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Monitoring *in situ* stress changes in a mining environment with coda wave interferometry

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SUMMARY

Coda waves are highly sensitive to changes in the subsurface; we use this sensitivity to monitor small stress changes in an underground mine. We apply coda wave interferometry to seismic data excited by a hammer source, collected at an experimental hard rock mine in Idaho Springs, CO. We carried out a controlled stress-change experiment in a mine pillar and we show how coda wave interferometry can be used to monitor the *in situ* stress change with modest hardware requirements.

Key words: coda waves, mining, scattering, time-lapse monitoring.

1 INTRODUCTION

The coda is the tail (Latin *cauda*) of a waveform, it consists of that part of the signal after the directly arriving phases Aki (1969). Its late part is dominated by multiply scattered waves (Aki & Chouet 1975), but in an area with free surfaces and strong reflectors it may also be due to reverberations. Coda waves are used in a multitude of geophysical applications, such as earthquake-magnitude estimation (Lee *et al.* 1972), earthquake prediction (Aki 1985; Sato 1986), volcano monitoring (Fehler *et al.* 1998; Aki & Ferrazzini 2000; Grêt *et al.* 2005; Wegler *et al.* 2006), monitoring of temporal changes in the subsurface (Chouet 1979; Poupinet *et al.* 1984; Robinson 1987), and earthquake location (Snieder & Vrijlandt 2005). Laboratory applications include diffusive wave spectroscopy (Cowan *et al.* 2002), medical imaging (Li *et al.* 1997) and cavity ring-down spectroscopy (O'Keefe & Deacon 1988).

Yamamura *et al.* (2003) measured tidally-induced changes in seismic velocity and attenuation in a mine using transmitted waves. In many applications, changes in a medium can be so weak that they have no detectable imprint on direct-arriving waves that sample the area only once. They are, however, amplified by the multiple scattering and may be readily seen in the coda. *Coda wave interferometry* (Roberts *et al.* 1992; Ratdomopurbo & Poupinet 1995; Nishimura *et al.* 2000; Weaver & Lobkis 2000; Snieder *et al.* 2002; Yamawaki *et al.* 2004), uses the sensitivity of coda waves to changes in the medium to extract these changes in the medium. The general theory of coda wave interferometry is described by Snieder (2006). We have previously used coda wave interferometry to study the non-linear temperature dependence of ultrasonic velocity in granite (Snieder *et al.* 2002), the dependence of ultrasonic velocity on stress change and water saturation in Berea sandstone (Grêt *et al.* 2006), and monitoring of a rapid temporal change in volcanoes (Grêt *et al.* 2005; Wegler *et al.* 2006).

In this work we demonstrate the usefulness of coda wave interferometry for monitoring stress changes in an underground mining environment. It is known that the sensitivity of the seismic wave velocity to stress changes in rocks is low (Nur 1971) and detection of temporal variations has been difficult (Niu *et al.* 2003). Attempts have been made to monitor stress changes in underground mines with traveltime tomography, where tomographic images based on one-way traveltimes of seismic waves for different stress states are compared (Friedel *et al.* 1995, 1996). Since a major stress change in rocks only leads to a small change in velocity, those methods are prone to large uncertainties. Instead of sampling the regional stress once, as in the tomography approach, we make use of the fact that repetitive sampling of the same area by waves that makes these waves increasingly sensitive to stress changes. Waves can sample a region of space repeatedly by bouncing back and forth between boundaries, or by being multiply scattered. In both cases we refer to these waves as *coda waves*, despite a different physical cause for the extended tails of the waveforms. Since coda waves are sensitive to small changes in the medium, and are a suitable tool to monitor stress changes in rocks.

