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DEEP SUBSURFACE STRUCTURE ESTIMATED BY MICROTREMORS ARRAY OBSERVATIONS AND GRAVITY SURVEYS IN KASHIWAZAKI AREA, JAPAN

Hiroyuki Goto[i], Chika Takahashi[ii], Yayoi Ishii[iii], Su-Qun Ling[iv], Ken Miyakoshi[v],
Hitoshi Morikawa[vi], Yoshiki Sato[vii], Sumio Sawada[viii], Yoshikazu Shingaki[ix],
Yuto Suzuki[x], Daisuke Takabatake[xi] and Masato Joshima[xii]

ABSTRACT

Kashiwazaki City and Kariwa Village, located in the north central part of Japan, were the major areas damaged during the 2007 Niigataken Chuetsu-oki Earthquake (M_w 6.6). Ground motion records for the area indicate the effects of a deep subsurface structure. Microtremor array observations were performed at three sites and gravity surveys at 120 sites in the Kashiwazaki area. The gravity basement estimated from the gravity surveys exists at a depth of about 1000 m around Kariwa Village Hall, and becomes shallow in the southwestern direction around Kashiwazaki City Hall. The gravity basement is compared with the results estimated by the microtremor array observations and validated at two sites, namely, KST and KVH (Kariwa Village Hall). The peak period of the calculated response spectra at Kashiwazaki Village Hall corresponds to the peak period of the observed *H/V* spectra during the main shock.

Key words: (2007 Niigataken Chuetsu-oki Earthquake), earthquake, geophysical exploration, (gravity survey), microtremor, seismic response (**IGC**: C2)

INTRODUCTION

The 2007 Niigataken Chuetsu-oki Earthquake (M_w 6.6) struck the north central part of Japan on July 16, 2007, its epicenter being located offshore of the Chuetsu region in Niigata Prefecture. In this earthquake, 14 people were killed and more than 1000 buildings were reported to have been destroyed or severely damaged (Cabinet Office, 2007). Most of the damage was concentrated in the Kashiwazaki area, which includes Kashiwazaki City and Kariwa Village. Several records of strong ground motion records during the main shock, e.g., K-NET records provided by the National Research Institute for Earth Science and Disaster Prevention (NIED), are available for this area. Longer peak periods and the larger peaks of pseudo velocity response spectra (pSv) were observed in the plain region of the Kashiwazaki area (Fig. 1). Several reports (e.g., Irikura et al., 2007; Koketsu et al., 2007; Yoshida et al., 2007) explain the reason for these peaks with combinations of the effects generated by the multi-asperity source process, the subsurface structure, and liquefaction.

In particular, a peak period of 2.6 seconds and a peak value of 616 cm/s for the pseudo velocity response spectra were obtained from the ground motion records at Kariwa Village Hall. From a comparison with the *H/V* spectrum, which is a Fourier spectrum of the horizontal motion normalized by the vertical motion, for the main shock and the aftershock records, the peak period during the main shock was seen to last longer than that during the aftershock, as shown in Fig. 2. The dominant period of the dynamic response for the subsurface structure shifted to a longer period because the *H/V* spectra emphasized a horizontal dynamic response by eliminating the source and the pass effects from the observed spectra. These effects imply the nonlinear behavior of the soil.

[i] Assistant Professor, Disaster Prevention Research Institute, Kyoto University, Japan (goto@catfish, dpri.kyoto-u.ac.jp).
[ii] Graduate Student, Interdisciplinary Graduate School of Science and Engineering, Tokyo Institute of Technology, Japan.
[iii] Graduate Student, Department of Urban Management, Kyoto University, Japan.
[iv] Geo-Analysis Institute, Japan.
[v] Geo-Research Institute, Japan.
[vi] Associate Professor, Interdisciplinary Graduate School of Science and Engineering, Tokyo Institute of Technology, Japan.
[vii] Graduate Student, Department of Urban Management, Kyoto University, Japan.
[viii] Professor, Disaster Prevention Research Institute, Kyoto University, Japan.
[ix] Graduate Student, Department of Urban Management, Kyoto University, Japan.
[x] Graduate Student, Interdisciplinary Graduate School of Science and Engineering, Tokyo Institute of Technology, Japan.
[xi] Graduate Student, Department of Urban Management, Kyoto University, Japan.
[xii] Advanced Industrial Science and Technology, Japan.

