Construction and Waveform Testing of the Crustal and Basin Structure Models for Southwest Japan

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SUMMARY:
For simulation of the long-period ground motions due to hypothetical subduction earthquakes we constructed three-dimensional crustal and basin velocity structure models of Southwest Japan. Crustal model is compiled from numerous physical explorations conducted in land and offshore areas and observational studies of natural earthquakes. Low-velocity uppermost crustal layers have added on the seismological basement. Thickness of the layers has been calibrated through a full waveform inversion iterative analysis. For the case of a model having thin surface layers, we proposed and used numerically effective inversion method without time consuming calculation of the Jacobian. Then, crustal model is combined with a finely elaborated sedimentary/basin structure model of Koketsu et al. (2008) and revised in this study using R/V spectral ratio method, geological information and gravity anomaly data. The final result is found to have good agreement with the results from other physical explorations; e.g. seismic refraction.

Keywords: velocity structure model, full waveform inversion, long-period ground motion

1. INTRODUCTION

Long-period ground motions (2 - 20 s) are responsible for observed damage of structures like high-rise building, oil tanks, and elevators. Basin response is effective in amplifying and extending of long-period ground motions (e.g. Aoi et al., 2008). Long-period ground motions are sometimes observed even when the earthquake source is located far from the target region, because due to cylindrical geometrical spreading of surface waves long-period ground motion is weakly attenuated. In combination with extended source of subduction earthquakes it enables to propagate over a great distance almost without attenuation. Huge sedimentary basins in accretion prism of subduction zones provide another mechanism of surface waves generation through leakage of basin waves on the landward edge of the accretion basin (e.g., Yamada and Iwata, 2005). Thus, generation and propagation of long-period ground motions from subduction earthquakes is extremely complicated process that involves 3D features of velocity structure and cannot be simulated by a simplified 1D or 2D method.

Accurate basin and crustal velocity structure model is necessary for simulation of long-period ground motions. Basin models were intensively generated by many researchers in Japan during last decade: e.g. (Koketsu et al., 2009) for the Kanto basin, (Kagawa et al., 2004) for Osaka basin and surrounding basins. Various methods were employed, e.g. seismic reflection/ refraction profiling, gravity anomaly analysis, surface geology analysis, borehole *PS*-logging, microtremor array and H/V measurements, and earthquake waveform simulations. These and other results are compiled into a single structure model by Koketsu et al. (2008).

Koketsu et al. (2009) systemized various approaches to the sedimentary structure modeling into a recipe. For crustal structure model, which is an important link between earthquake source model and sedimentary basin model, similar work were done by Petukhin and Kagawa (2006) and realized in

(Iwata et al., 2008). Here we will use these approaches for construction and calibration of 3D velocity structure model for simulation of long-period ground motions from Nankai earthquake. Fig. 1 shows the model area of the crustal velocity structure that includes the source area of the Nankai and Tonankai earthquakes. The model covers Southwest Japan between 129.5–138°E and 31–36°N.

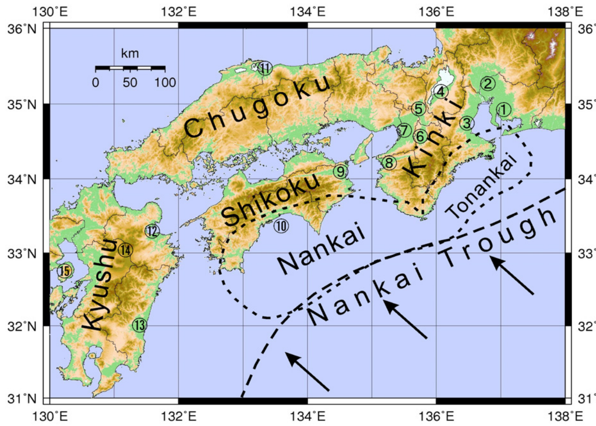


Figure 1. Target area for velocity structure modeling in western Japan. Dashed lines – Nankai Trough and source area of Nankai and Tonankai earthquakes. Arrows indicate direction of the Philippine Sea plate subduction. Circled numbers are major sedimentary basins and structures: 1 – Mikawa, 2 – Nobi, 3 – Ise, 4 – Ohmi, 5 – Kyoto, 6 – Nara, 7 – Osaka, 8 – Wakayama, 9 – Tokushima, 10 – Kochi, 11 – Yonago, 12 – Oita, 13 – Miyazaki, 14 – Aso-Kushu volcanic area, 15 – Unzen volcano.

