Temporal changes of the crustal structure associated with the M6.1 earthquake on September 3, 1998, and the volcanic activity of Mount Iwate, Japan

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Abstract. We detected temporal changes of crustal structure by cross spectrum analyses of seismic waves excited by two artificial explosions that were carried out about one month before and two months after a M6.1 shallow earthquake. Phase spectrum analysis shows that seismic velocity of the upper crust around the focal area of the M6.1 earthquake and Mount Iwate decreased 0.3 – 1.0 % during the three months. Since lager decreases of the seismic velocity were observed at the region close to the focal area and a volcanic pressure source, we conclude that stress change in the upper crust is a most plausible candidate to cause the velocity changes among other candidates. We further apply a coherence analysis to the observed seismic coda to locate regions where short-wavelength heterogeneity temporally changed. The result indicates a possibility of temporal changes of the heterogeneity beneath the volcano, where a magmatic fluid intrusion is inferred.

Introduction

Temporal changes of the structure associated with seismic and volcanic activities are one of the most interesting topics in the earth science, and have been investigated by many previous studies around the world. For example, Poupinet *et al.*(1984), analyzing seismic waveform data of earthquake doublets, and detected a temporal change of the seismic wave velocity related to the Coyote Lake earthquake (M=5.9), and estimated the velocity change in the crust to be 0.2 %. Sato (1988) summarized a lot of studies on temporal changes of the crustal structure in view of seismic-wave attenuation and scattering properties. However, most of the past studies used seismograms of natural earthquakes so that there often remained questions on temporal changes of seismic sources.

Mount Iwate, which is a stratovolcano located in the northeastern Honshu, Japan, has been active since the end of December 1997. Significant crustal deformation and many volcanic earthquakes have been observed since then, and several geophysical studies clarified ascent of magmatic fluid from a deeper region of the eastern portion of the volcano (Tanaka *et al.*, 1999, Ueki *et al.*, 1999). However, no eruptions have yet been reported until now. Under these conditions, a M6.1 earthquake took place near the southwest flank of the volcano (140.910°E, 39.796°N) on September 3, 1998. The earthquake formed a reverse fault with a dip of 41 degree and a strike of 216 degree (Earthquake Information Center, Univ. of Tokyo, 1998), and a part of the fault reached the ground surface at the northeastern part of the focal area (Koshiya and Ohtani, 1999).

About one month before the M6.1 earthquake, a seismic active-experiment using an artificial chemical explosion was carried out at the south of the focal area by a joint work of two groups; the Research Group for Intra-Plate Earthquakes, and the Research Group for Explosion Seismology, Japan. This experiment was designed for exploring the shallow crustal structure of the northeastern Japan Island arc. Since the earthquake occurred very close to the exploring field, the two groups conducted again a similar active-experiment nearly at the same place two months after the earthquake to examine whether the crustal structure around the earthquake temporally changed or not.

In this study, we investigate temporal changes of the crustal structure by analyzing seismograms obtained for the two active-seismic experiments. The seismic data we analyze have merits in the following points: (1) very similar seismic waves were excited from the controlled sources; (2) the experiments were conducted just before and after the M6.1 earthquake; (3) the explosion points and seismic stations are located very close to the focal area of the M6.1 earthquake and Mount Iwate. Applying the cross spectrum analysis to the high-quality seismograms, we estimate changes of averaged seismic velocity in the upper crust and locate regions where short-wavelength heterogeneity temporally changed.

Observation and Data

Figure 1 shows a location map of explosion site, seismic stations, the fault area of the M6.1 earthquake, and peaks of Mount Iwate. The artificial explosion site is located near the south edge of the M6.1 earthquake fault as indicated by a star mark. The first seismic active-experiment was conducted on August 10, 1998, and artificial seismic waves were radiated by a chemical explosion of 100 kg dynamite embedded at a depth of 30 m. The second experiment was conducted on November 2, 1998, which was about three months since the first experiment. The same amount of dynamite was exploded at the same depth, although the explosion point of the second experiment was located at 30 m to the north of the first one.

