ASPERITIES AND SEISMIC VELOCITY DISTRIBUTION ALONG THE FAULT PLANE
FOR TWO LARGE EARTHQUAKES IN JAPAN

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Abstract

We applied the double-difference tomography method (Zhang and Thurber, 2000) to two large shallow earthquakes that occurred recently beneath the land area to compare slip distribution of the mainshock rupture with seismic velocity distribution along the fault plane. The results show that the fault plane was imaged as a zone of steep velocity change for the M6.4 2003 Northern Miyagi earthquake, while both sides of the fault plane did not have any systematically different velocities for the M7.3 2000 Western Tottori earthquake. For both events, large slip areas of the mainshock rupture (asperities) are distributed in high velocity areas on the fault plane, avoiding low velocity areas. The present observations provide important information on the cause of the formation of asperities, and strongly suggest the possibility of imaging asperities of large earthquakes before their occurrences.

1. Introduction

Great progress has been made in the last few years in understanding the stress concentration mechanism causing interplate earthquakes beneath the Pacific Ocean off the northeastern Japan arc. Asperities are distributed in patches surrounded by stable sliding areas on the plate boundary. Aseismic slip in the surrounding stable sliding areas results in the accumulation of stress at the asperities, and
earthquakes occur when the strength limit of an asperity is reached leading to sudden slip. It has gradually become clear that this kind of asperity model (Lay & Kanamori, 1981) represents an accurate description of the mechanism of such earthquakes (Nagai et al., 2001; Yamanaka & Kikuchi, 2004; Matsuzawa et al., 2002; Okada et al., 2003, Hasegawa et al., 2004).

Although we have to continue examining the applicability of the asperity model to the earthquake occurrence, it is important to investigate what is the cause of the asperity. The only way to understand the mechanism of asperity formation is to look into the deep earthquake fault zone in detail that lies below the surface. Conventional seismic tomography cannot give us deep underground images at a sufficient spatial resolution that can be compared with the asperity; we are unable to obtain clear images of the earthquake source fault itself by the conventional tomography method, either. Double-difference tomography method which has been recently developed by Zhang & Thurber (2003) has the capability of imaging 3D inhomogeneous velocity structure of hypocentral area at a higher spatial resolution.

Dense temporary seismic observation networks were constructed after the two large shallow earthquakes; the 2003 M6.4 Northern Miyagi earthquake in northeastern Japan (Umino et al., 2003) and the 2000 M7.3 Western Tottori earthquake in southwestern Japan (Joint Group for Dense Aftershock Observation of the 2000 Western Tottori Earthquake, 2001). A number of aftershocks of the two events were observed by these dense seismic networks deployed above the source regions, which provided good opportunity to obtain a detailed seismic velocity structure near the fault plane of the mainshocks by applying the double-difference tomography method.

We applied the method to the data of aftershocks of the above two large inland earthquakes (Okada et al., 2004a, b) in order to answer the following questions: (1) Are there inhomogeneous velocity structures which correspond to (a) the earthquake fault plane and (b) the asperity? (2) If so, is it possible to image them?

In the followings, research results from Okada et al. (2004a, b) are briefly described.

2. Double-difference tomography

The double-difference tomography uses both absolute and relative arrival times of P- and S-waves, and can produce improved seismic velocity images, especially in and around the earthquake source region, by reducing systematic errors using the more accurate relative arrival times.
The misfit (travel time residual) between the observed and predicted arrival times is expressed as

\[ r'_i = \sum_{m=1}^{3} \frac{\partial T'_k}{\partial \chi'_m} \Delta \chi'_m + \Delta \tau'_i + \int_0^k \delta u ds \tag{1} \]

where \( T'_k \) is P- or S-wave arrival time from earthquake \( i \) to station \( k \), \( T'_i \) is origin time of event \( i \), \( u \) is slowness field, \( \chi'_m \) is source coordinate of event \( i \), and \( ds \) is an element of path length. Subtracting a similar equation for event \( j \) at the same station \( k \) from equation (1), we have,

\[ r'_k - r'_k = \sum_{m=1}^{3} \frac{\partial T'_k}{\partial \chi'_m} \Delta \chi'_m + \Delta \tau'_j + \int_0^k \delta u ds \]

\[ -\sum_{m=1}^{3} \frac{\partial T'_k}{\partial \chi'_m} \Delta \chi'_m - \Delta \tau'_j - \int_0^k \delta u ds \tag{2} \]

This can also be written as

\[ r'_k - r'_k = \left( T'_k - T'_k \right)^{bs} - \left( T'_k - T'_k \right)^{al} , \tag{3} \]

and this is called the double-difference.