2. CRUSTAL STRUCTURE MODEL

2.1 Initial Model

3D crustal structure model for this study were combined from the results of (1) Iwata et al. (2008) for Moho, Conrad, Seismic basement, and lower interface of Accretion prism layer, and (2) Baba et al. (2006) for the soft oceanic sediments (oceanic sedimentary layer 1), and for the subduction plate interfaces. Fig. 2 shows a schematic cross-section of the crustal velocity structure model. It reflects the main principles of velocity modeling: (1) intensive use of the results of the ocean bottom seismometers (OBS) refraction profiles and the multi-channel seismic (MCS) reflection profiles in offshore areas, for the accretion prism (sedimentary wedge) and oceanic crust modeling; (2) employing seismicity data for tracing the seismogenic slab and seismogenic upper crust; (3) intensive use of the receiver function inversion results for tracing Moho, and Conrad and slab in the aseismic areas; (4) employing deep seismic profiling results for inland areas whenever possible; (5) waveform inversion for modeling of crustal uppermost layers; and (6) employing 1D velocity models used for the hypocenter locations and generalized seismic tomography results for the remaining areas not covered by the data. Fig. 3 shows the depth distribution maps of the lower interface of Seismic Basement/Oceanic Sediments 2 layer, Conrad interface, Moho interface and the upper interface of the Oceanic Layer 2 in the studied area.

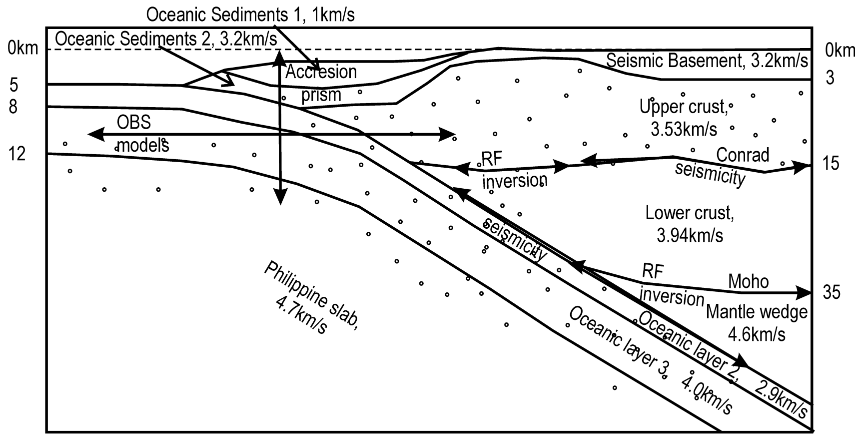


Figure 2. 3D crustal velocity structure model (schematic). *S*-wave velocity values and approximate layer depth are given for reference.

