomography surveys in a deep nickel mine showed that major events (moment magnitude > 1.0) are associated with the state of stress in the surrounding rock mass. Travel time picks from monitored seismicity at the Creighton Mine in Sudbury, Ontario, Canada between June 22nd and July 24th, 2011 were used to generate tomograms. During this time period, four major events were observed associated with 13630 microseismic events. Two different grid spacings were designed for double-difference tomographic inversion. A large-scale velocity model was used to examine the general trend of velocity distribution before and after the major events. A small scale velocity model with finer grid spacings gave rise to a higher resolution result. By comparing the results obtained from tomographic inversion on consistent time frames, it was found that the velocity in the region near major events started increasing before the occurrence of major events. High-velocity anomalies tended to expand in the region adjacent to the major events. In addition, the accumulating number of seismic events was computed to show correlations with major events. It was shown that the seismic rate increased dramatically during the time period when the major events occurred.

4.2 Introduction

Ground fall is the leading cause of injuries in underground mines. Ground control issues accounted for over 40% of fatal accidents at underground metal mines in the USA between 2008 and 2012 (MSHA report, 2013). Reliable ground control is an essential consideration in mine safety. In order to detect potential seismic hazard, it is necessary to assess the seismic potential that might lead to failures in rock masses. Specifically, seismic hazard analyses require identification of anomalously high stress associated with major events in mines. It is concluded that microseismicity distributes along significant fractures in the mine sites and stress inversions from seismicity are consistent with in-situ measurement of stresses (Urbancic, Trifu et al. 1993). Seismic tomography has been widely applied for imaging and mapping Earth's sub-surface characteristics (Vanorio et al., 2005). Further, tomography was proven to be a greatly useful tool to examine stress distribution in a rock mass by generating images of its interior using mining-induced seismicity (Young and Maxwell 1992; Westman, 2004). It allows for an examination of

stress distribution remotely and noninvasively. By using underlying framework of earthquake analyses, numerous studies related to the use of seismic arrays were carried out to improve mining safety (Friedel et al., 1995; Friedel et al., 1996).

Lab tests were conducted to validate that microseismic events are important precursory signatures of rock failure (Lockner, 1993). The explanation of this effect is that microseismicity is triggered by the development of microcracks, which reduces the velocity of wave propagation in the rock mass. The effects of microcracks on wave propagation were confirmed by rock physics studies (Mavko et al., 2009). Additionally, pressures orthogonal to the direction of microcracks tend to close the fractures, causing a rock healing effect to reinforce wave propagation. Considerable lab studies were devoted to show how changing the stress field affects velocity of wave propagation in rocks as well. It is concluded that the velocity at which a seismic wave travels through a rock mass is related to the applied stresses (Mavko et al., 2009). By extending the knowledge of lab studies to field scale, velocity tomograms can be studied to infer the stress redistribution around an underground mine (Luxbacher, 2008). The stress redistribution related to mining-induced seismicity was discussed to interpret the mechanisms of ground failures by imaging the velocity structures of rock mass in underground mining (Westman, 2008).

The goal of this work was to investigate change of stress distribution associated with the occurrence of major events in a hard-rock underground mine using double difference tomography. Temporal changes in the velocity structure reflected the stress distribution evolving through time and its response to major events. Specifically, significant tendency of stress change before the major events could provide insights to forecasting major events and mitigating seismic hazards. Due to the high seismicity rates and a good coverage of seismic monitoring system in the mine, thousands of well-recorded microseismicity can be used to infer stress distribution in the seismically active regions at the mine.

4.3 Data and methods

Microseismic events result from strain energy which has accumulated in the rock during loading and are the sources used to generate the tomographic image of the velocity distribution (Westman, 2003). Velocity tomography is conducted by using travel-time measurements of mining-induced seismicity. Tomographic inversion can be used to relate travel time picks from events to stations to the velocity distribution within the rock mass. Once velocity is estimated, the

stress distribution can be inferred by considering the velocity-stress relation, which is, high-velocity bodies representing stress concentration and low-velocity bodies shown in the stress relief regions. The travel times from microseismic events to recording stations were recorded by a microseismic monitoring system. Major events were recorded by a strong ground motion system in Creighton Mine. We analyzed 188250 travel times of P-wave from 13630 microseismic events, the depth of which approximately ranged from 1450 to 2700 m, occurring from June 22nd to July 24th, 2011 at Creighton Mine. These data, along with the locations of the events and stations, were required for tomographic inversion. Tomograms were used to compare the velocity change before and after the period that included all major events (Table 4.1).

Table 4.1. Times, Locations, and Magnitude of Major Events

Major Events	Date	Time	North (m)	East (m)	Depth (m)	Magnitude
1	July, 6 th 2011	8:41 AM	1927	1399	2332	3.1
2	July, 6 th 2011	8:46 AM	1914	1478	2333	1.2
3	July, 6 th 2011	8:47 AM	1879	1456	2284	1.3
4	July, 10 th 2011	2:44 AM	1853	1385	2392	1.4

Velocity tomography is an effective analysis method to assess the change of stress distribution over space and time. Tomograms are associated with velocity distribution, and are capable of imaging the distribution of stress of the whole rock mass. It is well known that mechanical properties of rocks are significantly influenced by the anisotropy, which is an important physical characteristic of a rock mass. Rock properties including porosity, permeability, and elastic moduli are sensitive to pressures that are applied in the rock (Mavko et al., 2009). The enhanced stress can lead to an increase in bulk modulus, which is positively correlated with P-wave velocity. After overcoming the dilatancy effect by increasing stress, the closure of preexisting fractures can increase the elastic moduli. A higher elastic moduli directly causes an increase in bulk modulus. Also, the closure of fractures offsets the elastic volume increase, giving rise to higher bulk modulus (Holcomb 1981). This explains why bulk modulus increase with a higher stress. Assuming that stress is the main factor leading to variations of velocity structure, a comparison between different tomograms can provide a reference to the stress change over a spatial range or time. On the premise that a rock mass is viewed as grossly linear, isotropic and elastic in rock physics, the empirical relationship between density and velocity is given by

$$V_p = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} \quad (4-1)$$

where V_P is the P-wave velocity, ρ is the density of rock, K is the bulk modulus, and μ is the shear modulus (Mavko et al., 2009).

TomoDD, a double-difference seismic tomography code, was used to perform velocity inversion (Zhang and Thurber, 2003). It requires the travel time difference for two similar raypaths. The process of subtracting the travel times of nearby events can significantly improve the accuracy of the location results and velocity distribution [11]. Equation (2) shows the computation for arrival time T_k^i , expressed as a path integral,

$$T_k^i = \tau_i + \int_k^i u ds \quad (4-2)$$

where τ_i is the origin time of event i , k is the seismic sensor that record event i , u is the slowness field, and ds is an element of path length. The event location is defined as the centroid of the hypocenters.

The double difference residual, dr_k^{ij} , is defined as:

$$dr_k^{ij} = (t_k^i - t_k^j)^{obs} - (t_k^i - t_k^j)^{cal} \quad (4-3)$$

where $(t_k^i - t_k^j)^{obs}$ is the difference in the observed travel-times between events i and j , and $(t_k^i - t_k^j)^{cal}$ is the difference in the calculated travel-times between events i and j . The tomographic inversion calculation develops a velocity model that minimizes the residual for all event-receiver pairs.

A velocity model, which consists of grid points with assigned initial velocity values, is required to perform velocity inversion. An initial background velocity was assigned to each node of the velocity models. Two velocity models were created, one covering the entirety of the studied rock mass and the other only covering the mining excavation region.

As can be seen in Figure 4.1, the nodes range beyond the limit of the mining area. This larger velocity model covered all of the sensors that monitor seismicity. This ensured that the model contained all relevant raypaths to provide sufficient information for velocity inversion. Considering the ray distribution of the selected data, a grid with 24×24 m node spacing covering active mining area, combined with a horizontal grid with 48×48 m node spacing covering external

areas beyond active mining region, were designed for the large inversion model. By using this arrangement of grid spacing, we obtained a coarse but reliable images of the velocity structure.

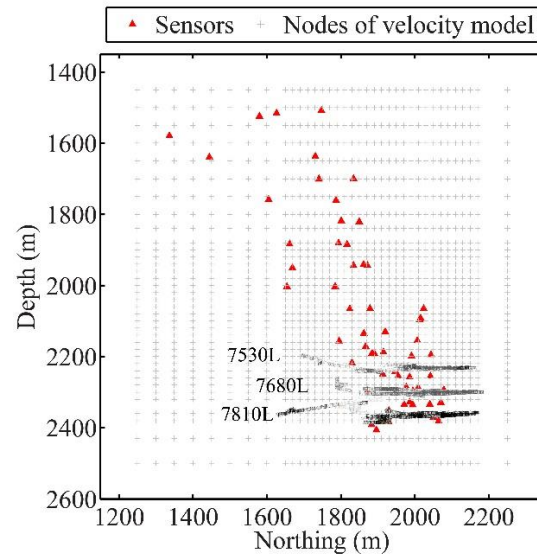


Figure 4.1. The large velocity model

A small-scale velocity model was constructed to generate a detailed velocity distribution only emphasizing the excavation region. As shown in Figure 4.2, grid points are evenly distributed in a rectangular pattern enclosing the area from 7530L (2295 m depth) to 7810L (2380 m depth). There were 9 m between adjacent nodes of the small-scale velocity model.

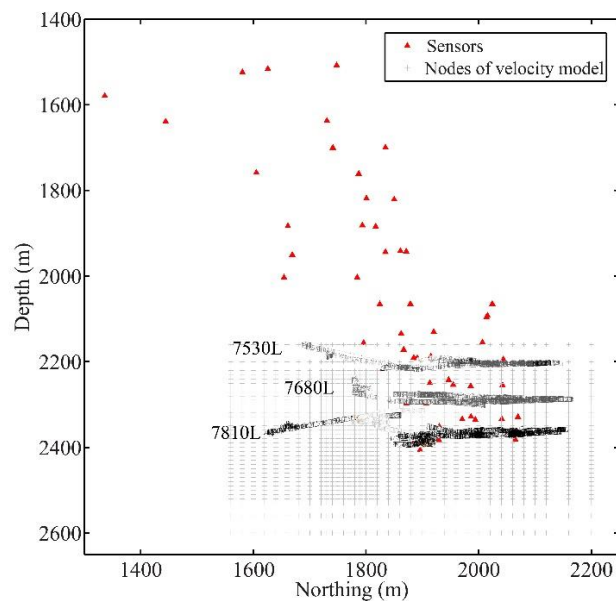


Figure 4.2. The small velocity model

4.4 Results

The correlation between microseismic events and the major events (moment magnitude > 1.0) was analyzed. The accumulating number of microseismic events was plotted from July 4th to July 15th (Figure 4.3). Approximately 2500 microseismic events were recorded on July 6th, 2011, the day of Major Events 1, 2 and 3. Also, the amount of microseismic events was considerable on July 10th, 2011 when Major Event 4 occurred, whereas less than July 6th, 2011. As can be seen in Figure 4.3, major events were associated with an increased amount of microseismic events.

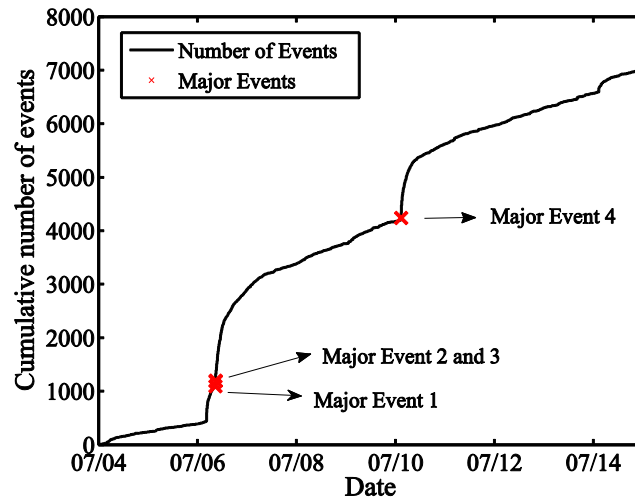


Figure 4.3. Cumulative number of events

It is well known that microseismic events arise from the release of stored elastic energy in a rock mass. The release of stored energy overcomes the force of crack resistance (a function of the plastic behavior of the material at the crack tip and of its fracture characteristics) and creates the surface of a new fracture (Griffith, 1921). Meanwhile, the energy is consumed with the development of a new crack surface. The surplus energy would turn into kinetic energy after a crack propagation (Ba and Kazemi, 1990). Seismicity, including vibration or trembling, are caused by the quick release of stored energy into kinetic energy. Consequently, microseismic events are associated with fracture development and fractures appear to be near them.

We compared the difference in velocity structures before and after the major events. Then the observed difference in velocity structure could potentially be used for forecasting the occurrence of major events. By investigating the velocity change associated with the major events, passive seismic imaging could contribute to improving safety in underground mining.

A. Coarse Grid Results

Microseismic events were divided into four data sets on the same time length (one week) to investigate the possible velocity change associated with the major events. As shown in Table 4.2, time frames were arranged consistently and included the same duration without overlapping before and after major events.

Table 4.2. Time frames with major events occurrence

Time Frames	Start	End
First week	June 22 nd , 2011	June 29 th , 2011
Second Week	June 29 th , 2011	July 6 th , 2011
Major events period	July 6th, 2011	July 10th, 2011
Third week	July 10 th , 2011	July 17 th , 2011
Fourth week	July 17 th , 2011	July 24 th , 2011

The velocity distribution during each of these time frames was determined by performing double difference tomographic inversion. The major events were located close with horizontal drifts with excavations and a vertical cross section crossed the center region of drifts (Figure 4.4).

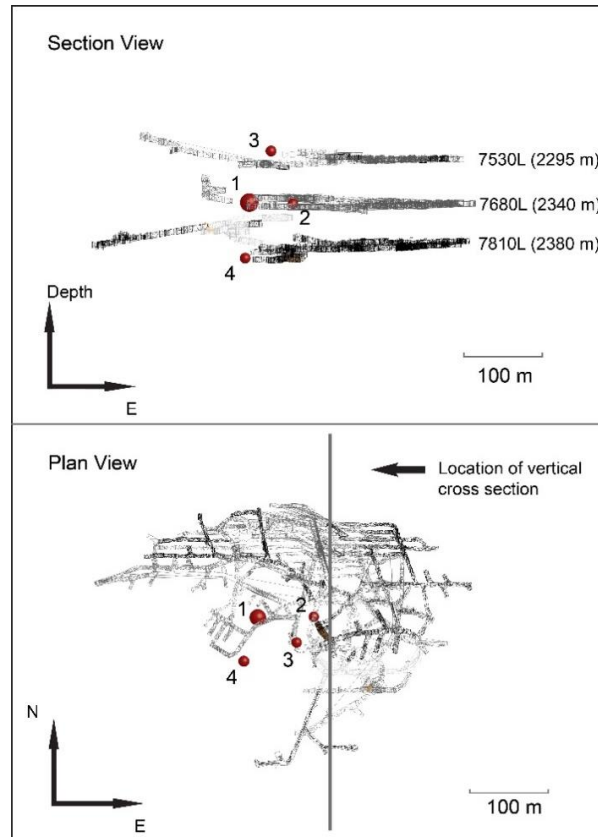


Figure 4.4. The side view (top) and plan view (bottom) of the major events in the study area. The main drifts are plotted and noted with depth. Red balls: major events.

The cross-sectional images of the velocity model generated from tomographic inversion of the travel time data for each of the four time periods are shown in Figure 4.5. The velocity distribution changed through the different time frames involved with the occurrence of the major events. There were a total of two weeks of travel time measurements before the major events. These two weeks remained similar to the general velocity distributions (Figure 4.5). A low-velocity anomaly is horizontally exhibited in the images for both weeks along the 7530L (2295 m), which had a lower density due to drift and stope excavations. The other low-velocity areas, originally scattered below the ends on north and south of the 7530L, connected and developed to a large low-velocity region in 7530L from the first week to the second week. The existence of fractures might impede the wave propagation of this region. Also, a high-velocity region was growing below the south end from 7810L (2380 m) to 7680L (2340 m), approaching the regions of major events. Note that the shift of the low velocity to high velocity indicates stress increases over that area.

The velocity distribution over the two weeks before the major events showed that stress increased above and below the excavation region, which, in contrast, was originally dominated by the modest velocity. The images from the first week to the second week indicated a decent repeatability, providing a measure of confidence in the tomography results without influence by occurrence of major events.

All of the major events occurred on July 6th and July 10th. As shown in Figure 4.5, the velocity distribution changed significantly after major events, especially on the regions adjacent to major events.

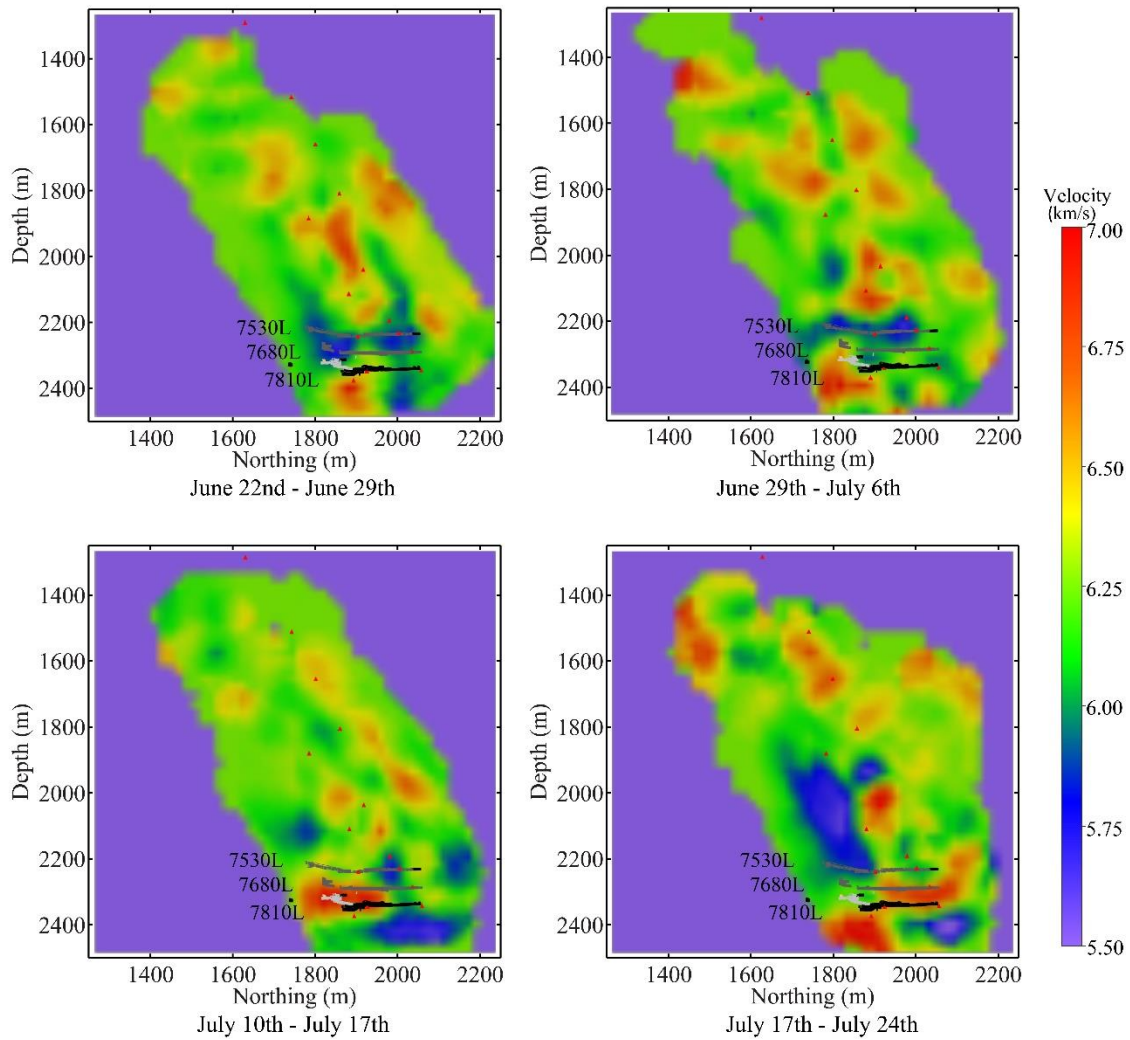


Figure 4.5. Tomography studies on large velocity model. Red trigangles: sensors in the seismic

First, the low velocity band around the 7530L shown on tomograms of the first and second week diminishes to a small spot on the north end of 7530L in the third week. Second, the high velocity anomaly originally located on the south end of the 7680L before the major events moved upwards vertically and was shown on the third week. Also, it stretched out to the north following the drift of 7680L on the fourth week. The overall velocity eventually achieved a greater level in the two weeks around drifts before the major events. It is inferred that stress concentrated near the coming major events prior to the occurrence of them. With approaching to the occurrence of Major Event 1, stress concentration overcame stress relief in the principal part around Major Event 1. The intensified stress concentration and relief in different areas are the main changes associated

with the occurrence of the major events. Observations imply that the major events are more likely to occur within the highly-stressed area.