In the conventional seismic tomography, the seismic velocity structure is estimated by solving only equation (1). On the contrary, both equations (1) and (2) are simultaneously solved as an observation equation in the double-difference tomography (Zhang & Thurber, 2003). In equation (2), as the portions of ray paths from two nearby events that are far away from the source region will substantially overlap, the velocity structure parameters only in and around the source region will be solved. Meantime, in equation (1), all of the velocity structure parameters on the ray from hypocenter to station are solved, in common with the conventional tomography.

Setting up 3D grid in the study area, the velocity structure is calculated from estimating the unknown velocity value at each grid node. In order to find rays and to calculate travel times between events and stations, we use the pseudo-bending algorithm (Um & Thurber, 1987). The derived observation equation, along with the smoothing constraints, is solved with the use of the LSQR algorithm (Paige & Saunders, 1982).
3. 2003 M6.4 Northern Miyagi earthquake

After 26 July 2003 Northern Miyagi earthquake of M6.4, we deployed a temporary observation network directly above its hypocentral area in order to monitor aftershock activity (Umino et al., 2004). Precise hypocentral distribution of aftershocks was obtained by using data collected by this temporary observation network as well as data from stationary observation networks of Tohoku University, National Research Institute for Earth Science and Disaster Prevention, and Japan Meteorological Agency (Umino et al., 2003, 2004; Okada et al., 2003). Using hypocenters of aftershocks located by Okada et al. (2003) by the double-difference location method (Waldhauser & Ellsworth, 2003), we applied the double-difference tomography method to the dataset of travel times of P- and S-waves collected both at the temporary and stationary observation stations. We utilized 853 earthquakes, 11,893 data of absolute P-wave arrival times, and 10,154 data of absolute S-wave arrival times observed at 18 stations. Also, 188,033 P-wave catalog differential arrival times and 143,081 S-wave catalog differential arrival times are used. Observation stations and earthquakes we used are shown in Fig.1. Grid points are also shown by pluses in the figure. The grid points are set at intervals of 2-3 km east and west, 4-5 km north and south, and 2-4 km in depths.

To assess the reliability of the estimated velocity structure, the restoring resolution test (RRT; Zhao et al., 1992) is carried out. Assuming that the inhomogeneous velocity structure obtained by the inversion is accurate, we create a synthetic dataset of travel times based on this structure. Then random noises are added to the synthetic travel times. By applying the tomographic inversion to this synthetic dataset, we can verify the accuracy of the restored velocity structure comparing to the original one. EW vertical cross-sections of P- and S-wave velocities along line BB’ in Fig.1 are shown in Figure 2. We can see that the velocity structure of both P-wave (above) and S-wave (below) are restored very precisely.

The estimated velocity structure of P-wave and that of S-wave are shown in Figure 3 and Figure 4, respectively. Those figures are vertical cross-sections of P- and S-wave velocity perturbations along lines AA’, BB’, CC’, DD’, EE’ and FF’ which are indicated in Figure 1. Figures 3 and 4 clearly show that the aftershock alignment corresponds with a zone of steep velocity change and that the velocity is lower in the hanging wall than in the footwall on any of the six vertical cross-sections (AA’ FF’). This trend is especially clear in the S-wave velocity images (Figure 4).

Aftershocks generally occur along the mainshock fault plane, and so the aftershock alignment in the figures represents the fault plane of the M6.4 event. Therefore, the results presently obtained using the double-difference tomography method (Figs.3 and 4) indicate that: 1) inhomogeneous structure corresponding to the earthquake fault plane was detected, and 2) it was imaged as a zone of steep
velocity change.

According to the precise analysis of the aftershock distribution, we inferred that the 2003 Northern Miyagi earthquake was caused by the reactivation of Ishinomaki-wan fault as a reverse fault (Umino et al., 2004; Okada et al., 2003), which was acted as a normal fault under an extensional stress regime when the Japan Sea was opening in the early Mio cene (Geological Survey of Japan, 1990). If the Ishinomaki-wan fault was a normal fault in the past, the characteristic distribution that the velocity is slower in the hanging wall than in the footwall can be explained. In other words, the distinctive distribution of seismic velocity in Figs.3 and 4 reflects the difference between the hanging wall and the footwall caused by the repeated displacements of the past normal faulting, especially by the presence of thick sedimentary layers in the hanging wall.