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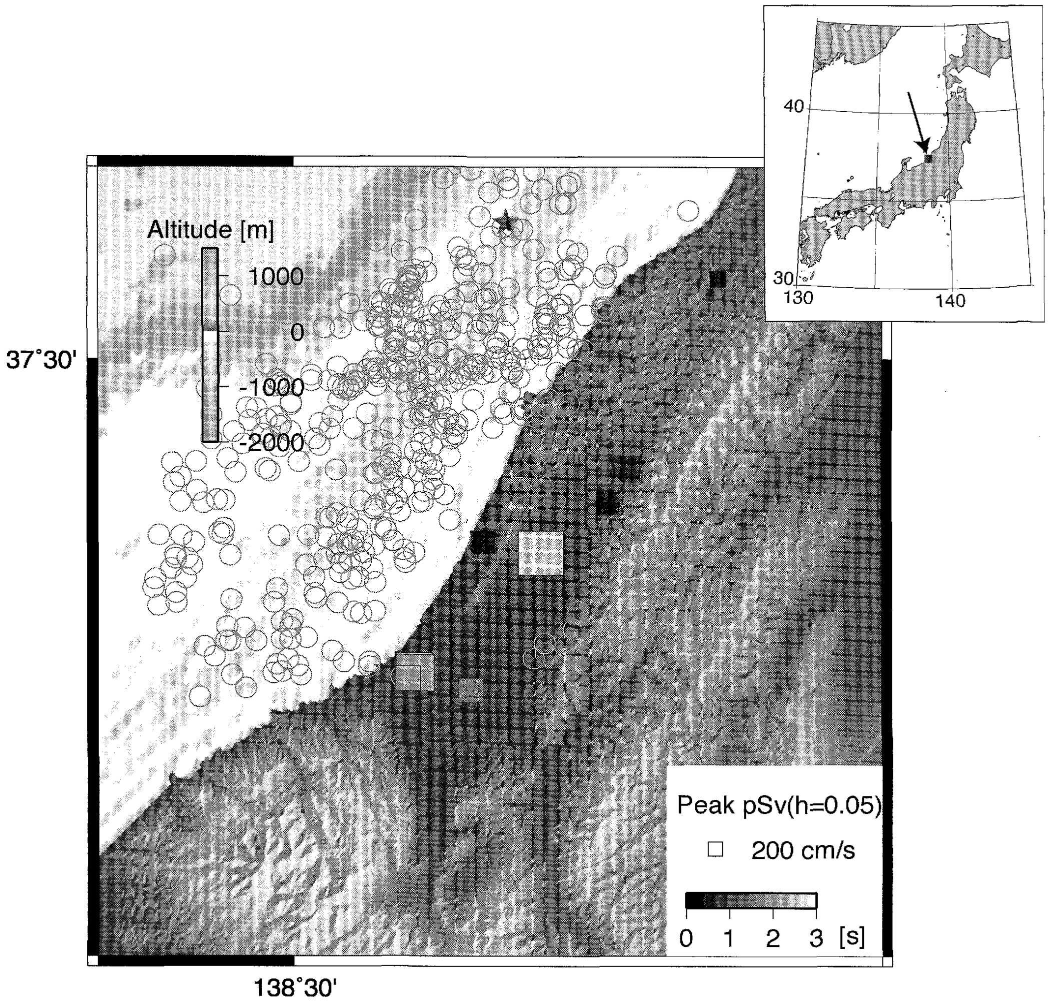


Fig. 1. The hypocenter of the 2007 Niigataken Chuetsu-oki earthquake [★] and the hypocenter of the aftershocks on July 16–17 [○] are plotted on this map of the Kashiwazaki area. The larger peak values (size of the squares) and the longer peak periods (brightness of the squares) of the pseudo velocity response spectra (pSv, damping factor 5%) calculated from the strong ground motion records during the main shock are obtained in the plain area of Kashiwazaki area

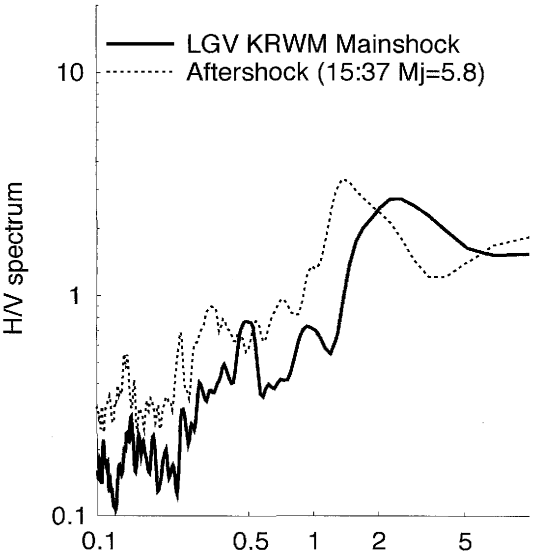


Fig. 2. The spectrum ratios of the horizontal component over the vertical component are compared between the main shock and the aftershock ($M_j=5.8$) at Kariwa Village Hall

However, the nonlinear effects alone do not explain the reason for such large pSv values. It is necessary to discuss not only the nonlinear effects, but also the effects of the deep subsurface structure which lies below the engineering basement.

In addition, during the 2004 Niigataken Chuetsu Earthquake, a strong ground motion observed at K-NET site NIG018 of NIED, which is located in downtown Kashiwazaki City, contained a peak period of the *H/V* spectra of about 1.9 seconds and a longer duration than the peak periods at the other sites located 40 km from the epicenter. The phases may consist of a basin-edge-induced surface wave which depends on the arrival direction of the seismic wave. This is because such a long duration was not observed at the other sites or during the 2007 Niigataken Chuetsu-oki Earthquake and its aftershocks. The characteristics of the records imply the effects of the deep subsurface structure. Information on the deep subsurface structure in the Kashiwazaki area is important to the investigation of the amplification of the strong ground motions brought about by the deep subsurface structure.