Besides pronounced changes of the velocity structure with major events, it continued to alter from the third week to the fourth week. The results of the third and fourth weeks (Figure 4.5) showed this tendency. The high-velocity areas of the second week diminished significantly and a high-velocity anomaly was observed between 7680L and 7810L in the third week (Figure 4.5). A high-velocity region below 7680L, which appeared in the third week, expanded on both the north and lower zones from the third week to the fourth week. Most of the largest low-velocity region below 7810L in the third week was replaced by the extending of neighboring moderate-velocity body and high-velocity body in the fourth week. In addition, a large low-velocity area was present above the south end of 7530L in the fourth week. By comparing the first week to the fourth week, we investigated that the occurrence of major events was associated with an increase in velocity from 7530L to 7680L. It is inferred that highly-stressed areas eventually extended on the zones adjacent to 7680L and 7810L in the fourth week. Stress relieved on the zones above the south end of 7530L in the last week. The high-velocity anomalies around the excavations suggested that stress concentration was enhanced during the third and fourth weeks.

B. Fine Grid Results

To evaluate the effect that major events had on the state of velocity in sufficient detail, we performed tomographic inversion using a finer velocity grid. A small velocity model only concentrating on the area from 7530 L to 7810 L was developed using the identical time frames as the coarse grid spacing analyses. The distance between neighboring grid nodes was 9.1 m in every principal direction for the small-scale model. By using finer grid spacing, the tomographic images had better resolution and included additional details.

The set of tomograms corresponding to the small-scale velocity model was shown in Figure 4.6. It was arranged in the same pattern as that of the results computed on the large-scale velocity model. As with the initial, coarser model, the similar velocity distribution around the mining area maintained from the first week to the second week.

The region around the drifts was mainly governed by the growing high-velocity portion after the occurrence of the major events. Also, the low-velocity zone around the 7810L drift of the second week disappeared and was replaced by a large high-velocity region, causing an imbalanced

distribution of velocity. Another velocity change was observed on 7530L from the second week to the third week. The low-velocity area of the south end of 7530L increased and exceeded the average velocity level in the third week. The changes between these two weeks gave the evidence that the stress concentrated on the region of excavation with the occurrence of major events.

There was a substantial velocity distribution change from the third week to the fourth week, as shown in the tomographic images (Figure 4.6). Regions around the mining area experienced a significant increase in velocity from the third week to the fourth week. Some regions still maintained low velocity, but most of the area around the drifts was dominated by high velocity. The significant change from the third week to the fourth week suggested that stress redistribution might continue over several weeks after major events. The additional noticeable change was that stress concentrated in the north end of the region from 7530 L to 7680 L, where there was originally an area of low stress in the first week. We compared the tomograms of the first week and the last week and concluded that the overall stress of the last week was higher than the first week.

For the results of the small-scale velocity model, there was a similar trend with the change for the large-scale velocity model. The average velocity of the mining area increased substantially from before the major events to after the major events, as shown in Figure 4.6. The principal drifts in the mining area experienced the most dramatic change. It took multiple weeks of monitoring to observe the new stability. During the third week after the occurrence of major events, a high-velocity area dominated near the major events locations. Low-velocity areas appeared on the north of drifts and were away from the location of major events. In the fourth week, the high velocity region experienced a trend of moving away from the center of drifts and low velocity areas distributed more extensively around the drifts.

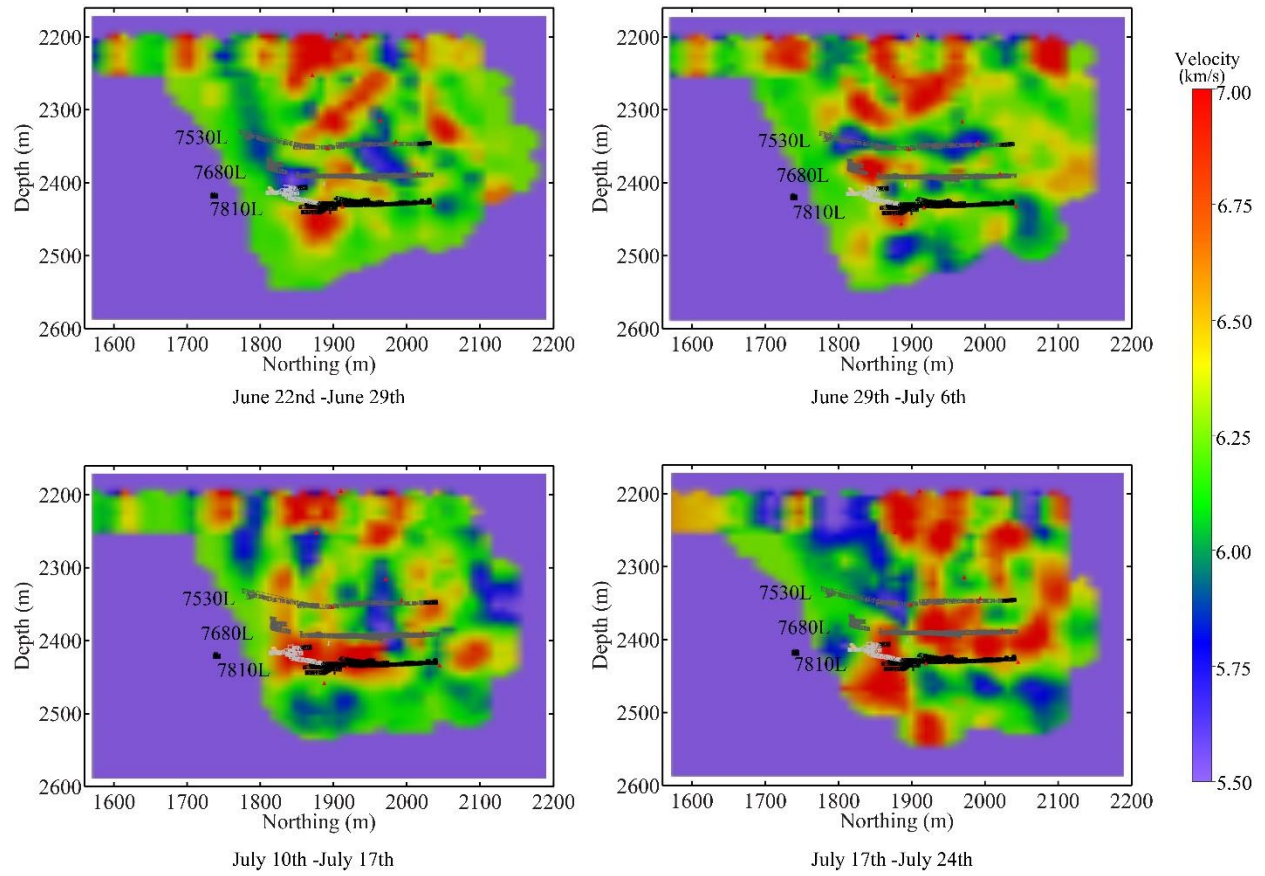


Figure 4.6 Tomography studies on small velocity model. Red triangles: sensors in the seismic

In the early period after the major events, a low-velocity area formed because the wave propagation was impeded by the fractures resulting from major events. Two possible reasons were responsible for the overall velocity level increase after several weeks. First, the newly formed fractures in the rock mass compacted and closed, which was the crack healing effect. Second, the modulus of the rock mass increased due to the compaction effects.

4.5 Summary and discussion

The tomograms generated from different time periods showed the effect of major events on the velocity structure, which reflected stress distribution in a mine site. By comparing them, significant change of the velocity structure was observed associated with the major events, indicating that stress concentrated to a greater extent with the approaching of the major events and

continued to maintain a high level near excavations in mines after major events. These findings confirm that highly stressed regions exist surrounding the major events in rock bursts. These regions experience significant stress redistributions before and after occurrences of major events.

The regions around the mine drifts indicated a great contrast between low and high velocities associated with the occurrence of major events and the active microseismicity. The change in velocity structure validated the effect of redistributed stresses. For example, the switch from low velocity to high velocity suggested stress concentration. An increase in fracture density of a region decreased the velocity where the seismic waves traveled through. Consistencies on the velocity distributions before the occurrence of major events proved the stability of stress distribution without the influence of major events.

Stress concentration and stress relief usually occurred simultaneously in close proximity. They both were intensified, affected by the active microseismicity and the major events. We inferred that the occurrence of major events significantly altered the stress distribution. In addition to the main consistency of stress distribution, the highly-stressed region appeared in the nearby of major events. After the major events, pronounced stress concentration appeared in the vicinity of major events. It was observed stress relief around the drifts and low stress areas were monitored close to the drifts within two weeks after the major events. We determined that stress concentration preceded the occurrence of major events from the results. However, whether major events were the main factor that caused the growth of stress concentration after the major events still needs further investigation. One limitation of the analysis is that the time frames used to generate tomograms might be arbitrary. The time scale of one week is reliable to provide enough raypaths coverage for double-difference tomographic surveys. However, the different amount of microseismic events in each data set could cause a potential bias in the tomographic surveys. In order to offset such bias, we propose to compare the results from data sets including the same amount of seismic events in future studies. Also, one proposed solution is to use overlapping time frames to exhibit better consistent change of the stress distribution combined with using the same amount of seismic events in each data set.

We also noticed that seismic rate was significantly boosted with the occurrence of major events, but it cannot be simply used for accurately predicting major events. Whereas a growing high seismic rate was observed preceding Major Event 1, 2 and 3, no significant seismic rate

change was found before Major Event 4. It is possible to apply tomographic studies in seismic hazard analyses in underground mines. There is a potential for forecasting major events by combining tomographic studies with the monitoring of abnormal seismic rates.

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5 Stress redistribution with the change of frequency of mining-induced seismicity

5.1 Abstract

Mining-induced seismicity in hard-rock underground mines occurred in response to stress generation due to excavations. The pattern of seismicity rates and how it changes with stress transfer remains unknown. Double-difference tomographic studies were used to determine the velocity structures, which served to illustrate stress distributions within the rock masses. Seismic events from Creighton Mine and Kidd Mine were obtained and temporally divided into a number of different groups, each of which including 1000 seismic events. By interpreting tomograms from every group of seismic events, stress and mining-induced seismicity were compared with velocity structures in the mining region. Velocity anomalies were examined with the distribution of seismicity. Stress concentrations were assessed by combining the information from the velocity imaging and seismicity. It has been observed that crustal earthquakes have a significant Gutenberg-Richter relationship on frequency and magnitude of seismic events. By investigating the frequency and magnitude of mining-induced seismicity, it is recognized that mining-induced seismicity of Creighton Mine and Kidd Mine exhibited a similar pattern on frequency and magnitude with earthquakes.

5.2 Introduction

The Creighton Mine and Kidd Mine are the sites of significant seismicity due to the excavation of great depths. Large magnitude events in the underground mine cause destruction in rock mass and infrastructures. Integration of seismic P-wave velocity imaging from microseismicity is useful for evaluating geomechanical response of the rock mass (Young, Maxwell et al. 1992). In natural earthquakes, it is believed that seismicity rate is enhanced by stress increases and suppressed by stress decrease (Toda, Stein et al. 2005). An earthquake is able to enhance or suppress subsequent seismicity, depending on locations and orientations of seismicity. Static stress changes influence the space-time pattern of seismicity before and after earthquakes (Bowman and King 2001). It has been generally agreed that events produce regional stress perturbations (King, Stein et al. 1994). Seismic studies of the Creighton Mine have been conducted by Malek et al. (2008). They provided evidence that most of the seismicity occur in proximity to excavations at Creighton Mine due to the mining induced stress fracturing (Malek, Espley et al. 2008). It was concluded that there is no distinct relationship between geological structure and seismicity. In addition, it is not recognized

that seismicity is correlated with shear zones in Creighton Mine as well (Snelling, Godin et al. 2013). Spatial correlations between stress and seismicity can be assessed by velocity imaging. Velocity imaging is a useful tool to evaluate the correlation between stress and seismicity. In velocity imaging techniques, double-difference tomography demonstrated substantial improvement on velocity inversion and relocating seismic events. It is generally agreed that the recorded seismicity is more likely to be confined to the high velocity areas (Wenzel, Sperner et al. 2002). It is necessary to note that seismicity locations and stress distributions in mining are not exactly the same as the situations in earthquakes. Case studies combining velocity images and induced seismicity of mining excavation were shown that the induced seismicity were evidently located in an area of velocity transition between the high-velocity (high-stress) and low-velocity (fractured zones) (Maxwell and Young 1996).

The goal of this work is to use double-difference tomographic studies for investigating and evaluating the stress distribution and seismic rate change. Seismicity rate changes were correlated with the regional stress change in underground rock mass. Spatial and temporal evidences from Creighton Mine and Kidd Mine proved that seismicity was more likely to be located in transition areas between high stress and low stress in seismicity active periods. The other purpose of this study is to identify the relationship between frequency of events and magnitude. Frequency-magnitude relationships of microseismic events were assessed based on the seismicity of two deep underground mines.

5.3 Data and methods

The double-difference tomography code *tomoDD* was used to image 3D velocity structure (Zhang and Thurber 2003). This technique incorporates the travel time difference of two paired events and the locations of events. In Creighton Mine and Kidd Mine, the microseismic system consists of triaxial sensors and uniaxial sensors, which can provide sufficient coverage of the mining region. Figure 5.1 shows the sensors arrangements and raypaths coverage on some microseismic events. Double-difference tomography performed velocity inversion using the locations of seismic events and travel time from events to sensors. Data from Creighton Mine were recorded between April and September, 2011. The selected data set consisted of 651,436 P-arrival times from 47,428 microseismic events. The 47,428 microseismic events used in double difference tomographic inversion were selected using the criteria that sensors picks were more than 8.

Similarly, data from Kidd Mine were recorded between March, 2007 and March 2012. The selected data included 190,568 P-arrival times from 39,160 microseismic events.

The tomoDD uses initial velocity models to perform velocity inversion. The initial velocity model was a 3D grid of points with a uniform P-wave velocity. We arranged 40 layers on depth in the seismic tomographic model. Each layer was subdivided into 40×40 grids. To include all the raypaths of Creighton Mine, the velocity model was applied over a 3300 ft (1000 m) $\times$ 3300 ft (1000 m) area and the range of depth of the velocity model was from 4700 ft (1433 m) to 7800 ft (2377 m). The distance between grid points was 82.5 ft (25 m) in the Northing and Easting directions and 77.5 ft (24 m) in the Depth direction. The velocity model of Kidd Mine was applied over a 680 m $\times$ 760 m area. The range on depth of the velocity model for Kidd Mine was from 400 m to 1400 m. The distance between grid points was 17 m in Easting direction, 19 m in Northing direction and 25 m in Depth direction. The initial velocity models were assigned 6250 m/s for Creighton Mine and 6000 m/s for Kidd Mine. Velocity inversion and location of events inversion were conducted until achieving the reliable accuracy. Velocity structures were displayed after interpolating 3D grid of points on last iteration of P-wave velocities.

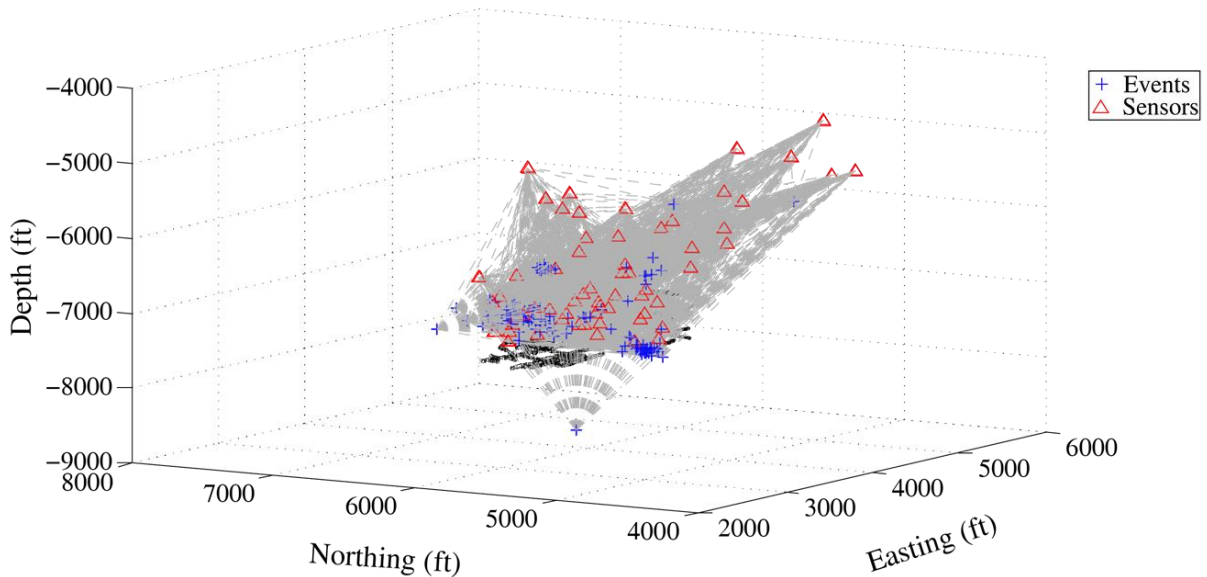


Figure 5.1. Sensors location and raypaths coverage in Creighton Mine

The seismicity rates were uneven in both Creighton Mine and Kidd Mine due to the stress change, which was affected by excavations and production blasts. The cumulative number of microseismic events changed with the date in Creighton Mine as shown in Figure 5.2. There was

an average of 250 events per day in Creighton Mine. The periods with abundant microseismicity can be recognized by the trend of the curve. It was identified that July 6th, July 10th, and August 4th experienced high seismicity rates. In tomographic studies on Creighton Mine, microseismicity data was divided chronologically by number of groups and every group included 1000 microseismic events. The same number of events in every group provided the close density of raypaths. Furthermore, the groups of 1000 microseismic events guaranteed that abundant raypaths were accessible to generate the tomograms on high resolution.

