Time-lapse monitoring of velocity changes in Utah

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Abstract
The Wasatch fault region is an actively deforming region characterized by prominent seismicity and has the potential for large magnitude events in the near future. The present day deformation of the region, which extends into the Basin and Range, motivates the need for continuous monitoring of the region. One of these continuous monitoring is a time-lapse characterization of velocity properties of the region. In this study we monitor time-lapse velocity changes within Utah and eastern Nevada using coda waves generated by explosive events. This monitoring characterizes velocity changes within the region from June to September of 2007. We observe, both temporally and spatially, variable velocity changes within the monitored region, with a maximum path-average velocity changes of 0.2%. This suggests a significant change in the velocity within the region given the short monitoring duration and provides enormous implications to the characterization of the deformation within the region.

Key words: time-lapse monitoring, coda wave interferometry

1 INTRODUCTION
Coda wave interferometry is an effective tool to monitor time-lapse changes within a medium, especially for weak changes within the medium (Snieder et al., 2002). Coda wave interferometry allows us to extract subsurface changes from scattered seismic waves generated from repeated sources. With identical sources and negligible noise, differences within the seismic coda (such as time shift and amplitude decay) provide information about the changes within the monitored medium through which the coda waves travel. Due to the redundancy in the coda waves and possibly increased illumination of the subsurface by the scattered waves, the sensitivity of the scattered waves to perturbations within the subsurface usually increases with increasing travel time. This sensitivity of the multiply scattered waves has allowed for monitoring weak changes in velocity, in the order of 0.1% (Snieder, 2006). Coda wave interferometry has successfully been used to monitor velocity changes along fault regions (Schaff and Beroza, 2004; Poupinet et al., 1984), detect in situ velocity changes due to stress changes in mining sites (Grt et al., 2006), characterize near-surface velocity changes (Nakata and Snieder, 2012), monitor temporal changes within volcanic regions (Matsumoto et al., 2001), and detect far-field stress-induced velocity changes (such as solid earth tide) (Spane, 2002).

The monitored region in this study covers western and central Utah, eastern Nevada, and the southern part of Idaho. Some of the notable features within this region include most of the Wasatch fault region, the north-western Colorado plateaus, the eastern part of the Great Basin, and the southern Rocky Mountain region (Prodeul, 1970) (figure 1). The Great Salt Lake and the Wasatch front are situated in the eastern section of the monitored region. The Wasatch fault system is a 370-km normal fault system (Armstrong et al., 2004) cutting across Utah from the north to the southern part of Utah. The fault defines the eastern boundary of the Basin and Range province of western North America. With a northward striking direction and a dip range of 45°-60° toward the west, the Wasatch fault characterizes a broad area of active deformation which consists of more than 10 segments, and the regions connecting the fault segments are characterized by gaps in seismicity (Simpson and Richards, 1981; Gori and Hays, 1992). The Wasatch fault separates the Wasatch Range to the east and Salt Lake Basin to the west. The Wasatch range consists of the Rocky Mountains characterized by late Paleozoic and Mesozoic rocks (Hunt, 1956). To the west of the fault is the Salt Lake Basin which is an extension of the Basin and Range. Unlike the Wasatch range, this section of the western US is actively deforming which has been attributed to a variety of factors including crustal shearing due to the north-
western relative movement of the Pacific Plate (Wise, 1963; Hamilton and Myers, 1966) and the elevated heat characteristics of the Western US (Eaton, 1982). Some places within the Salt Lake Basin consist of soft sediments which are more than 1-km deep (Roten et al., 2011). The soft sediments are potentially susceptible to both near-field and far-field stress loadings.

In this study, we monitor temporal velocity changes within the crustal sub-surface around the Wasatch fault region. We use coda waves generated by time-lapse active sources conducted in the summer of 2007. These coda waves were recorded by the equally spaced US-Array stations which cover most of the Wasatch fault cutting across Utah. In the following section, we describe in detail the data processing routine we use in this study. In sections 3 and 4 we present the results from the time-lapse monitoring of the Wasatch fault region and interpretation of the results, respectively. Section 5 both concludes and discusses the assumptions we made in this study.