2 INTRODUCTION TO CODA WAVE INTERFEROMETRY

A detailed description of coda wave interferometry is given by Snieder (2006). Here we give the main results that are needed to understand the technique. For a change in the medium that consists of a constant change δv in seismic velocity and unchanged locations of the scatterers and reflectors, the path length l that the seismic waves cover is given by $l = vt$, where v is the seismic velocity and t the propagation time. For a constant velocity change δv and an unchanged path, $l = vt = (v + \delta v)(t + \delta t)$, hence to first order in the perturbations δv and δt

$$\frac{\delta v}{v} = \frac{-\delta t}{t}, \tag{1}$$

where δt is the traveltime difference (in a given time window) caused by the velocity change δv. We extract δt from the data by means of the cross-correlation function, that is defined as

$$R^{(t,t_w)}(t_s) \equiv \frac{\int_{t-t_w}^{t+t_w} u_{\mathrm{unp}}(t')\,u_{\mathrm{per}}(t'+t_s)\,dt'}{\left(\int_{t-t_w}^{t+t_w} u^2_{\mathrm{unp}}(t')dt' \int_{t-t_w}^{t+t_w} u^2_{\mathrm{per}}(t')\,dt'\right)^{\frac{1}{2}}}, \qquad (2)$$

where the time window is centered at time t with duration $2t_w$, t_s is the time shift used in the cross-correlation, u_{unp} is the unperturbed (before the velocity change) wave field and u_{per} the perturbed wave field (after the velocity change). δt is the arrival time difference between the waves recorded before the velocity change and the waves recorded after the velocity change, for all the waves that arrive in the employed time window. Hence, we can estimate δt for multiple non-overlapping time-windows independently and from eq. (1) we calculate the relative velocity change for each time window. Thus, coda wave interferometry features it's own consistency check. Pacheco & Snieder (2005, 2006) derived the relation between δt and a heterogeneous change in velocity.

3 EXPERIMENTAL SETUP AND LOCAL GEOLOGY

The Edgar Mine, Idaho Springs, CO, is located about 55 km west of Denver, CO, and is owned and operated by the Colorado School of Mines. The mine portal is at an elevation of 2405 m. Workings consist of 305 m of crosscuts and drifts that access several silver-gold veins. Widths of crosscuts and drifts average about 3 m, and overburden above the mine is estimated to be 120 m. We performed the experiment on a 43 m long and 23 m wide pillar (Fig. 1). In order to introduce a controlled stress change in the pillar, a pressure cell was installed into the rock by Scott *et al.* (1999). The slot cut for the pressure cell is 1 m high and 0.2 m wide and is cut 3 m deep into the pillar. The slot was formed by drilling closely spaced holes into the pillar, and the pressure cell is installed to a depth of 2.1 m. Grout was pumped around the pressure cell to allow it to press against the walls of the slot without expanding excessively. The pressure cell that is used to exert a normal stress to the walls of the slot, measures 81.3 cm^2 by 1 cm (inset of Fig. 1). Finite element modeling shows that the stress change extends into the pillar over a distance comparable to the size of the plates that apply the stress from the jacks to the pillar. This means that a volume of the order of 10 m^3 is affected by the stress change.

The mine is developed in Precambrian metamorphic and granitic rocks of the Colorado Front Range. Specifically, the rocks are assigned to the Idaho Springs Formation. Rocks in the pillar include biotite schist, biotite microcline pegmatite, biotite-hornblende schist, quartz-feldspar-biotite gneiss and migmatized gneiss. Ultrasonic core velocity measurements on rock samples from the Edgar mine where taken in the laboratory by Scott *et al.* (1999). Ultrasonic *P*-wave velocities for the biotite-hornblende-schist range between 5.00 and 5.45 km s^{-1}, and for the pegmatite between 4.72 and 4.75 km s^{-1}. Carmichael (1982) lists *P*-wave velocity changes for similar rocks (not from the Edgar Mine) for a pressure difference of 6.89 MPa in schist and pegmatite that range between 0.03 and 0.045 per cent, depending on the specific kind and area of origin. The jacks in our experiment apply a uni-axial stress to the mine pillar. The associated change in the *P*-wave velocity is much larger for this uni-axial stress change than for a hydrostatic stress change. For example, Sarkar *et al.* (2003) measured a 4 per cent change in *P*-wave velocity in Berea sandstone that is exposed to a uni-axial stress of 6 MPa.