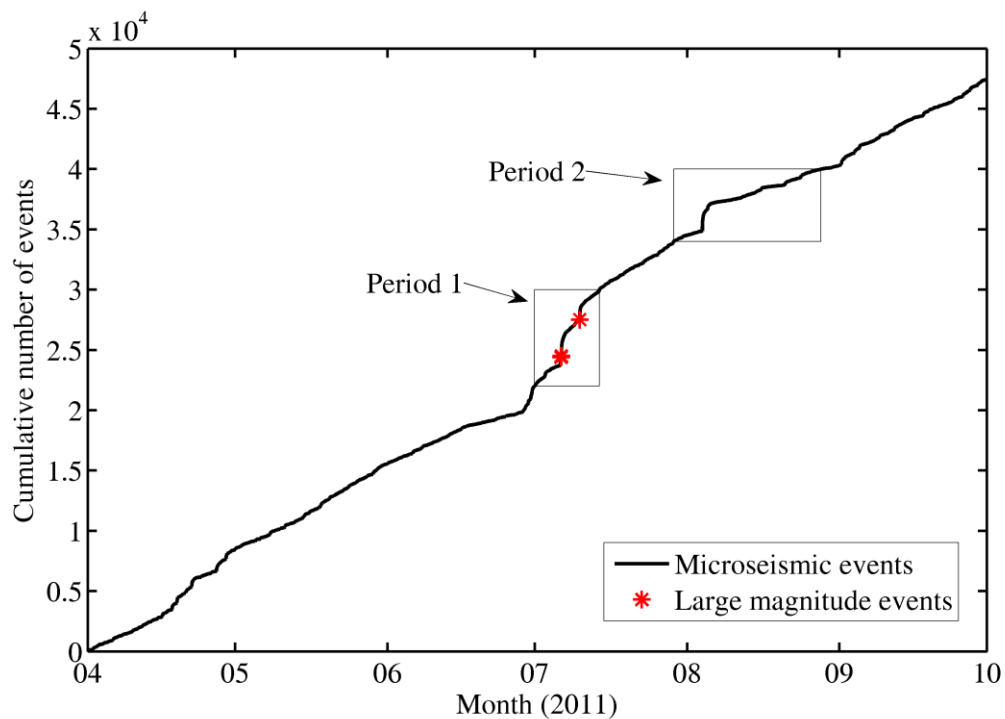


Figure 5.2. Cumulative number of microseismic events in Creighton Mine

The seismicity rate in Kidd Mine was much lower than that in Creighton Mine (Figure 5.3). The average number of microseismic events of Kidd mine was less than 20 per day from April, 2007 to March, 2012. Similarly, the data of Kidd Mine was divided into multiple groups, each one of which includes 1000 microseismic events. Combinations of tomograms and locations of seismicity from different groups are shown for comparison.

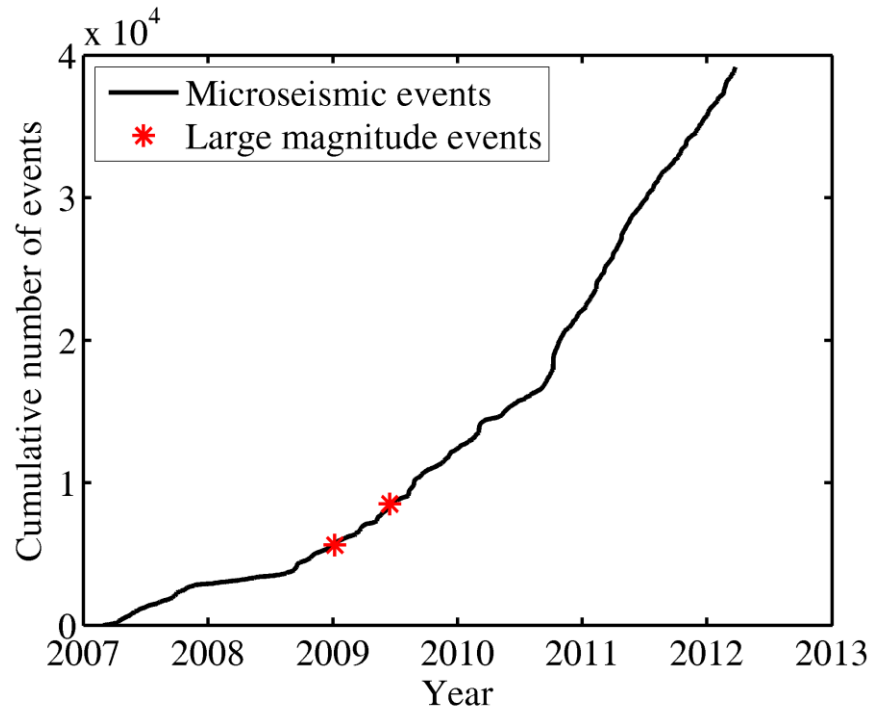


Figure 5.3. Cumulative number of microseismic events in Kidd Mine

The Gutenberg-Richter law (Gutenberg and Richter 1944) described a model that demonstrates the number of earthquakes with a magnitude $M > m$. It was applied in seismic activities of Creighton Mine and Kidd Mine to explore frequency-magnitude relationships of seismicity. The number of eligible seismic events and moment magnitude M_w for the time interval between April and September, 2011 of Creighton Mine are analyzed and shown in Figure 5.4. Similarly, the number of seismic events and moment magnitude M_w for the time interval between April, 2007 and March, 2012 of Kidd Mine were analyzed (Figure 5.5). Both frequency-magnitude relationships showed that mining-induced seismicity agreed closely with the Gutenberg-Richter law. The frequency of seismicity decreased with the rise of moment magnitude of seismic events.

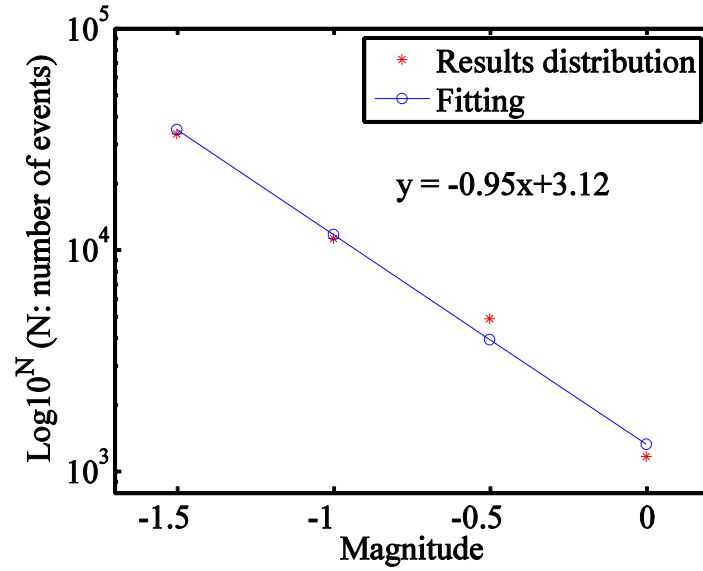


Figure 5.4. Frequency-magnitude relationship for the Creighton Mine

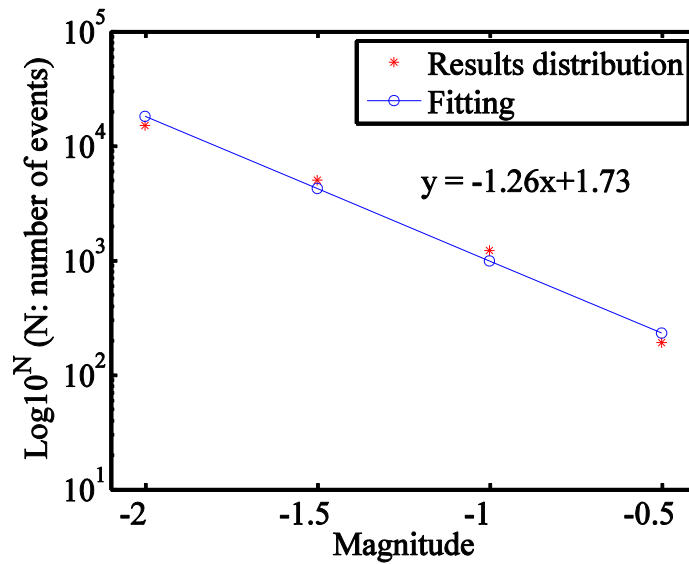


Figure 5.5. Frequency-magnitude relationship for the Kidd Mine

5.4 Results

Seismicity rates in both Creighton Mine and Kidd Mine were computed. The seismicity rate of each group in Creighton Mine ranged from 5 to 291 per hour (Table 5.1 and Table 5.2). Seismicity rate of Kidd Mine ranged from 29 to 221 per day (Table 5.3). Plan view tomograms with seismicity rates during July and August were compared.

Creighton Mine

The seismicity rate changed significantly during the time intervals between March, 2011 and September, 2011. It achieved high levels on July 6th and August 6th. The cross-section of Creighton Mine shows the location of the series of tomograms (Figure 5.6). Seismicity distribution and tomograms of level 7700 are shown in Figure 5.7 and Figure 5.8. Each tomogram was generated based on 1000 events in the same group, which consists of microseismic events that occurred within different time intervals. It is illustrated that microseismic events mainly concentrated on the area with a center at Easting 4500 ft and Northing 6200 ft from June 30th to July 11th. Some microseismic events scattered outside the drift area. It is noted that microseismic events were more concentrated when seismicity rate increased. The peak seismicity rate was achieved (193 events per hour) between 04:43 and 09:54, July 6th events (Figure 5.7. C). Also, the seismicity rate stayed high (101 events per hour) during the next events group from 09:54 to 19:48, July 6th (Figure 5.7. D). Meanwhile, the velocity images show that high velocity occupied the studied areas when they reached the large seismicity rates. It is inferred that the maximum principle stress strikes from NW to SE from the velocity images. Furthermore, stress increased when the seismicity rate of the rock mass increased. The location of event groups failed to move with the high stress region changing. Also, induced seismicity was not located in the high stress region, but was more likely to concentrate on the area between the high stress and low stress zones. With the decrease of seismicity rate after 19:49, 6th, July (Figure 5.7.E), the high velocity area started declining until the seismicity rate switched to recover on 10th, July (Figure 5.7.G).

Table 5.1. Seismicity rate of July in Creighton Mine

Events Group	Events Number	Start	End	Hours	Seismicity Rate (per hour)
C1	1000	6/30/2011 22:51	7/3/2011 3:20	52.97	18.88
C2	1000	7/3/2011 3:24	7/6/2011 4:43	73.32	13.64
C3	1000	7/6/2011 4:43	7/6/2011 9:54	5.18	192.93
C4	1000	7/6/2011 9:54	7/6/2011 19:48	9.90	101.01
C5	1000	7/6/2011 19:49	7/8/2011 19:57	48.12	20.78
C6	1000	7/8/2011 19:57	7/10/2011 3:38	31.68	31.56
C7	1000	7/10/2011 3:39	7/11/2011 4:41	25.03	39.95
C8	1000	7/11/2011 4:41	7/14/2011 3:42	71.02	14.08

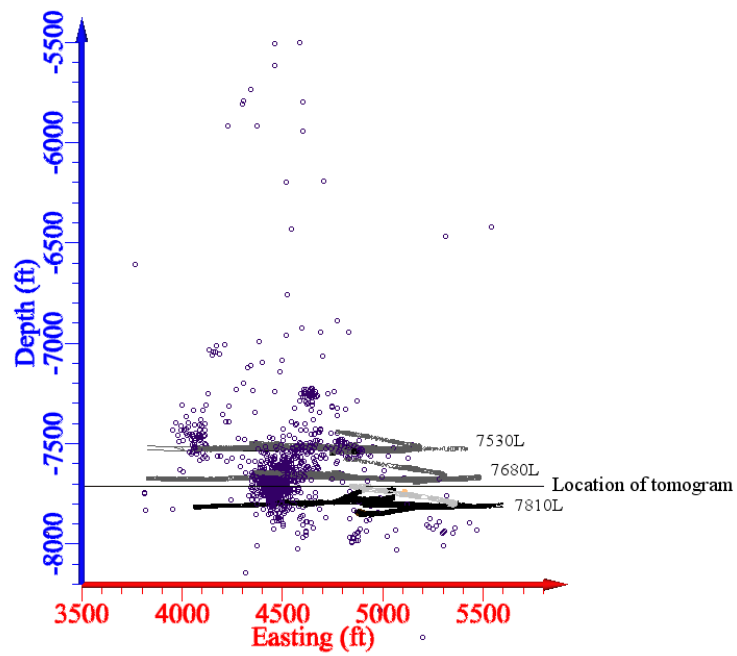
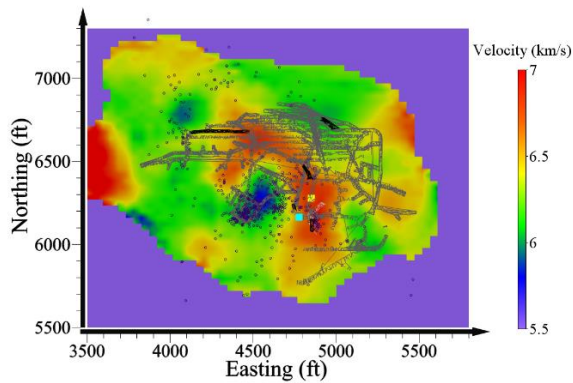
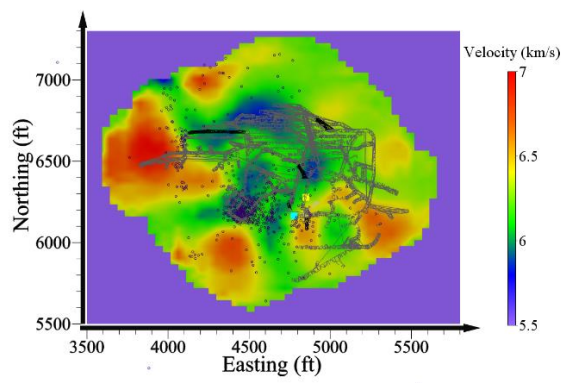


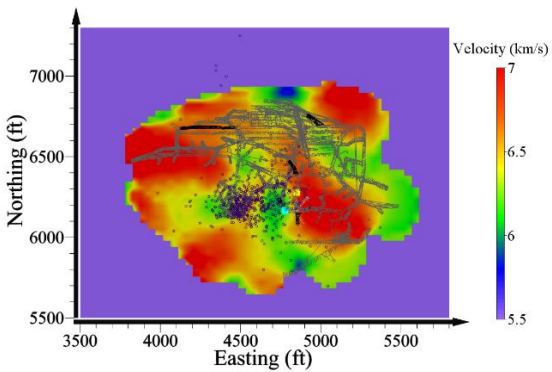
Figure 5.6 Cross-Section of Creighton Mine



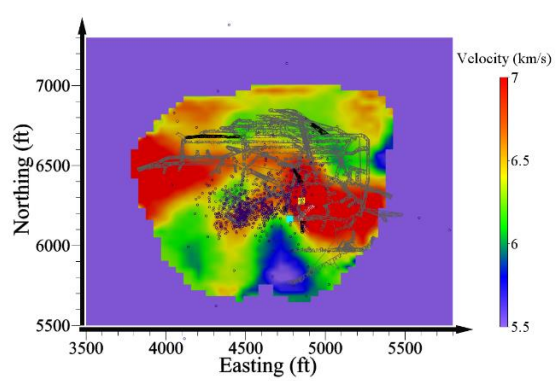
A. Event Group C1



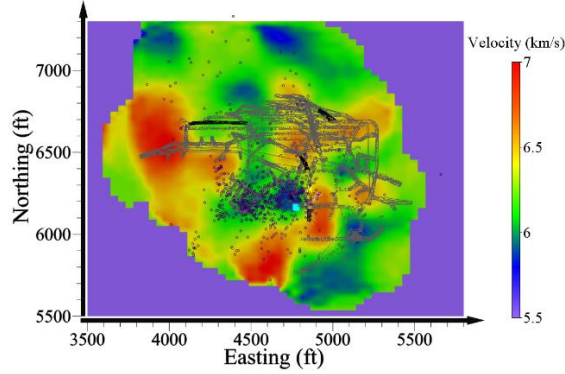
B. Event Group C2



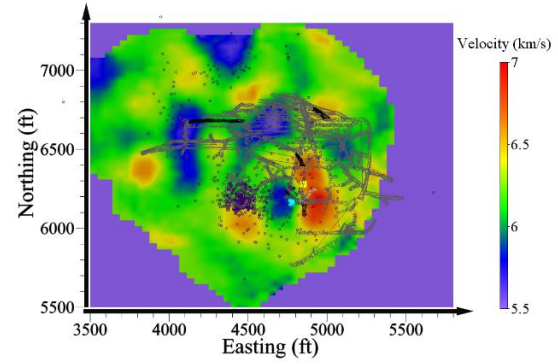
C. Event Group C3



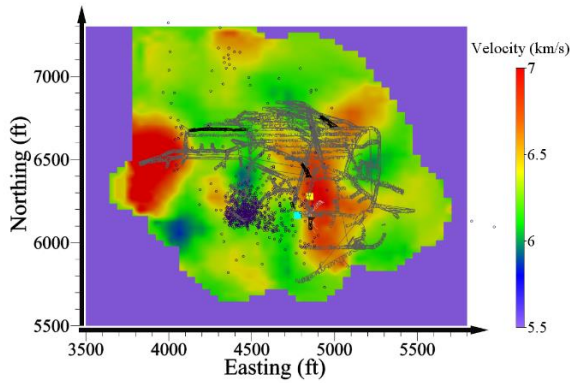
D. Event Group C4



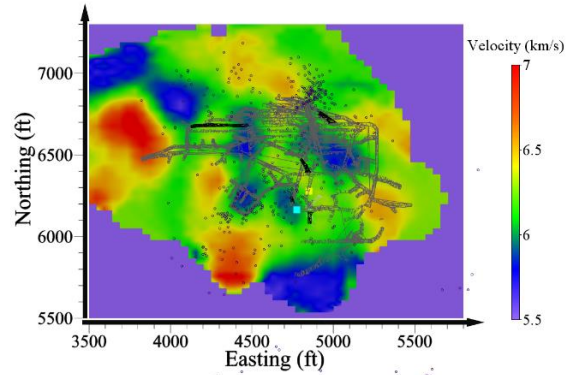
E. Event Group C5



F. Event Group C6



G. Event Group C7



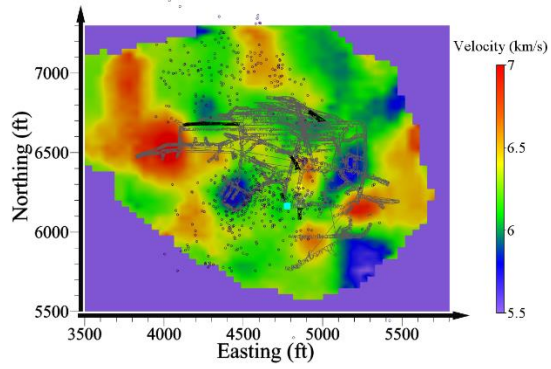
H. Event Group C8

Figure 5.7. Velocity on level and seismicity distribution in July of Creighton Mine

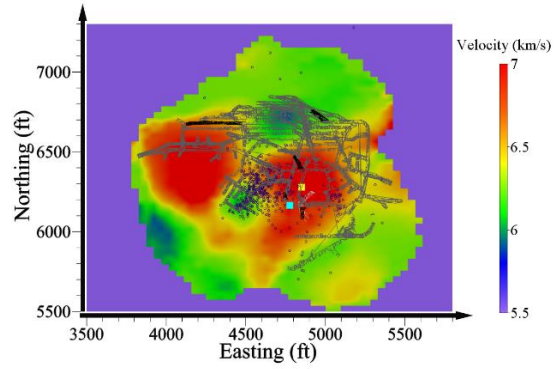
To evaluate whether the stress increases with larger seismicity rate, the other period with active seismicity was also examined. The other active seismicity with high seismicity rate appeared on 4th, August in Creighton Mine. The tomograms and microseismicity distribution around the seismic active periods were analyzed. The seismicity rate increased significantly and reached 291 events per hour on April, 4th, 2011. The tomograms of the same layer apparently reflect the velocity change. In particular, it indicates that higher velocity occupied the region when the seismicity rate was at its maximum (Figure 5.8.B). Events of this period were located on the intermediate velocity area between the two high velocity areas. With the lowering of the seismicity rate, low-velocity areas started to increase. The low stress is associated with the low-velocity area. After 14th, August, low-velocity (< 5.7 km/s) areas occupied most regions of the tomographic studied area.

