Seismic velocity reductions caused by the 2003 Tokachi-Oki earthquake

Justin L. Rubinstein,[1,2] Naoki Uchida,[3] and Gregory C. Beroza[1]

Received 11 April 2006; revised 21 November 2006; accepted 26 January 2007; published 26 May 2007.

[1] We use four repeating earthquake sequences located near Hokkaido to identify velocity changes caused by the M_w 8.0 2003 Tokachi-Oki earthquake. Using a moving window cross-correlation technique, we identify delays in the arrival time of seismic waves that accumulate linearly with time into the seismogram. This behavior is indicative of multiple scattering within a medium where the seismic velocity has been reduced. For all of our earthquake-receiver geometries, we find evidence of significant velocity reductions close to the receiver. The correlation of the size of the velocity reductions with both strong shaking and site characteristics suggests that these velocity reductions are caused by damage to near-surface materials created by nonlinear strong ground motion. For earthquake/receiver geometries where the seismic waves cross the Tokachi-Oki rupture zone, we identify particularly large reductions in velocity as a result of the earthquake. For these geometries, we believe that the rupture zone of the Tokachi-Oki earthquake or the shallow crust above it represents a second region where seismic velocities are reduced as a result of the main shock.

Citation: Rubinstein, J. L., N. Uchida, and G. C. Beroza (2007), Seismic velocity reductions caused by the 2003 Tokachi-Oki earthquake, *J. Geophys. Res.*, *112*, B05315, doi:10.1029/2006JB004440.

1. Introduction

[2] For years, seismologists have searched for changes in wave propagation both prior to and as a result of large earthquakes. Studies have suggested temporal changes in coda Q [e.g., *Su and Aki*, 1990], anisotropy [e.g., *Saiga et al.*, 2003], scattering [e.g., *Baisch and Bokelmann*, 2001], and seismic velocity [e.g., *Uchida et al.*, 2002] following earthquakes. Recently, seismic velocity changes caused by earthquakes have been intensively studied, as one only needs a repeating source, either natural (for example, repeating earthquakes) [e.g., *Poupinet et al.*, 1984; *Rubinstein and Beroza*, 2004a; *Schaff and Beroza*, 2004; *Peng and Ben-Zion*, 2006] or artificial (for example, explosions) [e.g., *Li et al.*, 1998, 2003; *Nishimura et al.*, 2005], to identify delays of one event relative to another, which are indicative of a change in seismic velocity. Authors typically use a moving window cross-correlation method to identify the aforementioned delays.

[3] The results of this type of cross-correlation analysis have been used to study delays found in direct body phases [e.g., *Li et al.*, 2003], fault-zone-guided waves [e.g., *Li et al.*, 1998], and the coda [e.g., *Schaff and Beroza*, 2004]. For studies that examine specific arrivals, assessing the influence of velocity changes is relatively straightforward, as one only needs to examine the change in traveltime (if the origin time is known precisely) or the change in lapse time between two arrivals (if the origin is not precisely known). Studies of the coda require more complicated processing as there are no distinct, consistent arrivals with which one can dependably examine arrival times. Many coda studies rely on the fact that as later phases of the coda spend progressively longer in the scattering medium producing them, the accumulated delays (assuming the velocity has been reduced in this region) will increase linearly as a function of time into the coda [*Niu et al.*, 2003; *Schaff et al.*, 2004; *Snieder et al.*, 2002]. From this, one can compute the change in slowness (1/velocity) in the medium by fitting a line to the delay function [*Schaff and Beroza*, 2004]. In addition to recovering changes in velocity, *Snieder* [2006] and *Snieder et al.* [2002], using a theory called "coda wave interferometry", have shown that changes in scatterer distribution and earthquake source movement can be identified by examining the results of moving window cross correlation.

[4] Authors have interpreted the velocity changes observed using these techniques in a number of ways. Some have appealed to strong-motion-induced damage to explain velocity reductions following large earthquakes [e.g., *Schaff and Beroza*, 2004; *Rubinstein and Beroza*, 2004a, 2004b, 2005; *Peng and Ben-Zion*, 2006]. Others have interpreted the velocity changes to be a result of the closure/opening of cracks caused by the change in static stress field following an earthquake [e.g., *Nishimura et al.*, 2000]. For third set of observations, a damage to the fault zone itself is believed to be responsible for velocity

[1]Department of Geophysics, Stanford University, Stanford, California, USA.
[2]Now at Department of Earth and Space Sciences, University of Washington, Seattle, Washington, USA.
[3]Research Center for Prediction of Earthquakes and Volcanic Eruptions, Tohoku University, Aoba-ku Sendai, Japan.

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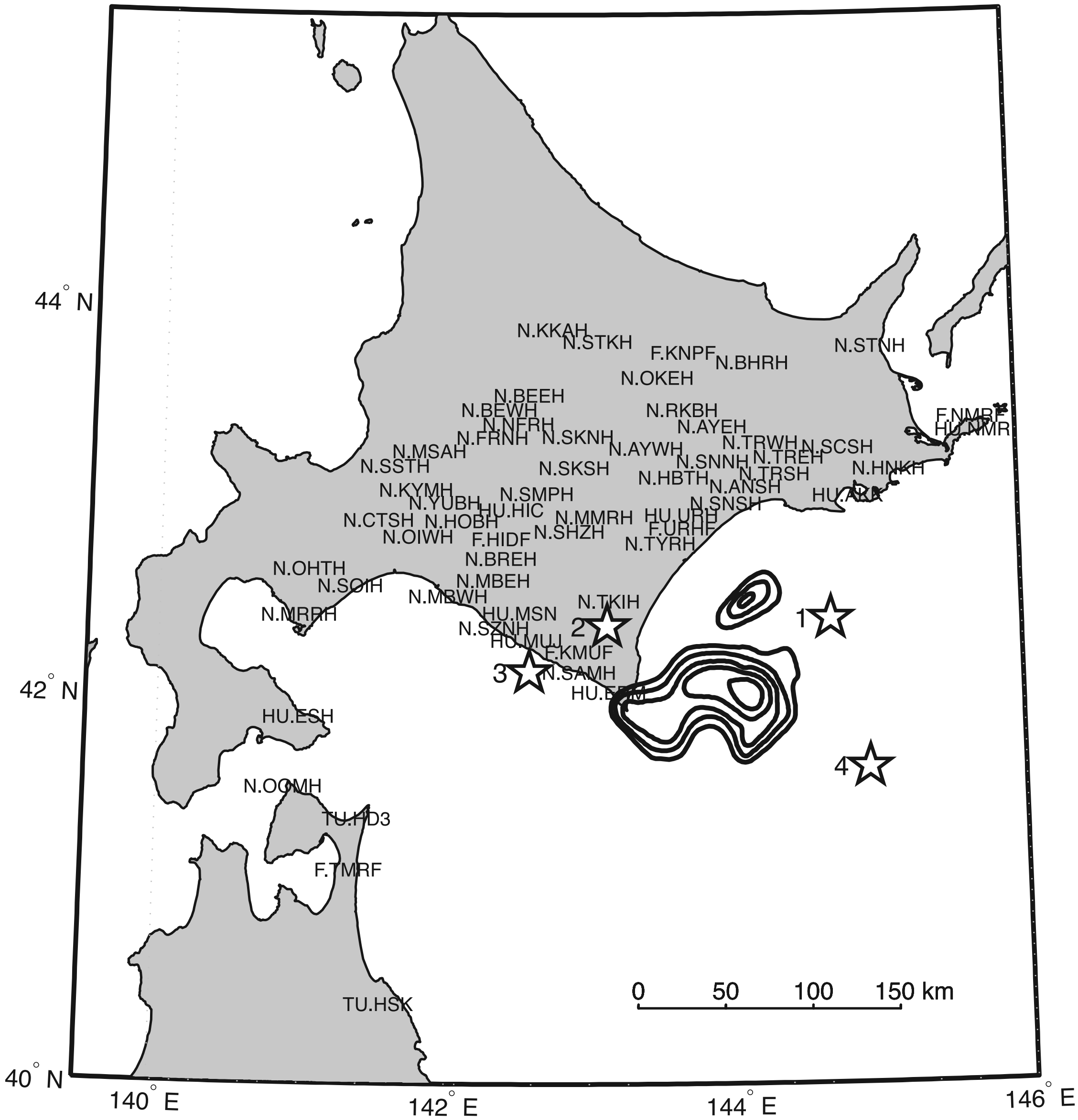

