VELOCITY STRUCTURE AND AFTERSHOCK DISTRIBUTION OF THE 1982 URAKAWA-OKI EARTHQUAKE

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Aftershock distribution of the Urakawa-oki earthquake (Ms 6.8) of 1982 is investigated. A simultaneous inverse method is applied to P-wave arrival time data from the aftershocks and quarry blasts in order to estimate a three-dimensional velocity structure in the aftershock region. Strong correlations between the obtained velocity structure, geology, and gravity anomalies are observed.

Hypocentral determination of the aftershocks by using the three-dimensional velocity structure shows that the epicentral distribution is a triangle and its area is 790 km². The detailed aftershock distribution indicates that three dipping planes with high aftershock activities appeared in the aftershock region: two of the planes are almost parallel to each other with a dip angle of 50° in a depth range of 3 km to 27 km, and these planes clearly form double aftershock planes, which have been previously supposed to be an identical plane; the other plane is located perpendicularly to the double aftershock planes in a depth range of 13 km to 27 km and seems to be a conjugate plane with the double aftershock planes. The aftershocks occur in a region with P-wave velocities higher than 5 km/s, and the conjugate plane is located along the structural boundary. It is found that the rupture of the main shock took place on the conjugate plane and also might have been caused by a tectonic force derived from a collision between the Kurile and Northern Honshu Japan Arcs.

1. Introduction

The Urakawa-oki earthquake (Ms 6.8) occurred off the coast of Urakawa, Hokkaido, Japan, on March 21, 1982 (Fig. 1). According to Suzuki and Motoya (1983), the hypocenter of the earthquake was located at 42.16°N, 142.53°E and at a 27 km depth. A few thousands of aftershocks took place around the hypocenter of this earthquake. Moriya et al. (1983a, b) and Iwasaki et al. (1983) have investigated the aftershock activity of the earthquake: Moriya et al. (1983a, b)

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installed fourteen temporary seismic stations with high-gain seismometers in the land area of the aftershock region, and IWASAKI et al. (1983) deployed four ocean bottom seismographs (OBS) in the sea around this region. On the other hand, the Research Center for Earthquake Prediction in Hokkaido University (RCEP) operates six permanent seismic stations in the Hidaka district in which the aftershock region is included. Based on these observations, MORIYA et al. (1983 a, b), SUZUKI and MOTOYA (1983), and IWASAKI et al. (1983) obtained time and spatial distributions of the aftershocks. The spatial distribution they obtained is summarized as follows:

(1) Spatial extent of the aftershock region is about 35 km × 25 km.
(2) The aftershock distribution is spatially separated into two groups forming a plane approximately; one has earthquakes which occurred on a plane with an area of 20 km × 25 km to 15 km dipping 20°–30° northeast and the other has earthquakes which occurred on a plane with an area of 10 km × 10 km to 15 km which inclines southwest with a degree of 60°–70°.
(3) These two planes formed by the aftershock distribution may be conjugate with each other.

The spatial distribution of aftershock hypocenters is based on a laterally homogeneous velocity structure with some station corrections. Such an unrealistic assumption may often cause a spatially vague distribution of the aftershocks. In
order to understand the more detailed aftershock distribution, more accurate velocity structure for the hypocentral determination must be applied. For instance, a three-dimensional (3-D) velocity structure should be included to locate the hypocenters.

The velocity structure around the aftershock region of the Urakawa-oki earthquake has been investigated: DEN and HOTTA (1973) and ASANO et al. (1979) carried out refraction and reflection measurements off Urakawa, and found the existence of a sedimentary layer 10 km to 17 km thick with a P-wave velocity of 4.2 km/s beneath the sea bottom off Urakawa. FUJII and MORIYA (1983) observed quarry blasts at 102 temporary stations in the Hidaka Mountains and obtained a sedimentary layer with a P-wave velocity of 3.7 km/s which thickens abruptly from the land near Urakawa to the coastal line on the west side of the Hidaka Mountains. TAKANAMI (1982) and MIYAMACHI and MORIYA (1984) also estimated a 3-D P-wave velocity structure beneath the Hidaka Mountains using the arrival time data of local earthquakes. The 3-D velocity structure they obtained showed that there is strong lateral heterogeneity around Urakawa. Furthermore, MIYAMACHI and MORIYA (1984) derived a structural model with an inclined low velocity zone in a depth range of 10 km to 65 km beneath the Hidaka Mountains.