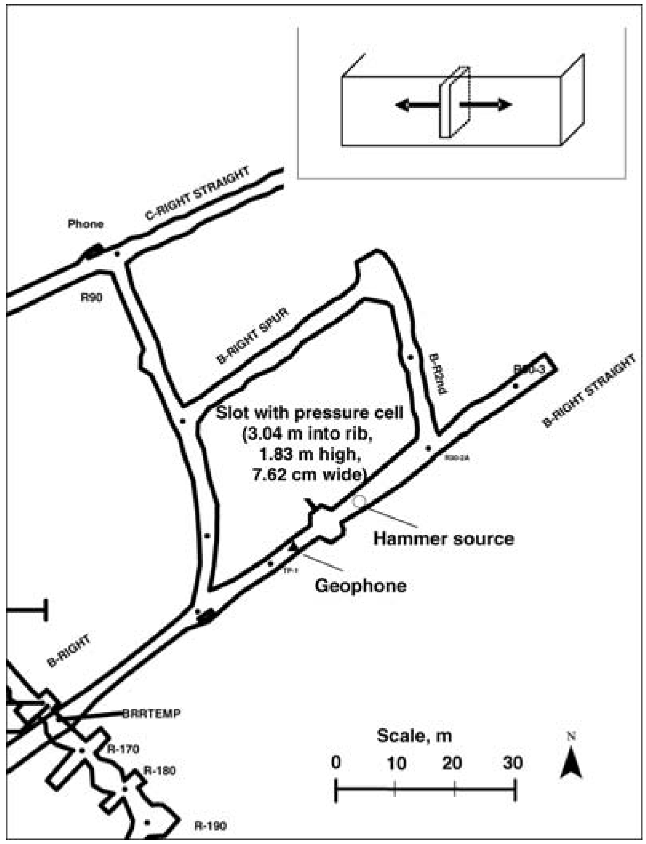

Figure 1. Plan view of the Edgar Mine pillar. Geophone and hammer source locations are labeled in the plan. The pressure cell in the slot is indicated by the small line between geophone and source location. The inset (top right) sketches the pressure cell installed in the pillar.

We used a hammer source (5 kg sledge hammer) and two vertical component geophones bolted to the wall on the left and the right of the slot in our experiment. One geophone is used as a trigger channel and the other for data acquisition with a Tektronix digital oscilloscope. The pressure in the cell is increased in steps of 1.51 MPa, from 0 to 15.1 MPa. For every increase of pressure, we record three repeated seismic waveforms (three hammer strikes), and stacked these waveforms in order to reduce random noise in the data. In addition, we apply a bandpass filter with corner frequencies of 20 Hz and 10 kHz to further reduce the electronic noise from the instruments. The stacked and filtered waveforms, recorded at different pressures in the cell are then compared to extract information about the stress change in the pillar.

4 REPRODUCIBLE SEISMIC WAVEFORMS

Coda waves commonly have a noisy appearance (Fig. 2). Not only are the first-arriving waves reproducible (upper panel), but the coda of two repeated measurements is reproducible as well (lower panel). The high repeatability shows that coda waves are different from noise and carry information about the medium they propagated through. In the late coda, the noise level in the un-stacked data (vibration noise, instrument noise, air movement) is ten times smaller than the signal amplitude. We estimate the noise of the seismic records from that part of the measurement before the first arriving wave.

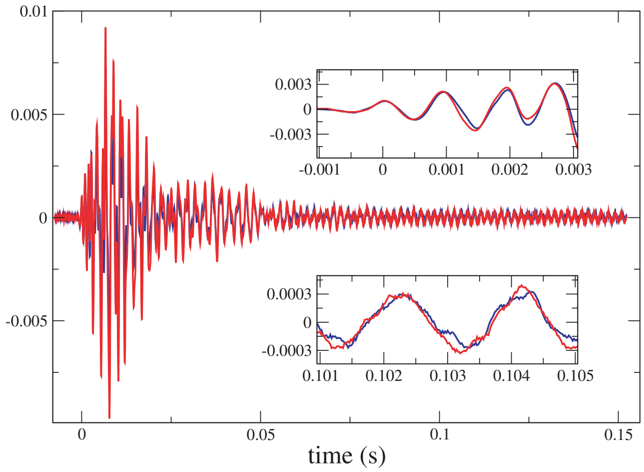

Figure 2. Waveforms recorded from different hammer strikes for the same stress. The blue line represents the first measurement and the red line the repeated measurement. The upper panel shows a time window of the early part of the measurement and the lower panel shows a time window of the late part (coda).