Figure 1. Map of Hokkaido and Northern Tohoku showing station locations, location of repeating earthquakes (stars: number indicates repeat number), and approximate slip distribution of the Tokachi-Oki earthquake (black contours) [*Yamanaka and Kikuchi*, 2003]. For the stations, the prefix before the period indicates the network: N (Hi-Net), F (F-Net), TU (Tohoku University), and HU (Hokkaido University).

reductions coincident with an earthquake [e.g., *Li et al.*, 1998, 2003].

[5] In this study we have a unique opportunity to observe multiple regions of velocity reductions associated with M_w 8.0 2003 Tokachi-Oki earthquake. We use search for a linear increase in delays in the coda, as described above, because accurate origin times are unavailable and the *S* arrivals are often emergent, making differential *S-P* times undependable. With this method, we observe the influence of velocity reductions at the near surface and also near the rupture zone of the Tokachi-Oki earthquake. Our results are consistent with previous observations of nonlinear strong ground motion (site effects). We also see evidence of similar velocity reductions due to either fault zone damage or multiple scattering near the Earth's surface along the source-receiver path.

2. Data

[6] We study four repeating microearthquake sequences (multiplets) in the Hokkaido region of Japan to monitor the influence of the 26 September 2003 Tokachi-Oki earthquake on seismic velocity (Figure 1, Table 1). These four repeating earthquake sequences represent a subset of 349 repeating earthquake sequences identified near the Tokachi-Oki earthquake *Uchida et al.* [2005]. All the repeating earthquake sequences we study are characterized by high coherence values that exceed 0.95 over a broad range of

Table 1. Timing of Repeating Earthquake Sequences Used

Repeating Earthquake Sequence	Event Number[a]	Date	Days After Tokachi-Oki	Magnitude (M_{JMA})	Depth, km	Stations Studied/ Stations Available[b]
1	1r	27 September 2003	1	3.9	39	16/77
	2	31 December 2003	96			
2	1r	14 February 2001	−954	3.2	51	17/56
	2	29 October 2003	33			
3	1	23 August 2001	−764	3.6	60	53/78
	2r	10 September 2003	−16			
	3	03 November 2003	38			
4	1r	24 September 2003	−2	3.5	45	15/69
	2	25 October 2003	29			
	3	04 April 2004	191			

[a]The "r" in "event number" column indicates that the event was used as a reference event. "Stations studied" indicates the number of stations for which we computed a velocity change induced by Tokachi-Oki for the corresponding repeating earthquake sequence.
[b]"Stations available" indicates the number of stations that recorded both the reference event and the "data" event (i.e., the event just after Tokachi-Oki for sequences 2–4, and event 2 for sequence 1). The difference in the number of stations available and stations studied represents the number of stations that were eliminated due to poor signal-to-noise or low correlation values.

frequencies (1–8 Hz) for 40 s long seismograms at a minimum of two stations. The high coherence values ensure that the events are colocated. Were the events not colocated, the coherence values for the 40 s long windows that contain both the *P* and *S* arrivals would be low because the *S-P* times would be different. A second line of reasoning supports our argument that the evens are colocated. The strict criteria used to determine the catalog of repeating earthquakes ensure that the member events of every repeating earthquake sequence have near-identical waveforms. This, in turn, guarantees that the earthquakes are colocated, as one needs identical source processes and Green's functions (which depend strongly on location) to produce identical waveforms [*Poupinet et al.*, 1984]. All these events are located on or near the plate interface, which is dipping to the northwest. The depths determined from the Japan Meteorological Agency (JMA) catalog for events 1–4 are: 39, 51, 60, and 45 km, respectively. JMA reports the uncertainties in absolute earthquake locations in this region to be on the order of 3 km. We have identified repeating events that, according to the JMA catalog, are 10 km apart. This suggests that the uncertainty in the absolute locations determined by JMA in this region is closer to 10 than 3 km, which is not surprising given the difficulty inherent in locating earthquakes offshore.

[7] We examine the recordings of these earthquakes using seismometers from four different seismic networks. Two networks are university networks: the Hokkaido University seismic network and the Tohoku University seismic network. We also use data from the high sensitivity seismograph network (Hi-Net) [*Obara et al.*, 2005; *Okada et al.*, 2004] and the broadband seismograph network (F-Net) [*Okada et al.*, 2004], which are administrated by the National Research Institute for Earth Science and Disaster Prevention (NIED). The instruments within the university networks are high-gain, short-period instruments and are typically located within vaults although some instruments are placed at the surface. Hi-Net is a network of high-gain, short-period velocity seismometers placed in deep boreholes (>100 m depth). F-Net is a network of broadband seismometers placed in shallow vaults (<50 m depth). All of the instruments within these networks record at a rate of 100 samples per second.

[8] Previous studies of velocity changes induced by large earthquakes have shown that healing of velocity changes is most rapid immediately after the earthquake and slows with time [*Li et al.*, 2003; *Vidale and Li*, 2003; *Schaff and Beroza*, 2004; *Rubinstein and Beroza*, 2004a, 2004b, 2005; *Peng and Ben-Zion*, 2006]. Therefore to ensure that the influence of the Tokachi-Oki earthquake on seismic velocities would be large, we select the multiplets that repeat within 2 months after the Tokachi-Oki main shock. The timing of the events in each repeating earthquake sequence is shown in Table 1 and is described in the following section.

3. Method

3.1. Moving Window Cross-Correlation Analysis

[9] In this paper we examine only the vertical component of the seismograms as we have found them to be more stable, likely because there is less noise on this channel. Prior to any data analysis, the seismograms in this data set are normalized and causally filtered with a band-pass window of 1–10 Hz. For each repeating earthquake sequence at each station, we then use cross correlation to align the seismograms to subsample precision to a manually picked *P* arrival. Once all the events within an individual sequence are aligned, we use a moving window cross-correlation technique to measure the relative arrival time of seismic phases of one event relative to a reference event in the same repeating earthquake sequence (Figure 2). As input to the moving window correlation, we use 128 sample windows of the vertical component seismogram weighted by a Hanning function. The windows are stepped forward at increments of 10 samples (0.1 s). We obtain subsample precision to the correlation data by fitting a parabola to the peak of the cross-correlation function [*Deichmann and Garcia-Fernandez*, 1992]. To ensure data quality, we enforce a minimum correlation coefficient of 0.9 for a window around the *P* arrival, a minimum correlation coefficient of 0.8 for the first 20 s of the seismogram, and a minimum signal-to-noise ratio of 4:1. Should a trace fail any of the previous three criteria, we dispose of it and do not analyze it further to ensure high-quality correlations throughout the seismogram. For any individual window that we