2 DATA PROCESSING

We process coda signals generated by 10 rocket motor explosions, occurring between June 4 and September 10 2007, which were surface rocket explosions at the Utah Test and Training Range (UTTR) of Hill Air Force Base, Utah (Stump et al., 2007). Each of the explosions was carried out at the same location with latitude and longitude of N41.141° and W112.9065°, respectively. These explosions are identical. The similarity in the source properties of the monitoring signals prevents errors in the estimated fractional velocity changes due to changes in the properties of the sources (such as shift in source location or source mechanism) generating the coda signals (Weaver et al., 2011; Kanu et al., 2013). The onset times of the explosions are given in Table 1. The blast signals were recorded on the USArray transportable array (TA) shown in Figure 1 and the TA consists of 54 stations surrounding the blast location (red star) shown in Figure 1.

2.1 Processing of blast doublets for velocity changes

In this study we process three (N-S, E-W and vertical) components of the recorded explosive signals. Figure 2 shows an example of the three-component blast signals. The signals in figure 2 are unfiltered having an amplitude spectrum shown in Figure 3. We use a 1-5Hz frequency band in this time-lapse monitoring and the blast signal for the 1-5Hz frequency band is given in Figure 4.

In this time-lapse study we use the coda section of the signals (i.e. the section of the signal with exponential amplitude decay) (Figure 5). The onset time of the

![Figure 1. USArray transportable array given by the green squares. The locations of the stations are given relative to the blast location (red star).](image1)

![Figure 2. Typical recordings of the blast events. The three components, N-S (blue), Vertical (red) and the E-W (black) are all used in the time-lapse analysis.](image2)
Table 1. Descriptions for the blast events used in this studies.

<table>
<thead>
<tr>
<th>Event (Data name)</th>
<th>Month</th>
<th>Day</th>
<th>Hour</th>
<th>Minutes</th>
<th>Seconds</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference (20070604)</td>
<td>06</td>
<td>04</td>
<td>19</td>
<td>52</td>
<td>21.22</td>
</tr>
<tr>
<td>Signal 1 (20070611)</td>
<td>06</td>
<td>11</td>
<td>19</td>
<td>49</td>
<td>24.10</td>
</tr>
<tr>
<td>Signal 2 (20070626)</td>
<td>06</td>
<td>26</td>
<td>19</td>
<td>43</td>
<td>19.78</td>
</tr>
<tr>
<td>Signal 3 (20070709)</td>
<td>07</td>
<td>09</td>
<td>21</td>
<td>38</td>
<td>37.13</td>
</tr>
<tr>
<td>Signal 4 (20070716)</td>
<td>07</td>
<td>16</td>
<td>17</td>
<td>33</td>
<td>31.13</td>
</tr>
<tr>
<td>Signal 5 (20070801)</td>
<td>08</td>
<td>01</td>
<td>20</td>
<td>01</td>
<td>24.26</td>
</tr>
<tr>
<td>Signal 6 (20070806)</td>
<td>08</td>
<td>06</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Signal 7 (20070813)</td>
<td>08</td>
<td>13</td>
<td>19</td>
<td>38</td>
<td>30.56</td>
</tr>
<tr>
<td>Signal 8 (20070827)</td>
<td>08</td>
<td>27</td>
<td>20</td>
<td>43</td>
<td>11.45</td>
</tr>
<tr>
<td>Signal 9 (20070910)</td>
<td>09</td>
<td>10</td>
<td>17</td>
<td>33</td>
<td>01.56</td>
</tr>
</tbody>
</table>

Figure 3. Typical amplitude spectrum of the blast events. The three components, N-S (blue), Vertical (red) and the E-W (black) are all used in the time-lapse analysis. The gray rectangle bounds the used frequency range.

Figure 4. Typical recordings of the blast events (filtered) bandpass filtered between 1-5Hz. The three components, N-S (blue), Vertical (red) and the E-W (black) are all used in the time-lapse analysis.

Figure 5. The coda section of the blast signals. The three components, N-S (blue), Vertical (red) and the E-W (black) are all used in the time-lapse analysis. The gray rectangle bounds the coda section of the signal.

coda section after the S-waves varies between stations. We filter the signals between 1-5Hz. We also normalize each signal with its maximum amplitude and resampled each of the signals from a time interval of 0.025s to 0.001s. The resampling increases the resolution of the estimated velocity changes. The data is processed using three signal referencing cases. Case 1 consists of comparing each of the recorded signals to the first blast signal (Signal 20070604 - Table 1). In Case 2, each signal is compared to the previous signal (in time). For Case 3, we use only the signals from 08/01/2007 to 09/10/2007. Each of the signals in this Case is compared to the blast signal on 08/01/2007 (Signal 20070801 - Table 1). The purpose for these three cases is to check for consistency in our estimate of velocity change.