The waves shown in Fig. 2 decay rapidly with time around 0.02 s and have a monochromatic character with a slowly decaying amplitude for $t > 0.05$ s. The envelope of these waves does not have the smooth shape that is characteristic of multiply scattered waves whose energy spreads diffusively through space with time (e.g. Sato & Fehler 1998; Wegler 2004). This indicates that for $t > 0.05$ s the coda is caused by a reverberation rather than by multiple scattering.

The correlation coefficient of the repeated waveforms of Fig. 2 is about 0.96. If all the recording parameters (source, receiver, instrumentation) stay constant and nothing changes in the rock-mass, we can reproduce the same seismic measurement to a high degree of precision. If there is a change in the rock (e.g. stress, modulus of deformation) the seismic measurements will change too.

The velocity perturbation inferred from coda wave interferometry is averaged over the region sampled by the coda waves, as well as over shear and compressional waves. However, an examination of the coda shows that after the transient has died out ($t > 0.05$ s) we have excited a narrow-band signal with a frequency of about 540 Hz. This reverberation is not due to an air wave; the air wave that bounces between the walls of the tunnel a period of about $f = (340\,\mathrm{m\,s^{-1}})/(6\,\mathrm{m}) = 57$ Hz, and the air wave that propagates through the tunnel around the pillar has a resonant frequency of about $f = (340\,\mathrm{m\,s^{-1}})/(145\,\mathrm{m}) = 2.3$ Hz. This observed frequency of 540 Hz is consistent with the fundamental mode of a surface wave bouncing up and down between the base and roof of the pillar (about 3 m) along the surface near the source and receiver. It is also consistent with a surface wave that propagates around the tunnel. With our limited sampling of the wavefield we cannot identify the wavepaths that most strongly contribute to the coda.

5 DECORRELATION OF CODA WAVES WITH STRESS CHANGE

An increase of the pressure in the cell induces a change the seismic velocity, which leads to changes in the waveforms. When the change in the medium is small, the coda waves are more affected by this change than the early-arriving waves are. Fig. 3 shows two waveforms, one recorded at 4.14 MPa and one at 12.41 MPa. The early arriving waves (upper panel) are unaffected by the stress change, and cannot be used to monitor a stress difference of 8.27 MPa in the slot. Consequently, tomographic images based on one-way traveltimes, before and after the stress change, would be identical and cannot be used to monitor the stress difference. The coda waves, however, show a decorrelation of the waveforms (lower panel). Due to their increased sensitivity, the coda waves can be used to monitor the velocity change that is induced by a stress difference of 8.27 MPa.

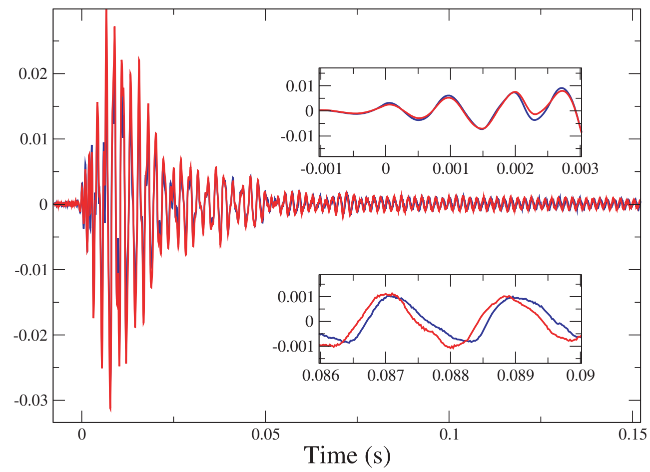

Figure 3. Two waveforms, one measured at 4.14 MPa of pressure in the cell (blue) and one measured at 12.41 MPa (red). The upper panel shows the early time window and the lower panel part of the coda.