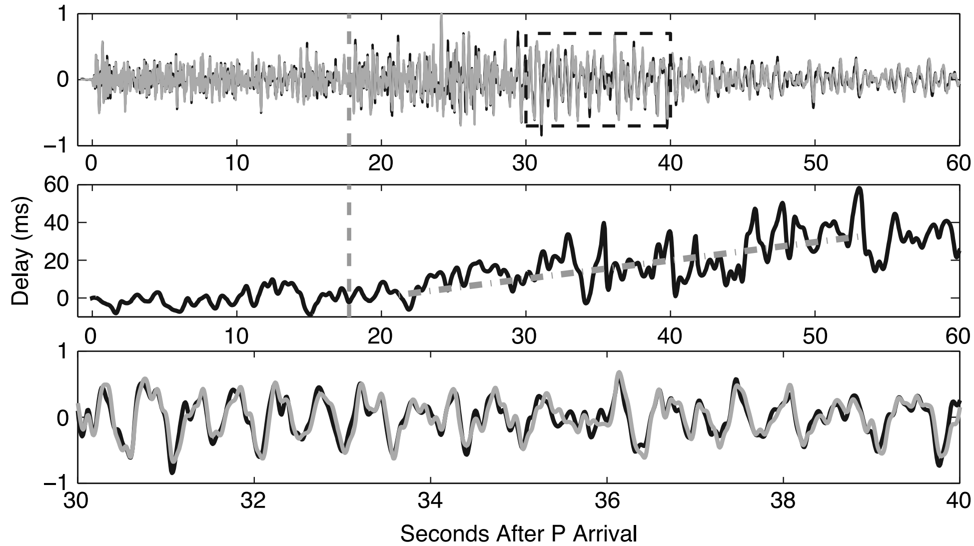

Figure 2. (a) Events 2 (black) and 3 (gray) of repeating earthquake sequence 3 recorded by Hi-Net station SNSH. *S* arrival is indicated by the vertical dashed line. Note that the seismograms align nicely at the *P* arrival but separate further into the seismogram, in particular, after the *S* arrival. Region within the dashed black box is shown at larger scale in Figure 2c. (b) Delay of event 3 relative to event 2 (black). Line fit to delay within a manually selected window (dashed gray).

correlate on, we use 0.7 as a minimum threshold for the correlation coefficient. Any delay values associated with correlation coefficients less than 0.7 are not included in our further processing. We choose a threshold of 0.7 because previous studies have shown that delay values determined using cross correlation provide useful information for earthquake location for correlation coefficients as low as 0.7 [*Schaff et al.*, 2004].

3.2. Determining Velocity Changes

[10] Using moving window cross correlation, we find that, following the Tokachi-Oki earthquake, the delays at many stations increase linearly as a function of time into the seismogram shortly after the *S* wave arrives (for example, Figure 2b). *Schaff and Beroza* [2004] also identified a linear increase in delays with time into the coda for the 1984 Morgan Hill and the 1989 Loma Prieta earthquakes. *Niu et al.* [2003] and *Snieder et al.* [2002] have shown that if the velocity within a medium has been reduced, as later phases within the coda spend progressively longer amounts of time within this altered medium, the seismogram should be stretched and delays should increase linearly as a function of time into the seismogram. To interpret these delays, we follow the method of *Schaff and Beroza* [2004]. We first manually select a window of the delay function, specifically where the delays appear to increase/decrease linearly as a function of time. For those stations where significant delays are not apparent, we still select a window of the delay function, such that we can quantitatively identify those stations where there is an insignificant accumulation of delays.

[11] The data within these windows must satisfy the signal-to-noise criterion discussed above and have a minimum length of 5 s. The windows average approximately 30 s in length but exceed 50 s for some stations. We then fit a line to the delays within this window, the slope of which has been shown to represent the fractional change in slowness (1/velocity) (Figure 2b) [*Schaff and Beroza*, 2004]. This fractional change in slowness represents a path-averaged value, meaning that these changes may be localized and much larger. Throughout the rest of this paper, for the ease of the reader, we refer to velocity changes instead of changes in slowness. To convert our measurements of slowness changes into velocity changes, we must assume a nominal path-averaged velocity in the medium where the velocity has been reduced. Because we believe that the velocity changes accumulate in the near surface, we choose a nominal reference velocity of 1000 m/s.

[12] Our method of fitting a line to the delays allows us to estimate error bars on the velocity change based upon the formal standard errors of the slope from the line fit. These errors do not take into account all the potential sources of error. For example, a major source of uncertainty arises from window selection. Windows are selected where delays are seen to vary linearly with time, i.e., where the data is most consistent with our model. We also see significant deviations from the linear trend in the form of peaks and valleys in delays (for example, Figure 2). Numerical experiments have shown that sudden, short-lived peaks in the delay function may represent a movement of scatterers rather than a change in velocity [*Niu et al.*, 2003]. These sharp peaks and valleys are often observed when the correlation coefficient is particularly low, which suggests that the delay values they provide may be unreliable.

3.3. Choice of Reference Events and the Temporal Interpretation of Observations

[13] The delay values that are determined by cross correlation describe changes in the medium, which occurred at some point in time between the two repeating events being correlated. Because we are interested in changes induced by the Tokachi-Oki earthquake, we try to minimize the time between our reference events and our data events, such that any changes we observe are likely a result of the

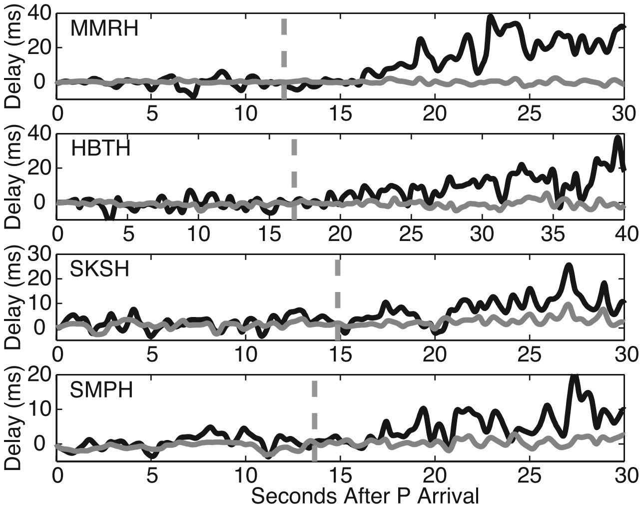

Figure 3. Delay plotted as a function of time for events in repeating earthquake sequence 3 relative to event 2 in the same sequence at Hi-Net stations MMRH, HBTH, SKSH, and SMPH. The gray line indicates event 3 (03 November 2003). The black line indicates event 1 (23 August 2001). *S* arrival is indicated by the vertical dashed line.

Tokachi-Oki earthquake. For repeating earthquake sequences 3 and 4, our reference events are events 2 and 1, which occur 16 and 2 days, respectively, prior to the Tokachi-Oki earthquake. In sequences 3 and 4, the next event occurs 5 and 4 weeks, respectively, after the Tokachi-Oki earthquake. The short window of time between the reference and data events for these sequences indicates that the Tokachi-Oki earthquake is likely responsible for any large differences observed between the reference events and these data events. There is more uncertainty in the timing of delays observed using repeating earthquake sequence 2. For this sequence, there is an event that occurs approximately 5 weeks after Tokachi-Oki, but its reference event occurs in 2001, over 2 1/2 years prior to Tokachi-Oki. If seismic velocities between 2001 and 2003 remained constant, the measurements from repeating earthquake sequence 2 should reflect Tokachi-Oki-induced changes. Considering that there were no major tectonic events within the region from 2001 to 2003 (prior to Tokachi-Oki), we expect that the velocities remained constant and that any changes in wave propagation that we observe using this repeating earthquake sequence are a result of the Tokachi-Oki earthquake. The delays of event 1 (23 August 2001) relative to event 2 (10 September 2003) in repeating earthquake sequence 3 offer further evidence that regional velocities remained constant from 2001 to 2003, as we do not observe significant delays between these two events (Figure 3). The lack of changes in seismic velocity from 2001 to shortly prior to Tokachi-Oki indicates that the measurements made using repeating earthquake sequence 2 should be comparable to those from multiplets 3 and 4, which had data events at approximately the same time as multiplet 2. The effect that we observe with repeating earthquake sequences 2–4 should be considered a lower bound on the coseismic reduction of seismic velocity induced by the Tokachi-Oki earthquake. It is a lower bound because the velocity reductions are shown to heal with time after the main shock (Figures 4 and 5), and these repeating earthquake sequences have no events within the first month after the Tokachi-Oki main shock, so velocity reductions induced by the earthquake probably healed significantly by the time any of these repeating earthquake sequences repeated after the Tokachi-Oki earthquake.