We performed microtremor array observations and gravity surveys around the Kashiwazaki area in order to model both the velocity structures and the gravity basement. In this paper, we make independent estimations of the models for the ground structure from the microtremor array observations and from the gravity surveys. To verify the obtained models, we compare them in a kind of blind test.

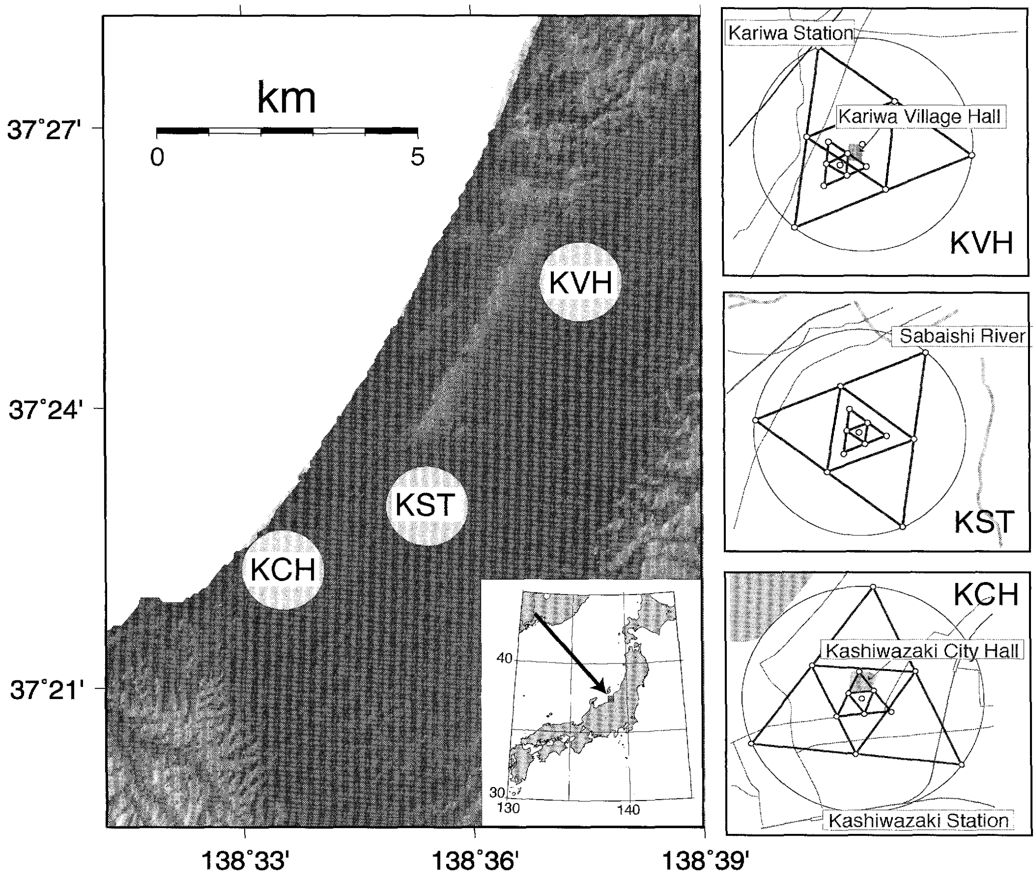
MICROTREMOR ARRAY OBSERVATIONS

Microtremors have been observed at three sites in the Kashiwazaki area, through observations performed on August 12–13, 2007, in order to investigate 1-D velocity structures. Figure 3 shows the locations of the observation sites, namely, KCH, KST and KVH, and a detailed layout of each array. KCH (138°34′E, 37°22′N) is set around the Kashiwazaki City Hall, where a seismic intensity of 6+ on the JMA seismic intensity scale (almost equivalent to a seismic intensity of 10 on the Modified Mercalli Intensity Scale) was observed. Another seismometer, belonging to NIED and named K-NET NIG018, located near Kashiwazaki City Hall, also observed the waveform of a strong ground motion of about PGV = 100 cm/s. KVH (138°37′E, 37°25′N) is set around Kariwa Village Hall, where a seismic intensity of 6+ was observed. KST (138°35′E, 37°23′N) is set between the KCH and KVH sites, where a deep geological layer structure was estimated from the deep borehole data (Kobayashi et al., 1995).

Each microtremor observation site included two arrays with different radii. Hereafter, we call these arrays S-array and L-array. Both arrays consist of seven seismographs. Three of the seismographs formed equilateral triangles, and one is set on the centroid. The other three seismographs are set on the middle points of each edge. The maximum radii of the equilateral triangles are 200 m for the S-array and 800 m for the L-array.