For both of the explosions, we obtained good waveform data with broad-band seismometers at three nearby stations, MTI, GNB and ANS (see Fig.1), which had been deployed for detecting volcanic earthquakes associated with the activity of Mount Iwate since April to July 1998 (Nishimura *et al.*, 1999). The seismometer we used is a three component set (UD, NS, EW) of STS-2 (Streckeisen), which has a flat response in velocity from 0.02 s to 120 s. At each station, signals are continuously recorded by a digital data recorder LS8000WD (Hakusan Co.) with a sampling rate of 0.02 s and an A/D resolution of 24 bits. The internal clock (precision is higher than 10^{-6}) is corrected by a GPS clock system every two hours so that error of the recording time is less than 8 ms.

Figure 2 shows velocity seismograms of UD component observed at the three stations. Upper and lower traces of each station are seismograms for the first and second explosions, respectively. The signals are band-pass filtered for 1 to 16 Hz to eliminate long-period noises. In gross features, the waveforms of the two explosions are quite similar at all the stations, which indicate a satisfactory reproducibility of explosion condition. We use three component data at these three stations for the analysis.

Paper number 1999GL005439.
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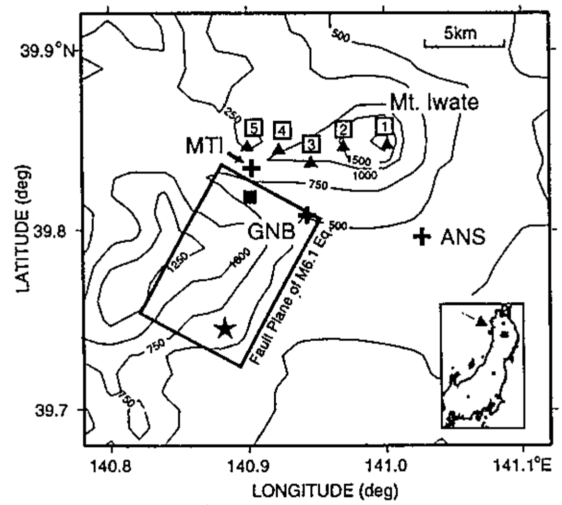

Figure 1. Location map of broad-band seismic stations (large plus symbols), an explosion-point (star), the epicenter of the M6.1 earthquake (solid square), and GPS stations (small plus). Horizontal projection of the fault plane of the M6.1 earthquake (September 3, 1998) estimated from the aftershock distribution (Umino *et al.*, 1998) are denoted by a rectangular. Peaks 1 to 5 of Mount Iwate (solid triangles) are aligned in the east-west direction, and peak 1 is the highest (2036m). A spherical pressure source is estimated at a depth of a few kilometers beneath peaks 4 to 5 (Ueki *et al.*, 1999).

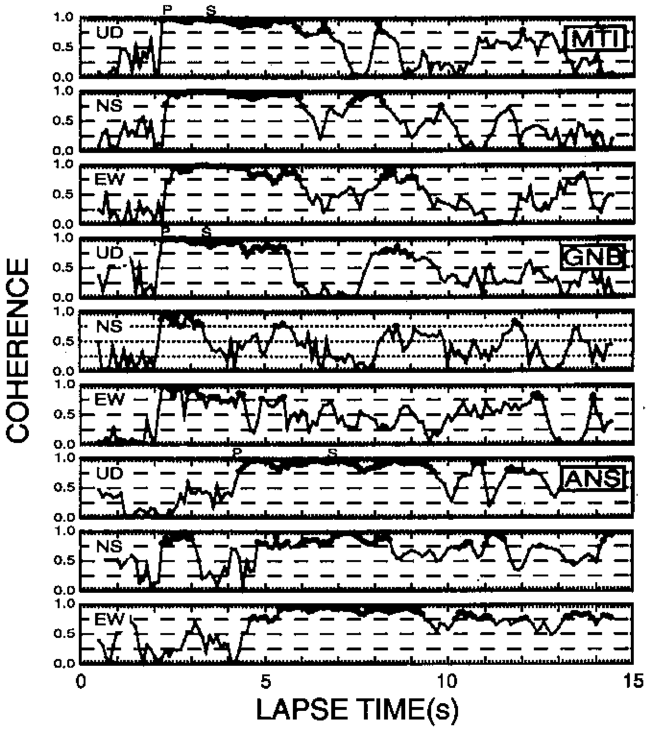

Figure 3. Temporal variation of coherence at the frequency of 4.7 Hz. Coherence larger than 0.75 are marked by a large circle. Arrival times of P and S-waves are denoted by P and S symbols, respectively, on the top of trace for each station.