The purpose of the present study is to obtain the aftershock distribution more precisely and to discuss a relation between the spatial distribution of the aftershocks and the crustal structure. To determine the crustal structure, the 3-D inverse method based on AKI and LEE (1976) and THURBER (1983) is applied.

2. Geology and Bouguer Anomalies

For the studied area, very complicated features in geology have been reported by OKADA (1983), KIMINAMI and KONTANI (1983), and CADET and CHARVET (1983). Figure 2 shows the geological map simplified by the author, on the basis of the reports. The studied area is composed of three major geological belts, the Ishikari Belt, Kamuikotan Belt, and Hidaka Belt. The Ishikari Belt is made up of a thick sedimentary sequence mainly composed of late Mesozoic and Tertiary strata. The Yezo Group consists of terrigenous elastic sediments, while the Sorachi Group is mainly composed of an ophiolitic rock associated with chert, micritic limestone, basic pyroclastics, basaltic pillow lava and diabase. The Kamuikotan Belt consists mainly of Kamuikotan metamorphic rocks which comprise glaucochane schists, ultramafic and mafic rocks, partly of the Sorachi Group and the Yezo Group. The Hidaka Belt is characterized by the Hidaka metamorphic rocks and the Hidaka Supergroup. The Hidaka metamorphic rocks are composed of gneisses, migmatic, green schists, hornfelses, basic plutonic rocks, ultramafic rocks, and granitic rocks, whereas the Hidaka Supergroup consists of slate, chert, limestone, basic tuff, and flysch-type sediments of great thickness. As shown in Fig. 2, the Hidaka main thrust runs between the Hidaka metamorphic rocks and the Hidaka Supergroup.
Fig. 2. A simplified geological map in the study region, modified from Okada (1983). I.B., S.K.M., Y.G., H.S., and H.M. indicate the Ishikari Belt, the Sorachi Group and the Kamuiotan metamorphic rocks, the Yezo Group, the Hidaka Supergroup, and the Hidaka metamorphic rocks, respectively.

Fig. 3. Bouguer anomalies in the land area of the study region. (after Ookawa and Kasahara, personal communication, 1983). The solid and broken lines indicate the positive and negative Bouguer anomalies in mgal, respectively.
Figure 3 shows Bouguer gravity anomalies in this region (after Ookawa and Kasahara, personal communication, 1983). As shown in the figure, the inland area is characterized with positive Bouguer anomalies, while the coastal area is rather characterized with negative Bouguer anomalies. The positive Bouguer anomalies correspond to the Hidaka Belt and the Kamuikotan Belts. At the left end of the region, there are extremely negative Bouguer anomalies which may be caused by the sediment with the substantial thickness corresponding to the sedimentary sequence of the Ishikari Belt.

3. Method of Analyses

To estimate the three-dimensional (3-D) P-wave velocity structure, we apply a 3-D simultaneous inverse method based on Aki and Lee (1976) and Thurber (1983) to the arrival time data derived from aftershocks of the Urakawa-oki earthquake. The travel time is calculated along a circular ray path developed by Thurber (1981). Such a ray path can provide a good approximation in a calculation of travel times (Thurber, 1981, 1983). Therefore, the inversion using the circular ray path can provide a good approximation of the shallower part of the velocity structure, if the arrival time data from local shallow earthquakes are used. In this calculation, the partial derivatives of travel times with respect to hypocentral parameters are directly evaluated by variation of the earthquake location, and those with respect to velocity parameters are approximated by the "block method" described in Thurber (1983). After an estimation of the 3-D velocity structure, we determined locations of aftershocks and the main shock by using the 3-D velocity structure.