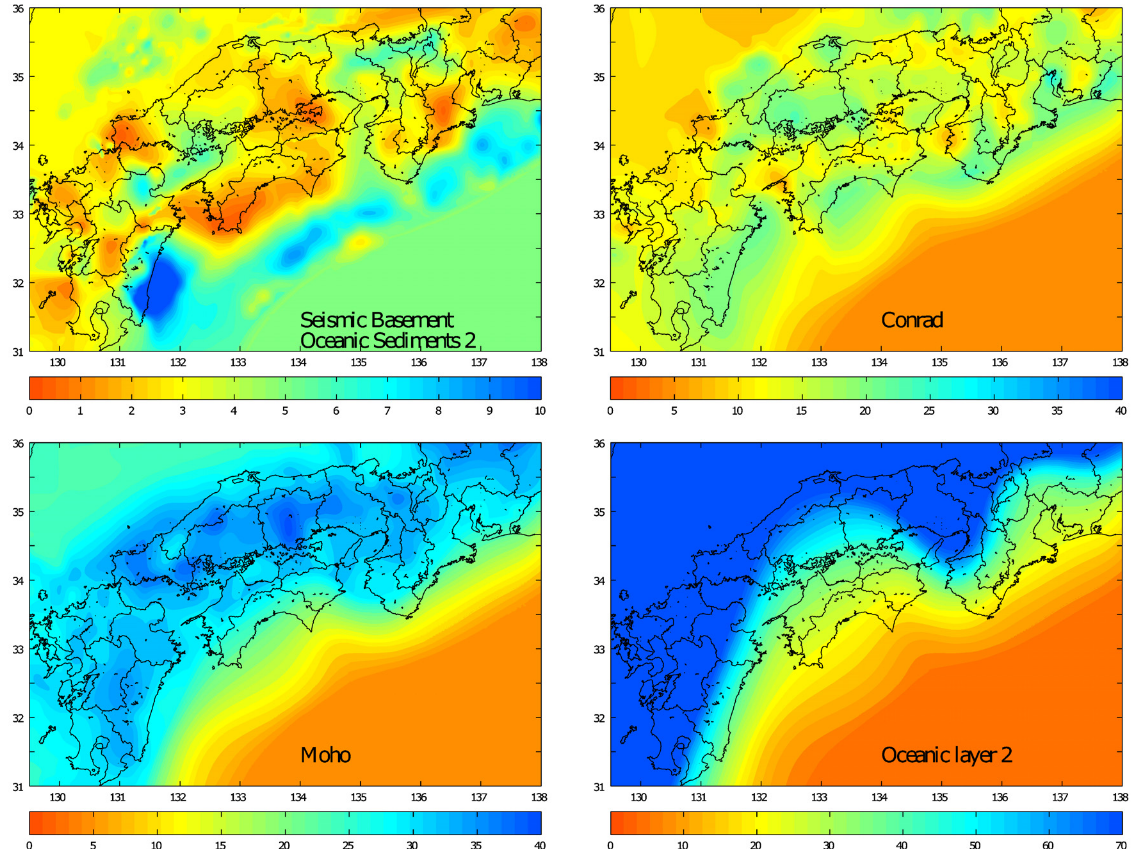


Figure 3. Depth distribution of crustal velocity structure interfaces: lower interface of Seismic Basement/Oceanic Sediments 2 layer, Conrad, Moho, and upper interface of Oceanic Layer 2. For Oceanic Sediments 1 see Fig. 11 below.

2.2. Uppermost Crustal Layer

For calibration and validation of crustal velocity structure model by waveform simulation we choose 12 medium size earthquakes (M4.5-7). Source locations are shown in Fig. 4. Source mechanism, depth and rise time T_r are critical parameters for getting good waveform fit. We used F-net or JMA determinations of source mechanism and depth. For estimation of T_r multi-time window source inversion results were used when it was possible. For the rest of sources we estimated T_r from the displacement pulse width for records nearest to epicenter. The shortest target period in the simulations was 2.8 s. The wave fit is good except of a general tendency for underestimation of amplitudes, which is critical point for the strong ground motions prediction.

For outcrops of plutonic rocks (e.g. granite) it is generally assumed high P- and S-velocities, like in the Seismic Basement layer. In contrast, results of seismic exploration show that short distance P-arrivals from artificial sources at the surface, are better explained by model having Uppermost Crustal Layer (UMC) with P-velocities considerably lower than value typical for the seismic basement, 5.5 km/s: Vp = 4.1-4.3 km/s (Vs = 2.4 km/s) for granite provinces in studied region (Ito et al., 2005). In contrast to sedimentary layers, for which Vs velocities clearly change with sediments type, age and depth of layer, it is difficult to associate velocity values in rock layers with rock type and age. For modeling of UMC we use self-consistent approach and rely only on the waveform amplitude fit. Based on the seismic exploration results and borehole PS-logging results we associate UMC layer with layers having Vs = 2.0 and 2.4 km/s respectively. As a preliminary test of effectiveness of this layer, we introduced into our model 1.2 km thick UMC layer having velocity Vs = 2.4 km/s. Fig. 5 show example of simulated waveforms compared with observed waveforms and waveforms for model without UMC. There is clear tendency to increase peak amplitude, delay time of wave phases become equal to observed delay, amplitude of later phases increased accordingly.