[14] The fourth repeating earthquake sequence we consider in this paper does not have a repeat prior to Tokachi-Oki. This means that we cannot directly observe the delays caused by Tokachi-Oki. Instead, we observe the healing of Tokachi-Oki-induced velocity reductions by comparing an event the day after Tokachi-Oki and a repeat that occurred 3 months later (Figure 4). As a result, we see negative delays that reflect a velocity increase. The percentage increase in velocity from September to December 2003 that we observe using repeating earthquake sequence 1 is often larger than the percentage decrease in velocity we observe at the same stations using repeating earthquake sequences 2–4 (Figure 6). It is not surprising that we observe larger changes in velocity with multiplet 1 than the others because the repeats within events 2–4 occur 1 month after the Tokachi-Oki earthquake, allowing for a significant amount of healing before the effects of the velocity changes can be measured. As a result, the velocity change measurements we compute for repeating earthquake sequence 1 are useful for understanding the spatial behaviors of the delays, but they should not be compared in amplitude to velocity change measurements made for repeating earthquake sequences 2–4. For this reason, when we compute the mean velocity changes at a station, we do not include those determined from repeating earthquake sequence 1 in these computations.

4. Observations

[15] The Tokachi-Oki earthquake produces a distinct signal that is detected at many stations. Nearly all of our stations show a systematic increase in arrival times in the *S* coda following the Tokachi-Oki earthquake. For those repeating earthquake sequences where we have multiple

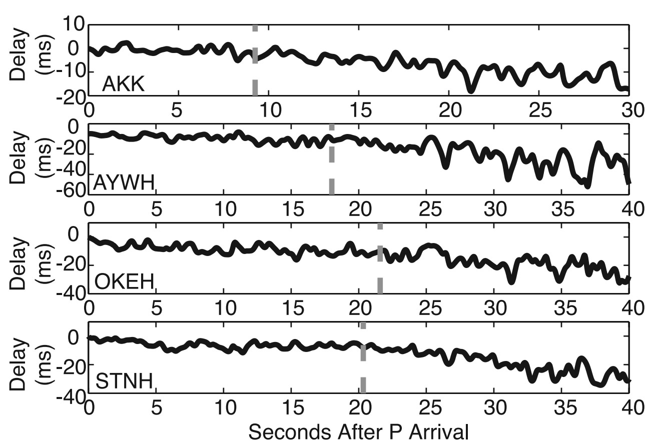

Figure 4. Delay plotted as a function of time for event 2 relative to event 1 within repeating earthquake sequence 1 at Hokkaido University station AKK and Hi-Net stations AYWH, OKEH, and STNH. *S* arrival is indicated by the vertical dashed line. Negative delays indicate that phases are arriving sooner in December than they are in September.

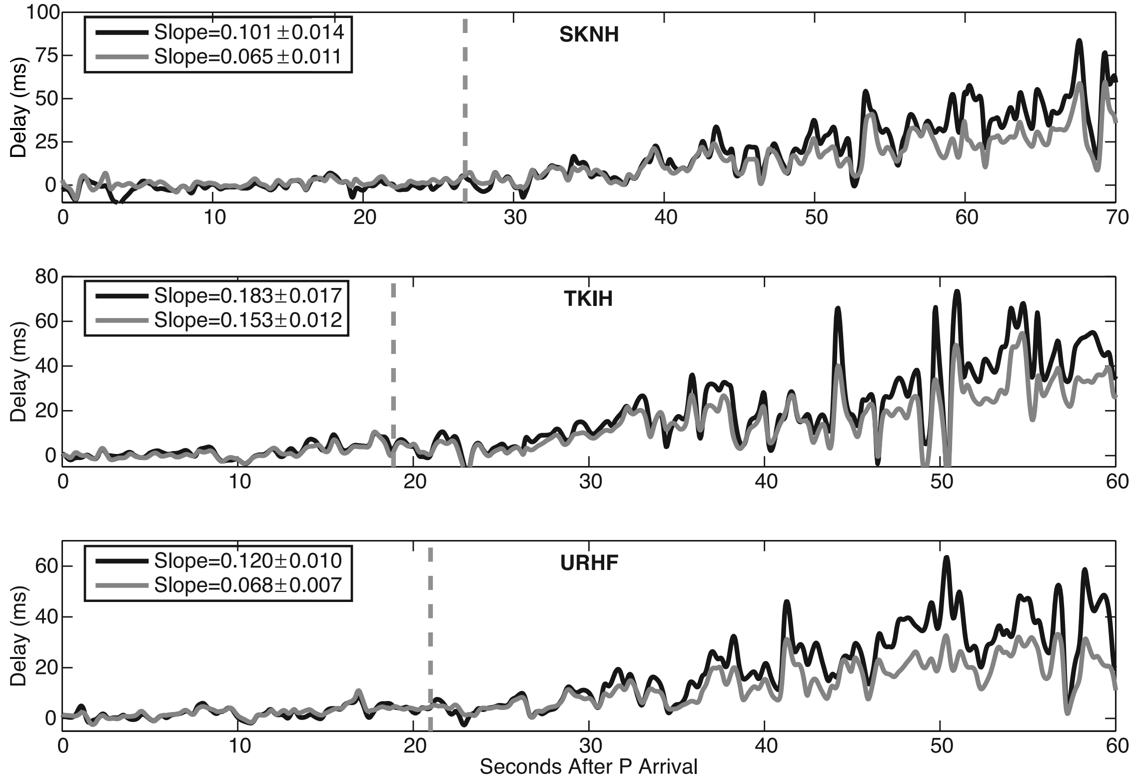

Figure 5. Delay plotted as a function of time for events 2 (gray) and 3 (black) of repeating earthquake sequence 4 at Hi-Net stations SKNH and TKIH and F-Net station URHF. *S* arrival is indicated by the vertical dashed line. The much smaller delays and gentler slope of event 3 relative to event 2 indicate that the velocity reductions caused by Tokachi-Oki are healing. The slopes of the delay function for all of the events are listed in the upper left-hand corner of each panel.

repeats after the Tokachi-Oki earthquake (1 and 4), we see that the delays decrease with time after the main shock. This suggests that the seismic velocity reductions caused by Tokachi-Oki heal with time after the main shock (Figures 4 and 5). Interestingly, we do not typically observe any significant change in the arrival times of the *P* coda (Figures 2 and 3). Although the general response to the Tokachi-Oki earthquake is consistent from station to station, in that they reflect a decrease rather than an increase in velocity, the size of the velocity changes varies significantly from station to station and from repeating earthquake sequence to repeating earthquake sequence (Figures 6 and 7). Below we search for systematic controls on this variation to localize where the velocity reductions are occurring.