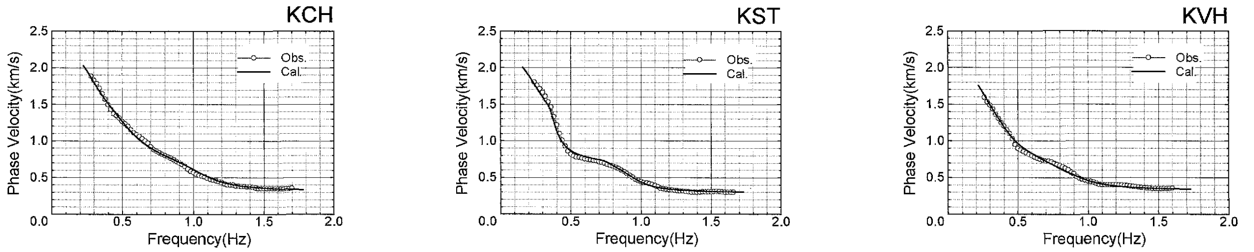
Three-component velocity-type seismographs with a natural period of seven seconds were used. We have made simultaneous observations for 60 minutes at seven sites for each size of array. The microtremors were recorded by digital recorders with a resolution of 16 bits. The records were synchronized with the slow code generated by Global Positioning System (GPS) clocks. A sampling rate of 100 Hz and a low-pass filter of 3 Hz were adopted.

We estimated the phase velocities for the fundamental mode of the Rayleigh waves from the vertical compo-



(a) Locations of the microtremor array observations (b) Array layout

Fig. 3. The microtremor array observation sites are plotted in the Kashiwazaki area. The circles represent the circumscribed circles of the L-array (radius 800 m). The KVH array contains the Kariwa Village Hall, and the KCH array contains Kashiwazaki City Hall. KST is located between KVH and KCH. The figures on the right show detailed layouts of the S-array and the L-array for each site



V_s (m/s)	V_p (m/s)	ρ (t/m^3)	top depth of the layer (m)
100*	300*	1.82*	0*
210*	900*	1.90*	3*
190*	1600*	1.80*	5*
350*	1740*	1.81*	20*
700	2080	1.96	145
1000	2380	2.07	295
1400	2860	2.19	460
1750	3350	2.28	525
2050	3770	2.36	1685
2600	4650	2.48	2785

(a) KCH (Kashiwazaki City Hall)

V_s (m/s)	V_p (m/s)	ρ (t/m^3)	top depth of the layer (m)
310	1730	1.79	0
775	2170	2.00	150
920	2300	2.05	435
1700	3280	2.27	1075
2000	3700	2.35	1575
2500	4420	2.46	2775
—	—	—	—
—	—	—	—
—	—	—	—
—	—	—	—

(b) KST

V_s (m/s)	V_p (m/s)	ρ (t/m^3)	top depth of the layer (m)
350	1740	1.81	0
700	2080	1.96	165
1000	2380	2.07	450
1400	2860	2.19	925
1700	3280	2.27	1355
2000	3700	2.35	1945
2400	4270	2.43	2975
—	—	—	—
—	—	—	—
—	—	—	—

(c) KVH (Kariwa Village Hall)

Fig. 4. The microtremor array observation results for the three sites, KCH (left), KST (middle), and KVH (right), are displayed. The estimated phase velocities from the microtremor array observations and the calculated fundamental modes of dispersion curves of the Rayleigh waves are plotted across the top row. The model parameters, S-wave velocity V_s, P-wave velocity V_p, density ρ and the top depth of the layer, are listed at the bottom of each figure. At the KCH site, the shallower portion ($<$20 m) of the model parameters indicated by "*" was constrained by the well-logging data

nents at three sites by using the SPAC (Spatial Auto-Correlation) method (Aki, 1957). The estimated frequency ranges for the phase velocities were about 0.7–1.7 Hz for the S-array and 0.3–0.8 Hz for the L-array. The 1-D structure models down to a depth of about 2 km were estimated by fitting of the observed and the theoretical phase velocities through a nonlinear inversion using a genetic algorithm (GA) (Cho et al., 1999). At the KCH site, a shallow portion (shallower than 20 m at the K-NET stations) of the underground structure model was constrained by the well-logging data. We referred to the geological structural model estimated by JNES, Japan Nuclear Energy Safety Organization, (2005), in order to set the depth of the basement (V_S: 2.4–2.5 km/s) beyond 2 km. JNES (2005) also provided information on the S-wave velocities (V_S; 1.4 km/s, 1.7 km/s, and 2.0 km/s) at each layer corresponding to the geological structure. By referring to the S-wave velocities, we estimated the thickness of each layer using GA. In one estimation using GA, 5000 generations were simulated. We selected one of the results, namely, the one which had the smallest residuals in five estimations. The P-wave velocity and the density were assumed as a function of the S-wave velocity by using the relationship of Ludwig et al. (1970).

Figure 4 shows the comparisons of the estimated phase velocities and the dispersion curves calculated from the modelled 1-D structures listed at the bottom of each panel. The estimated phase velocities are distributed between 0.4–1.8 km/s for 0.3–1.7 Hz with a normal dispersion. The shape of the dispersion curve for KST is similar to that for KVH, while the phase velocities in the frequency range of 0.3–1.0 Hz for KCH are faster than those for both KST and KVH. As expected from the characteristics of the phase velocities, the top depth for the 1000 m/s S-wave velocity layer of the KCH model is shallower than that of the KVH model. The KST model has no 1000 m/s layers; however, the top depth for the 920 m/s layer is almost the same as the depth of the layer for the KVH model.