To extract the fractional velocity change from the time-lapse data, we use the stretching method (Hadziioannou et al., 2009). However rather than using the maximum correlation as the misfit function for the time-lapse misfit, we use the $L_2$ norm as the misfit function (Kamu et al., 2013). We initialize the time of the signals to the onset times of the explosive events in order
to apply the stretching algorithm. A doublet consists of time-lapse signals from two explosive events. The source properties of the explosive signals are the same; therefore any difference in the signals results from either changes along the propagation path of the signals or differences in noise properties of the signals. The major challenge we encountered while analyzing this explosive data was accounting for missing data. In some cases the data were either missing one of the components or all three of the components. We only used explosive signals that are recorded on all three components of the stations. Table 2 shows the stations with their missing data.

Because of the presence of noise in the data and because the noise becomes prominent with increasing coda time, we only use the codas where the correlation of the early part of the coda is greater than 0.75 and the correlation of the end of the coda is greater than 0.5 (similarity criterion). This section of the coda is submitted to the stretching algorithm. In cases where this criterion is not met, we assign zero to the relative velocity change and its error. We compute the error associated with the estimated velocity change using the method of Kanu et al. (2013) (equation 14).

The fractional velocity changes estimated in Cases 1 and 3 are cumulative velocity changes from the reference signals. However in Case 2, the changes estimated are interval velocity changes for each monitored time period. In order to express the estimated velocity changes of Case 2 in a way that is consistent with the other two cases, we sum the estimated velocity changes and their associated error values using the following equations:

\[ \langle \frac{\delta V}{V} \rangle_j = \sum_{i=1}^{j} \langle \frac{\delta V}{V} \rangle_i, \]

and

\[ e_j = \sqrt{\sum_{i=1}^{j} e_i^2}, \]

where \( \langle \frac{\delta V}{V} \rangle \) is the estimated average velocity change, \( e \) is the error of the relative velocity change, and \( i \) and \( j \) denote the time intervals we are monitoring with values 1 to 9.

### 3 TIME-LAPSE VELOCITY CHANGE

#### 3.1 Average velocity changes

Figure 6 shows the cumulative estimates of the average relative velocity changes for Case 1, Case 2 and Case 3. We compute the average velocity change using all the monitoring stations with non-zero estimated velocity change. In Case 2, we have zero velocity changes in the first two time periods (doublets) because the time-lapse doublets do not meet our similarity criteria in the data processing (section 3.1). This inability to meet our criteria is because the signal resulting from the explosion of June 11 (Signal 20070611 - Table 1) is different from the rest of the signals due to the presence of noise. In Case 3 only signals after August 1 are processed, and the reference signal for the estimation of velocity change in Case 3 is the signal of August 1 (Signal 20070801 - Table 1). Each of these cases show significant velocity changes between June 4 and July 9. However the subsequent estimated velocity changes (\( \delta V/V(\%) \)) for both Cases 1 and 2 are below the error-level while in Case 3 these estimated velocity changes are well above the error values. The error values are standard deviations of the estimated velocity changes.

Comparison of the estimated velocity changes from all three cases reveals that the estimated velocity changes for each of the components follow a consistent trend. The average velocity increases within the monitored region from the June 4 to June 26. This increase in velocity remains fairly constant until August 1 after which the average velocity decreases. The reduction in velocity occurs between August 1 and 6. As is indicated in Case 3, the average velocity remains constant between the August 6 and 13. After the August 13, the average velocities increases and decreases for the following time periods: August 13 to 27 and August 27 to September 10, respectively.

### Table 2. The stations with missing data and their missing data

<table>
<thead>
<tr>
<th>USArry Station</th>
<th>Missing Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>L15</td>
<td>20070604, 20070611, 20070626, 20070709</td>
</tr>
<tr>
<td>L16</td>
<td>20070604, 20070611, 20070626, 20070709</td>
</tr>
<tr>
<td>M16</td>
<td>20070604, 20070611, 20070626, 20070709</td>
</tr>
<tr>
<td>N16</td>
<td>20070604, 20070611, 20070626, 20070709</td>
</tr>
<tr>
<td>O15</td>
<td>20070604, 20070611, 20070626</td>
</tr>
<tr>
<td>O16</td>
<td>20070604, 20070611, 20070626</td>
</tr>
</tbody>
</table>
3.2 Spatial distribution of velocity change

Figure 7 shows a trend in velocity change similar to what is observed in the average velocity changes computed using all the stations. These average changes are the estimated velocity changes in Case 2 for each time-lapse period. The first two time-lapse periods (06/04/2007 - 06/11/2007 and 06/11/2007 - 06/26/2007) are missing due to presence of noise in signal 20070611 (Table 1). Figure 7, shows the spatial distribution of the velocity changes as changes seem to vary with source-receiver pairs; and the region of dominant velocity changes also varies from one time period to the other.