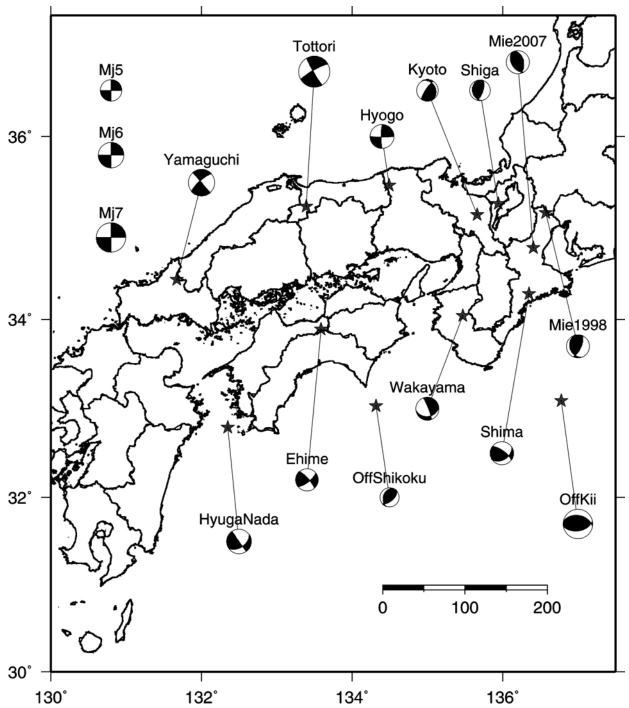


Figure 4. Location and source mechanisms of earthquakes used for testing of crustal velocity structure model.

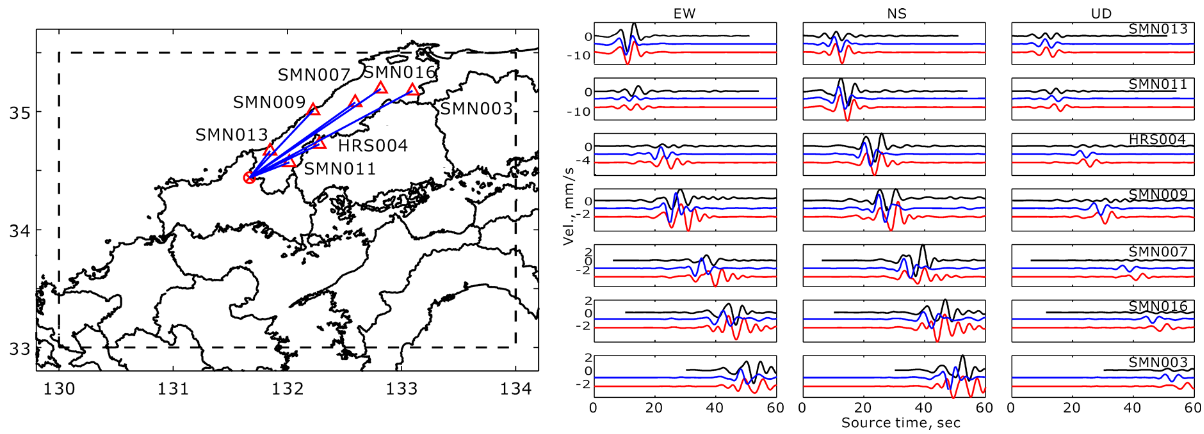


Figure 5. Example of comparison of velocity waveforms: observed (black), synthetic for the model without uppermost crustal layer (blue) and synthetic for the model with 1.2 km uppermost crustal layer (red). Left plot shows location of source and sites.