GRAVITY SURVEY

The area for the gravity observations was located 37°11′N to 37°34′N and 138°28′E to 138°42′E: 45×26 km of NS-EW. The observation sites were set at 2 km intervals, but in some cases when the gravity anomaly changed rapidly, further observation sites were set at 1 km intervals. We spent five days for the observations, from July 28 to August 1, 2007, and obtained the values of gravity at 120 sites.

We used an Automated Burris Gravity Meter (Serial No.: B026) made by the ZLS Corp. and a CG-3M Automated Gravity Meter (Serial No.: 408258) made by Scintrex Ltd. The positions of the observation sites were determined by a differential survey by GPS. Errors in the positions were less than 1 m in the horizontal and vertical directions.

To ensure the stability of the analysis, we introduced the gravity data obtained by the Geological Survey of Japan (2004), Takahashi et al. (2006), and Okamura et al. (1994). We calculated the Bouguer anomaly using the gravity data from 746 sites in the area under consideration whose locations are indicated by circles (○ and ●) in Fig. 5. The open circles (○) indicate the 120 sites where

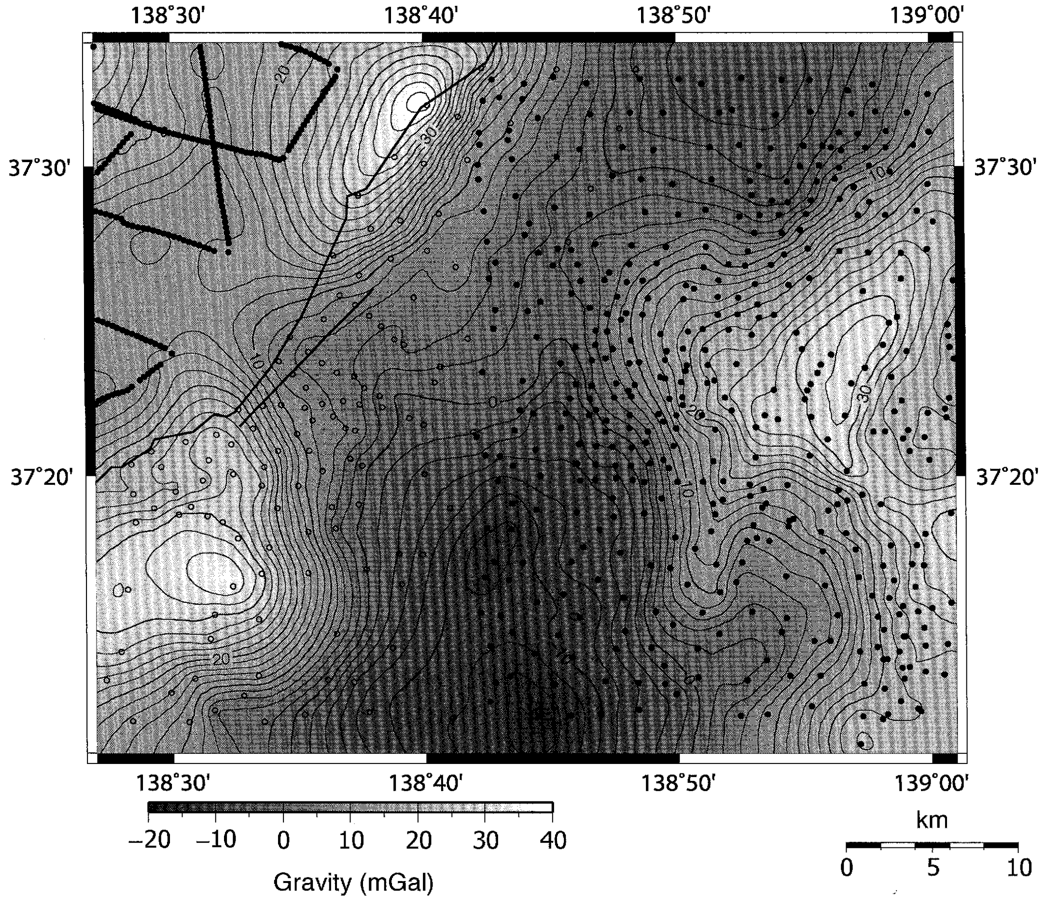


Fig. 5. The Bouguer anomaly distribution obtained from the gravity surveys is presented. The assumed density is 2.40 t/m^3, and the contour interval is 2 mGal. The open circles (○) indicate the observation sites of the gravity anomaly in this survey, and the dark circles (●) indicate the other gravity anomaly data

the gravity anomaly was observed in this survey, and the dark circles (●) indicate the locations of the other gravity anomalies that are available. On the basis of some established methods (Rikitake et al., 1965; Nettleton, 1976), the density of the basement was determined as 2.40 t/m^3, which corresponds to the density of the Teradomari formation shown in Furumura et al. (2005). After corrections were made to the height and the drift of the instrument, the tide, the free-air, the Bouguer and the terrain, the Bouguer anomaly map was obtained as shown in Fig. 5.