Cross Spectrum Analyses

If the crustal structure temporally changed for the three months, the seismic waves excited by the first artificial explosion are expected to be different from those by the second one in travel time and/or waveform. To quantitatively evaluate such difference, we calculate cross spectra (i.e., coherence and phase difference) of the waveforms from the two explosions. We first band-pass filter the seismograms for 1 to 16 Hz to eliminate long-period noises. Then, coherence and phase difference as a function of time are calculated by shifting a time window every 0.1 s from the explosion time. Length of the time window is set at 1.5 s for waves of 2.0 Hz and 1.0 s for those of 3.1 Hz and 4.7 Hz. Figure 3 shows temporal variations of coherence for the three components of the three stations. The origin of time axis is set at the explosion time. In the figure, results for 4.7 Hz, which is the dominant frequency of the seismic waves, are plotted. For all of the stations, the coherence shows large values of more than 0.75 for a few to several seconds just after the P-arrivals. Although the coherence sometimes indicates significant dips, large values continue from the P-wave arrival to a lapse time of about 8 - 10 s for most of the components. These high coherence signatures suggest that the two artificial explosions excited nearly equal seismic waves.

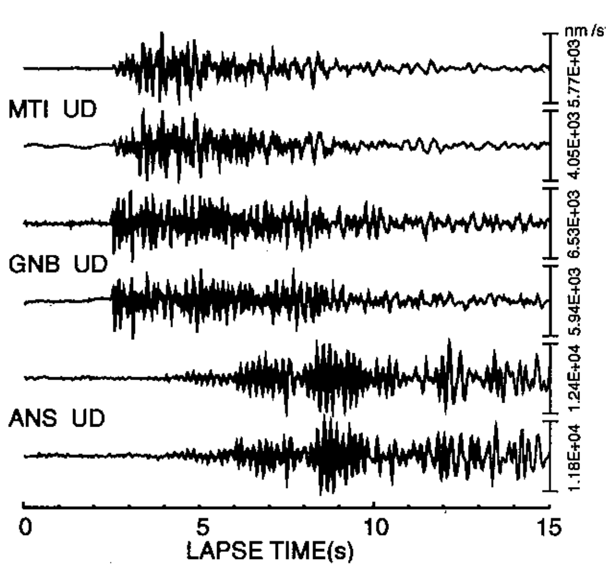

Figure 2. Vertical velocity seismograms for the two artifical explosions. Upper and lower traces for each station are the seismograms for the first (Aug. 10, 1998) and the second (Nov. 2, 1998) explosions, respectively. The records are band-pass filtered from 1 to 16 Hz.

Temporal Changes of the Average Velocity as Inferred from Phase Difference

Figure 4 shows temporal variations of phase difference for the three components of the three stations at the frequency of 4.7 Hz. Minus sign represents a phase delay of the waveforms for the second explosions to that for the first one. In Figure 4, we find that phase difference, ϕ, gradually decreases with time when we place the focus on the portions having a large coherence. Similar characteristics are also observed for horizontal components at the three stations. We fit a straight line to the temporal variation of phase difference for about 10 s from the onset of P-waves. Then, we estimate the time derivative of phase differences (slope of the straight line), $d\phi/dt$, to be -10.3 degree/s, -16.9 degree/s, and -2.6 degree/s for the data of UD component at MTI, GNB and ANS, respectively. The observed gradual decreases of phase difference are credible enough judging from the precise sampling timing of 10^{-6} of the digital data recorders. For the onset of direct P wave, however, we could not detect a travel time difference that exceeds the absolute time accuracy of 8 ms. Direct S wave onsets are too ambiguous to read the travel times.