Table 5.2. Seismicity rate of August in Creighton Mine

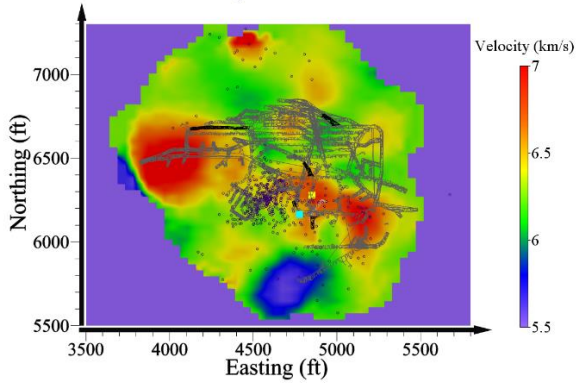
Events Group	Events Number	Start	End	Hours	Seismicity Rate (events/hour)
C9	1000	7/29/11 6:13	8/4/11 2:39	164.42	6.08
C10	1000	8/4/11 2:39	8/4/11 6:05	3.43	291.26
C11	1000	8/4/11 6:05	8/5/11 9:18	27.20	36.76
C12	1000	8/5/11 9:31	8/14/11 11:55	242.40	4.13
C13	1000	8/14/11 12:49	8/21/11 16:32	195.72	5.11
C14	1000	8/21/11 16:55	8/28/11 6:29	181.55	5.51



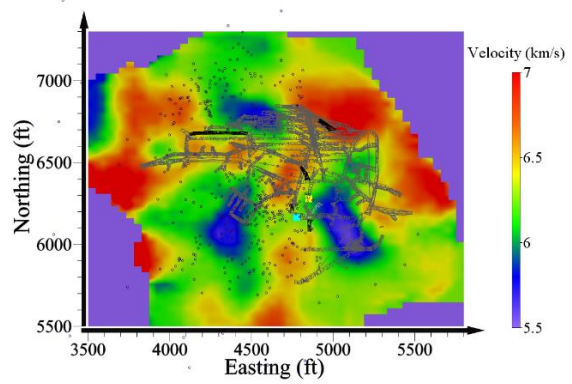
A. Event Group C9



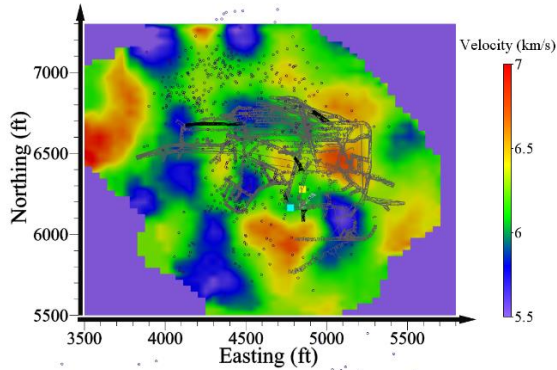
B. Event Group C10



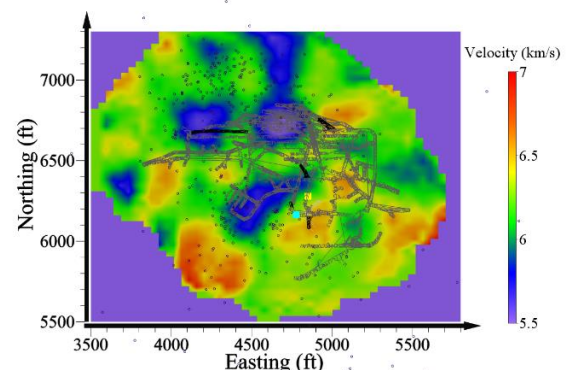
C. Event Group C11



D. Event Group C12



E. Event Group C13



F. Event Group C14

Figure 5.8. Velocity on level and seismicity distribution in August of Creighton Mine

Kidd Mine

The rate of the seismicity of Kidd Mine was significantly lower than that of Creighton Mine. A total of 17,859 seismic events occurred from April, 2007 to March, 2012 and were provided to be analyzed. Like the seismicity of Creighton Mine, seismicity is divided into multiple event groups, each of which consists of 1000 seismic events. The seismicity rate per day is calculated and listed in Table 5.3. There were roughly 10-20 events each day. Velocity structure of each events group was generated by performing double difference tomographic inversion.

Table 5.3. Seismicity rate in Kidd Mine

Event Group	Events number	Start	End	Days	Seismicity rate (per day)
K1	1000	4/1/2007 5:08	6/16/2007 4:37	75.98	13.16
K2	1000	6/16/2007 5:06	9/28/2007 4:45	103.99	9.62
K3	1000	9/28/2007 4:47	3/24/2008 10:30	178.24	5.61
K4	1000	3/24/2008 11:15	9/16/2008 5:45	175.77	5.69
K5	1000	9/16/2008 5:48	12/3/2008 1:34	77.82	12.85
K6	1000	12/3/2008 2:27	2/19/2009 22:30	78.84	12.68
K7	1000	2/20/2009 0:44	5/2/2009 14:32	71.57	13.97
K8	1000	5/2/2009 19:14	6/8/2009 18:52	36.99	27.04
K9	1000	6/8/2009 19:04	8/17/2009 3:56	69.37	14.42
K10	1000	8/17/2009 14:46	11/16/2009 3:08	90.52	11.05
K11	1000	11/16/2009 3:15	2/4/2010 5:28	80.09	12.49
K12	1000	2/4/2010 5:28	3/5/2010 18:23	29.54	33.85
K13	1000	3/5/2010 18:23	5/28/2010 4:37	83.43	11.99
K14	1000	5/28/2010 4:46	10/19/2010 12:09	144.31	6.93
K15	1000	10/19/2010 12:27	2/12/2011 4:50	115.68	8.64
K16	1000	2/12/2011 4:50	5/6/2011 13:58	83.38	11.99
K17	1000	5/6/2011 15:10	8/10/2011 11:12	95.83	10.43
K18	859	8/10/2011 13:42	3/18/2012 22:29	221.37	3.88

Figure 5.9 shows the locations of tomograms to be observed in Kidd Mine. Stress distribution can be inferred through velocity distribution of the tomograms and locations of microseismicity is indicated on tomograms. Furthermore, tomograms calculated from different events groups can be compared to estimate the stress transfer as time periods changed. According to the seismicity rates (Table 5.3), Event Group K8 (Figure 5.8.D), Event Group K9 (Figure 5.8.E) and Event Group K12 (Figure 5.8.H) experienced the highest seismicity rates.

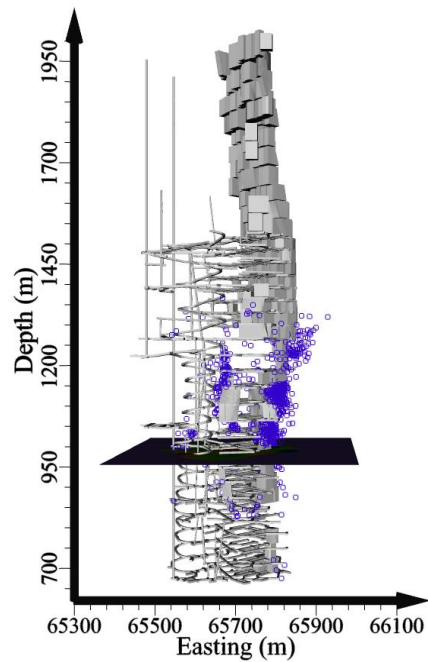
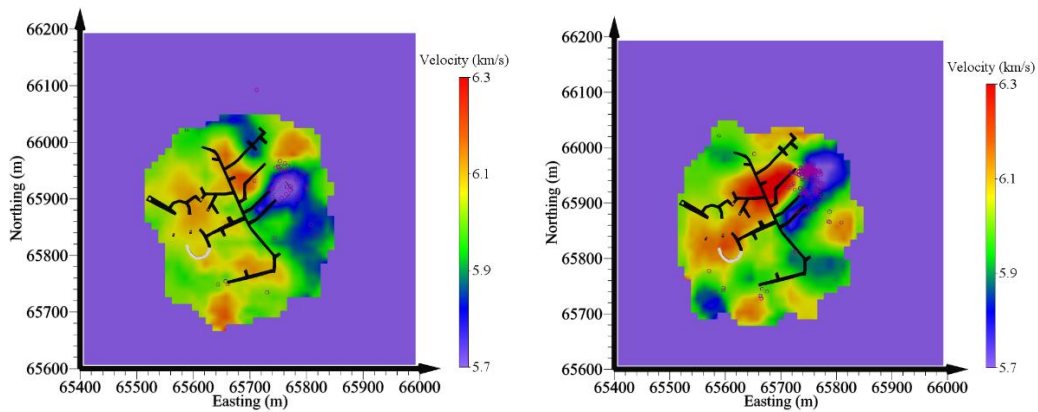


Figure 5.9. Cross section of Kidd Mine

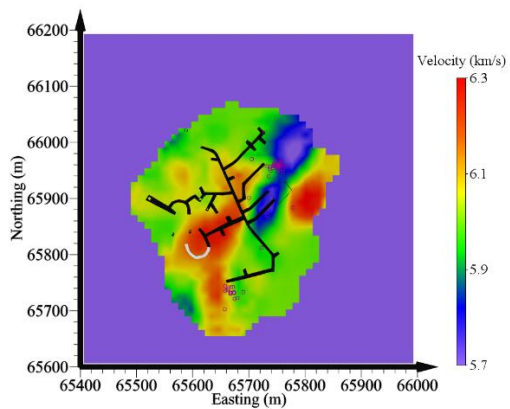
Tomograms calculated from Events Group K5 (Figure 5.10.A) to Events Group K14 (Figure 5.10.J) following the time sequence are shown in Figure 5.10. The tomograms are illustrated with the microseismic events close to the level that tomograms are located on. From Events Group K5 (Figure 5.10.A) to Events Group K8 (Figure 5.10.D), the low velocity region diminished with the raising of the seismicity rate. Although it is generally recognized that seismic events are more likely to occur within a high stress region, microseismic events in this study close to the level of tomogram are located in the vicinity of low velocity regions, and there are a few events located between the borders of high velocity regions and low velocity regions. The largest change of velocity distribution is observed from Events Group K8 (Figure 5.10.D) to Events Group K9 (Figure 5.10.E). The rise of high velocity regions is greatest. The areas adjacent to the entry of NW-SE drift were enhanced. It is shown that the group of events within selected ranges of depth

was mainly located in areas of low and moderate velocity rather than high velocity (Figure 5.10). However, the areas of high velocity grew with the increase of the total seismicity rate. It is inferred that the increase of seismicity rate was associated with enhanced stress, implying that high stress promoted the occurrence of these seismic events. In studies of natural earthquakes, it is recognized that small events enhanced stress at the epicenters of them (Stein, King et al. 1992). Result of this study agrees with the recognition in natural earthquakes that the small events increased stress. However, whether the majority of seismicity occurred in regions where the stress had increased needs further investigations. In addition to the change of the high velocity region, the low velocity region displayed on NE of the drift reduced with the growth of the seismicity rate from Event Group K5 to Event Group K8. Although the seismicity rate of Events Group 9 is lower than that of Events Group 8, both tomograms (Figure 5.10. D and E) suggest that stress concentration significantly enhanced along the NW-SE axis of the drift. Stress kept relieving until the time period of Events Group K12 (Figure 5.10.H). The last high seismicity rate appeared over Events Group K12. As a result of stress reduction before the time of Events Group K12, high velocity regions on the center of drift decreased and low velocity regions developed to the east of the drift. Compared with other Events Groups, the low velocity regions initially displayed in other event groups are replaced by moderate velocity regions. For the event groups occurring after Events Group K12, stress relieved on the east outside of the drift and low velocity regions recovered with the drop of seismicity rate.

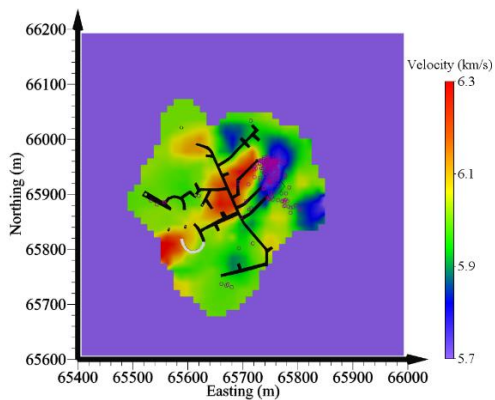


A. Event Group K5

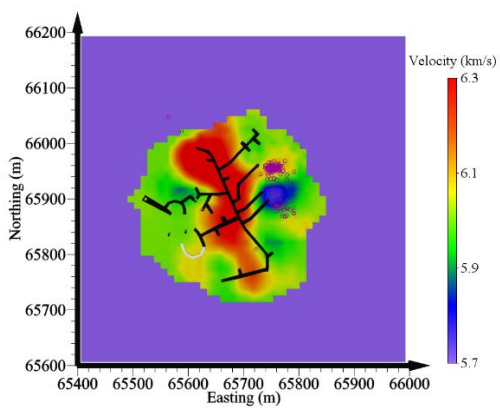
B. Event Group K6



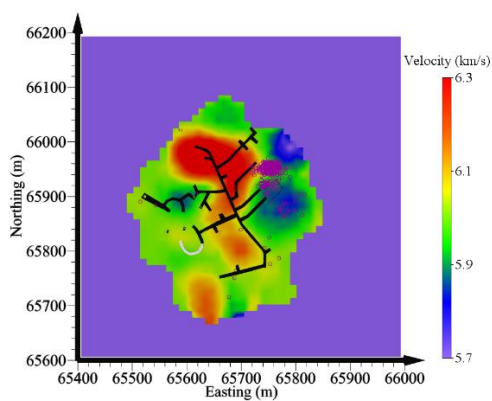
C. Events Group K7



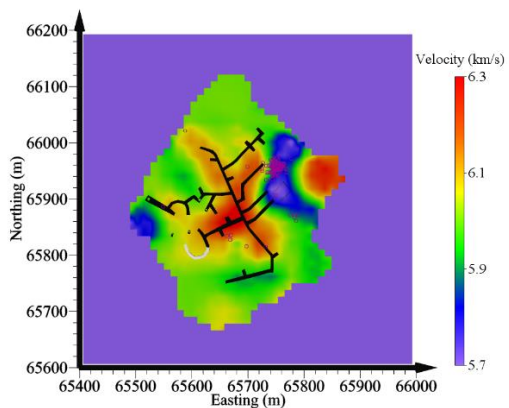
D. Events Group K8



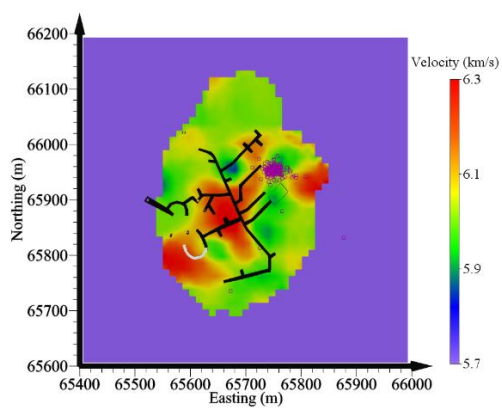
E. Events Group K9



F. Events Group K10



G. Events Group K11



H. Events Group K12

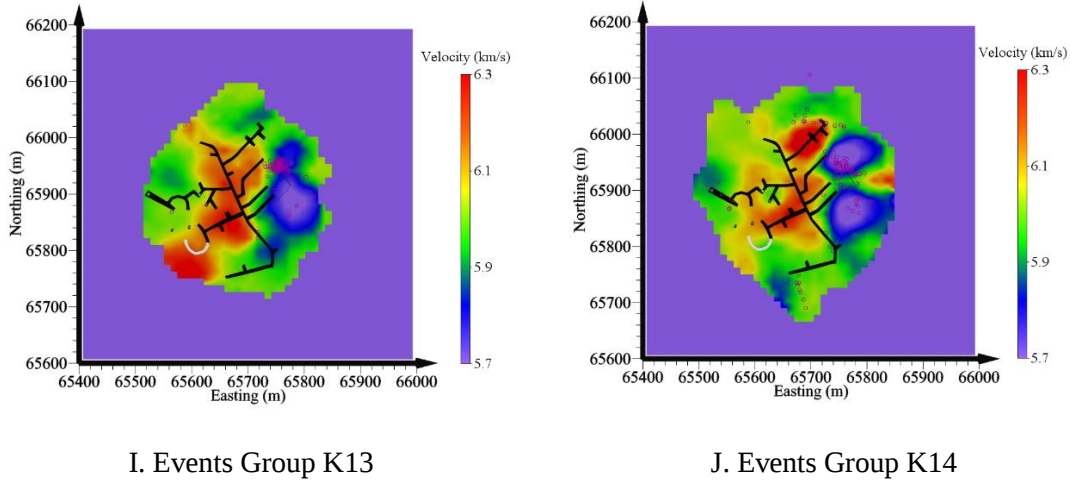


Figure 5.10. Tomograms of Kidd Mine

5.5 Summary and discussion

Mining-induced seismicity rate has shown to be correlated with stress changes on horizontal cross sections. It was observed that high velocity region expands with the growth of the seismicity rate. These findings confirm that the tomographic survey is a useful tool to investigate the relationship between stress distributions and seismic rate changes in underground mines. Measurements on the stress distribution with seismic rate changes provide a continuous estimation on seismic risks since large magnitude events are not triggered in some circumstances even if there is a high seismic rate in underground mines.