Applying an upward-continuation filter with a band-pass of 50–6000 m to the Bouguer anomaly, we estimated the 3-D gravity basement under the assumption that the ground consists of two layers and that the sediment is homogeneous with a density of 2.00 t/m^3, which is the average density of the sediment above the Teradomari formation.

To obtain a realistic model of the gravity basement, we considered removing the contribution for the Bouguer anomaly from the upper mantle and constraining the depth to the bedrock. For the former, a band-pass filter was applied to the Bouguer anomaly. For the latter, we gave some control points, that is, one site from the array observations of the microtremors (Yamanaka et al., 2006), 19 sites from the deep boreholes which reach the bedrock (Japan National Oil Corp., 1970, 1997, 1999; Kobayashi et al., 1986, 1989, 1991, 1995; Yanagisawa et al., 1986), one site from P- and S-wave reflection experiment (Furumura et al., 2005), and four sites from the mountainous area where the bedrock appeared on the surface (Kobayashi et al., 1986, 1991). The upper boundary of the Teradomari formation was used as the constraint of the gravity basement at the control points, which are indicated by triangles in Fig. 6. The 3-D shape of the gravity basement, which was calculated using the method proposed by Komazawa (1995), is shown in Fig. 6. To show the sensitivity of the density contrast between the sediment and the basement, we present some different densities for the sediment. As a result, the depth to the basement varies by 10% at the maximum in the case of a 10% variation in sediment density.

DISCUSSION

Both the microtremor observations and the gravity surveys are focused on deep subsurface structures. A comparison of these estimated structures is useful when discussing the validity because the models were independently constructed.

Firstly, we compare the profile of the 1-D models with the depth to the gravity basement. Figure 7 shows the

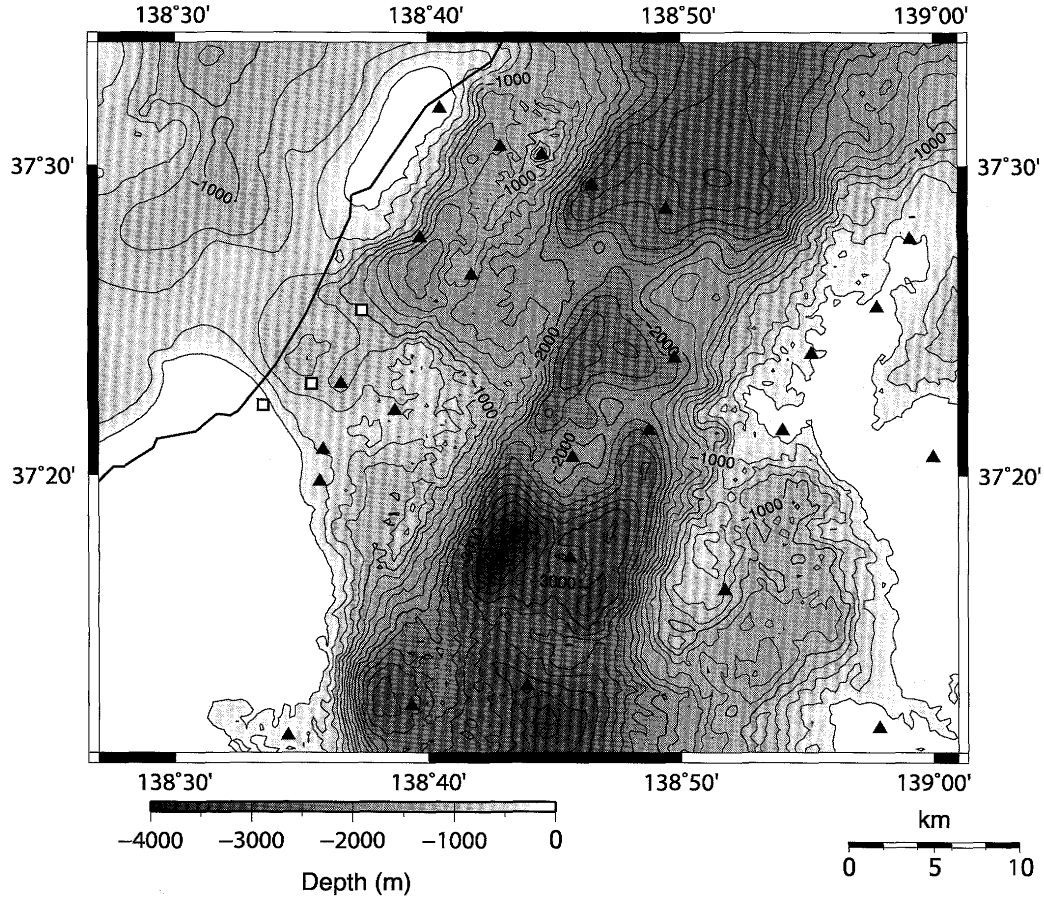


Fig. 6. The depth of the gravity basement estimated from the Bouguer anomaly is given along with some information at the control points indicated by triangles (▲). The density of the sediment is set to be $2.0\ \mathrm{t/m^3}$, and the contour interval is 200 m. The open squares (□) indicate the array observation sites of the microtremors shown in Fig. 3