[16] If we examine a map showing the average decrease in velocity caused by the Tokachi-Oki earthquake plotted at all our stations, there is a very striking feature in that the velocity reductions are largest in the east and west of Hokkaido, while the reductions in velocity in the central region of Hokkaido are much smaller (Figure 7). This correlates very nicely with the topography of Hokkaido; the central region of Hokkaido is very mountainous, while the regions to the east and west of it are plateaus (Figure 7). In fact, there appears to be a threshold of approximately 100 m altitude; almost all the largest velocity changes are observed at stations of 100 m altitude or less (Figure 8). This correlation may be a result of our measurement technique, which relies on scattering within a medium with reduced seismic velocities. Any differences in the media where the scattering occurs (for example, location, rock type, age, etc.) will translate into differences in the observed velocity changes. For our study there are significant differences wherein the coda is generated. The majority of the scattered energy in the Hidaka Mountains (southern end of the central mountain region of Hokkaido) appears to be scattering at depths of 50 km or greater [*Taira and Yomogida*, 2004], while the scattering in the eastern plateau region (where the delays are consistently large) has been shown to be occurring at shallow depths and is locally heterogeneous [*Taira and Yomogida*, 2006]. Laboratory and theoretical studies have shown that the susceptibility to damage by strong shaking decreases with increasing compressive stress (increasing depth) [*Ostrovsky et al.*, 2000; *Zinszner et al.*, 1997]. This indicates that the strong shaking of the Tokachi-Oki earthquake is unlikely to significantly change the elastic properties of rocks at 50 km depth (those generating the coda for the mountain region), while near-surface materials will be much easier to damage (those generating the coda for the plateau regions). There is an alternative explanation for the correlation of the velocity reductions and the station altitude, which relates to the local geology. It is probable that the rocks within the mountains are harder than those elsewhere and are thus less subject to damage from the Tokachi-Oki earthquake.

[17] We also note that the influence of Tokachi-Oki is particularly small in the north of Honshu (Tohoku area). The lack of changes in Tohoku is expected, as these stations are farther from the main shock than the majority of the

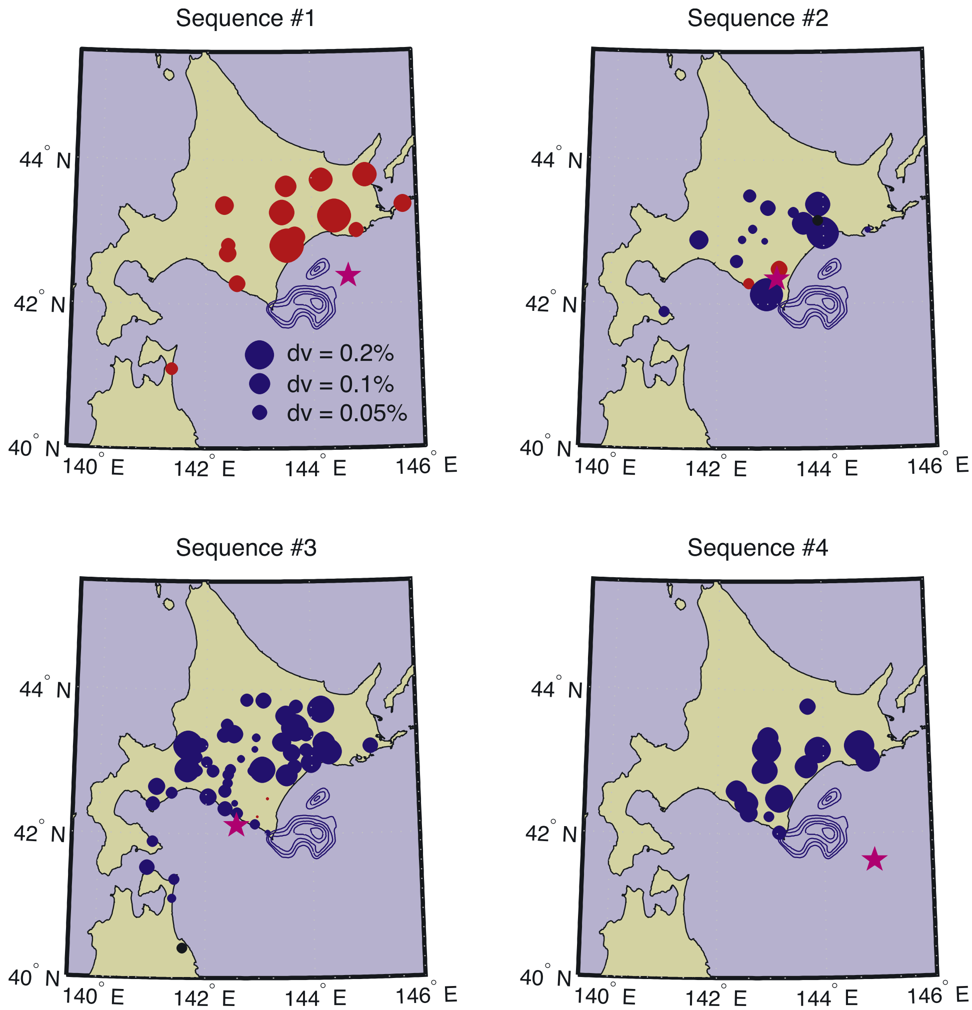

Figure 6. The reduction in velocity at each station caused by the Tokachi-Oki earthquake is plotted as a circle at the location of each station. Each panel represents a different repeating earthquake sequence and the associated observations with it. The size of the circle indicates the amount by which the path-averaged velocity decreased (increased) at that station for stations marked in blue (red). The scale is shown in the upper left panel, which depicts the delays observed using repeating earthquake sequence 1. The black circles represent stations where the error bars to the change in velocity indicate that the change in velocity is not significantly different from zero at the 95% confidence level. Pink stars indicate the location of the repeating earthquake sequences. The circles for repeating earthquake sequence 1 are all red because these measurements reflect the healing from 27 September to 31 December 2003, and therefore the velocity should be increasing.

stations on Hokkaido, and the influence of the Tokachi-Oki earthquake is likely to be smaller there.

5. Constraining the Source Region of the Delays

[18] Before determining a mechanism for the observed decreases in seismic velocity associated with Tokachi-Oki earthquake, we attempt to localize the region in which the delays are accumulating. The possible source regions for these delays are near the earthquake sources, in the path between the earthquakes and the receivers, near the receivers, or some combination of the three. We have previously argued that a significant percentage of the delays are a result of scattering within a volume near our receivers. Below, we offer further evidence that a large percentage of the delays are accumulating near the receivers. We also find evidence that substantial delays are accumulating in the path between the sources and the receivers. We rule out near source effects as a significant source of the delays that we observe. If delays were accumulating near the repeating earthquake sources, we would expect that they would be approximately constant at all of our stations for any individual repeating earthquake sequence. We see very large variations in the velocity reductions at different stations for the same repeating earthquake source (Figure 6). This wide range in the velocity reductions for any repeating earthquake sequence indicates that any delays that are accumulating near the repeating earthquake sources are much smaller than those accumulating elsewhere. As such, velocity reductions near