2.3. Non-Jacobian Inversion of the Uppermost Crustal Layer

In order to get better waveform fit we apply waveform calibration in accordance with recipe of Koketsu et al. (2009), Step 7. For this we use iterative waveform inversion method proposed in (Petukhin et al., 2011). Next assumptions are made: (1) For velocity model having a thin surface low-velocity layer over half-space or over a thick layer, it is assumed that increasing/ decreasing depth of the thin layer under a site increases/decreases amplitude of waveform at this site, respectively, while variations of layer depth at another sites have no effect. This is realistic assumption that helps to skip time-consuming calculation of sensitivity matrix (Jacobian); (2) The degree of the layer depth correction is proportional to the misfit value; (3) Waveforms of a small-to-medium size earthquake at rock sites have simple pulse shape, which allows to use ratio of peak ground velocities $v_{max.sim}/v_{max.obs}$ as the misfit value. These assumptions lead to a simple iteration scheme shown in Fig. 6. We apply "geological" constrain on the inversion results; which means that layer depth should agree with geological, seismic exploration, gravity anomaly and etc. information.

We made 5 iterations. Mean misfit for Iteration 3 already is close to saturation. Moreover, thickness of UMC is consistent with the results of Shingu-Maizuru and other reflection profiles (geological constrain). We assumed Iteration 3 as the result of the iteration inversion process. Results for UMC initial models, Iteration 3 and final model after tuning are shown in Fig. 6. Fig. 7 shows comparison of *P*-velocity cross-section for model without UMC layer (Koketsu et al., 2008); final model and Shingu-Maizuru reflection profile (Ito et al., 2005).

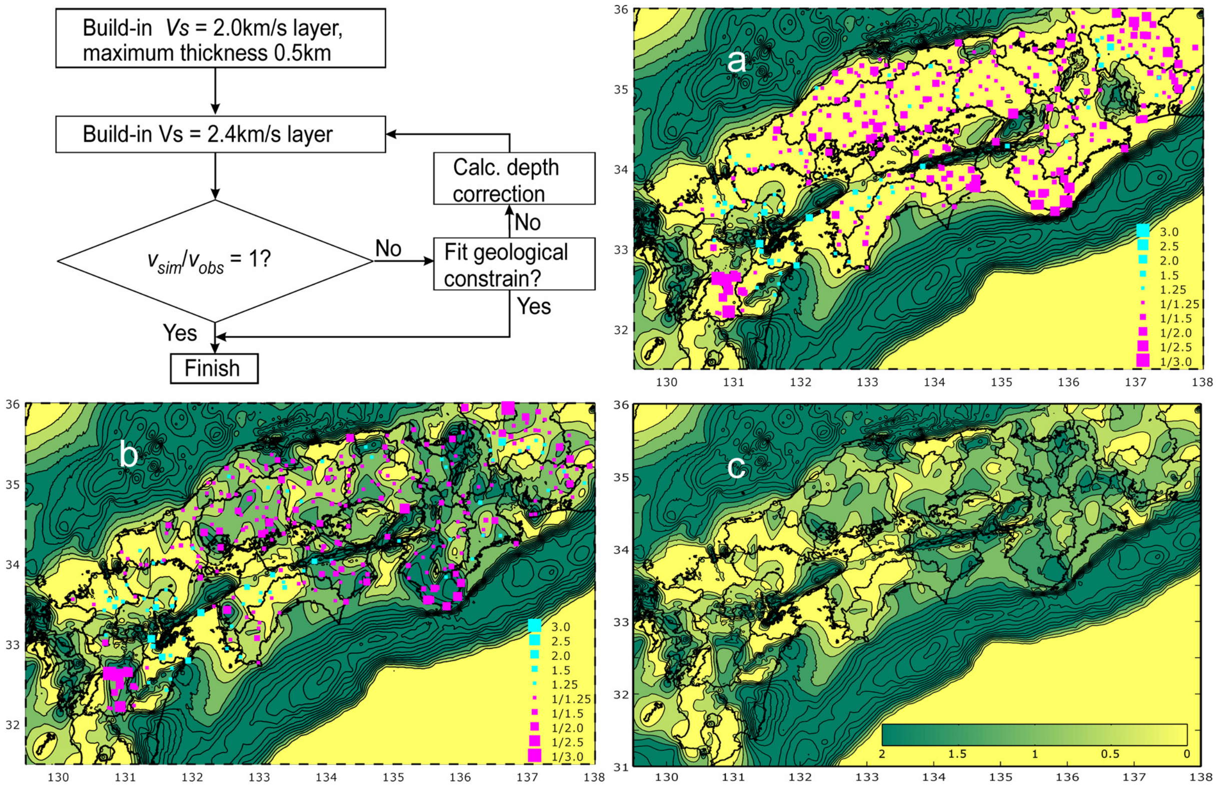


Figure 6. Iteration method (upper left) and results: distribution of the $v_{max.sim}/v_{max.obs}$ values (squares) and depth of UMC layer for model without the uppermost crustal layer (a), Iteration 3 model (b) and final model (c).