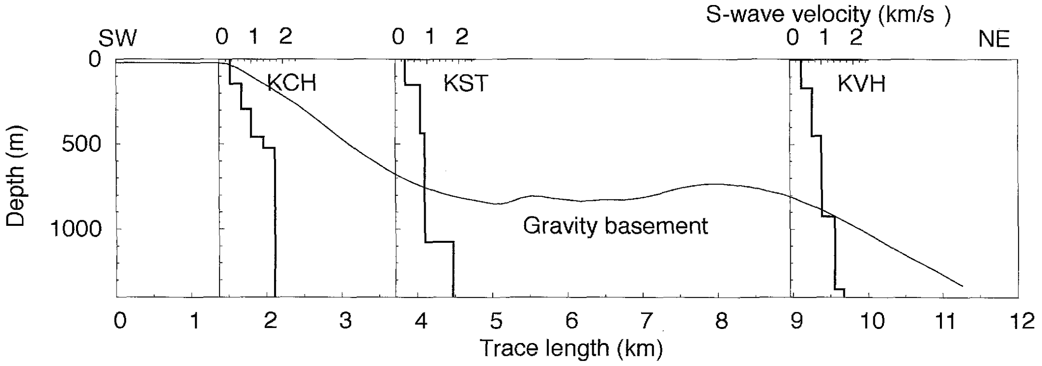
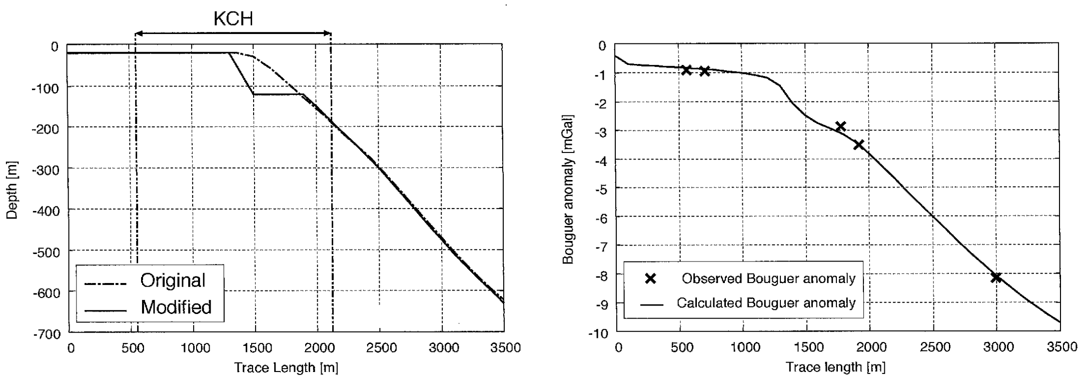


Fig. 7. A comparison of the gravity basement and the 1-D profile obtained from the array observation of the microtremors on the line from (138° 33′E, 37°22′N) to (138°38′E, 37°26′N) is shown. It is seen that the gravity basement corresponds to the trend of the upper boundary of the layer with an S-wave velocity of 1700 m/s

cross section of the gravity basement along the line from (138°33′E, 37°22′N) to (138°38′E, 37°26′N), where the line passes close to the points of the center of the L-array at KVH and KCH. In the south-western part, between KCH and KST, it is found that the depth to the gravity basement becomes shallow and its shape is very steep. On the other hand, the gravity basement in the north-eastern part between KST and KVH is deep with a moderate slope. We can say roughly that the macroscopic 3-D shape of the gravity basement is consistent with the 1-D models obtained from the microtremor array observations. From Fig. 7, it is observed that the top depth of the layer with an S-wave velocity of 1400 m/s is close to the depth of the gravity basement at site KVH. There is, however, no layer with a velocity of 1400 m/s at the KST site. The top depth of the layer with a velocity of 1700 m/s is chosen for the gravity basement, because the layer with a velocity of 1700 m/s commonly appears at all three



(a) Gravity basement

(b) Bouguer anomaly

Fig. 8. An alternative shape for the gravity basement is given which satisfies the observed Bouguer anomaly data indicated by cross marks (×) in the right panel. In the left panel, the original and the modified shapes of the gravity basement are shown by dot-dashed and solid lines, respectively. The solid line of the right panel denotes the Bouguer anomaly calculated analytically. From this figure, the depth of the gravity basement can be more than 100 m around the KCH site