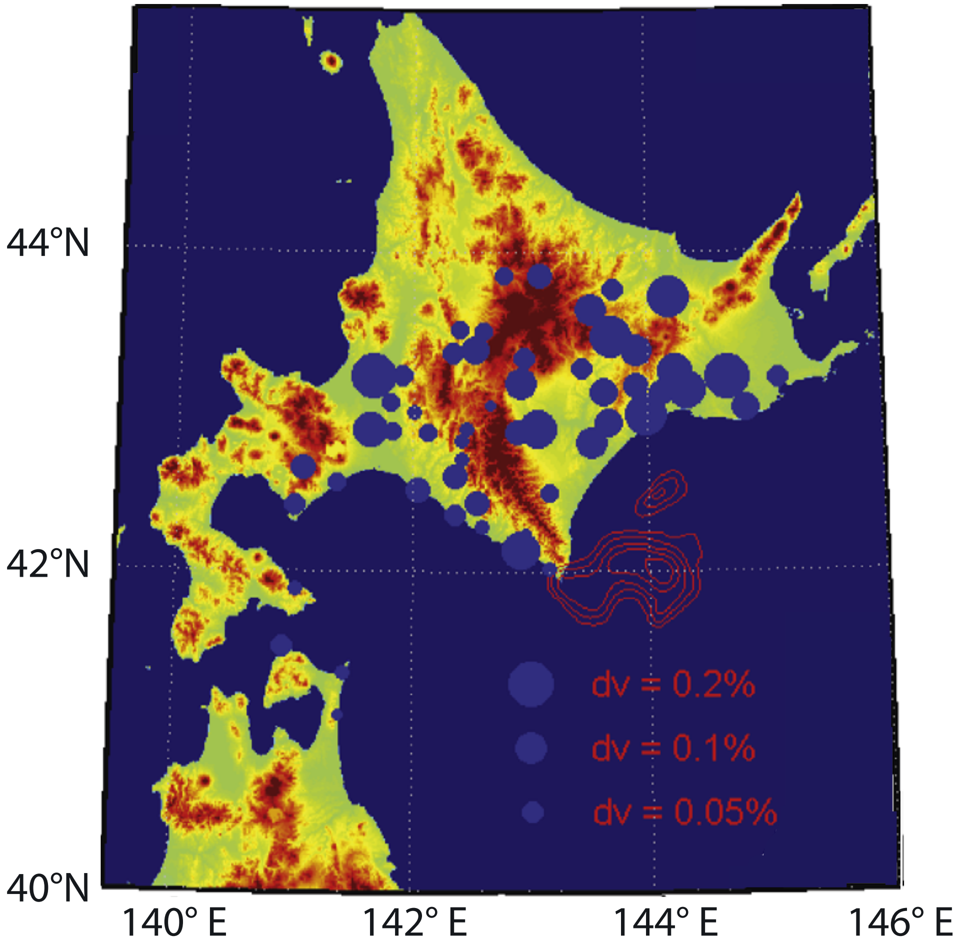

Figure 7. The average decrease in velocity at each station caused by Tokachi-Oki. This averages over all repeating earthquake sequences (within multiplets 2–4) available at any individual station. The number of observations for each station varies, so this may affect the relative values of the delays.

the earthquake sources should be considered an insignificant source of the delays that we observe.

5.1. Site Effects

[19] We believe that a large percentage of the delays are accumulating very close to the sites that we are observing them at. We have shown previously that the delays we observe are much smaller in the mountainous regions than they are in the plateau regions (Figure 7). We believe this to be the result of much of the coda being generated at great depths for the mountainous regions, where high compressive stresses prevent damage induced by strong ground motion. Mountainous regions are also more likely to have harder rocks, which are less likely to be damaged (and have their seismic velocities temporarily reduced) than the softer rocks found in plateaus and elsewhere. We can observe this effect if we compare site-specific amplification factors to the observed velocity reductions (Figure 9). The most convincing evidence, though, that a large percentage of the delays are accumulating near the site is the strong variation in velocity reductions with location. We observe very large difference in the velocity changes observed at nearby sites (Figures 6 and 7). At nearby sites, the difference in path and source effects should be negligible, so any

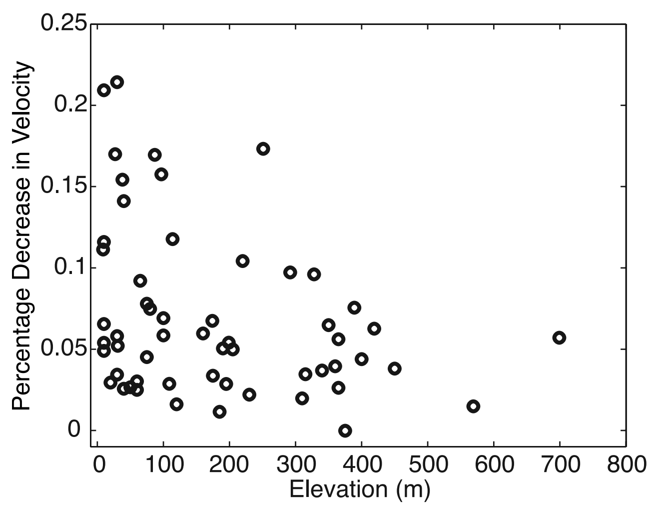

Figure 8. The average decrease in velocity at each station plotted against elevation. Elevation is determined at the borehole adit (for borehole stations) or surface elevation (surface stations). The intent is to understand the influence of local topography on velocity reductions and our method.

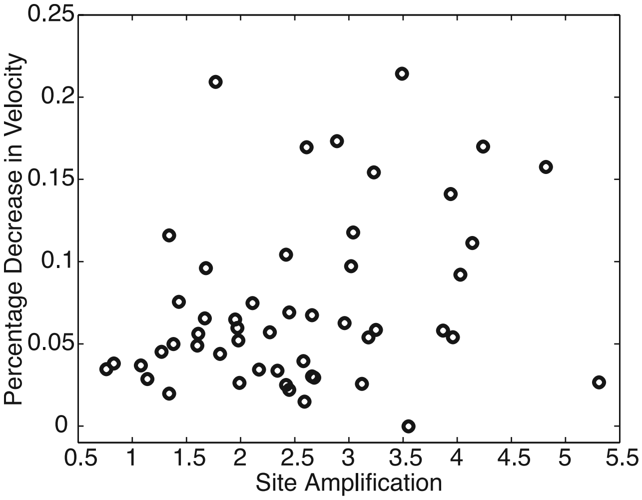

Figure 9. Mean velocity decrease at each station plotted against peak ground velocity (PGV) amplification factors. Site amplification factors were determined based on the ratio of observed peak amplitudes versus theoretical peak amplitudes computed based on magnitude and hypocentral distance [*Watanabe*, 1971] for a suite of 64 $M > 3.5$ earthquakes within 900 km of Hokkaido. Amplification factors were not determined for the six F-Net stations used in this study, and so these six data points are not plotted.

difference between sites should reflect differences in their site responses.

5.2. Path Effects

[20] As noted above, the strength of the observed velocity reductions is strongest in the plateau regions in the east and west of Hokkaido, while velocity reductions in the central mountainous region of Hokkaido are significantly smaller. The strength of observed velocity reductions in the central mountainous region of Hokkaido appears to vary depending upon the source locations. Using the multiplets that are located downdip of the Tokachi-Oki rupture (sequences 2 and 3), we find that the velocity reductions in the central region of Hokkaido are much smaller than they are in the plateau regions to the east and the west (Figure 6). Making this same comparison using the multiplets that occur to the east and southeast of the main shock rupture (repeats 1 and 4), we find that the velocity changes in the central region are much more comparable in magnitude to the plateau areas. We believe that the difference in velocity changes for the central mountainous region for repeating earthquake sources in different locations is related to the path that the waves take to the stations. For multiplets 1 and 4, the paths to the stations in the central region cross the rupture zone of the Tokachi-Oki earthquake, while for sequences 2 and 3, the paths to the central region do not cross regions of high slip (Figure 1). The difference between the two paths is illustrated clearly if we compare the velocity changes for repeating sequences 2–4 observed by stations in this region where some paths cross the rupture zone (paths from repeat 4) while others do not (paths from repeats 2–3) (Figure 10). Velocity changes for paths that cross the rupture zone of Tokachi-Oki are consistently larger than those that do not, strongly suggesting that delays are not just accumulating near the receiver for paths that cross the rupture zone. For those paths that cross the rupture zone of Tokachi-Oki, we expect that a significant percentage of the delays that we observe are accumulating in the region immediately surrounding the main shock rupture.