4. Velocity Structure

The study area in Fig. 1 is divided by 9 grid points at an interval of every 8 km on the x-axis directed toward N120°E, and by 8 grid points at an interval of every 7 km on the y-axis, while on the z-axis, 6 grid points are taken at depths of −5, 2, 7, 12, 19, and 35 km. The initial model for the inversion is assumed to be laterally homogeneous as listed in Table 1. This model is based on Moriya et al. (1983 b) and is slightly modified by the authors.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Position in depth (km)</th>
<th>P-velocity (km/s)</th>
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<tbody>
<tr>
<td>1</td>
<td>−5</td>
<td>3.6</td>
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<tr>
<td>2</td>
<td>2</td>
<td>4.1</td>
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<td>3</td>
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<td>4</td>
<td>12</td>
<td>6.0</td>
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<td>5</td>
<td>19</td>
<td>6.1</td>
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<tr>
<td>6</td>
<td>35</td>
<td>6.5</td>
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</table>
For the three-dimensional inversion, we used 1,462 P-wave arrival times obtained from 136 aftershocks well located in the study area with focal depths less than 35 km. These data were obtained by 20 seismic stations consisting of 2 permanent stations operated by RCEP, 14 temporary stations by Miyamachi and Moriya (1984), and 4 ocean bottom seismographs (OBS) by Iwasaki et al. (1983). Locations of these stations are presented in Fig. 4. The initial locations of these aftershocks are shown in Fig. 5. In addition to these data, 63 P-wave arrivals obtained from quarry blasts in the northeastern part of the study area were used (Fujii and Moriya, 1983). Temporary stations for the observation of quarry blasts are also shown in Fig. 4, and the location of the blasts is denoted in Fig. 5 with a star. Reading errors of all observed data are less than 0.05 s.

We give the damping factors in the inversion as follows: for the velocity 3.5 s, for the x- and y-coordinates of aftershocks 0.10 s/km, for the z-coordinate 0.05 s/km, and for the origin times 0.30 s/s. The standard derivation of travel time residuals in the initial model is 0.44 s, and after 4 iterations it becomes 0.16 s. The final value of 0.13 s is obtained after 6 iterations. This value is larger than the reading errors (0.05 s). It would be necessary to build a model by a finer grid points spacing for interpreting more accurately the velocity distribution in the study region. Figure 6 shows the inverted P-wave velocity distribution and the diagonal elements of the resolution matrix at each grid point. The standard error of the P-wave velocity is less than 0.2 km/s for all points. However, we cannot obtain reliable solutions for the first (-5 km depth) and sixth (35 km depth) layers, because the ray paths passing through these layers are sparse. The diagonal elements of the resolution matrix with respect to the hypocentral parameters are also more than 0.7 and the mean standard errors are 0.4 km for the x-coordinate, 0.4 km for the y-coordinate, 0.6 km for the z-coordinate, and 0.08 s for the origin times.

Figure 7 shows the contour maps of the velocity distribution in each layer based on the P-wave velocity values with resolution more than 0.5. However, for example, the velocity 7.0 km/s at a grid point near the station HOU in Fig. 7(a) is not taken into consideration, although the corresponding resolution is greater than 0.5, because the velocity is very different from those obtained at the surrounding grid points. This velocity difference may be caused by some misidentification of the onset times of first P-wave arrivals recorded at HOU located near the aftershock region. The misidentification is caused mainly by the traffic noise, and partly by the noise produced by a water flow in rivers. However, referring to a result of an auxiliary inversion excluding the arrival time data obtained by HOU, the velocity distribution is almost the same as that in Fig. 7, though the velocity contrast around HOU becomes weak. Accordingly, it can be said at least that P-wave velocities near HOU are certainly faster than those in the surrounding area. The near-surface velocities at 2 km depth (Fig. 7(a)) correlates strongly with the geology (Fig. 2) and the gravity anomalies (Fig. 3). High velocities around HOU and beneath the inland area may be related to the Kamuikotan Belt and the Hidaka Supergroup, respectively. This high velocity zone also corresponds to a region of positive gravity.
Fig. 4. Map showing the locations of stations used in this study. Open squares denote the temporary stations for the observation of quarry blasts (FUJI and MORIYA, 1983). Seismic stations KMU and MSN belong to the Research Center for Earthquake Prediction in Hokkaido University (RCEP). Ocean bottom seismographs P1, P2, P3, and P4 were operated by IWASAKI et al. (1983). Other stations with solid squares were deployed by MIYAMACHI and MORIYA (1984).