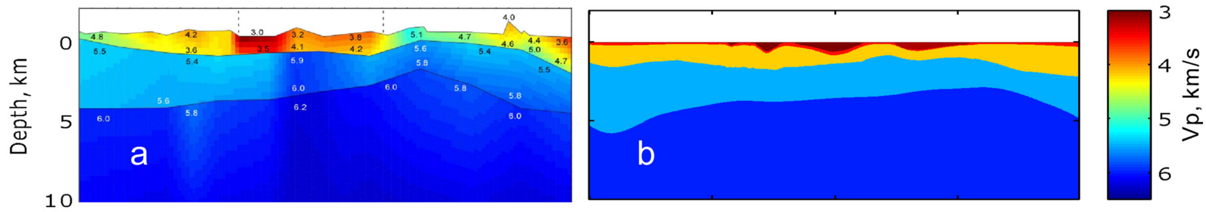
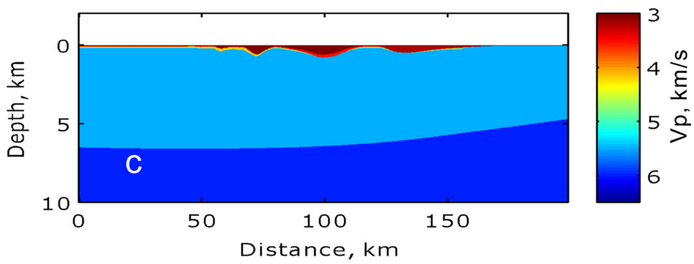


Figure 7. Comparison of *P*-velocity cross-sections: (a) for the Shingu-Maizuru reflection profile (Ito et al., 2005), (b) for final velocity model, and (c) for (Koketsu et al., 2008) velocity model.



3. SEDIMENTARY LAYERS (BASIN) MODEL

In order to improve accuracy of waveform simulation, crustal velocity structure above were combined with the sedimentary layers model. According to study of Koketsu et al. (2009), especially deep sedimentary layers below and near engineering bedrock (Vs=400-700m/s) are most important for accuracy of long-period waveform simulation. We used deep sedimentary structure of the (Koketsu et al., 2008) model that generally follow recipe principles and interpolates numerous seismic exploration, borehole and microtremor array data, using geologic information. It was calibrated then using H/V spectral ratio method applied to coda waves (Rayleigh waves) of natural earthquakes recorded at KiK-net and K-NET strong motion sites. Final model consists of 14 layers. Physical parameters of layers are listed in Table 1.

For this study we continue calibration of sedimentary model using estimates of 1D velocity structure by radial/vertical (R/V) spectral ratio method applied to coda waves of natural earthquakes recorded at numerous strong motion sites, and interpolated then using geology and gravity anomaly data. Our target area was Chugoku and Shikoku, not covered by previous studies. Calibration procedure includes next steps: (1) selection of data: records of middle-to-large crustal earthquakes having M = 4.5-7 (26 earthquakes) at sedimentary strong motion sites (151 sites) having epicentral distance D > 100km; (2) calculation of R/V spectral ratios of coda waves; (3) 1D velocity models under the selected sites were extracted from the (Koketsu et al., 2008) model and theoretical R/V ratios for Rayleigh wave calculated; (4) sites having large difference of 1st peak frequencies were selected for further calibration; (5) 1D structures were nonlinearly inverted using Genetic Algorithm, example of observed R/V ratio, R/V for initial 1D model and R/V result of inversion are shown in Fig. 8 (left), results for all sites are shown in Fig. 8 (right); (6) for Kochi basin in southern part of Shikoku (see Fig. 1), we additionally employed results of microtremor array measurements and single site microtremor H/V measurements scattered through the basin (Yamamoto et al., 2010), shown in Fig. 9; (7) at final step we extrapolated changes of depth of the layers into surrounding area; for this we used geological and gravity anomaly information.