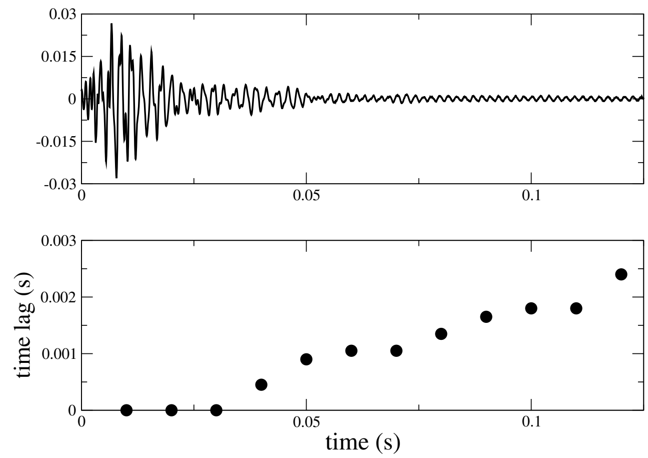

Figure 4. The seismic record (top panel), and estimates of δt for multiple time-windows (bottom panel). The first three points are computed from the early part of the waveform and are not sufficiently sensitive to detect the velocity change.

The velocity change in the medium does not lead to an erratic decorrelation of the coda but rather a coherent time lag or phase shift (lower panel of Fig. 3). We compute this time lag δt by the lag of the maximum of the cross-correlation function given by eq. (2). By computing δt for multiple, non-overlapping, time windows of length $2t_w = 0.01$ s, we obtain increasing values for δt for increasing total traveltimes t (Fig. 4). These are independent estimates of traveltime perturbations for given total traveltimes.

Because of the low sensitivity of early arriving waves to changes in the velocity, δt is for early times smaller than the sampling interval of the seismic record. This leads to large uncertainties in the estimated velocity changes as inferred from eq. (1). For that reason,

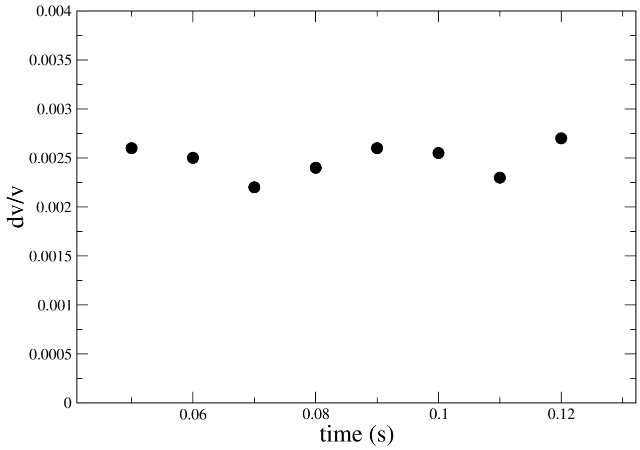

Figure 5. Velocity change estimates as a function of the center time of the employed time windows. The mean velocity change is 0.25 per cent and the standard deviation is 0.02 per cent.

we don't estimate velocity changes from the early part of the signal (δt smaller than sampling interval). Furthermore, we assume a constant velocity change over the whole area sampled by the waves. If that assumption holds, then the time lag δt grows linearly with traveltime t (Grêt *et al.* 2006). If the velocity change is localized, however, then there is a non-linear dependence of δt on t (Pacheco & Snieder 2005, 2006). In our experiment we introduce a localized stress change around the pressure cell and would therefore expect a non-linear dependency of δt on t. Nonetheless, we measure a linear increase of δt with time t (Fig. 4). In order to explain this contradiction, we speculate that the energy that leaks out of the pillar does not come back. Hence, all the energy measured for the late coda is trapped in the pillar sampling only the rock mass in the pillar. In other words, the induced stress change is locally confined but so is the area sampled by the waves in the late coda.