Fig. 5. Map showing the initial locations of the earthquakes (open circles), the quarry blasts (star), and the main station locations (solid squares) used in the inversion.
anomalies. On the other hand, low velocities around SPP on the Ishikari Belt correspond obviously to a region with negative gravity anomalies where the sediment is very thick. Low velocities around TAK, which seem to run to the land from the sea, can be related to the Yezo Group. At 7 km depth (Fig. 7(b)), the velocity distribution in the land is different from that in the coast and sea—velocities in the land are high and those in the sea are low. Distinctly, the regions of low velocities around MUT and IKD correspond to those in the 2 km depth layer.

Fig. 7. Contour maps of the P-wave velocity in km/s of each layer. The small numbers indicate the values of the P-wave velocity at grid points with resolution of more than 0.5. (a) 2 km depth, (b) 7 km depth, and (c) 12 km depth.
Fig. 7
Accordingly, it is found that the strong lateral heterogeneity at this depth may be caused by a thickness of the sedimentary layers of the Ishikari Belt and the Yezo Group. At 12 km depth (Fig. 7(c)), the velocities less than 5 km/s almost disappear and the P-velocities of the basement are revealed. Accordingly, the sediment with low velocities beneath the coast and the sea in the study area seems to extend down at least to a depth of 10 km. MIYAMACHI and MORIYA (1984) investigated a 3-D velocity structure beneath only a land part of this study area by an inverse method similar to that of this study. Their obtained velocity structure is consistent with that in the land part obtained in this study. Also, two-dimensional velocity structure obtained by FUJII and MORIYA (1983) from observations of quarry blasts shows the same pattern of the velocity distribution as that by this study.

The relation between the initial and the inverted locations of the aftershocks used is shown in Fig. 8, where the initial locations were determined in a one-dimensional model by MORIYA et al. (1983 b). There are two distinct features in the relation as follows: (1) as shown in Fig. 8(a), there is a tendency that the aftershocks are so relocated as to move to the sea area; (2) most of aftershocks become deeper from the initial depths. However, a few aftershocks which were determined to be at
the greater depths in the 1-D model, become certainly shallow as shown in Fig. 8 (b) and (c).

5. **Main Shock and Aftershock Distribution of the Urakawa-Oki Earthquake**

In order to investigate the main shock location and the aftershock distribution of the Urakawa-oki earthquake, we construct the 3-D P-wave velocity model beneath the study area based on the results obtained in Sec. 4. We also add the seventh layer with laterally homogeneous P-wave velocity of 7.7 km/s at 45 km depth to the 3-D model for the sake of deep aftershocks with reference to the results of Miyamachi and Moriya (1984). For the hypocentral determination, theoretical travel times and their partial derivatives with respect to hypocentral elements are calculated in the same way as that used in Sec. 4.

![Aftershock distribution](image)

*Fig. 9. Aftershock distribution revealed by using the 3-D velocity model obtained in the inversion during the period from 24 March to 24 June, 1982. (a) x-y Plane, (b) x-z plane, and (c) y-z plane. We call this a long-term distribution. The location of the main shock is denoted by a star. In (a), lines at the top and the bottom in the figure denote a range of x-axis, (a), (b), (c), and (d), used in the y-z cross sections in Figs. 10 and 12.*
We estimated the location of the main shock using three stations, MSN, KMU, and IKD, which could observe the arrival times from the main shock. In the estimation, a non-linear method of Tarantola and Valette (1982) was used, because the least squares method commonly used cannot be applied to the data obtained by only 3 stations. The location of the main shock is shown with a star in Figs. 9–11. The standard errors in the hypocentral elements are 1.1 km for the x-coordinate, 7.6 km for the y-coordinate, and 6.2 km for the z-coordinate. Taking the large errors into consideration, the hypocenters of the main shock obtained in this study are almost the same as reported by Suzuki and Motoya (1983) (42.16°N, 142.53°E, and 27 km depth).