The estimated velocity changes generally decrease with increasing source-receiver distance (Figure 8). The dependence of the estimated velocity change on the source-receiver distance explains the high error levels in the average of the estimated velocity change shown in section 3.1. The spread in the velocity changes estimated at the stations is not due to random error but due to variations in the estimated velocity change with the source-receiver distance. The spatial distribution of the velocity changes and the dependence of the velocity change on the source-receiver distance might suggest that the velocity changes in the medium are localized. During some of the time intervals shown in Figure 8, the estimated velocity changes are distributed over broad values of velocity changes for the same source-receiver distance (for example, for period 07/16/2007 - 08/01/2007 at the source-receiver distance of 100km). This might be an indication of lateral variation of the velocity change within the monitored region.

3.3 Parameter Estimation:

The travel time shifts (and therefore the estimated velocity changes) depend on the magnitude of the subsurface velocity change $\Delta v$, the volume of the velocity change $\Delta V$, the diffusion coefficient $D$ of the scattered wave intensity (Pacheco and Snieder, 2005), and the source-receiver distance $R$. Assuming a uniform lateral velocity change which extends to the earth’s free surface (Figure 9), we can formulate the parametric dependence of the estimated relative velocity change $\langle \epsilon \rangle$ as follows:

$$
\langle \epsilon \rangle = F(\Delta v/v, h, D, r),
$$

where $h$ is the depth extent of the velocity change.

The function $F$ can be found from the theory of
Pacheco and Snieder (2005), which relates the estimated relative velocity change \( \langle \epsilon \rangle \) to the model parameters \((\Delta v/v, h, D, r)\) as follows:

\[
(\epsilon(t)) = \int \frac{\Delta v}{v} K(r, t; D) \, dV.
\]

The estimated relative velocity change seems to depend directly on the parameter \( \zeta \), which is defined as

\[
\zeta = \frac{h}{D} \frac{\Delta v}{v}.
\]

Figure 10 shows the apparent (estimated) velocity changes using the source-receiver distance of station J13 for values of \( \zeta \) from 0 to 0.1 using equation (4). These values of \( \zeta \) correspond to values of \( \Delta v/v, h \), and \( D \) ranging from 0-10\%, 0-10km, 1-10x10^3 m/s^2, respectively. Figure 10 suggests that the three parameters - the magnitude of the subsurface velocity change \( \Delta v \), the depth of the velocity change \( h \), and the diffusion coefficient \( D \) - only influence the estimated velocity change through the combination that defines \( \zeta \) in equation 5. The diffusion coefficient, however, shows a weaker trade-off with the other two parameter, the magnitude of subsurface velocity change \( \Delta v/v \) and the depth of velocity change \( h \). Station J13 is chosen arbitrarily because the trade-off of the three model parameters applies to all the stations.

The strong trade-off between the \( \Delta v/v \) and \( h \) suggests that determining each of these parameters independently will be difficult. According to equation (A12) in Appendix A, we can use a power law relation to describe the estimated velocity changes (Figure 11) as:

\[
\langle \epsilon \rangle = a R^b,
\]

where \( a = 3 \frac{h}{D} \Delta \Sigma \) and \( b = -1 \) (See Appendix A). \( \Sigma \) is a dimensionless constant of the monitored region. Table 3 gives the estimated parameters for the fitting solutions of the estimated velocity changes, the qualities of the fitting solutions in the adjusted R squared (a measure of regression) values, and the inverted values of \( h \Delta v/v \). We
Figure 6. The cumulative relative velocity changes expressed in percentages for N-S (red), Vertical (green) and the E-W (black) components. Here the average is computed using all the stations in TA. Missing first two estimates in Case 2 is due to influence of noise in the signal of June 11. Missing first five estimates in Case 3 is because only signals from 1st of August is processed in Case 3.

Figure 9. Schematic description of the layer of near-surface fractional velocity change. The red star is the source and the black triangles are the receivers.