It was reported that a mining-induced seismicity sequence appeared on a high seismic velocity region (Young and Maxwell 1992). However, this study's findings show that microseismic events are more likely to occur close to low velocity regions and the areas between low velocity regions and high velocity regions. Although the location of the microseismicity groups is not the same as the high velocity region, the stress accumulation still contributes to the increasing density of seismicity groupings. It can be attributed to the fact that stress relaxes after the energy dissipation, which is associated with the opening and closing of microcracks in the process of forming mining-induced seismicity. Microseismic events are more widely distributed when the seismicity rate is low. Conversely, microseismic events are more spatially concentrated during the time interval of high seismicity rate.

Based on the mining-induced seismic data, Gutenberg-Richter law is applied to demonstrate the correlation between frequency and moment magnitude of seismic events. It is concluded that the mining-induced seismicity can be fitted by Gutenberg-Richter law as well. In order to mitigate seismic risks, future studies will investigate whether there is a specific magnitude range of seismic events, which significantly precede the major events. Then, detecting seismic events with this specific magnitude can play a role in forecasting potential occurrences of major events.

Chapter 5 References

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6 Passive tomographic study on velocity changes in underground mines

6.1 Abstract

Double difference tomographic inversion on measurements of travel time and location are performed to analyze the velocity structure within rock mass in underground mining. Residuals of each iteration are estimated to evaluate the conversion of computation. Average wave propagation speeds in tabulation areas are assessed to compare the velocity change affected by large magnitude events. It is summarized that velocity increases before the occurrence of large magnitude events and then reduces with temporal evolution after them. Possible explanations include static stress buildup enhances the waveform propagation before the large magnitude events and static stress reduction weakens the waveform propagation. Moreover, waveform propagation is attenuated by the dynamic-shaking induced fractures and ruptures within rock masses. Velocity change is shown to be of importance in assessing the stress redistribution and stability of rock masses.

6.2 Introduction

Passive tomographic model is widely used to characterize the structure of earth crust by minimizing difference between simulated and observed seismic waveforms (Korenaga, Holbrook et al. 2000, Tape, Liu et al. 2010). It is estimated that a velocity ranges from 6.0 to 7.0 km/s traveling through the continental crust without deformation (Korenaga, Holbrook et al. 2000). Iterative inversions on both locations of seismic events and the velocity structure are performed till achieving the minimal error (Zhang and Thurber 2003). Analyses of velocity change associated with earthquakes provide evidence that significant velocity tends to decrease after earthquakes (Schaff and Beroza 2004). It is revealed that P and S wave velocities decrease with damaged rock in the earthquakes and velocity recovered due to the healing effect with time (Li, Vidale et al. 2003, Li, Chen et al. 2006). Seismic imaging and microseismic monitoring are used to detect highly stressed and failed regions of underground mines, especially in hard rock mines (Young and Maxwell 1992). In situ stress redistribution influenced by mining excavation can be analyzed by velocity anomalies, which are displayed by tomography. Studies suggested that reliable microseismic monitoring systems are playing a key role for safety in mining (Urbancic and Trifu 2000); however, the knowledge of how to predict rock bursts by using microseismic events is still not enough. It is found that velocity structure of rock masses can reveal the fracture-induced anisotropy and burst-prone regions in mining (Young and Maxwell 1992, Wuestefeld, Kendall et

al. 2011). The aim of this paper is to investigate the velocity change affected by the occurrence of seismic events with large moment magnitude ($M_n > 1$) in underground mining. A three-dimensional tomographic model of the mining region is established to display the velocity change due to the occurrence of large magnitude events.

6.3 Large magnitude events

In the underground mines, sensors of microseismic monitoring systems are installed at a great depth to ensure accurate monitoring of microseismic events. Referring to the record of seismic events in the data set, a considerable amount of seismic events occurred from April, 2011 to March, 2012. It exhibits that the seismic events at the depth of 7000 – 8000 ft take up over 90% of all the seismic events (Figure 6.1). Creighton Mine experienced four large magnitude events in July, 2011 (Table 6.1). The location of microseismic events and large magnitude events provided the reference to the dimension and spatial location for the velocity model.

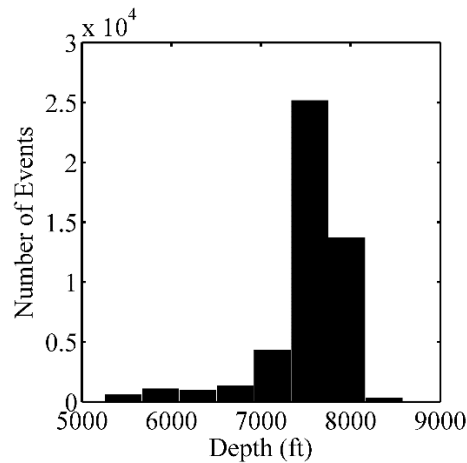


Figure 6.1. Depth range of seismic events in Creighton Mine

Velocity distribution prior to large magnitude events and redistribution after them are compared to discuss the velocity change affected by large magnitude events. The seismic network of Creighton Mine provides a good quality seismic dataset for high resolution tomographic inversion. Velocity profile within a certain spatial and temporal scale is accomplished based on waveform propagation from seismicity to receivers. The period from July 6th to July 10th, 2011 is emphasized because the knowledge on whether some pronounced changes exhibit around the occurrences of four major events would benefit seismic hazard assessment in underground mines.

Table 6.1. Times, Locations, and Magnitude of Major Events in Creighton Mine

Events	Time	North (ft)	East (ft)	Depth (ft)	Magnitude
1	7/6/2011 8:41	6321	4589	7654	3.1
2	7/6/2011 8:46	6284	4850	7653	1.2
3	7/6/2011 8:47	6186	4782	7495	1.3
4	7/10/2011 2:44	6106	4540	7843	1.4

Seismic events were picked and compiled to groups of event-receiver pairs based on whether they occurred prior to or after the large magnitude events. Tomographic results of them are compared to evaluate the change between velocity distributions before and after the large magnitude events. Each group includes 2000 events. Events in all groups are arranged sequentially in time. A double difference tomography method is used to invert simultaneously the location of seismicity and distribution of velocity. The comprehensive description of the tomographic method and software package manual is given by Zhang (Zhang and Thurber 2003).

Table 6.2. Events Grouped for tomographic studies in Creighton Mine

Group	From	To	Number of Events
First group before large magnitude events	6/29/2011 6:23:17	7/1/2011 16:58:54	2000
Second group before large magnitude events	7/1/2011 17:18:11	7/6/2011 8:40:32	2000
First group after large magnitude events	7/10/2011 2:44:58	7/12/2011 2:58:59	2000
Second group after large magnitude events	7/12/2011 21:08:51	7/19/2011 16:11:40	2000

Unlike all large magnitude events of Creighton Mine occurring within a short period range, the two large magnitude events in Kidd Mine are in January and June. As a result of two different periods including large magnitude events, 2000 events are split into four groups. As shown in Table 6.2 and Table 6.4, each group of Creighton Mine consists of 2000 microseismic events, while 500 events are included in each group of Kidd Mine. Two main factors are considered to determine the amount of events in each group. First, the events could provide enough raypaths traveling through the target areas. Then, periods based on the choice on the number of events should be in reasonable ranges of several days in this tomographic study. The seismic rate of Creighton Mine is higher than that of Kidd Mine, thus more microseismic events of Creighton Mine and less microseismic events are included for the period range of several days. Monitoring of seismicity at Kidd Mine shows that two major events occurred respectively in January and June, 2009 (Table 6.3). The number of events distribution on corresponding depth is shown in Figure 6.2.

Table 6.3. Times, Locations, and Magnitude of Major Events in Kidd Mine

Events	Date	Time	North (m)	East (m)	Depth (m)	Magnitude
1	January, 6 th 2009	4:40 AM	65733	65686	1150	3.8
2	June, 15 th 2009	7:01 PM	65861	65737	1035	3.1

Table 6.4. Events Grouped for tomographic studies in Kidd Mine

Events Groups	From	To	Number of Events
First group before large magnitude event 1	10/07/2008 19:20:20	11/20/2008 04:23:44	500
Second group before large magnitude event 1	11/20/2008 05:56:53	01/06/2009 04:07:31	500
First group after large magnitude event 1	01/06/2009 04:57:40	02/05/2009 09:06:37	500
Second group after large magnitude event 1	02/05/2009 15:06:36	03/21/2009 03:50:20	500
First group before large magnitude event 2	05/10/2009 05:58:16	05/28/2009 16:44:05	500
Second group before large magnitude event 2	05/28/2009 21:18:27	06/15/2009 17:40:13	500
First group after large magnitude event 2	06/15/2009 19:19:49	08/10/2009 06:46:51	500
Second group after large magnitude event 2	08/10/2009 06:57:17	08/24/2009 16:53:46	500

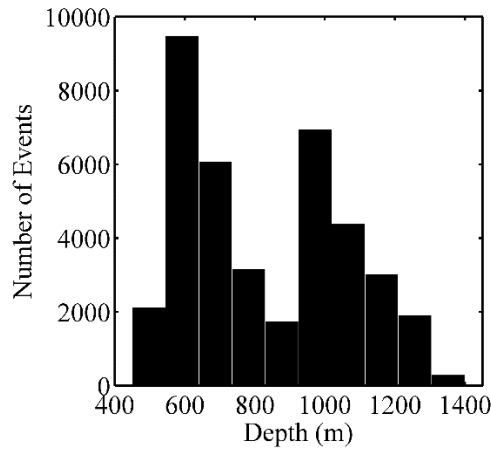


Figure 6.2. Depth range of seismic events in Kidd Mine

6.4 Average velocity analysis in tabulation areas

To analyze the influence on velocity posted by large magnitude events, velocity results computed by double difference tomography are visualized in tabulation areas. TomoDD is used to perform velocity inversion on 3D nodes, which consists of $40 \times 40 \times 40$ nodes (Figure 6.3). All nodes are assigned with a same initial velocity value (Creighton Mine 6000 m/s; Kidd Mine 6025 m/s). The initial velocities are estimated by linear fit on travel time and travel distance of raypaths traveling through the rock mass.

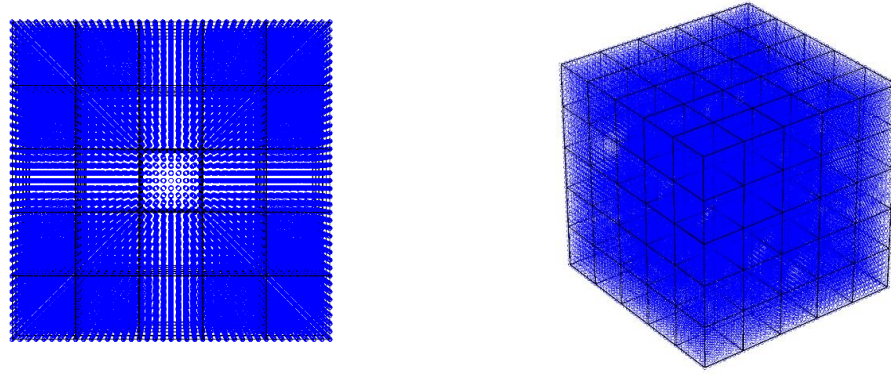


Figure 6.3. Mesh grids of velocity model

The accuracy of tomographic studies for seismic events is determined by the density of raypaths traveling through the geometry of velocity model. Travel time measurement and locations of seismic events are combined using double difference tomographic inversion to compute updated velocity models and relocations of seismic events at each iteration by LSQR algorithm (Zhang and Thurber 2003). The conversion of the velocity models are validated by quantifying the residual of each iteration in the inversion. It is indicated that the residuals of travel time estimation keep decreasing with more iterations and converge after a certain amount of iterations for tomographical studies of both Creighton and Kidd Mine (Figure 6.4). More accurate velocity distributions are generated interacting with ultimate relocations of seismic events.

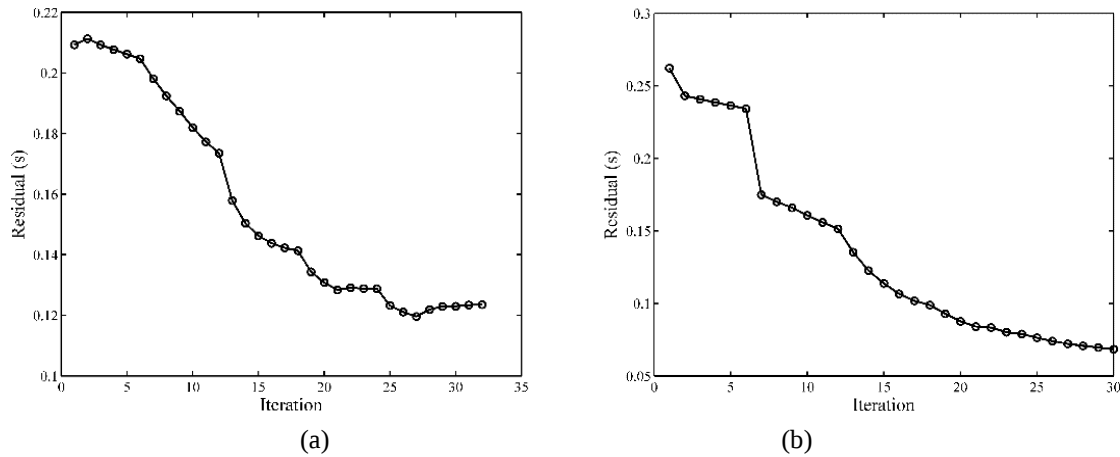


Figure 6.4. Residuals of Creighton Mine (a) and Kidd Mine (b)

Creighton Mine

Velocity inversion is performed at each node. All nodes are divided by 5×5×5 mesh grids, as shown in Figure 6.3. Average velocity of each unit cube is computed by all the velocity of nodes

in the same unit cube. Results of velocity distribution for mesh grids are exhibited in Figure 6.5 and Figure 6.6. Events 1, 2 and 4 are located in cubes on the level that ranges from 7546 to 7940 ft (Figure 6.5). Event 3 is located in cubes on the level ranges from 7152 to 7546 ft (Figure 6.6). The most conspicuous feature is that significant velocity changes around the large magnitude events.

It is observed in Figure 6.5 that the velocities within the objective region surrounding Event 1, Event 2 and Event 4 are higher than the background velocity for all the results from all event groups. High velocity anomalies are identified around Events 1, 2, and 4 before the large magnitude events periods and the velocity increases significantly with the closer period to the occurrence of large magnitude events (Figure 6.5a. and Figure 6.5b.). However, velocity trend changes along with the occurrence of large magnitude events. The velocity of the tabulation area especially including high velocities with events 1, 2 and 4 experienced a pronounced drop of velocity in the postseismic periods (Figure 6.5b. and Figure 6.5c.). Eventually, the area with events 1, 2 and 4 continues to decrease to the level of background velocity. In addition to the fact that the region with events 1, 2, and 4 indicates the most striking velocity change, increasing before the seismic periods and decreasing in the coseismic and postseismic periods, other adjacent areas experienced similar velocity changes and reaches low velocities as well. Similarly, a significant velocity rising is manifested in the vicinity of Event 3 before the period ranges of large moment magnitude (Figure 6.6a. and Figure 6.6b.). The region including Event 3 is identified by reductions in average wave speed comparing the velocity distribution before the large magnitude events with that after the large magnitude events, forming a slow velocity anomaly (Figure 6.6b. and Figure 6.6c.). Two possible causes, including static stresses concentration and shaking damage, could be responsible for the velocity change. Static stress change can explain why the velocity anomalies are located nearby the large magnitude events. The areas with static stress concentration are likely to cause the large magnitude events. The high velocity areas before large magnitude events suggest that a rock mass might be compacted due to the load force. A rock mass with a greater density is easier for waveform propagation. The most likely explanation for the vicinity of large magnitude events appearing close to high velocity anomalies is that stress concentration enhances density of the rock mass. Bulk modulus of a rock mass is capable of increasing with volume decrease under the pressure. Volumetric hardening is likely to be invoked if the isotropic pressure causes irreversible volumetric compaction. Production blast activities are performed before the

occurrence of large magnitude events. It is inferred that production blast activities cause static stress change. As a result of static stress change, the stress field fails to keep the originally balanced state. Static stress change leads to uneven distribution in some regions. Abrupt energy release might be triggered to form the source of large magnitude events.

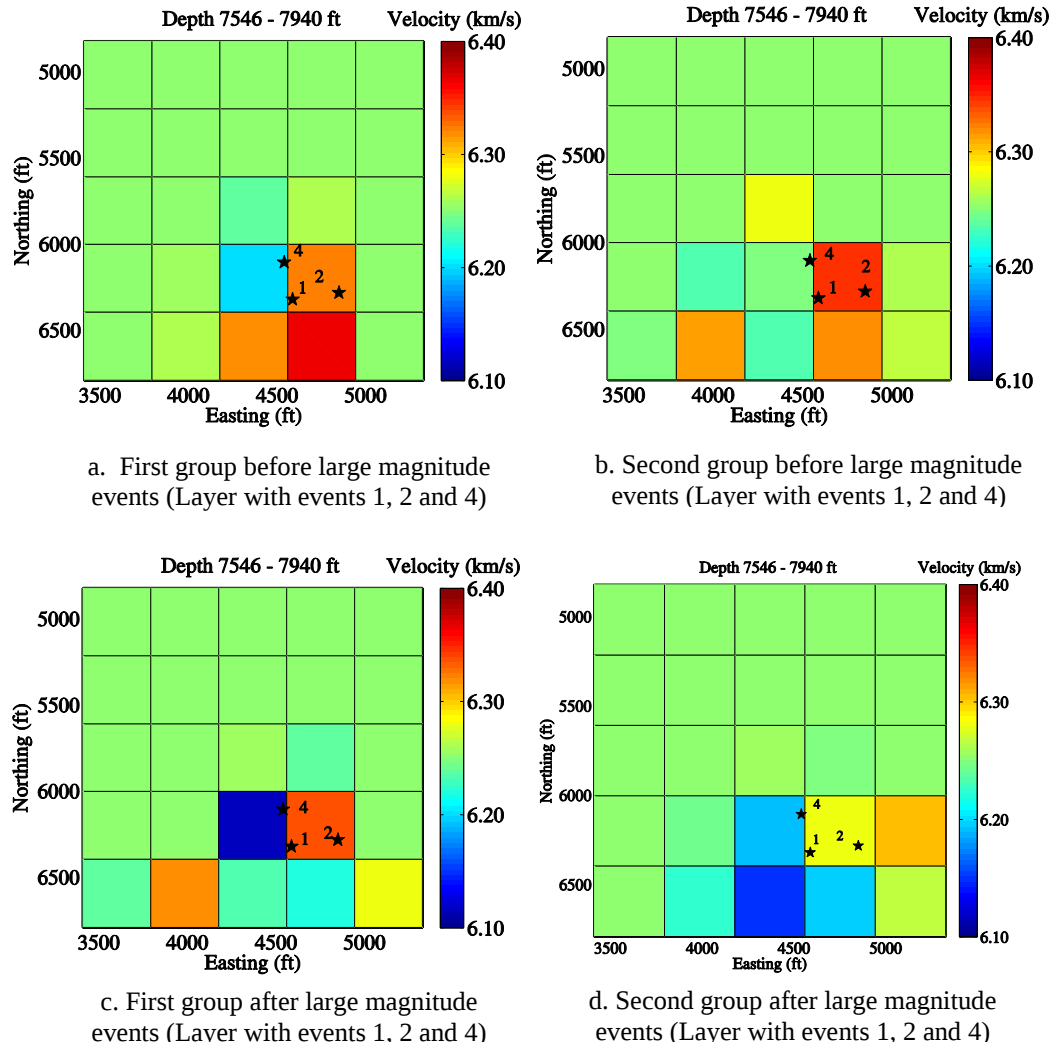


Figure 6.5. Average velocity of cubes by velocity inversion of Creighton Mine

After the large magnitude events, velocities in the vicinity of large magnitude events experienced a significant reduction associated with decreasing in varying extent of the other areas. A reasonable explanation is that the shaking effect by the large magnitude events leads to ruptures and damage in nearby regions of rock mass. Opening of fractures by the shaking-induced damage impedes the propagation of waveform in rock mass. Shaking-induced damage is inversely correlated with the distance between its location and the hypocenter of large magnitude events.