[21] Further evidence supporting the argument that some stations are seeing delays that accumulate both at the site and in the region surrounding Tokachi-Oki can be found by comparing repeating earthquake sequences 1 and 4. For repeating earthquake sequence 1, the velocity reductions in the central region are approximately 50% the size of the velocity reductions to the east, while for repeating earthquake sequence 4, the velocity reductions in the central region of Hokkaido are equivalent in amplitude with those to the east (Figure 6). The waves from repeating earthquake sequence 4 traverse the entire width of the rupture zone and the region of highest slip en route to the central region of Hokkaido, while the waves from repeating earthquake sequence 1 cross a smaller portion of the rupture zone (Figure 1). This suggests that both the amount of time the waves spend crossing the rupture zone of the Tokachi-Oki earthquake and the amount of slip on the rupture zone in which the waves are traveling are important in determining the amplitude of velocity changes observed at the stations in the central region.

6. Physical Mechanism for Seismic Velocity Reductions

[22] As discussed in the previous section, we appear to be observing velocity reductions in the near surface near our receivers. For those paths that cross the rupture zone of the Tokachi-Oki earthquake, we also see velocity reductions somewhere in the path between our repeating earthquakes and receivers, likely near the Tokachi-Oki earthquake. In

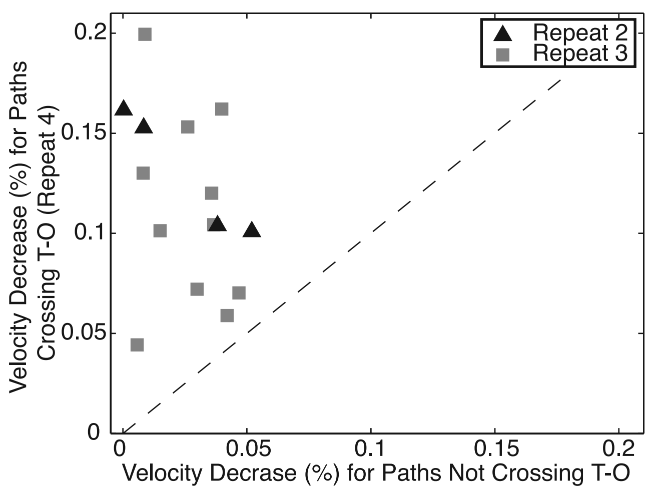

Figure 10. Velocity decrease for repeating earthquake sequence 4 for stations where the path crossed the Tokachi-Oki earthquake plotted against the velocity decreases observed at those same stations using repeats 2 (black triangles) and 3 (gray squares). We consider paths for repeating earthquake sequence 4 to cross the Tokachi-Oki earthquake if the station is located to the west of 144°E. A one-to-one line is provided as reference.

this section we attempt to determine the physical mechanisms responsible for both sets of velocity reductions. We believe that the velocity reductions that are accumulating near the receivers are a result of nonlinear strong ground motion in the Tokachi-Oki earthquake, i.e., strong ground motion damaging near-surface materials resulting in velocity reductions. The cause of the velocity reductions that we believe to be in the region of the Tokachi-Oki rupture zone is less clear. To explain the velocity reductions near the Tokachi-Oki rupture zone, we posit two hypotheses: (1) the linear increase in delays that we observe for these paths may result from delays accumulated by phases that reflect multiple times off the surface (where velocities will have been reduced by nonlinear strong ground motion), and (2) the linear increase in delays results from multiple scattering within the plate interface in the rupture zone itself.

6.1. Site Effects

[23] To explain the reductions in seismic velocity caused by the Tokachi-Oki earthquake, which appear to be localized to the near surface near our sites, we appeal to nonlinear strong ground motion as the cause of the velocity reductions. Nonlinear strong ground motion has been widely documented for many earthquakes, including the 1994 Northridge earthquake [*Field et al.*, 1997], the 1999 Chi-Chi earthquake [*Huang et al.*, 2005], the 1995 Kobe earthquake [*Aguirre and Irikura*, 1997], and the 2003 Tokachi-Oki earthquake [*Yamanaka and Kikuchi*, 2003]. The physics and the effects of nonlinear strong ground motion have been thoroughly reviewed in the study of *Beresnev and Wen* [1996] and more recently in the work of *Ostrovsky and Johnson* [2001]. Nonlinear strong ground motion is characterized by the strong shaking of large earthquakes resulting in damage (the growth and/or opening of microcracks) to geomaterials. This damage results in a decrease in the elastic moduli of the medium and, therefore, a decrease in the seismic velocities.

[24] The strength of nonlinearity and its damage are expected to be positively correlated with the strength of shaking [*Guyer et al.*, 1998; *Ostrovsky et al.*, 2000; *Rubinstein and Beroza*, 2004a, 2004b, 2005; *Peng and Ben-Zion*, 2006]. Our observations parallel this expectation, as there only appear to be large velocity reductions at stations that experienced strong shaking from the Tokachi-Oki earthquake with peak accelerations exceeding 100 cm/s^2 (or approximately 10% the acceleration of gravity) (Figure 11). The shaking parameter we cite is a simple interpolation of the NIED Kyoshin Network (K-net) [*Kinoshita*, 1998; *Okada et al.*, 2004] strong motion observations of the Tokachi-Oki earthquake, and any site effects specific to the K-net sites are included and those specific to our observation sites are not; hence they are subject to considerable uncertainty. Further uncertainty arises from the fact that many sites are quite distant from the closest K-net site. We have previously shown that the velocity reductions caused by the Tokachi-Oki earthquake correlate with site amplifications (Figure 9). This behavior is also predicted by nonlinear strong ground motion, as softer materials are more susceptible to damage by nonlinear strong ground motion than hard materials [*Van Den Abeele and Van de Velde*, 2000; *Rubinstein and Beroza*, 2004a, 2004b].

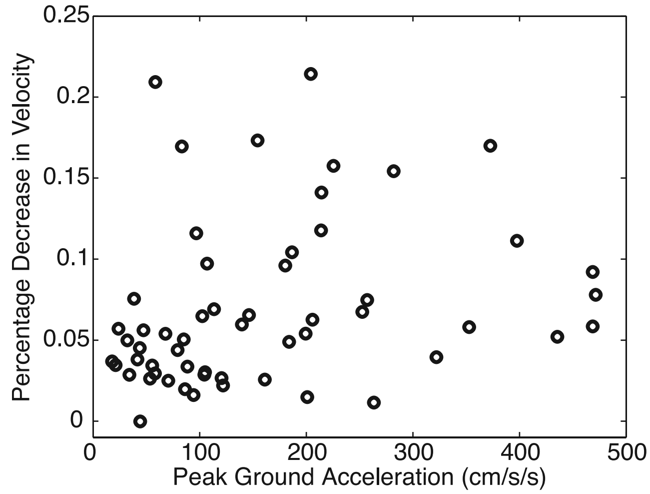

Figure 11. Mean velocity decrease at each station plotted against peak ground acceleration in Tokachi-Oki earthquake. Site-specific PGA values are interpolated from the PGA values determined at K-net stations. These values reflect the peak acceleration on one component. The peak vector acceleration, although probably more directly correlated to our observations, is not readily available. Furthermore, the value we examine and the peak vector acceleration are likely to be closely related.

[25] One other prediction of nonlinear strong ground motion is that damage caused by it should be limited to the very near surface as the susceptibility to nonlinearity decreases with increasing compressive stress and, hence, with increasing depth [*Zinszner et al.*, 1997; *Ostrovsky et al.*, 2000]. We do not find such a correlation (Figure 12). We expect that there is a depth dependence to the damage for our sites, but our method of measuring velocity changes cannot resolve this detail. Our method is unable to resolve the depth dependence of the damage caused by Tokachi-Oki because the scattering volume that is contributing to the coda likely extends deeper than our deepest stations. Therefore the delays at any site should be relatively depth independent.