Figure 10. Dependence of the estimated velocity change on the dimensionless parameter $\zeta$. The definition of $\zeta$ is given in equation 5. The green region corresponds to the range of velocity changes estimated with station J13.

use the weighted nonlinear least-squares method with the inverse of the estimated errors as weights. We ignore estimated velocity changes less than 0.02% for period 07/16-08/01 in the fitted solution, because the values do not fit the power law relation. Using the diffusion model of scattered intensity in the 3D semi-infinite medium we can only invert for values of $h\Delta v/v$ from the estimated velocity changes. Based on the surface layer model, the inverted $h\Delta v/v$ values are in the order of 1%km, which implies that the percentage velocity change is of the order 1% over a depth in the order of 1km. These values seem large for a region such as Utah with negligible precipitation and given that the durations of the monitoring are only about 2 weekly basis.
Figure 11. Power law fitting of the estimated velocity changes in Utah Wasatch fault region. Fit 1, 2, and 3 are the fitting solutions for the estimated velocity changes of E-W, N-S, and Z-U components, respectively.

4 CAUSES FOR THE VELOCITY CHANGES

4.1 Rainfall and groundwater level

Utah is a region that receives little precipitation. Figure 12 shows the amount of precipitation observed within the time period we are monitoring using the following two stations: Salt Lake County Government Center (N40.7284 W111.8877) (Figure 12.1A and 12.2A) and Magna Station (N40.7109 W112.1002) (Figure 12.1B and 12.2B). The figures show that the amount of precipitation is small (amounting to a cumulative precipitation of less than 2 inches) during the monitored period. This may explain why there is a negligible relation between the precipitation and estimated velocity changes.

The groundwater level can vary temporally, not only due to precipitation but also to stress loading resulting from large earthquakes (Montgomery and Manga, 2003) and solid earth tides (Spane, 2002). Figure 13 shows the cumulative groundwater subsidence over the monitored period in this study. The recording wells of the groundwater subsidence are shown in Figure 14. Figure 13 shows that the cumulative groundwater subsidence generally increases and decreases with the average estimated velocity change. However, the maximum cumulative groundwater subsidence is less than 10m. This amount of groundwater subsidence is unlikely to totally explain the observed velocity changes.

4.2 Local seismicity

Figure 15 shows the peak horizontal acceleration (PHA) due to local seismicity and the estimated velocity changes. We compute peak horizontal acceleration using the PHA relationship with event magnitude proposed by Campbell (1981) based on world-wide data (equation 7):

$$\ln PHA(g) = -4.141 + 0.868M - 1.09 \ln[R + 0.606 \exp(0.7M)],$$

(7)

where $M$ is the event magnitude, $R$ is the distance to the fault rupture location in km, and $g$ is the accelera-
Table 3. Estimated model parameters using the inverse R relation.

<table>
<thead>
<tr>
<th>Event period (mm/dd)</th>
<th>( a (% \text{km}) )</th>
<th>Adjusted ( R^2 )</th>
<th>( h \frac{\Delta v}{v} (% \text{km}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Period 1 (06/04-06/11)</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Period 2 (06/11-06/26)</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Period 3 (06/26-07/09)</td>
<td>E-W</td>
<td>8.272 ±0.6536</td>
<td>0.3448</td>
</tr>
<tr>
<td></td>
<td>N-S</td>
<td>8.223 ±0.6651</td>
<td>0.3593</td>
</tr>
<tr>
<td></td>
<td>Z-U</td>
<td>8.450 ±1.1125</td>
<td>-0.3393</td>
</tr>
<tr>
<td>Period 4 (07/09-07/16)</td>
<td>E-W</td>
<td>-3.382 ±0.3990</td>
<td>0.2096</td>
</tr>
<tr>
<td></td>
<td>N-S</td>
<td>-3.342 ±0.4005</td>
<td>0.2440</td>
</tr>
<tr>
<td></td>
<td>Z-U</td>
<td>-3.574 ±0.5839</td>
<td>0.2823</td>
</tr>
<tr>
<td>Period 5 (07/16-08/01)</td>
<td>E-W</td>
<td>4.677 ±0.3941</td>
<td>0.4538</td>
</tr>
<tr>
<td></td>
<td>N-S</td>
<td>4.677 ±0.3941</td>
<td>0.4538</td>
</tr>
<tr>
<td></td>
<td>Z-U</td>
<td>4.382 ±0.4513</td>
<td>0.2150</td>
</tr>
<tr>
<td>Period 6 (08/01-08/06)</td>
<td>E-W</td>
<td>-9.001 ±0.6926</td>
<td>0.09855</td>
</tr>
<tr>
<td></td>
<td>N-S</td>
<td>-8.932 ±0.7020</td>
<td>0.1131</td>
</tr>
<tr>
<td></td>
<td>Z-U</td>
<td>-7.889 ±0.8025</td>
<td>-0.0293</td>
</tr>
<tr>
<td>Period 7 (08/06-08/13)</td>
<td>E-W</td>
<td>4.003 ±0.5133</td>
<td>-0.3547</td>
</tr>
<tr>
<td></td>
<td>N-S</td>
<td>3.937 ±0.4923</td>
<td>-0.4684</td>
</tr>
<tr>
<td></td>
<td>Z-U</td>
<td>4.019 ±0.4722</td>
<td>0.06362</td>
</tr>
<tr>
<td>Period 8 (08/13-08/27)</td>
<td>E-W</td>
<td>-5.143 ±0.9411</td>
<td>0.118</td>
</tr>
<tr>
<td></td>
<td>N-S</td>
<td>-5.602 ±1.1110</td>
<td>0.1414</td>
</tr>
<tr>
<td></td>
<td>Z-U</td>
<td>-6.328 ±3.2270</td>
<td>-0.355</td>
</tr>
</tbody>
</table>