This explains why the velocity drop is more evident on the regions that are closer to the location of large magnitude events. The velocity around Event 3 changed more evidently than the region of the large magnitude Event 1, 2, and 4 during the whole process. The region including Event 3 especially achieves a low velocity level right after the large magnitude events in postseismic period.

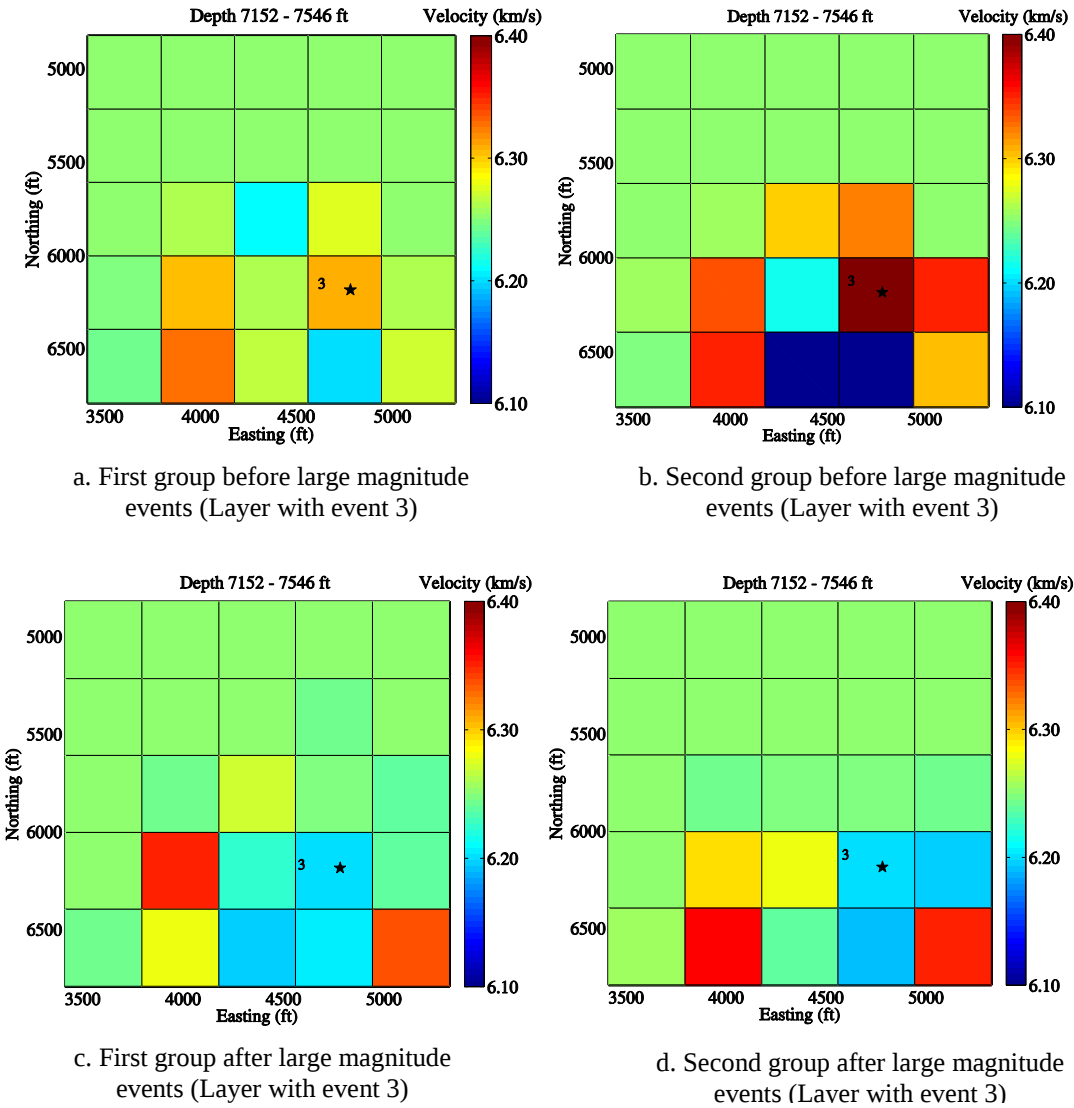


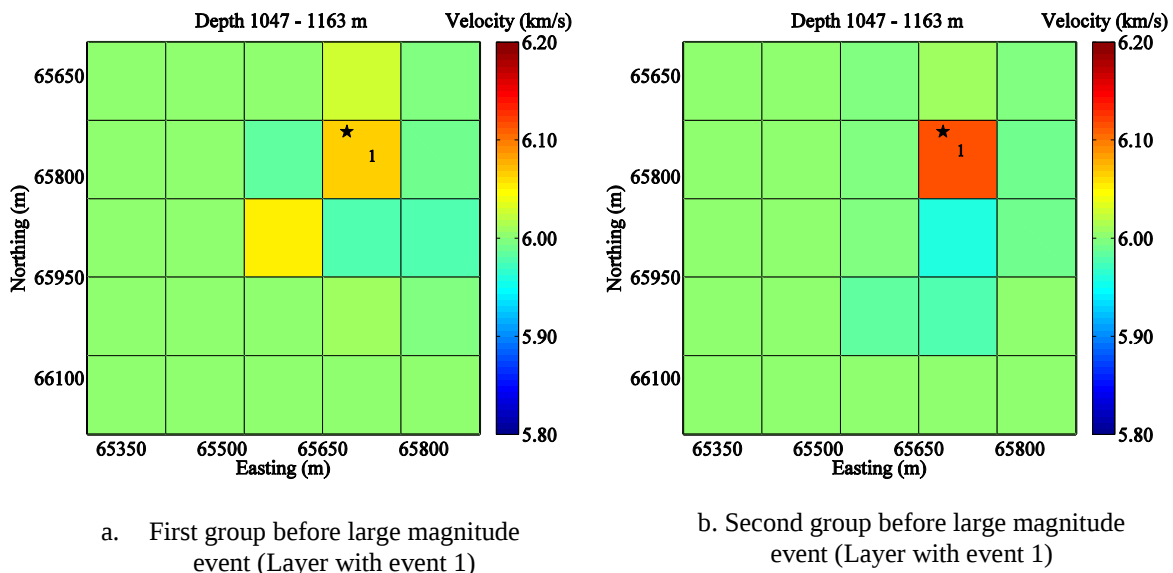
Figure 6.6. Average velocity of cubes by velocity inversion of Creighton Mine

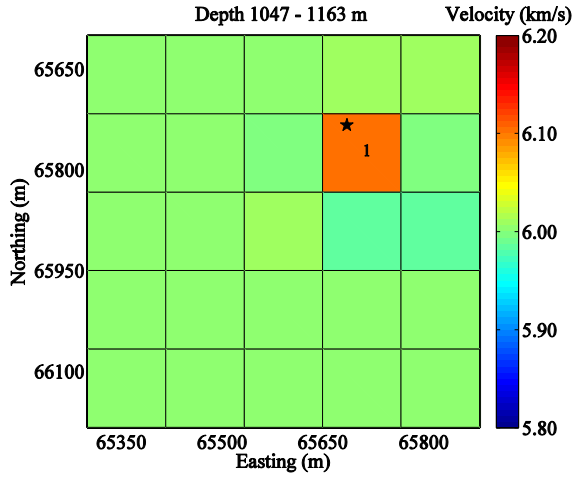
KIDD MINE

According to the large magnitude events of Kidd Mine, events group before and after the first large magnitude event (Event 1) and the second major event (Event 2) are analyzed, respectively.

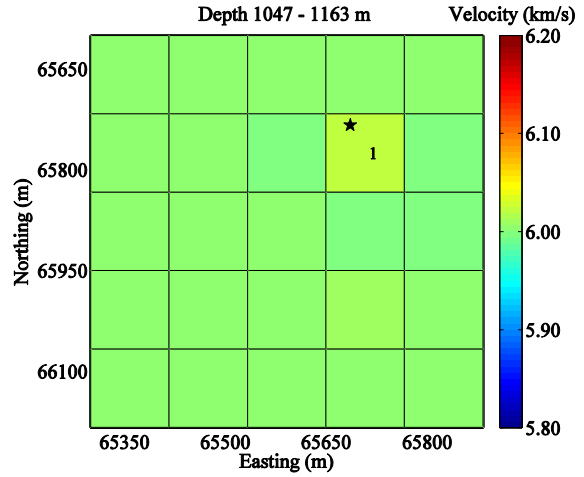
Analyses on microseismic events triggered before and after the seismic period in Kidd Mine are conducted to further illustrate the velocity change associated with the large magnitude events. There is a strong similarity of velocity change pattern between Creighton Mine and Kidd Mine. It demonstrates that velocity around Event 1 grows and forms a high velocity anomaly in the region before its occurrence (Figure 6.7a. and Figure 6.7b.). It is inferred that the stress in the vicinity of major events increases to a higher level and the waveform propagation of seismic events is strengthened before Event 1. The average velocity in the tabulation region of Event 1 also experiences a significant drop after Event 1 (Figure 6.7b. and Figure 6.7c.). The average velocity of the region of Event 2 keeps higher than the background velocity (6.3 km/s) all the time. There is not a pronounced velocity change in the region of Event 2. However, the total high velocity areas expand before the seismic period (Figure 6.8a. and Figure 6.8b.) and then they diminish significantly in the postseismic periods (Figure 6.8c. and Figure 6.8d.). It might be because the static stress is released and the propagation of dynamic shaking-induced fractures impair the waveform propagation after the occurrence of either Event 1 or Event 2.

The most noticeable difference between the velocity change with Event 1 and Event 2 is that the velocity change with Event 1 is strongly concentrated and intense in comparison of the region of Event 2. The velocity changes with Event 2 are smoother and spread over larger areas. The reasonable explanation is that the magnitude of Event 1 is greater than the magnitude of Event 2. The hypothesis that the dynamic shaking-induced effect from Event 1 is stronger than that from Event 2 is supported by the fact that the seismic energy of Event 1 is larger than Event 2.



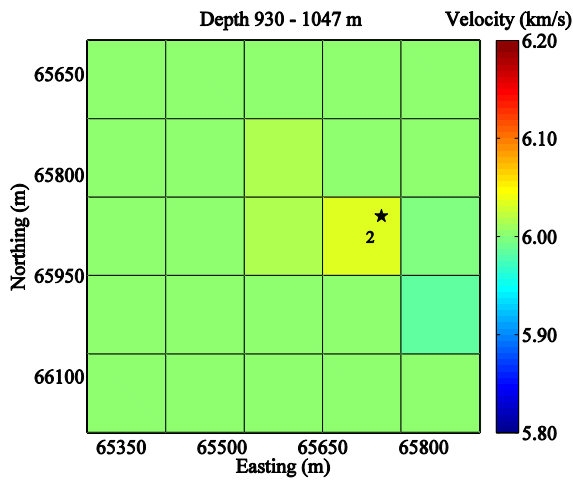


c. First group after large magnitude event
(Layer with event 1)

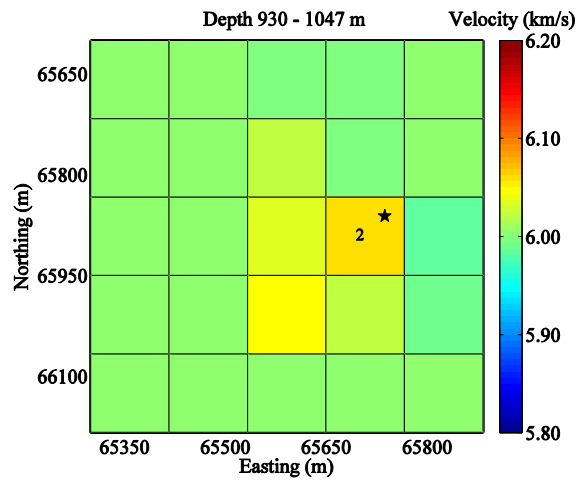


d. Second group after large magnitude event
(Layer with event 1)

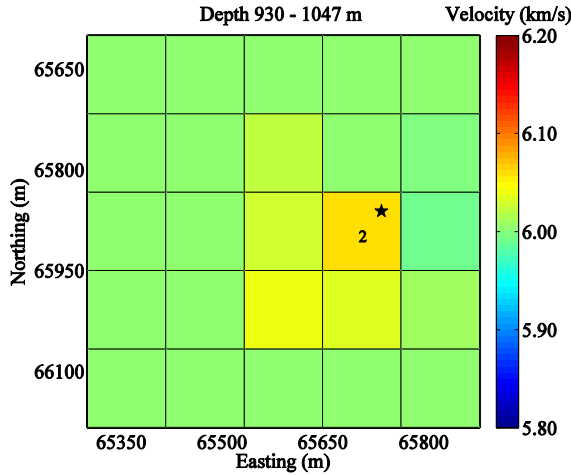
Figure 6.7 Average velocity (Event 1) of 500 microseismic events of Kidd Mine



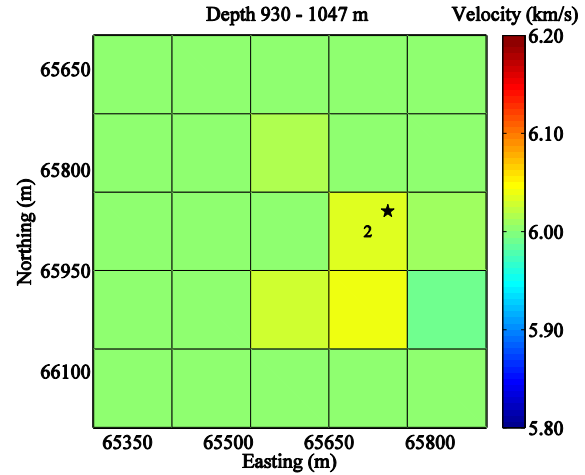
a. First group before large magnitude event
(Layer with event 2)



b. Second group before large magnitude event
(Layer with event 2)



c. First group after large magnitude event
(Layer with event 2)



d. Second group after large magnitude event
(Layer with event 2)

Figure 6.8. Average velocity (Event 2) of 500 microseismic events of Kidd Mine

6.5 Summary and discussion

It is exploited that the velocity of waveform propagation experienced changes within rock masses caused by large magnitude events in underground mining. It is useful to have a simplified model that allows prediction of seismic risks in terms of large magnitude events. Data sets from two hard rock underground mines are recorded and investigated. High seismic rates of each mine provide a good raypath coverage for objective regions. Groupings, comprised of one thousand events each, underwent velocity inversion in a double difference tomographic model to produce the velocity structures. In the comparison of velocity distribution of multiple periods close to large magnitude events, some findings are summarized.

The waveform propagation is enhanced in the large magnitude events prone regions before their occurrence. It is inferred that high velocity anomalies are caused by the accumulation of static stress in the regions including potential large magnitude events of the rock mass.

Velocity reduction after the large magnitude events in the vicinity of them implies that static stress drop and dynamic-shaking induced fractures mutually lead to the weakness of waveform propagation. The rate and magnitude of velocity change seems related to the depth of occurrence and magnitude of events.

Double difference tomographic studies benefit the development of seismic hazard assessment in underground mining. The results presented indicate that velocity anomalies within rock mass in

underground mining are associated with the occurrence of large magnitude events. It is possible to predict the large magnitude events by detecting whether the wave speed changes in a relatively large amplitude compared with the historical data set. To explore the application of passive tomographic studies for seismic hazard assessment, tomographic investigation on microseismic data set during a long term period would be attempted to accurately compute the ranges of velocity anomalies. Crucial thresholds of secure average velocity for a mine can be derived from multiple data sets of mining-induced seismicity. Seismic hazards in terms of large magnitude events can be alerted when a velocity anomaly exceeds the threshold of secular velocity to improve the safety of underground mines.

Chapter 6 References

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7 Statistical analyses of mining-induced seismicity from deep hard-rock mines

7.1 Abstract

Previous studies have implicated changes of b-value associated with the occurrence of mainshocks. However, whether changes of b-value can be used as a precursory for mainshocks remains largely unknown. We compute the temporal changes of b-value based on a reliable estimation of magnitude of completeness using mining-induced seismicity from a deep hard-rock mine. The b-value analysis reveals a significant decrease of b-value with the occurrence of mainshocks. To investigate behavior of aftershocks in mining-induced seismicity sequence, we used a temporal decay model based on Generalized Omori's law to examine temporal decay of aftershock sequences. Results of temporal decay model indicated a close agreement between the modeled temporal decay process and practical cumulative number of events with time. The computed parameters conform to the empirical studies from crustal earthquakes, validating effectiveness for mining-induced seismicity. Taken together, these results have important implications for seismic hazard analyses in underground mines.

7.2 Introduction

Using mining-induced seismicity data to detect potential danger and mitigate hazard is a long-term quest in mining safety research. Seismic monitoring system is a feasible and practical way to monitor and record the seismic activities. Data describing the triggered time, locations and magnitude of mining-induced seismicity contain the spatial and temporal information on the occurrence of seismic events through periods of time. Whereas the average occurrence of rate of seismicity is an important estimate of the potential dangers in underground mining, the average occurrence of rate is insufficient to define the secular rate of seismicity. Seismologists have devoted a significant effort on seismic hazard analysis by applying statistical scaling methods in mainshock and aftershock sequence. Further, frameworks from crustal earthquake have been proved that they can be used in mining-induced seismicity for improving safety (Young and Maxwell 1992, Wuestefeld, Kendall et al. 2011).