Figure 5 shows frequency dependence of $d\phi/dt$ for the all components at the three stations. The frequency higher than

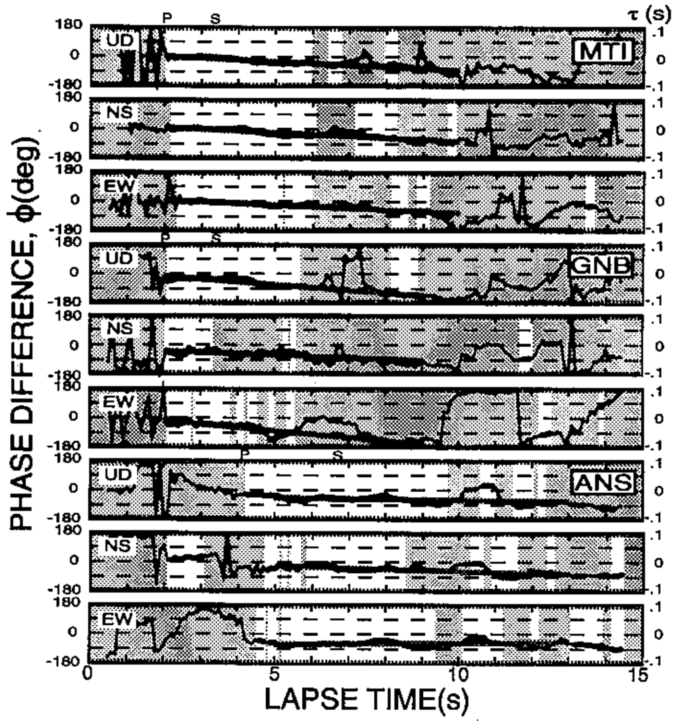

Figure 4. Temporal variation of phase difference at the frequency of 4.7 Hz. Positive phase difference corresponds to the phase advance of the second explosion data to the first one. Data shaded by light-gray color represent the data having coherence less than 0.75. Dark gray lines are best-fitted lines (see text). Time difference τ is estimated from a relation of $\tau=\phi/2\pi f\ (f=4.7\text{Hz})$.

4.7 Hz is not included because of poor coherence. It is found that $d\phi/dt$ decreases as frequency for the GNB and MTI stations. The data of ANS show a week dependence on frequency. These linear relations enable us to estimate velocity changes of the crust from the relation of $dV/V = (1/2\pi f)d\phi/dt$, where f is the frequency of the waves and V the averaged velocity in the medium where scattered seismic waves propagate. Averaging data of the three components at each station, we estimate the velocity changes to be -0.0069, -0.0096 and -0.0031 for MTI, GNB and ANS, respectively. This means that seismic velocity of the structure around the stations and the explosion point decreased by 0.3 % - 1.0 % on the average during the three months. We roughly estimate the corresponding region to be the upper crust within a distance of 15 km from the fault, assuming that coda waves consist of singly scattered waves in the crust having a wave velocity of 5 km/s.

The station MTI and GNB, where a larger velocity change (0.7 - 1.0 %) were observed than ANS, are located close to the focal area and a volcanic pressure source estimated by Ueki *et al.* (1999). This result strongly suggests that stress field changes associated with the occurrence of the M6.1 earthquake and/or the volcanic activity of Mount Iwate are closely related to the observed velocity decrease of the upper crust. Although regional stress changes due to the seismic and volcanic activity are not known in detail, we roughly estiamte whether or not the expected stress change of the crust are sufficiently large to cause the observed velocity decreases. Results of laboratory experiments summarized by Press (1966) show that seismic velocity changes on confining pressure (i.e., $(dV/V)/dP$, where V is the velocity and P the confining pressure) for granite are about 1-6 x 10^{-3} /MPa at a low confining pressure (<100MPa). The M6.1 earthquake brought about a stress drop of 4.9 MPa (Earthquake Information Center, Univ. of Tokyo, 1998) so that the velocity decrease is roughly estimated to be 0.5 – 3 percent by assuming a confining pressure decrease equal to the stress drop. The estimated value coincides with the velocity decrease (about 1%) observed in the vicinity of the focal area in the order of magnitude. Taking into account that the observed velocity change is an average over a larger area of about 15 km in linear dimensions, we conclude that our observation is reasonably accounted for by the stress release by the M6.1 earthquake. While this interpretation is most plausible, other possibilities such as changes of water table or permeation of rainwater into the ground are not completely ruled out.