Table due to gravity. We compute the PHA values using magnitude and hypocenter parameters of the detected seismic events during the monitored time period for station L11A. Station L11A is selected arbitrarily.

The following durations have anomalous PHA events: 06/04/2007 - 06/11/2007, 06/26/2007 - 07/09/2007, 08/01/2007 - 08/06/2007, 08/13/2007 - 08/27/2007, and 08/27/2007 - 09/10/2007. Among these time periods, only the following time periods – 06/04/2007 - 06/11/2007, 06/26/2007 - 07/09/2007 and 08/01/2007 - 08/06/2007 – have the anomalous PHA events occurring near the end of the period. Apart from the period 06/04/2007 - 06/11/2007 – in which we do not have the estimate of velocity change – these three periods correspond to periods with highest velocity changes (Figure 14). However, the periods 06/26/2007 - 07/09/2007 and 08/01/2007 - 08/06/2007 with anomalous PHAs show different signs in the velocity changes. To make a closer comparison of the PHA and velocity changes, we might need to incorporate the local site effects of the monitored region into the PHA computation and compute the spatial PHA across the monitored region.

The effect of the local seismicity is expected to be constrained near the Wasatch fault which is situated in the eastern section of the monitored region. This section of the monitored region is close to the source location and might in part explain the source-receiver dependence of the estimated velocity change.

4.3 Crustal deformation

A critical assumption we made in the estimation of \( h \frac{\Delta v}{v} \) is that the velocity changes are uniformly limited to a surface layer, which may be unrealistic. The monitored region covers the eastern flank of the Basin and Range, which has been shown to have elevated heat characteristics (Eaton, 1982) that might be responsible
for broad active deformation in the region. A tectonic deformation of the region will induce velocity changes within the crust; however, these velocity changes will be deeper, broader and more gradual than the changes due to other physical mechanisms such as changes in groundwater level or regional seismicity. Because of the broad and gradual properties of the crustal deformation-induced velocity changes, we expect larger values of $h$ (volume of change) and lower fractional velocity changes $\Delta v/v$. With a larger value in $h$, crustal deformation due to stress changes or elevated heat deep within the crust might explain, at least in part, the inverted values of $h\Delta v/v$ of an order of 1% km, however, discerning the depth profile of the observed velocity changes is challenging using the coda waves and the surface receivers.

5 DISCUSSIONS AND CONCLUSIONS

The estimated velocity changes and the inverted values of $h\Delta v/v$ suggest that the Wasatch front and, in extension, the eastern flank of the Basin and Range, are undergoing small but significant velocity changes within a short period of time. We observe path-averaged velocity changes that depend on the distance between the monitoring source and receivers. The magnitude of the estimated average velocity change varies from one time-lapse period to the other. The maximum of the absolute value of the estimated velocity change is 0.2%. To resolve the magnitude of the localized velocity change and the location of the observed velocity changes using the coda waves, we need to make some assumptions about the monitored region. In this study, to resolve the magnitude of the velocity changes and the region of change, we have made the following simplifying assumptions:

First, we assume that the coda waves we use for the time-lapse monitoring are described using the diffusion approximation of multiply scattered waves. This approximation is usually an over-simplification of the scattered waves especially in the early part of the coda. This approximation is likely to result in an overestimation of the inverted values of $h\Delta v/v$. Also we assume a uniform diffusion model for the monitored area defined by an average diffusivity constant. This diffusivity constant is related to the distance between source and receivers using the diffusion model of the recorded intensities. If the average diffusivity constant is unchar-
characteristic of the variation of the diffusion property of the monitored region, the inverted $h\Delta v/v$ will also be erroneous.