Seismic slip distribution of the 2003 Northern Miyagi earthquake estimated from geodetic data (Miura et al., 2004) and that from seismic waveform data (Yagi et al., 2003) both show that an extremely large slip area is located at a shallower depth in the northern part of the fault plane. This large slip area (asperity) is indicated by white dashed lines in the figures of the inhomogeneous velocity structure (Figures 3 and 4). We can see that this asperity is located in an area with relatively high velocities on the fault plane. This trend is especially noticeable in P-wave velocity images (Figure 3). According to the present analysis, we can conclude that:1) There exists inhomogeneous velocity structure that corresponds to the asperity, and 2) inhomogeneous structure corresponding to the asperity was imaged as a high velocity area.

4. 2000 M7.3 Western Tottori earthquake

After the occurrence of the 2000 Western Tottori earthquake, a group of researchers from many universities across the nation jointly conducted a dense seismic observation of its aftershocks (Joint Group for Dense Aftershock Observation of the 2000 Western Tottori earthquake, 2001). We applied the double-difference tomography method to P- and S-wave arrival time data of 960 aftershocks which were observed at 59 stations of the temporary dense observation network including a part of stationary stations of Kyoto University and of National Research Institute for Earth Science and Disaster Prevention. We used 51,922 P-wave absolute arrival times and 33,840 S-wave absolute arrival times. In addition, we used 208,166 P-wave arrival time differences and 161,382 S-wave arrival time differences. Observation stations and earthquakes used in the inversion are shown in Figure 5. Plus signs in the figure show the location of grids allocated. Grid intervals are 2.5-5 km along the strike of the mainshock fault, 1-7 km in the direction perpendicular to the fault and 2-4 km in depth. We utilized
Data from the earthquake catalog by the Joint Group for Dense Aftershock Observation of the 2000 Western Tottori earthquake (2001) for the initial earthquake locations.

Figures 6 (a), (b) and (c) show horizontal sections of P-wave velocity perturbations at depths of 4 km, 6 km and 8 km, respectively. Relocated aftershock hypocenters are indicated by circles in the figure. Figure 6 shows the tendency that low velocity areas are distributed in the north at depths of 4-6 km and in the middle of the fault plane at depths of 6-8 km close to the area of preseismic activities (Shibutani et al., 2002). However, the aberrational velocity change was not found along the earthquake source fault inferred from the linear distribution of aftershocks at any depths. In other words, the inhomogeneous velocity structure that corresponds to the earthquake source fault was not detected by this tomographic inversion study.

Next, we conducted a resolution test in order to examine the spatial resolution of the obtained velocity structure. Assuming that a 1-km width anomaly zone with 3-5% velocity drop exists along the fault plane inferred from the aftershock alignment, we examined whether it can be restored by applying the inversion to the datasets derived from the same combination of observation stations and earthquakes. The result shows that the anomaly zone was well restored except its northern and southern edges in any case.

The obtained vertical cross section of P-wave velocity perturbations along the fault plane is shown in Figure 7. Relocated aftershocks are indicated by circles. The figure shows that high velocity areas are distributed in the southeastern part of the fault plane and that low velocity areas are distributed in the northwestern part. Also, a low velocity area can be found near the mainshock hypocenter located at the central part close to the preseismic activities (Shibutani et al., 2002). We conducted RRT (Zhao et al., 1992) in order to assess the reliability of the estimated velocity structure; the result is shown in Figure 7(b). Comparison of (b) with (a) shows that the original velocity structure was restored very precisely. In other words, it indicates that high resolution images can be obtained along the fault plane.