Temporal Changes of Localized Velocity Structure Estimated from Coherence Analysis

The observed coherence occasionally indicates a sudden decrease (< 0.75) in the period of otherwise high coherence (see Fig.3). For example, the coherence at MTI keeps high values of more than 0.75 for the initial 3 – 4 s after the P-arrival. The coherence suddenly drops down to lower values (< 0.75) for the following a few seconds, but increase again before the gradual decrease with time. Occurrence times of such drops do not coincide with the expected arrival times of S wave and surface waves. Hence, such drops are interpreted to be originated from a temporal change of scatterers localized in small regions. In this section, we locate such anomalous regions from the time functions of coherence, based on the idea that seismic coda consist of the waves singly scattered from short-wavelength heterogeneity of the crust.

We set grid points around the explosion point and stations with a separation of 1 km in the north-south, east-west and vertical directions. We assume that the seismic waves radiated from the explosion point are singly scattered at each grid point, and then reach the stations to generate seismic coda. When seismic wave is scattered from a grid point where the crustal structure temporally changed, the coherence at the corresponding lapse time results in decrease. To locate such regions, we calculate averaged coherence coefficient ($\overline{C}$):

$$\overline{C}(x,y,z) = \frac{1}{3N}\sum_{j}^{3}\sum_{i}^{N} C_{ij}(t(x,y,z)), \qquad (1)$$

where $C_{ij}(t)$ represents the coherence for the j-th component of the i-th station at a time t, N the total number of stations, and $t(x,y,z)$ the travel time from the explosion point to the station for the seismic wave that are singly scattered at a location (x, y, z). Travel times are calculated from a 1-D velocity structure

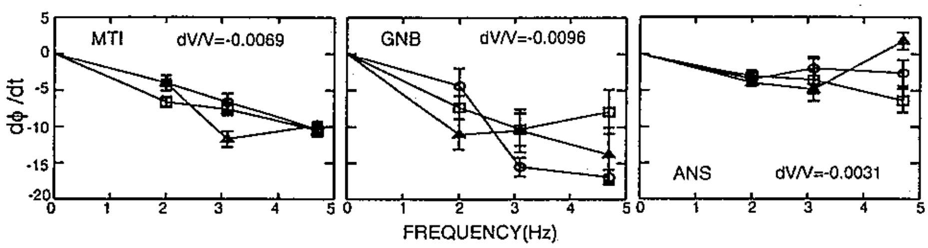

Figure 5. Frequency dependence of the time derivative of phase difference. Open circle, square, and triangular represent $d\phi/dt$ estimated from UD, NS, and EW components, respectively.