Second, we use a semi-infinite 3D subsurface with a fully reflecting free surface to invert for $h\Delta v/v$. The fully reflecting free surface may be a fair approximation of the free surface, but we ignore the presence of Moho discontinuity. This boundary will change the magnitude of the sensitivity kernel (equation A2) of the coda because the crustal thickness is much less than the source-receiver distance of the estimated velocity changes. Keller et al. (1975) suggest that the crustal Great Basin-Colorado Plateau transition has a thickness of about 25 km with a crustal thickness of 30 km for the Basin and Range (Gilbert and Sheehan, 2004). Assuming an absorbing boundary at the crust-upper mantle interface will likely increase the magnitude of the sensitivity kernel (equation A2) (Rossetto et al., 2011). This will result in a reduction of the value of the inverted $h\Delta v/v$.

Third, we made a 2D approximation of the equation A1 in equation A6. This is a valid approximation for $h/R \ll 1$. Therefore, the inverted $h\Delta v/v$ represents the true $h\Delta v/v$ within the subsurface if the velocity changes are restricted to the near surface. If the observed velocity changes result from a deeper region of the subsurface, then the true values of $h\Delta v/v$ might deviate from the inverted values. Because of the large values of $R$ (several hundred km), the $h/R$ approximation is not likely to significantly affect the inverted values of $h\Delta v/v$.

Fourth, due to the extent of the monitored region, the recorded explosive signals are expected to generate surface waves. The surface waves can arrive within the coda wave time window especially for short source-receiver distances. In the recorded explosive signals, the surface waves mostly arrive after the coda waves, however the relative arrival times of the coda wave and the surface wave are dependent on the source-receiver distances and the surface wave can contaminate the coda signal. In this study we have restricted the velocity changes analysis to the coda wave. However the presence of surface wave or surface dominated scattered waves will increase the sensitivity of the scattered waves to lo-
calized velocity change. This will result in lower values of $h\Delta v/v$. Therefore the inverted $h\Delta v/v$ values can be considered as the upper limit of the real $h\Delta v/v$ changes within the monitored region.

Finally, we assume a uniform lateral velocity change in a layer of velocity change across the monitored region. The reality is that the velocity changes are localized in space both laterally and in depth. The inverted $h\Delta v/v$ can be seen as the average $h\Delta v/v$ of the true $h\Delta v/v$ within the monitored region. This means that in some areas of the monitored region the true changes will be higher or lower than the inverted values.

The above assumptions notwithstanding, the estimated velocity changes suggest that the crust underneath the region around Wasatch fault is undergoing a significant velocity change within a short period of time. These velocity changes will have significant implications for the characterization of both the seismicity and deformation of the region. It would be useful to discern if these velocity changes are seasonal, i.e., if they might be linked to seasonal loadings (such as tides and precipitations) in both near- and far-fields.

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APPENDIX A: ANALYTICAL APPROXIMATION OF EQUATION 4:

The timeshift \( \langle \tau(t) \rangle \) extracted from repeating coda (assuming the diffusion model) can be related to the localized velocity changes by the model of Pacheco and Snieder (2005) as

\[
-\langle \tau(t) \rangle = \langle \epsilon(t) \rangle = \int_V K(s,x_o,r,t) \frac{\Delta v}{v}(x_o) \, dV,
\]

where \( V \) in the integration volume and \( s, x_o, \) and \( r \) are the source, arbitrary, and receiver locations, respectively. The sensitivity kernel \( K(s,x_o,r,t) \) is

\[
K(s,x_o,r,t) = \int_0^t P(s,x_o,t') P(x_o,r,t-t') \, dt',
\]

and \( P(x_1, x_2, t) \) is the normalized intensity recorded at a receiver location \( x_2 \) due to a source at \( x_1 \). Using the diffusion approximation for the normalized intensity in a 3D semi-infinite inhomogeneous medium with a full reflecting surface boundary and where the source and receivers are located on the boundary, the normalized intensity is

\[
P(R,t) = \frac{2}{(4\pi Dt)^{3/2}} \exp \left( -\frac{R^2}{4Dt} \right).
\]