[26] Since we are unable to constrain the depth to which damage is occurring using our data set, we are forced to appeal to evidence from other studies. On the basis of the studies of *Sawazaki et al.* [2006] and *Rubinstein and Beroza* [2005], we argue that the delays are accumulating in the upper 100–150 m. *Sawazaki et al.* [2006] found a time dependence of site response (which is indicative of nonlinear strong ground motion) in the upper 150 m coincident with the 2003 Tokachi-Oki earthquake. *Rubinstein and Beroza* [2005] examined the change in *S-P* times for surface and borehole sites caused by the 2004 Parkfield earthquake, and they showed that that damage caused by strong shaking was limited to depths less than 100 m. We are unable to use this technique with this data set because the *S* arrivals at many of our sites are very emergent.

6.2. Path Effects

[27] We are less certain of the physical mechanism responsible for the delays that seem to accumulate along the paths that cross the Tokachi-Oki rupture zone. We offer

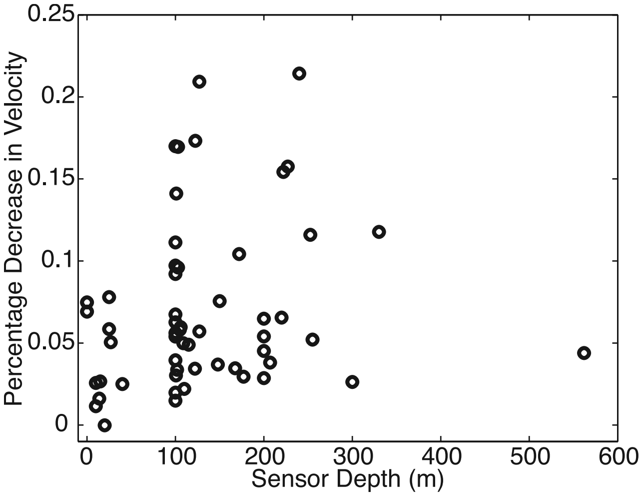

Figure 12. Mean velocity decrease at each station plotted against sensor depth.

two hypotheses for the source mechanism of these velocity reductions.

[28] We first suggest that a damage induced by nonlinear strong ground motion at the near surface above Tokachi-Oki is resulting in the delays that we observe. Most likely, the shaking of the Tokachi-Oki earthquake was strongest at the near surface, directly above the rupture. Because this region experienced the strongest shaking, the strong-motion-induced damage will likely be largest. This damage will likely be particularly large because the ocean sediments which overlie the rupture zone are probably poorly consolidated and therefore particularly subject to nonlinearity. Therefore we expect velocity reductions be quite large at the near surface above the Tokachi-Oki to. While it is particularly likely that the damage induced by the strong motion of the Tokachi-Oki earthquake was particularly strong directly above the rupture zone, for these velocity reductions to be observed as a linear increase in delays at our stations onshore, a significant percentage of the *S*-coda energy recorded at our seismometers onshore must be multiply reflected at the near surface in this region.

[29] In our second hypothesis we suggest that delays are accumulating within the rupture zone of the Tokachi-Oki earthquake. This model requires that there should be significant velocity reductions within the rupture zone. Studies of strike-slip earthquakes in California have provided evidence that earthquake rupture can result in significant velocity reductions within a narrow region surrounding the rupture [*Li et al.*, 1998, 2003, 2006]. For fault zone velocity reductions to result in the linear increase in delays that we observe, it requires that much of the energy passes through this region and scatters repeatedly. It is likely that the energy from these repeating earthquakes does pass through the rupture zone of the Tokachi-Oki earthquake, as both the repeating events and the Tokachi-Oki earthquake are located on the plate interface, with the main shock located downdip from the smaller, repeating events. The distances between the repeating earthquakes and receivers range approximately between 100 and 250 km, which suggests that a significant percentage of the energy arriving at the stations will have departed the earthquake in a down-going direction. We can then infer that a significant portion of the energy from the repeating earthquakes will cross the Tokachi-Oki rupture zone.

[30] The waveguide nature of fault zones could allow for multiple scattering, and a linear increase in delays could permit for fault-zone-trapped waves. There is an uncertainty, though, in how much energy will escape the waveguide and reach the surface. There is evidence that the subduction interface in Hokkaido is a large source of scattered energy in the coda albeit at very high frequencies (14–18 Hz) [*Taira and Yomogida*, 2003]. Thus the plate interface (and therefore the rupture zone of the Tokachi-Oki earthquake) is a potential source for the scattered energy we are observing late into the *S* coda and therefore is a potential location where the delays could be accumulating.

6.3. Fluids?

[31] While we find significant changes in the arrival times of the *S* coda for both path and site effects on the seismic velocities, we do not find significant changes in the *P* arrival times (for example, Figures 2 and 3). This may suggest that fluids have somehow been introduced, as the inability of fluids to support shear will result in large *S* velocity reductions (and therefore delays). The effect of fluids on *P* velocity is less straightforward; depending on conditions, the introduction of fluids can result in an increase or decrease in *P* velocity. Here we expect that the *P* velocities are reduced or unchanged, as an increase in *P* velocities would show an advance in the arrival times in the *P* coda, which we do not observe. Assuming that the *P* velocities are reduced by fluids in our case, the reduction in *S* velocity will, nonetheless, be larger, resulting in larger delays in the *S* coda than in the *P* coda. Our observations parallel this and suggest that introduction of fluids could be partially responsible for the observed phenomena.

[32] It may also be that the delays in the *P* coda are more difficult to observe. First, we have less time to observe an accumulation of delays in the *P* coda than we do for the *S* coda. We can only look for changes in the arrival times of the *P* coda until the *S* wave arrives (typically 5–20 s for this data set), whereas for the *S* arrival, we often have 50–60 s of coda to examine. The higher velocity of *P* also makes it harder to observe changes in arrival time than for *S* because the same percentage reduction in *S* and *P* velocities will result in significantly larger delays for the *S*. All this considered, we do occasionally see what appear to be linear increases in delay for the *P* coda caused by the Tokachi-Oki earthquake (for example, Figure 4), but they are minor when compared with the delays observed in the *S* coda.

7. Summary

[33] We have used repeating earthquake sequences near the Tokachi-Oki earthquake rupture to identify time-dependent changes in the seismic velocity. We identify reductions of up to 0.3% in path-averaged velocity caused by the Tokachi-Oki earthquake. Our varied recording geometry allows us to identify velocity reductions both near the receivers and in the region surrounding the main shock rupture. The near-receiver velocity reductions we expect are a result of nonlinear strong ground motion. The velocity reductions nearer the rupture we believe are either a result of

damage in the kilometers immediately surrounding the Tokachi-Oki rupture or of damage induced by nonlinear strong ground motion directly above the Tokachi-Oki main shock.

[34] **Acknowledgments.** The authors would like to thank Hiroe Miyake for the assistance with strong motion data and descriptions of site response. The waveform data were provided by Hokkaido University, Tohoku University, NIED Hi-Net, and NIED F-Net. We thank M. Kasahara and M. Ichiyanagi for their cooperation in using data from Hokkaido University seismic network. Strong motion parameterization was provided by NIED K-Net. The comments from A.E. Rodolfo Console and two anonymous reviewers improved this manuscript. This work was supported by USGS grant 05HQGR0007. J.R. was partially supported by the William K. Whiteford Fellowship.

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G. C. Beroza, Department of Geophysics, Stanford University, Stanford, CA, USA.

J. L. Rubinstein, Department of Earth and Space Sciences, University of Washington, Box 351310, 4000 15th Avenue NE, Seattle, WA 98195, USA. (justin@ess.washington.edu)

N. Uchida, Research Center for Prediction of Earthquakes and Volcanic Eruptions, Tohoku University, Aoba-ku Sendai 980-8578, Japan.