Seismic hazard analyses of crustal earthquakes are mainly based on earthquake frequency statistics on historical catalogues of seismicity. It is well known that aftershocks of a mainshock satisfy Gutenberg-Richter frequency-magnitude scaling (Gutenberg and Richter 1956). Numerous

global and regional surveys have been performed to assess the empirical constant in validating the Gutenberg-Richter law (Wesnousky 1999) (Shcherbakov, Turcotte et al. 2005). Isacks and Oliver (1964) claimed that it is reasonable to use the hypothesis of constant b value to predict the earthquakes by extrapolating to higher magnitudes based on frequency magnitude relations. The b -value provides a new insight that can be used as an indicator of failures in a rock mass in the laboratory and to forecast occurrences of earthquakes (Mogi 1963) (Smith 1981) (Lockner 1993). Observations on b -values before large earthquakes in New Zealand, California and Venezuela indicate that the mainshocks were preceded by periods of high b -values (Smith 1981). Evidences support the view that the seismicity and rock deformation are shown to be closely related in space and in time (McGarr and Green 1975). Merged with underground observations, seismic hazard analyses on seismicity data indicate that seismicity in mining (mine tremors) obey the same magnitude-frequency relation as crustal earthquakes (Boettcher, McGarr et al. 2009). Magnitude-frequency data for events over 1 years at Harmony Gold Mine agree Gutenberg-Richter relation very well (McGarr 1971). Accordingly, it is inferred that b -values of mining-induced seismicity can be useful for seismic hazard assessment for underground mines.

Although b -values could be the determinants of seismic hazard analyses in different scales, restraints and uncertainties involved in seismicity forecasting on historical data from different sites can yield imprecise results. The spatial and temporal distribution of aftershocks, and the dependence on the magnitude of the mainshock were assessed to provide reference for potential danger and mitigating hazards (Krinitzsky 1993). Quantifying mainshocks and aftershocks can improve the knowledge of correlations of seismic activities (Shaw 1993). The time dependence of earthquake aftershocks is described in Generalized Omori's law, which empirically gives the temporal decay in the rate of aftershock occurrence (Shcherbakov, Turcotte et al. 2005). Shcherbakov (2005) applied a scaling method on multiple aftershock sequences to scale the parameters of Generalized Omori's law. Based on two physical ingredients: rupture activation and stress transfer, it is measured that seismic decay rates linearly increase with the magnitude of the mainshock (Ouillon and Sornette 2005). Further, numerous studies are devoted to incorporating the understanding of Generalized Omori's law into seismic hazard analysis on mining-induced seismicity. Vallejos interpreted a link between the productivity of seismicity and decay time of seismicity to ensure the safe event rate for re-entry protocol in underground mines (Vallejos and McKinnon 2011). Based on the Generalized Omori's law, the aftershocks sequences from different

mine sites were assessed by a statistical analysis to establish the optimal re-entry protocol (Vallejos and McKinnon 2010).

The purpose of this article is to present the change of b-value associated with mining-induced seismicity sequences, to describe the relationships between b-values and magnitude of aftershocks, and to establish time decay model of mining-induced seismicity from two mines. It has been investigated and found that mining-induced seismicity from two mine sites agree the law of Gutenberg-Richter frequency-magnitude very well. Then, average b-values are computed using the seismicity that occurred in the same time length. The maximum likelihood method and least square method are used to perform the regression on data of frequency-magnitude, respectively. According to empirical studies on mining-induced seismicity, the decay rate of aftershocks is a function of time as well. Statistical analyses are involved to model the predicted seismicity rate since Generalized Omori's law is validated on the basis of empirical investigation. Specifically, secular rates of seismicity defined by the model can be used to assess long-term mining-induced seismic hazards.

7.3 Data of mining-induced seismicity and analysis procedure

It is well known that events are triggered associated with the nucleation of microcracks in rocks. Cracks are generated when the local stress exceeds the local strength (Kranz 1983). Mining works can affect the stress regime. As a result of mine excavations, the rock mass in the proximities of excavations loses its balance state of stress and stress concentration is generated in it. Rock failures in the proximities of maximum stress concentration are usually found in the edges of mining excavations. Associated with the occurrence of seismicity, fracture planes are formed parallel to the stope faces (Cook 1976). Seismic events in mines mainly include fault-slip events and strain-burst events. Whereas a fault stays in balance without change of load, it can be disturbed by a mine's excavation. Nucleation can be initiated when the stress is large enough. The length of cracks and stress increase slowly during the first regime and then increase faster to achieve the critical stress. A seismic event is triggered preceding the rupture development of the fault. Specifically, the stress drop arises on the fracture plane associated with the seismic event, which redistributes stress to vicinities (Shaw 1993).

Several mainshocks and aftershocks sequences recorded in the seismic networks of Creighton Mine and Kidd Mine are compiled for this study. The data in Table 7.1 and Table 7.2 serve to

illustrate the time, location, and magnitude of mainshocks from Creighton Mine and Kidd Mine, respectively. The moment magnitude of aftershocks is approximately from -2.5 to 0. The data presented in this article was obtained from microseismic monitoring system and strong ground motion system. Microseismic monitoring system is specifically for detecting the microseismic activities, most of which are with moment magnitude less than zero. Strong ground motion system is specially emphasized on recording mainshocks, there were approximately 40000 microseismic events from April to September, 2011, in the Creighton Mine data sets, which includes source parameters such as moment magnitude, focal mechanism, and ratio of E_s/E_p . Similar to the Creighton Mine data set, reliable information of seismicity is provided in the Kidd Mine data set. The Kidd Mine data set cover seismicity from September, 2007 to March, 2012. The first mainshock was a fault slip burst, but the second mainshock was a strain burst in Creighton Mine. Both of the mainshocks in Kidd Mine were fault slip bursts.

Table 7.1. Times, Locations, and Magnitude of Mainshocks in Creighton Mine

Mainshocks	Date	Time	North (m)	East (m)	Depth (m)	Magnitude
1	July, 6 th 2011	8:41 AM	1927	1399	2332	3.1
2	July, 10 th 2011	2:44 AM	1853	1385	2392	1.4

Table 7.2. Times, Locations, and Magnitude of Mainshocks in Kidd Mine

Mainshocks	Date	Time	North (m)	East (m)	Depth (m)	Magnitude
1	January, 6 th 2009	4:40 AM	65733	65686	1150	3.8
2	June, 15 th 2009	7:01 PM	65861	65737	1035	3.1

An important empirical observation in seismology is the proportional relationship between the magnitude M and cumulative number of seismic events with magnitude larger than M . In order to check the existence of similar patterns in mining-induced seismicity sequences, mainshocks are defined according to their magnitude. Then, aftershocks temporally following the mainshocks are determined. The validity of the Gutenberg-Richter relationship of mining-induced seismicity in both Creighton Mine and Kidd Mine is shown in Figure 7.1. The cumulative number of aftershocks is given as functions of magnitude. Due to the fact that the average rate of seismicity in Creighton Mine is approximately 60 times higher than that in Kidd Mine, different time scales are used for these two mines. For Creighton Mine, seismicity within a time period of $T = 3.5$ days after the mainshock are considered for the b -value measurements. For Kidd Mine, seismicity within a time period of $T = 90$ days after the mainshock are performed. Note that b -value measurement by using the least-square fit regressions are set to data with magnitude between -1.5 and 0. The cutoff magnitude level of -1.5 is determined because only the data sets with magnitude above -1.5 are

strongly correlated and directly related to the aftershocks. The process of selecting magnitude of completeness is discussed below.

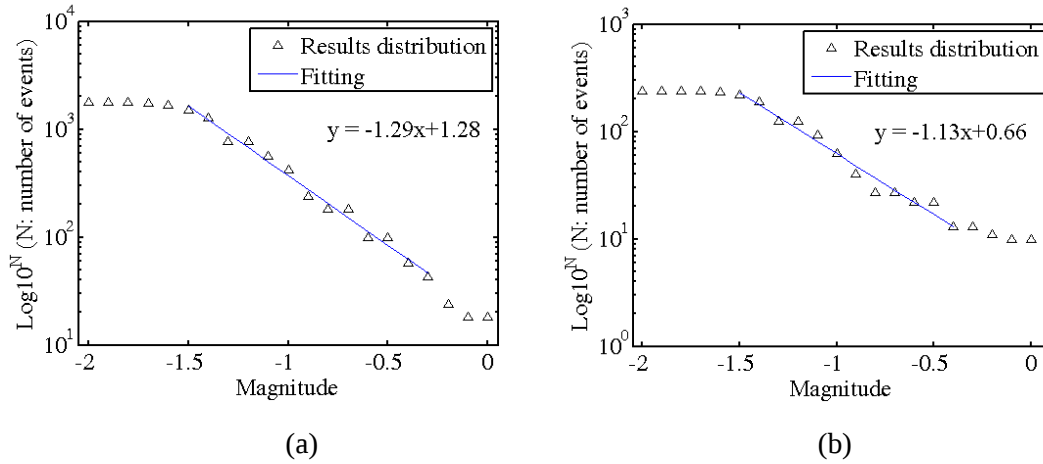


Figure 7.1. Cumulative numbers of aftershocks are given as functions of magnitude for (a) the first mainshock in Creighton Mine and (b) the first mainshock in Kidd Mine. The solid straight line are the best fit of Gutenberg-Richter relation.

The cutoff magnitude of aftershocks in these two mines is assessed by analyzing the magnitude of completeness, which is the lowest magnitude at which 100% of the events are detected in temporal and spatial scales (Woessner and Wiemer 2005). The estimate method of the magnitude of completeness presented by Woessner (2005) is based on the self-similarity of seismicity process, which implies a power-law distribution of seismicity. According to the implication of power-law distribution of seismicity, the estimation method from Woessner (2005) considers the estimate of magnitude of completeness and its influence on the b-value. The Monte Carlo approximation of the bootstrap method is used in this method for calculating b-values and the magnitude of completeness. In order to check the reliability of data regressions, maximum likelihood estimation method is used to compare the result of it with the result obtained from the least-square estimation method. Differences are found between the results of the least-square estimation and maximum likelihood estimation method. As discussed in the results section below, it is observed that the maximum likelihood method amplifies the variation of b-value, but the results of both methods conform to the similar trend of variation. Figure 7.2 and Figure 7.3 illustrates the results of the magnitude of completeness and b-value by both the least-square method and maximum likelihood method for aftershocks of Creighton Mine and Kidd Mine. It is verified that -1.5 is the optimum magnitude of completeness for both Creighton Mine and Kidd Mine (Figure 7.2 and Figure 7.3). First, it is well shown that better linearity is embodied at $m > -$

1.5 between cumulative number of aftershocks and magnitude of each aftershock sequence. Second, b-values given by these two methods at $m = -1.5$ are conspicuously in closer agreement than at most other magnitudes, whereas some results yielded by maximum likelihood method fluctuate on some range of magnitudes. It thus confirmed the claim that $m = -1.5$ is the optimum magnitude of completeness.

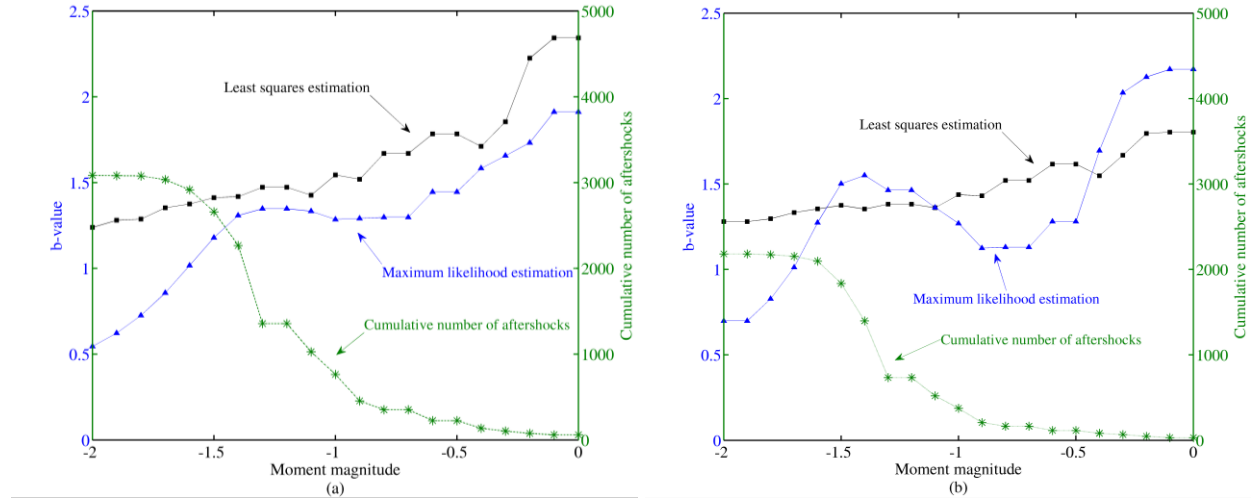


Figure 7.2. Change of b-values and number of aftershocks with the magnitude of completeness for (a) first mainshock of Creighton Mine (b) second mainshock in Creighton Mine. The cumulative magnitude distribution curve is approximately linear for $M > -1.5$ for both the aftershocks sequences. The b-values calculated by the least-square method and maximum likelihood method at magnitude = -1.5 are significantly closer than that with most other magnitudes.

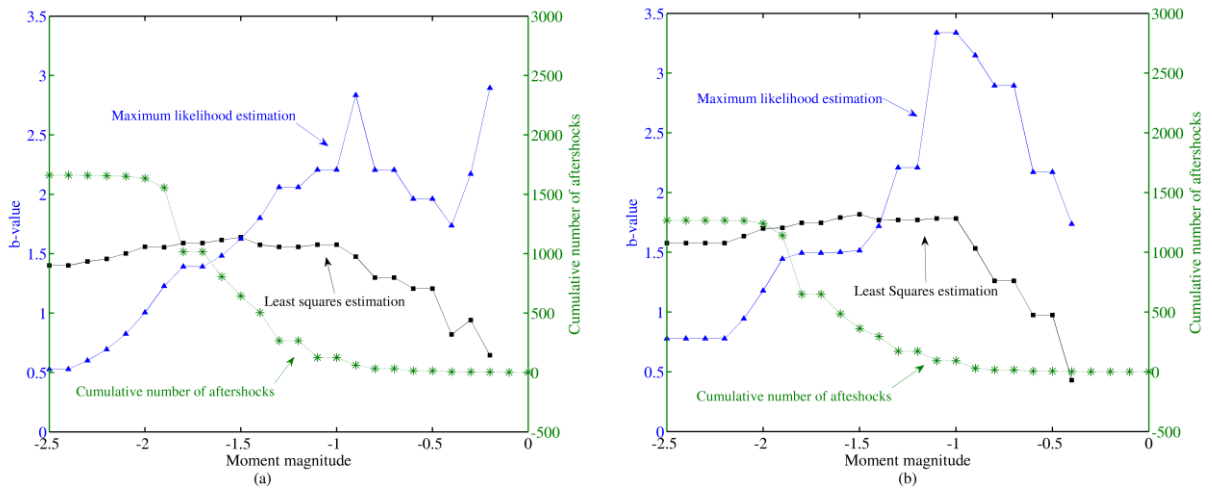


Figure 7.3. Change of b-values and number of aftershocks with the magnitude of completeness for (a) first mainshock in Kidd Mine (b) second mainshock in Kidd Mine. The cumulative magnitude distribution curve is approximately linear for $M > -1.5$ for both aftershocks sequences. The b-values calculated by the least-square method and maximum likelihood method at magnitude = -1.5 are significantly closer than that with most other magnitudes.

The temporal decay of aftershock activity is described by the modified Omori's law (Utsu and Seki 1954). The applicability of the modified Omori's law in mining-induced seismicity has been proven (Vallejos and McKinnon 2010) (Vallejos and McKinnon 2011).

$$\gamma \equiv \frac{dN}{dt} = \frac{K}{(c+t)^p} \quad (7-1)$$

where t is the time elapsed since the mainshock and K , p , c are empirical parameters. Omori's law manifests the temporal correlations in aftershock sequences, which are relax processes after mainshocks (Shcherbakov, Turcotte et al. 2005). The occurrence of a mainshock redistributes the stress. During a main shock, the stress and strain are enhanced in some regions neighboring the fault where the mainshock is located. The stress relaxation is associated with the occurrence of aftershocks. Aftershocks assist to relieve the stress concentration arising from mainshocks. All aftershocks contribute to the reduction in regional stress as a function of the magnitude of aftershocks (Shcherbakov, Turcotte et al. 2005).

Since the value of parameters in Omori's law analysis corresponds to different seismicity sequences, the value of parameters K , p , and c are empirically determined. The exponent p is the most important parameter among them. The dependence of p was investigated and found that the exponent p increases as a function of the magnitude of mainshock. The physical mechanism of p is explained in that aftershocks of the mainshock with a large magnitude decay at a faster rate than aftershocks of the mainshock with a smaller magnitude. Previous studies demonstrated that the average p value ranges from 0.9 to 1.2 for mainshocks with magnitudes going from 5 to 7.5 (Ouillon and Sornette 2005). By extending the Omori's law in mining-induced seismicity, it is measured that p is 0.4 for the aftershocks sequence that occurred in the Creighton Mine between October 1997 and March 1998 (Marsan, Bean et al. 1999).

7.4 Results

The b-value variation in Creighton Mine

It has been discussed that the temporal and spatial changes of b-value have potentially important implications for understanding seismicity patterns and forecasting. After selection of the magnitude of completeness, a series of b-values of seismicity during the period March to December 2011 are yielded. Figure 7.4 plots the variations of b-value with the occurrence of mainshocks, which includes both the first mainshock and second mainshock. Note that the b-values fluctuate significantly during the starting four weeks because of the large uncertainty of that period.

This phenomenon has also been found in the data set of arrival times of seismicity, noting that the data of the four weeks are flawed. Both the results from one week and two weeks scales exhibit the lowest b-value comparing with b-values of other periods.

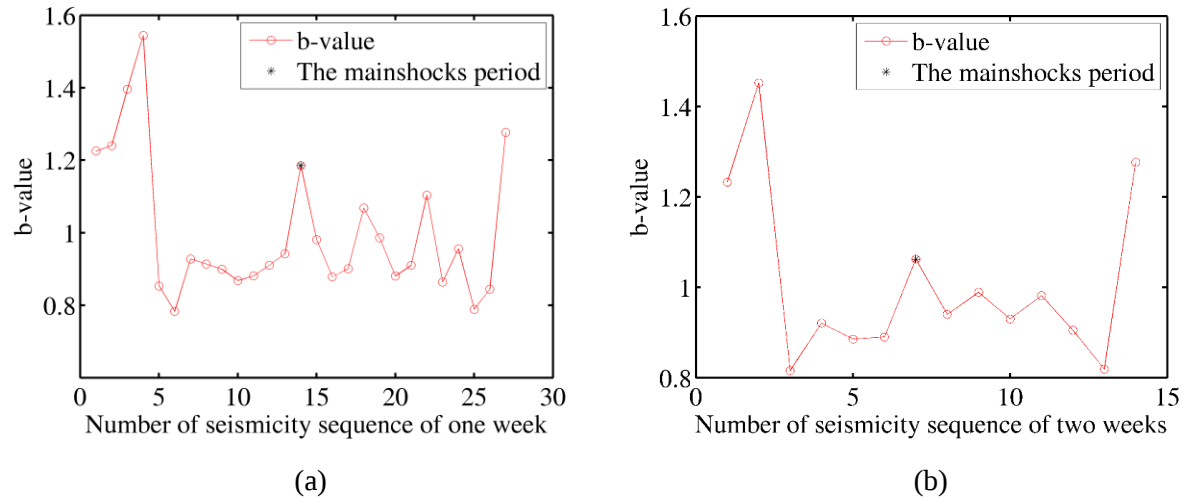


Figure 7.4. Temporal change of the b-values for (a) one week scale (b) two weeks scale. Continuous line with open circles: b-values through all time periods; Asterisk: the specific time period with occurrence of mainshocks. The period noted by an asterisk includes the time of the mainshocks in July, 2011.