Then the sensitivity kernel \( K(s,x_o,r,t) \) in a 3D medium with a full reflecting surface boundary is given by (Rossetto et al., 2011)

\[
K(s,x_o,r,t) = \frac{1}{2\pi D} \exp \left( \frac{R^2 - (r+s)^2}{4Dt} \right) \left( \frac{1}{s} + \frac{1}{r} \right),
\]

where \( R \) is the source-receiver distance, and \( r \) and \( s \) are distances from the receiver and source to \( x_o \), respectively. Therefore from equation (A1),

\[
-\frac{\langle \tau(t) \rangle}{\langle t \rangle} = \langle \epsilon(t) \rangle = \int_s \int_y \int_0^h \frac{1}{2\pi D(t)} \exp \left( \frac{R^2 - (r+s)^2}{4Dt} \right) \left( \frac{1}{s} + \frac{1}{r} \right) \frac{\Delta v}{v}(x_o) \, dz \, dy \, dx.
\]

Assuming \( h/R \ll 1 \) and that the fractional velocity change is constrained to the near surface slab such that \( \frac{\Delta v}{v}(x_o) = \frac{\Delta v}{v_o} \), then

\[
-\frac{\langle \tau(t) \rangle}{\langle t \rangle} = \langle \epsilon(t) \rangle \approx \frac{h}{v} \Delta v \int_s \int_y \int_0^h \frac{1}{2\pi D(t)} \exp \left( \frac{R^2 - (r+s)^2}{4Dt} \right) \left( \frac{1}{s} + \frac{1}{r} \right) \, dy \, dx.
\]

The travel time \( \langle t \rangle \) associated with the estimated average velocity \( \langle \epsilon(t) \rangle \) is

\[
\langle t \rangle = \int_{x'} \int_{y'} w(t) \, dt' \int_{x'} \int_{y'} w(t') \, dt.
\]

where \( w(t) \) is the intensity of the scattered waves. Because the intensity is higher in the early part of the coda rather than in the later, then the weighted average time \( \langle t \rangle \) lies close to the peak of the intensity. At the intensity peak,

\[
\frac{\partial P(R,t)}{\partial t} = 0.
\]

Based on equation (A3), \( R^2/6D(t) = 1 \).

To express the variables as dimensionless quantities, let \( x' = x/\sqrt[6]{D(t)} \), \( y' = y/\sqrt[6]{D(t)} \), \( R' = R/\sqrt[6]{D(t)} = 1 \), \( r' = r/\sqrt[6]{D(t)} \), and \( s' = s/\sqrt[6]{D(t)} \). Then,

\[
-\frac{\langle \tau(t) \rangle}{\langle t \rangle} = \langle \epsilon(t) \rangle \approx \frac{h}{\pi v} \Delta v \int_{x'} \int_{y'} \exp \left( \frac{3}{2} \left( 1 - (r'+s')^2 \right) \right) \left( \frac{1}{s'} + \frac{1}{r'} \right) \, dy' \, dx'.
\]

Let

\[
\int_{x'} \int_{y'} \exp \left( \frac{3}{2} \left( 1 - (r'+s')^2 \right) \right) \left( \frac{1}{s'} + \frac{1}{r'} \right) \, dy' \, dx' = \Sigma.
\]

Therefore,

\[
-\frac{\langle \tau(t) \rangle}{\langle t \rangle} = \langle \epsilon(t) \rangle \approx \frac{h}{\pi v} \Delta v \sqrt{\frac{3}{2D(t)} \Sigma}.
\]
Using $R^2/6D\langle t \rangle = 1$, equation A11 follows that

$$-\frac{\langle \tau(t) \rangle}{\langle t \rangle} = \langle \epsilon(t) \rangle \simeq \frac{3h}{\pi} \frac{\Delta v}{v} \frac{1}{R \Sigma}.$$  \hspace{1cm} (A12)

Equation (A12) allows for a parametric description of the apparent velocity change $\langle \epsilon(t) \rangle$ in terms of the model parameters $h$, $\frac{\Delta v}{v}$, and $R$. $\Sigma$ is a constant dimensionless parameter and is equal to $4.52 \pm 0.02$. The error in $\Sigma$ is a numerical error.
REFERENCES