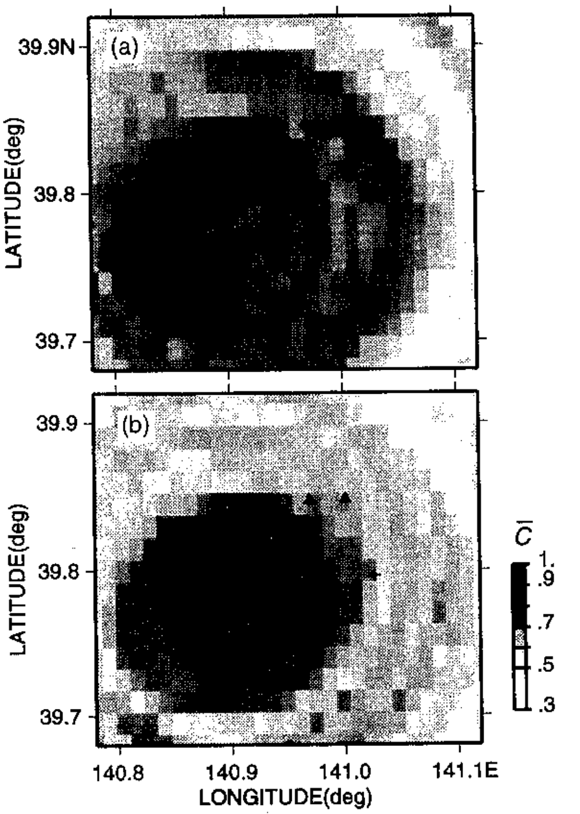

Figure 6. Spatial distribution of averaged coherence coefficient, $\bar{C}$ (see text), at a depth of 6km. (a) case 1: PP and PS scattering are assumed for vertical and horizontal components, respectively. (b) case 2: SP and PS scattering are assumed. Solid triangles and open rectangular are peaks of Mount Iwate and the fault plane of M6.1 shown in Figure 1.

that is used for hypocentral determination in this region (Tanaka *et al.*, 1999). Since all the stations are located on a low-velocity layer of volcanic edifices, PP and/or SP scattering waves are expected to be dominant in the vertical component while PS and/or SS waves are dominant in horizontal components. There are plural combinations to calculate $\bar{C}$ from vertical (denoted as V) and horizontal components (H): (1) PP for V and PS for H; (2) PP for V and SS for H, (3) SP for V and PS for H; (4) SP for V and SS for H.

Figure 6 show a spatial distribution of $\bar{C}$ at a depth of 6 km for the cases (1) and (3). Regions having high values of $\bar{C}$ (hereafter referred to as high-$\bar{C}$ region) horizontally extend about 5 – 10 km around the center of the fault plane of the M6.1 earthquake. Such high-$\bar{C}$ regions are observed at a depth of 2 to 8 km. Results for cases (2) and (4) also show similar characteristics. We, therefore, conclude that the uppermost crust around the M6.1 earthquake did not cause a significant change in short-wavelength heterogeneity for the three months. This may be because the fault is too thin to have changed the scattering property. In Figure 6, we further find a low-$\bar{C}$ zone that circularly surrounds the high-$\bar{C}$ regions. Since a little higher $\bar{C}$ value appears outside the low-$\bar{C}$ zone, the low-$\bar{C}$ cannot be attributed to the decrease of $\bar{C}$ value with distance. This suggests that the low-$\bar{C}$ zone is the candidate region where heterogeneous structure temporally changed. Although it is difficult to discuss errors and spatial resolution of $\bar{C}$-value due to a limited number of stations, it is noteworthy to mention that the low-$\bar{C}$ zone surrounds Mount Iwate where a dike intrusion is inferred from an analysis of borehole tilt, volumetric strain and GPS data by Sato and Hamaguchi (1999).

Summary

We detected temporal changes of crustal structure by detailed analysis of seismic waves excited by two artificial explosions that were carried out about one month before and two months after a M6.1 shallow earthquake. The results are summarized as follows;

(1) Seismic velocity of the upper crust decreased 0.3 - 1.0 % in a wide region within a distance of 10 – 15km which surrounds the M6.1 earthquake fault and Mount Iwate. The velocity decrease was probably originated from the stress change due to the occurrence of the M6.1 earthquake and a volcanic pressure source beneath Mount Iwate.

(2) Coherence analysis indicates a possibility of temporal changes of the heterogeneity beneath the Mount Iwate volcano region where a magmatic fluid intrusion is inferred from an analysis of geodetic data. However, such changes were not observed in and near the focal area of the M6.1 earthquake.

Acknowledgements. We are grateful to the Research Group for Intra-Plate Earthquakes Research Group for Explosion Seismology and the Research Group for Explosion Seismology, Japan, which organized the active seismic experiments. We thank Eisuke Fujita for his help to the observation at the GNB station. Thanks are also due to Takashi Sato for valuable discussions.

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(Received: September 29, 1999; revised: November 10, 1999; accepted: November 22, 1999.)