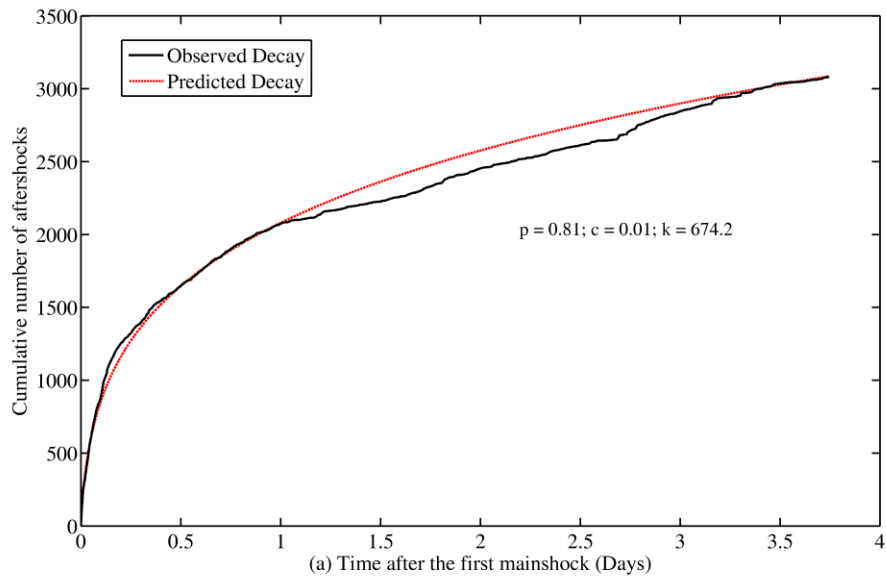
According to the evidence of the temporal variation of b-values in crustal earthquakes, there are mainly two patterns of b-value change associated with mainshocks (Smith 1981). First, the b-value increases to a peak and then decreases preceding the mainshock. The mainshock occurs during the decrease. Second, a high b-value appears prior to the mainshock. The b-value is thus influenced by either the previous mainshock or the mainshock after the peak of b-value. The investigation of b-value change in Creighton Mine demonstrates that b-value changing with the occurrence of mainshock agrees with the b value change in the crustal earthquakes study discussed above.

The b-value analyses of temporal change associated with the occurrences of mainshocks are not performed, because the low seismic rate cannot yield reliable b-values if seismicity is divided into small durations of the time periods.

Modeling fit for aftershocks in Creighton Mine using Omori's law

Figure 7.5A and Figure 7.5B show the modeling fit for the aftershocks for the first and second mainshocks, respectively. The statistical model used in this analysis is developed for aftershocks in natural earthquake sequences by Woessner (2005). Bootstrapping was originally developed for

the statistical accuracy estimation. Bootstrap samples of seismicity are generated by randomly picking the same amount of seismic events with the replacement from the other data sets (Woessner and Wiemer 2005). Accurate estimation of parameters p , c , and K are ensured since there are thousands of aftershocks for each magnitude interval on the premise of selecting reasonable magnitude of completeness. The modeling fit is performed within 4 days after the mainshocks. The modeled cumulative number of aftershock decaying with time exhibits close agreement with the distribution of the real cumulative number of aftershocks, especially during the beginning of the decay process (within one day). The discrepancy between the fitted model and real cumulative numbers primarily appears during the middle temporal section in the decay process. The fitted model and real cumulative numbers eventually integrate at the neighboring of the end in the decay process.



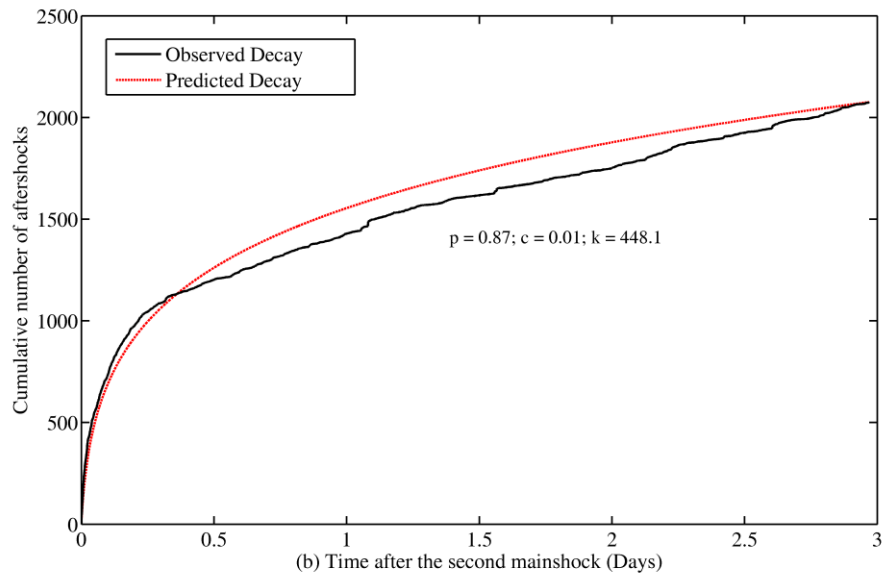
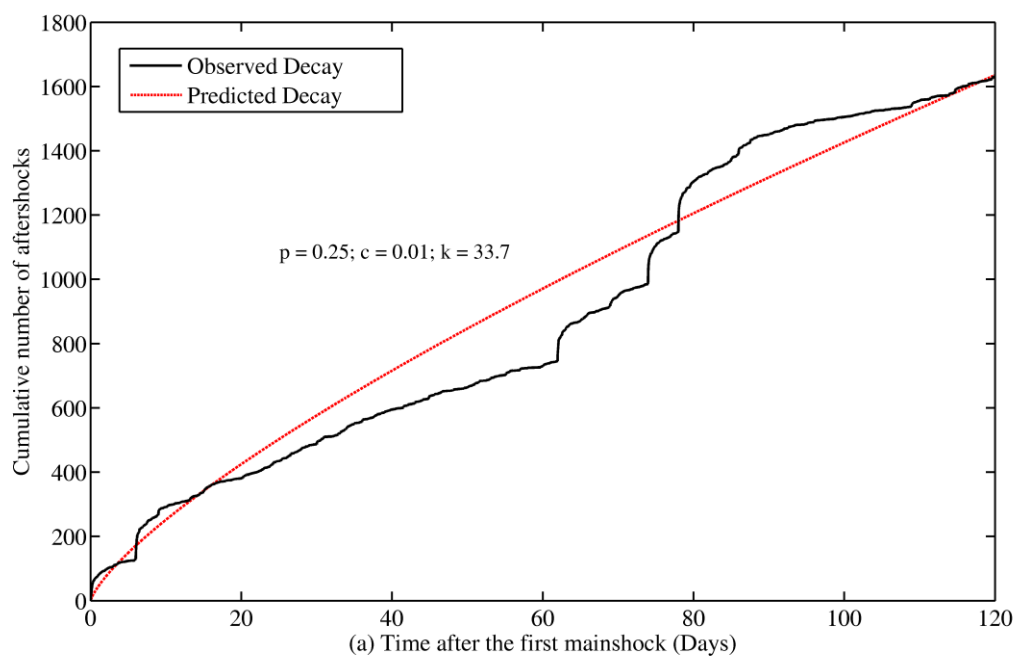


Figure 7.5. Creighton Mine cumulative number of aftershocks decay with time modeling for (a) the first mainshock (b) the second mainshock. Solid lines: practical cumulative number of aftershocks; Dash lines: fitted model using Generalized Omori's law.

The fitted parameters of modeling for Generalized Omori's law are computed. The p value is 0.81 and 0.87 for the first mainshock and the second mainshock, respectively. The fitted p values indeed conform to the reasonable range from the natural earthquake study, which gives $0.14 < p < 1.20$ for a considerable number of earthquakes between 1932 and 2003 (Ouillon and Sornette 2005). By comparing the yielded p value between the first mainshock and the second mainshock, it is found that p (0.87) of the second mainshock is significantly close with that (0.81) of the first mainshock. However, the p value usually increases with larger magnitudes in the study of crustal earthquakes. It is interpreted as the magnitude dependence of p values because aftershocks of large mainshocks decay at a faster rate than aftershocks of small mainshocks (Ouillon and Sornette 2005). However, a larger mainshock can trigger more events and then the decay of aftershocks needs more time. Thus, a larger mainshock does not imply a large p value due to uncertainties of the decaying process, whereas evidences are found in empirical studies of crustal earthquakes. As a consequence, magnitude dependence of the p value fails to apply to mining-induced seismicity. The close parameters for different aftershock sequences suggests that the fitted model is accurate for application in the mainshocks of the same mine site.

Modeling fit for aftershocks in Kidd Mine using Omori's law

Following the framework of modeling fit for Creighton Mine, similar analysis is performed for the aftershock sequences of two mainshocks in Kidd Mine. Comparisons of the fitted model and real cumulative number of aftershocks can be seen in Figure 7.6 to see how the fitted model agrees with the practical decay process. Close agreements are found near the beginning and end of the decay process. However, some evident differences are found between the fitted model and real results, which can be interpreted by the random distribution over the long time scale of the data set. The p values for these two aftershock sequences are 0.25 and 0.2, respectively. Close p values of different mainshocks suggest that magnitude dependence of p value is not compatible for mining-induced seismicity. Again, close p values prove that the fitted model is stable to be used in the same mine site.



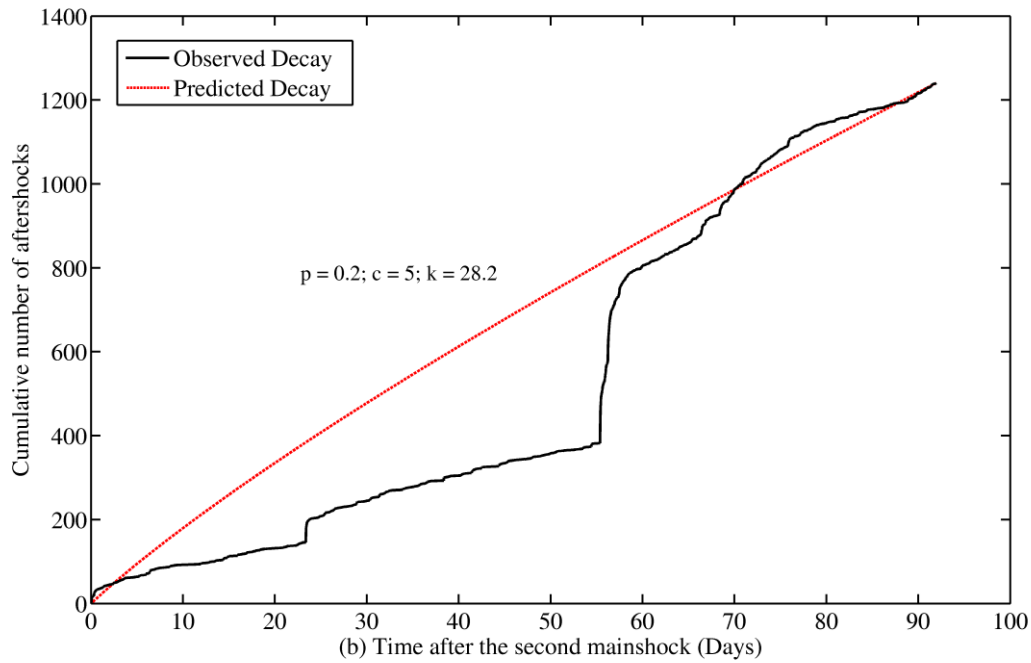


Figure 7.6. Kidd Mine cumulative number of aftershocks decay with time modeling for (a) the first mainshock (b) the second mainshock. Solid lines: practical cumulative number of aftershocks; Dash lines: fitted model using Generalized Omori's law.

7.5 Summary and discussion

Through b-values analyses and fitted model for aftershocks decay, we identified significant changes of b-value associated with mainshocks and the pattern of aftershocks of decay. These findings confirm the implication that b-value can be used to forecast potential mainshocks. In addition, the fitted model for aftershocks decay provides a reference of decay process from historical data. Indicators of potential hazards can be reported if the aftershocks fail to follow the decay model in seismic hazard analysis. The comparison between observed and predicted decay process may be applied to examine whether the aftershocks from rock bursts keeps a normal decay, ensuring safety for restarting work in underground mines.

A reasonable examination of completeness of magnitude ensures accurate regression between cumulative number of seismic events and the distribution of magnitudes. From our case studies, -1.5 is a reliable completeness of these two mines. It is recommended that -1.5 be used as future analyses of aftershocks in mining-induced seismicity.

Previous studies of b-value and fitted model of Generalized Omori's law furnish a comprehensive framework for seismic hazard analyses in mining-induced seismicity. However, it

should be noted that some properties of natural earthquakes and mining-induced seismicity are not completely identical. For example, magnitude dependence of p in Generalized Omori's law is not found in aftershocks decay of mining-induced seismicity.

Further improvement on ascertaining relationships between b -value change and mainshocks is still needed, because more confirmation from case studies is needed for using b -value as the precursory of mainshock. Also, the parameters estimation of Generalized Omori's law suggests that aftershocks from the same mine site share considerable similarities, which could be used to search for potential hazards in aftershocks decay.

Chapter 7 References

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8 Conclusions and Discussion

The study presented in this paper sought to develop adequate seismic hazard analysis for deep hard-rock mines based on mining-induced seismicity. In order to complete this objective, double difference tomographic studies, velocity fitting through raypaths, and statistical analyses on aftershock modeling are designed to provide a comprehensive means for evaluating and forecasting the potential seismic hazard in deep hard-rock mines. The fundamental concept used in this paper is to investigate the changes associated with the occurrence of mainshock, and then these changes discovered from historical data can be used as factors or indexes to assess stress distribution and forecast potential seismic hazards in future mining processes. Geophysics techniques used in crustal earthquakes provide underlying frameworks for mining-induced seismicity. Velocity structure of rock masses, b-value change with mainshocks and temporal decay of aftershocks give rise to comprehensive understanding of forecasting mainshocks in deep hard-rock mines.

We have applied the double-difference tomography in data sets of mining-induced seismicity within a certain space-time neighborhood of mainshocks (major events). It is observed that both high velocity and low velocity anomalies are exhibited for all the periods. Prior to the occurrence of mainshocks, high velocity body shifts toward the location of mainshocks. In addition, the discrepancy between high velocity anomalies and low velocity anomalies concentrates in a larger extent. Considerable consistence of the overall velocity distributions still exhibits before the occurrence of mainshocks. Associated with the mainshocks, dramatic change of velocity distribution is observed. A high velocity body dominates the vicinity of mainshocks, whereas low velocity areas shift away with the mainshocks. The high velocity body experiences a trend of moving beyond the center of drifts, and low velocity areas continue to approach the mainshocks around drifts. It is explained that mainshocks are a possible driving force to alter the stress distribution. Highly-stress approaches the potential mainshocks and high velocity anomalies are found before the occurrence of mainshocks. Also, stress relief is found north of drift. Within two weeks after the mainshocks, stress developed less concentrated around the drifts and low stress areas were observed close to the drifts. Appearance of the high-velocity body and comparing validity before and after the mainshocks proves the potential to use high velocity anomalies as a precursor of mainshocks.

Besides the tomography studies on the velocity structures, the average velocities of Creighton Mine and Kidd Mine are computed by robust linear fitting on distance and travel time pairs of all the raypaths. Results show that coseismic and postseismic velocity decreases caused by the mainshocks are observed in the rock mass adjacent to the mining region. The velocity change suggests that velocity decreases due to the fractures, caused by regional static stress and dynamic shaking from mainshocks and microseismicity, weakening the wave propagation in the rock mass. The velocity eventually returns to the background values assigned in velocity models after recovery due to the closure of the cracks (crack healing effects). Velocity reductions associated with the increasing number of seismic events reflect that considerable fractures weakened wave propagation. One possible explanation of velocity increases is an increase in bulk modulus due to crack healing effects. Two factors lead to the increase in bulk modulus. First, dilatancy-induced preexisting cracks are closed by enhanced stress, giving rise to an increase in elastic moduli. Second, the bulk modulus increases when dilatancy-induced crack closure offsets the elastic volume increase.

In addition to focusing the investigations on seismic analysis before and after mainshocks, the whole data set of recorded microseismicity for both Creighton and Kidd Mine are sequentially divided into different groups, each of which includes 1000 seismic events. Seismicity rates are computed to analyze how the seismicity rates change with the velocity distribution in corresponding tomograms generated from each group. It is summarized that microseismic events are more likely to occur close to the areas between low velocity regions and high velocity regions. Microseismic events spread to a larger extent when the seismicity rate is low. Conversely, microseismic events tend to be spatially concentrated during the time interval of high seismicity rate. Further, it is indicated that frequency and moment magnitude of seismic events in both Creighton Mine and Kidd Mine follow the Gutenberg-Richter law, which is a fundamental observation from empirical studies of crustal earthquakes.

By using a Matlab function “Pcolor”, the average velocities were calculated and plotted from the velocity of nodes distributed in the same unit cube and give a quantitative comparison of the velocity before and after mainshocks. The velocity structures of Creighton Mine and Kidd Mine are computed and inversed by TomoDD. It is found that high velocity anomalies are caused by the accumulation of static stress in the regions including potentially large magnitude events of the rock mass. Velocity reduction after the mainshocks in the vicinity of mainshocks implies that static

stress drop and dynamic-shaking induced fractures mutually lead to the weakness of waveform propagation. The results presented indicate that velocity anomalies within the rock mass in underground mining are associated with the occurrence of large magnitude events. It is possible to forecast the mainshocks by detecting whether the wave speed changes in a relatively large amplitude compared to the historical data set.

An important index b-value is computed to illustrate its temporal change associated with mainshocks on the premise that mining-induced seismicity conforms to the Gutenberg-Richter law. The b-value change patterns of crustal earthquakes are introduced and referenced for feasible discussion of b-value of mining-induced seismicity. The b-value change temporally around the mainshock indicates that the dramatic increase of b-value can be used as a precursory signature for assistance on forecasting the occurrence of mainshocks. In addition, another essential issue of how to determine the completeness of magnitude is discussed based on least-square method and maximum likelihood method. It is explained that -1.5 is a reliable completeness of magnitude for mining-induced seismicity. Further, aftershock decay temporal processes are modeled based on the statistical model of crustal earthquakes. Fitted models exhibit close agreements with real cumulative numbers temporal decay. Detailed parameters of Generalized Omori's law are generated for extended use on other aftershock sequences for the same mine site. It is also mentioned that the p value dependence of magnitude is not evident in aftershock sequences of mining-induced seismicity. However, the p value dependence of magnitude has been discovered in crustal earthquakes. Although developed specifically for the seismic risk analysis of individual sites, the methods applied to these mines can be extended to other deep underground mines. More case studies of mining-induced seismicity can be developed to ensure this property, which can be applied to improve the accurateness of mainshock forecasting.