Figures 8 (a) and 9 show P-wave velocity distribution along the fault plane compared with seismic slip distribution of the mainshock rupture (Iwata & Sekiguchi, 2001; Yagi & Kikuchi, 2000). White contours in the figures indicate seismic slip distribution of the mainshock rupture, a star sign is the mainshock hypocenter (starting point of the mainshock rupture). White circles in Fig.8(a) indicates hypocenters of preseismic activities located by Shibutani et al (2002). Although large slip areas of the mainshock rupture (asperities) shown in Figure 8(a) and Figure 9 are slightly different from each other, we can see the tendency that asperities are distributed in high velocity areas avoiding low velocity areas in both figures. Therefore, as in the case of the 2003 Northern Miyagi earthquake, we can assume that
asperities are distributed in high velocity areas on the fault plane. Also, Figure 8(a) suggests that the starting point of the mainshock rupture and the area of the preseismic activities are located in a relatively low velocity area surrounded by high velocity areas.

Figure 8(b) shows the distribution of seismic wave scattering coefficient along the fault plane which was estimated by Asano & Hasegawa (2004). Comparison with P-wave velocity distribution in (a) shows the approximate correspondence of low velocity areas to large scattering coefficient areas and of high velocity areas to small scattering coefficient areas.

5. Conclusions

We applied the double-difference tomography method to dense observation data of aftershocks of the 2003 Northern Miyagi earthquake and the 2000 Western Tottori earthquake in order to study 3D inhomogeneous velocity structure along the earthquake source fault and the surrounding areas. The findings are as follows.

1) We detected the inhomogeneous velocity structure corresponding to the earthquake fault plane as a zone of steep velocity change for the 2003 Northern Miyagi earthquake. However, we could not detect it for the 2000 Western Tottori earthquake.

2) On the contrary, the inhomogeneous velocity structure corresponding to asperities was imaged as high velocity areas on the fault plane for both of the two earthquakes.

The present tomographic inversion study has achieved important results for understanding the cause of the formation of asperities. However, since this study covered only two earthquakes, we need to continue examining more cases.


Shibutani T., S. Nakao, R. Nishida, F. Takeuchi, K. Watanabe, and Y. Umeda, Swarm-like seismic activity


Figure Captions

Fig. 1 Locations of observation stations (triangles), earthquakes (dots), and grids (crosses) used in the inversion. Bold broken lines show the location of the Ishinomaki-wan Fault. Thin broken lines show the locations of vertical cross-sections (AA’ through FF’) in Figs. 3 and 4.

Fig. 2 Result of RRT. (a) P- and (b) S- wave velocity structure are shown on a vertical cross section along line BB’ shown in Fig. 1. Original and restored velocity structures are shown on the left and right, respectively.

Fig. 3 Vertical cross-sections of P-wave velocity perturbations along lines AA’ to FF’ shown in Fig. 1. Perturbations from the average velocity at each depth are shown. White broken lines show the location of the asperity (large slip area; Miura et al., 2004). White stars denote hypocenters of the main shock (FF’), the largest foreshock (EE’), and the largest aftershock (AA’). Denser color means greater DWS value. Box on the top shows the location of the Ishinomaki-wan Fault at each cross section.

Fig. 4 Vertical cross-sections of S-wave velocity perturbations along lines AA’ to FF’. Others are same as Fig. 3.

Fig. 5 Locations of observation stations (triangles), earthquakes (dots), and grids (crosses) used in the inversion.

Fig. 6 Horizontal sections of P- wave velocity perturbations at depths of (a) 4 km, (b) 6 km, and (c) 8 km. Circles show aftershocks at each depth.

Fig. 7 (a) Vertical cross section of P-wave velocity perturbations along the mainshock fault plane. Circles are aftershocks. (b) Results of RRT. Restored velocity structure is shown.

Fig. 8 (a) Slip distribution of the mainshock rupture (white contours) by Iwata and Sekiguchi (2001) and the P-wave velocity perturbations on a vertical cross-section along the fault plane. White circles and colored circles show events of the preseismic activities (Shibutani et al., 2002) and relocated aftershocks, respectively. (b) Distribution of seismic scattering coefficients on a vertical cross-section along the fault plane. (Asano and Hasegawa, 2004).

Fig. 9 Slip distribution of the mainshock rupture (white contours) by Yagi and Kikuchi (2000) and the
P-wave velocity perturbations on a vertical cross-section along the fault plane. Others are same as Fig.8(a).
Figure 13

(a) \( V_p \)

Original

RRT

(b) \( V_s \)

Original

RRT
Figure 15
Figure 17

a) 4 km  
b) 6 km  
c) 8 km
Figure 18

a) Original

b) RRT