We compute the relative velocity change in the rock $\delta v/v$ from eq. (1) to obtain the phase shift δt from the cross-correlation function and the total traveltime from the time-location of the data window. Since we use non-overlapping time windows, we obtain multiple, independent estimates of $\delta v/v$ (Fig. 5). This makes it possible to compute the mean and variance of the relative velocity change. Note that we are measuring a small relative velocity change of the order of 0.2 per cent. By stacking over many shots, Yamamura *et al.* (2003) detected comparable velocity changes from transmitted waves, but in general it is difficult to detect such small velocity changes with transmitted waves.

6 MONITORING STRESS CHANGE

In order to test the accuracy of the method we measure the relative velocity change for four pressure differences of equal amount (8.27 MPa) but with different starting pressures (0, 4.14, 5.51 and 6.89 MPa, see Table 1). The measured velocity changes for a stress difference of 8.27 MPa each, agree well with the exception of the initial stress increase from 0–8.27 MPa. This stress change is of the order of the lithostatic overburden pressure (about 4 MPa). The large velocity increase with the initial stress change can be explained with the closure of void space, e.g. pores and cracks (Hubbert & Willis 1957; Friedel *et al.* 1995). The velocity change is greatest with an early increment increas in stress because during this stress change the voids are closed.

Table 1. The relative velocity change averaged over all employed time windows (middle column), and the associated variance (right column), for the stress change shown in the left column.

Stress difference (MPa)	Mean δv (per cent)	Standard deviation (per cent)
0–8.27	0.25	0.02
4.13–12.41	0.14	0.05
5.51–13.79	0.15	0.02
6.89–15.17	0.13	0.03

The relative velocity change, due to a pressure increase of 8.27 MPa, measured with coda wave interferometry is lower than what we would expect from the values of about 2 per cent given by Sarkar *et al.* (2003) for Berea sandstone under uni-axial stress. In the analysis we assume that the stress change occurs in the entire pillar, which is not the case in this experiment. What we measure with coda wave interferometry is a mean velocity change over the region of the pillar that is sampled by the reverberating coda waves. This results in a lower value of velocity changes, since some areas of the pillar are not affected by the stress change. An increase of 8.27 MPa in the entire pillar would result in a constant velocity change in the whole area sampled by the coda waves. In that case, coda wave interferometry would measure a larger average velocity change. In other words, coda wave interferometry would be even more sensitive to a stress change in the entire pillar and could monitor smaller stress changes than 8.27 MPa.

7 DISCUSSION

The pressure cell changes the stress in a limited area close to the slot. We can not tell from the coda waves alone where the stress change is located. In practice one could complement coda wave interferometry with local measurements of stress change or deformation. Commonly used instruments are strain gages (Hawkes & Bailey 1973), extensometers (Joass 1993; Sheppard & Murie 1992) tiltmeters (Joass 1993) and one-way traveltime tomography.

We found that the smallest pressure difference that can reliably be detected in our experiment with coda waves is 8.27 MPa. We introduced a localized stress change; a volume of about 10 m^3 is affected by the stress change. This constitutes a small fraction of the total volume of the pillar. The waves can only be sensitive to such a localized stress change, when they repeatedly sample the region that is affected by the change in stress. This is the case for waves that bounce along the tunnel wall between the top and the bottom of the tunnel, or for surface waves that propagate around the tunnel wall. This conclusion is supported by by the linear dependence of the time lag on time shown in Fig. 5. If the waves would sample an increasingly larger region of space with time, the time lag due to the localized stress stress change would not grow linearly with time (Pacheco & Snieder 2005).

We demonstrate that coda wave interferometry is a sensitive tool to monitor stress changes in an underground mining environment. It is a suitable method when classical tomography is not sensitive enough to measure a stress change. In addition, coda wave interferometry needs only one receiver (geophone) and a hammer source, hence the hardware requirements are modest. This makes coda wave interferometry inexpensive and easy to install. The data processing is basic and can easily be built in a simple, stand-alone instrument. An instrument can be developed that is a fully automated monitoring system and resists the rough mining environment. The method has potential for monitoring rates of stress change and deformation in the rock mass surrounding excavations both in underground and surface mines.

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