Figure 9 shows 387 hypocenters of the aftershocks which occurred during the period from March 24 to June 24, 1982. We call this distribution a long-term distribution. The mean standard errors in these hypocentral elements are 0.15 s for the origin times, 0.7 km for the x-coordinate, 0.8 km for the y-coordinate, and 1.7 km for the z-coordinate. The map (Fig. 9 (a)) shows that the epicenters are distributed in a triangular shape with an area of 790 km². The seismicity in the land

Fig. 10. Vertical cross sections of the long period distribution projected on the y-z plane. Range in the x-axis of the sections (a), (b), (c), and (d) is shown in Fig. 9. Broken lines denote Plane-A, Plane-B, and Plane-C estimated in this study, respectively. In (c), a star is the location of the main shock, and a region with especially low seismic activity on Plane-A is hatched.
Velocity Structure and Aftershock Distribution

Fig. 11. Aftershock distribution revealed by using the 3-D velocity model obtained in the inversion during the period from 29 March to 6 April, 1982, which is called a short-term distribution. (a) x-y Plane, (b) x-z plane, and (c) y-z plane. The location of the main shock is shown with a star.

is clearly more active than that in the sea. In the region, two active zones exist: one is along a line connected between KWK and P4, and the other is in the area surrounding by lines connected by P1, P2, and P3. Figure 9 (b) and (c) show that most of the aftershocks occur in the depth range of 3 km to 30 km. To understand the aftershock distribution more precisely, we divide the distribution into four parts by 10 km width along the x-axis as shown in Fig. 9 (a), and represent each vertical cross section projected on the y-z plane in Fig. 10. In Fig. 10 (b), three dipping planes can be identified on which the seismic activity is very high. We call these planes Plane-A, Plane-B, and Plane-C, respectively: Plane-A has a depth range of 3 km to 20 km with a dip angle of 50° and Plane-B a depth range of 6 km to 27 km, which is almost parallel to Plane-A with about 7 km distance, while Plane-C is located at an angle of 80°-90° with Plane-A and Plane-B at a depth range of 13 km to 27 km. In Fig. 10 (a), a few aftershocks occur around Plane-A, but around Plane-B and Plane-C quite few aftershocks occur. As shown in Fig. 10 (b), Plane-A and Plane-B form double aftershock planes, and Plane-C seems to be a conjugate plane
with the double aftershock planes. Aftershock activities in the double aftershock planes are higher than that in the conjugate plane. Figure 10(c) shows that the aftershock activity in a deep area of Plane-A (a hatched area in the figure) is very low, while in the shallow portion of Plane-A and in a whole area of Plane-B the activity is high. The main shock appears to be located in Plane-C. Aftershock activity is very low in the vicinity of the main shock. In Figure 10(d), the aftershock activity on Plane-C is higher than that on Plane-A and Plane-B.

Next, in order to inspect the aftershock distribution just after the main shock, we select 195 hypocenters which occurred during the period from March 29 to April 6, 1982 when both the land stations and the OBSs were operated at the same time. Figure 11 shows the distribution which is called a short-term distribution. The figure shows the same seismic pattern as that of the long-term distribution. Therefore, it is found that the aftershocks just after the main shock took place on three dipping planes revealed in the long-term distribution.

6. Discussion and Conclusions

The aftershock distribution of the Urakawa-oki earthquake has been studied by MORIYA et al. (1983 a, b), SUZUKI and MOTOYA (1983), and IWASAKI et al. (1983). They applied laterally homogeneous velocity structures with some station corrections to the hypocentral determination, and showed the existence of two conjugate dipping planes in the aftershock distributions. However, the results of the inversion in Sec. 3 distinctively show that lateral velocity variations are very strong and the velocity distribution is well correlated with the geology and the Bouguer anomalies in the study region. Therefore, the hypocentral determination should be carried out under the 3-D velocity structure in order to obtain the detailed hypocentral distribution. As shown in Sec. 5, by using the 3-D velocity structure, we successfully separate two dipping parallel planes, Plane-A and Plane-B, from the aftershock distribution, which have been previously supposed to be an identical plane. Plane-C is identical to another conjugate plane in the previous works.