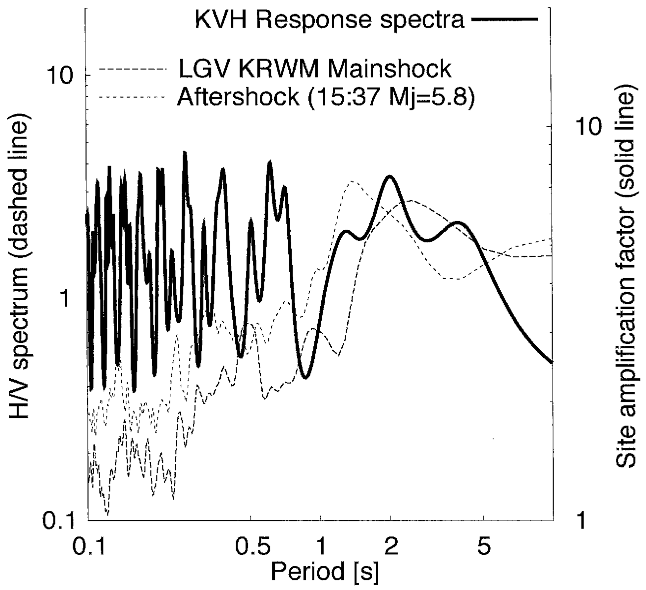


Fig. 9. The response spectra calculated from the estimated structure of KVH by assuming the SH wave field is presented. The spectrum ratios of the horizontal component over the vertical component for the main shock and the aftershock ($M_j=5.8$) at Kariwa Village Hall overlap. The peak period of the response spectra corresponds to the peaks of the *H*/*V* ration for the main shock

array sites. There are different opinions about the S-wave velocity for the Teradomari formation such as 1400 m/s (Suzuki et al., 2005), 2200 m/s (Furumura et al., 2005), and so on. From this, 1700 m/s seems to be an appropriate value for the Teradomari formation.

The depth to the gravity basement for KCH is much shallower than that shown in the microtremor results. The depth cannot be explained even if a considerably low density is adopted. From this, we have to say that the model from the gravity surveys does not agree with the model from microtremor array observations at KCH. It is considered that this discrepancy appears for several reasons. One is the lower resolution at the observation sites of gravity on the steeply sloping area. To confirm this first reason, we carry out a simple numerical experiment.

Figure 8 shows an example of an alternative shape for the gravity basement obtained from the 2-D analysis. The basement is the deepest case which satisfies the Bouguer anomaly. In the left panel of this figure, dot-dashed and solid lines indicate the original and the modified shape of the gravity basement, respectively. We set a step-like structure around the KCH site. The right panel shows the Bouguer anomaly calculated analytically from the modified one. It is recognized that the calculated Bouguer anomaly satisfies the observed data indicated by the cross signs.

The resolution of the gravity data may explain a part of the discrepancy at KCH, but it is not sufficient. Thus, we have to consider the other reasons for the discrepancy. One is the effect of the sea water which has very low density. For this purpose, a new technique should be developed to estimate the ground model consisting of media with three different densities. However, this kind of work is beyond the scope of this study. Thus, it will be left for future research. We wanted to highlight the importance of the dense observations around the steep structure and the possibility of the effect of sea water.

Another reason for the discrepancy is that the SPAC method assumes a horizontal layered structure beneath the observation sites of the microtremor arrays. As shown in Fig. 8, the gravity basement is not horizontal. Kudo et al. (2004) explained that the estimated structure models are affected by the 3-D shape of the basement. Uebayashi et al. (2009) noted that the depth of the basement estimated from the microtremor arrays became deeper than that in the borehole P-S logging data in the region with the inclined basement on the Osaka plane. They summarized that body waves and a higher mode of Rayleigh waves generated by the 3-D shapes of the basement are included in the microtremor observations, and that they apparently decrease the estimated phase velocity. For KCH, a deeper structure is estimated from the microtremor array observations than from the gravity basement. We just mention that the discrepancy may be caused by the 3D shapes of the basement.

At Kariwa Village Hall, a response spectra of the deep subsurface structure was calculated by the estimated structure of KVH by assuming an SH wave field. The waves were vertically input to the boundary between the fourth and fifth layers at a depth of 1355 m, and the damping coefficient was zero. The peak period of the H/V spectrum for the observed ground motion during the main shock clearly corresponds to the peaks of the amplification spectra, as shown in Fig. 9. Although we mentioned that the peak periods of the H/V spectrum during the main shock may contain the nonlinear effect, the amplification around the peak period does not come from only the nonlinear effect, but also is emphasized by the effects of the deep subsurface structure. Therefore, the nonlinear effects and the deep structure effects might have overlapped and have caused the large pSv peak of the strong ground motion at Kariwa Village Hall.

CONCLUSIONS

We performed microtremor array observations and gravity surveys in the Kashiwazaki area in order to identify the deep subsurface structure. The gravity basement estimated from the gravity surveys forms a deep structure around KVH and KST sites at a depth of about 1000 m, and becomes shallow in the south-western direction around the KCH site. The estimated models from both surveys are valid around the KST and KVH sites. The calculated response spectra at the KVH site correspond to the observed H/V spectra during the main shock. It is considered that the deep structure amplified the strong ground motion at Kariwa Village Hall, and might have caused the large peak values of pSv there.

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