Figure 12 shows a relation between the aftershock distribution in Fig. 10 and the velocity distribution obtained by the inversion. The figure shows that the aftershocks occur in a region with P-wave velocities higher than 5 km/s. In addition, the trace of Plane-A seems to correspond to the contour of 6 km/s velocity. Probably, aftershocks cannot take place in a region with P-wave velocities lower than 5 km/s which corresponds to the Ishikari Belt and the Yezo Group.

Next, we consider a relation between the location of the main shock and the three planes. TAKEO et al. (1983) showed that the main shock had a reverse fault mechanism from P wave first motions; the directions and the locations of two nodal planes they estimated correspond approximately to Plane-A and Plane-C in this study. They concluded that the levelling data measured along the coastal line in the aftershock region were comparatively more satisfied by the fault plane on Plane-A than that on Plane-C from an analysis of coseismic crustal deformations. However,
we consider that the levelling data do not have a resolving power to discriminate whether the fault plane of the main shock is Plane-A or Plane-C, for two reasons described below: one is that the area, where the levelling data were obtained, was only a part of the total area of the aftershock region, and the other is that the direction of the levelling survey was parallel to the null axis of the main shock mechanism, if Plane-C is the fault plane. Assuming that the aftershock distribution shows the fault plane, the top depth of Plane-A and Plane-B (a few kilometers as shown in Fig. 7) is much shallower than that of TAKEO et al. (1983) which is estimated to be 12 km depth. MORIYA et al. (1983a) indicated from an analysis of 100 focal mechanisms that most of the aftershocks had a reverse fault with the compressional axis perpendicular to the coastal line and 17% of them have a normal fault; the former was widely distributed over the aftershock region and the latter was concentrated only in Plane-A. If the mechanism solutions of the aftershocks located near the main shock are of the same type as that of the main shock, it can be said that the main shock may have taken place on Plane-C. As shown in Fig. 10(c), the hypocentral distribution seems to indicate that the main shock is located on Plane-C. SUETSUGU and NAKANISHI (1987) also investigated a relation between the
From a viewpoint of tectonics in the study area, MIYAMACHI and MORIYA (1984) showed that the low velocity zone (LVZ) is subducted, being strongly affected by a compression derived from a collision between the Kurile and Northern Honshu Japan Arcs beneath the area. Along a line perpendicular to the coastal line in the study area, a schematic model can be presented as shown in Fig. 13 on the basis of results from DEN and HOTTA (1973), MIYAMACHI and MORIYA (1984), MORIYA (1986) and this study. Plane-C is approximately located along the lower boundary of the LVZ, suggesting that the main shock took place along the
structural boundary. On the other hand, the double aftershock planes, Plane-A and Plane-B, cross the LVZ, and the area with the low seismicity on Plane-A corresponds to that of the LVZ. According to these facts, we consider that the occurrence of the Urakawa-oki earthquake in 1982 is closely related to the existence of the subducted low velocity zone derived from a collision between the Kurile and Northern Honshu Japan Arcs. The number of aftershocks in Plane-A and Plane-B is also greater than that in Plane-C. This fact may be explained by an assumption that a material of LVZ is more brittle than that of the surrounding region with a relatively high velocity.

Takeo et al. (1983) also pointed out that the Urakawa-oki earthquake is multiple shocks composed of at least five events from an analyses of the far-field P waveform. The hypocentral distribution obtained in this study suggests the fault model of the Urakawa-Oki earthquake to be complicated. This complexity is probably caused by the complicated velocity structure. Therefore, we conclude that the first event of the main shock is certainly triggered along Plane-C, which is the structural boundary, by a compression derived from a collision between two arcs, and there is a possibility that some of the subsequent events may take place along other planes (Plane-A and Plane-B) conjugate to Plane-C.

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