Effectiveness of used approach is demonstrated by waveform simulation at sites KOC005, TKS001 and TTRH04. For simulation we used two earthquakes in subduction zone: 2009 Off Shikoku M6.9 earthquake and 2007 Eastern Ehime pref. M5.3 earthquake. The computations here and below were carried out by the FDM method (Graves, 1996) with a non-uniform grid size (Pitarka, 1999). Fig. 10 shows comparison of observed and simulated waveforms. Main features, peak amplitudes and duration of waveforms, are reproduced much better for calibrated model than for initial model. Examples of resulting interfaces are shown in Fig. 11. Major basins mentioned in Fig. 1 are well reflected here.

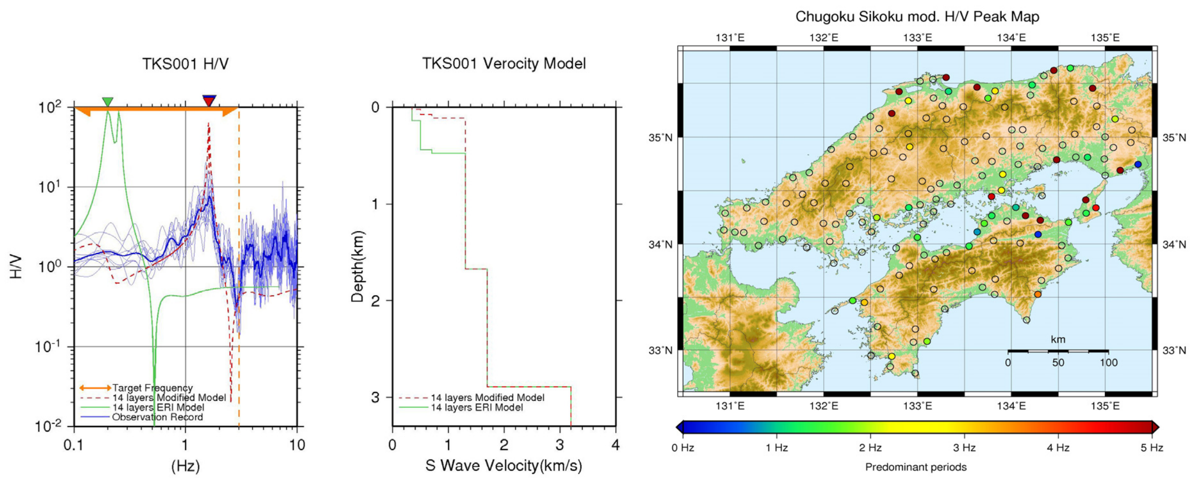


Figure 8. Example of 1D velocity structure inversion for site TKS001 and results for all studied sites. Left: observed R/V ratio (blue), R/V ratio simulated using initial 1D velocity model (green) and R/V ratio simulated using inverted 1D velocity model (red dashed). Middle: initial (solid green) and inverted (red dashed) 1D velocity models. Right: Location of all analyzed sites (open circles) and calibrated sites (colored circles). Color indicates frequency of peaks of inverted R/V ratios.

Table 1. Physical parameters of the deep sedimentary velocity structure model.

Layer number	*Vp*, km/s	*Vs*, km/s	Density g/cm^3	*Qp*	*Qs*	Layer number	*Vp*, km/s	*Vs*, km/s	Density g/cm^3	*Qp*	*Qs*
1	1.7	0.35	1.80	119	70	8	2.7	1.3	2.20	442	260
2	1.8	0.5	1.95	170	100	9	3.0	1.5	2.25	510	300
3	2.0	0.6	2.00	204	120	10	3.2	1.7	2.30	578	340
4	2.1	0.7	2.05	238	140	11	3.5	2.0	2.35	680	400
5	2.2	0.8	2.07	272	160	12	4.2	2.4	2.45	680	400
6	2.3	0.9	2.10	306	180	13	5.0	2.9	2.60	680	400
7	2.4	1.0	2.15	340	200	14	5.5	3.2	2.65	680	400

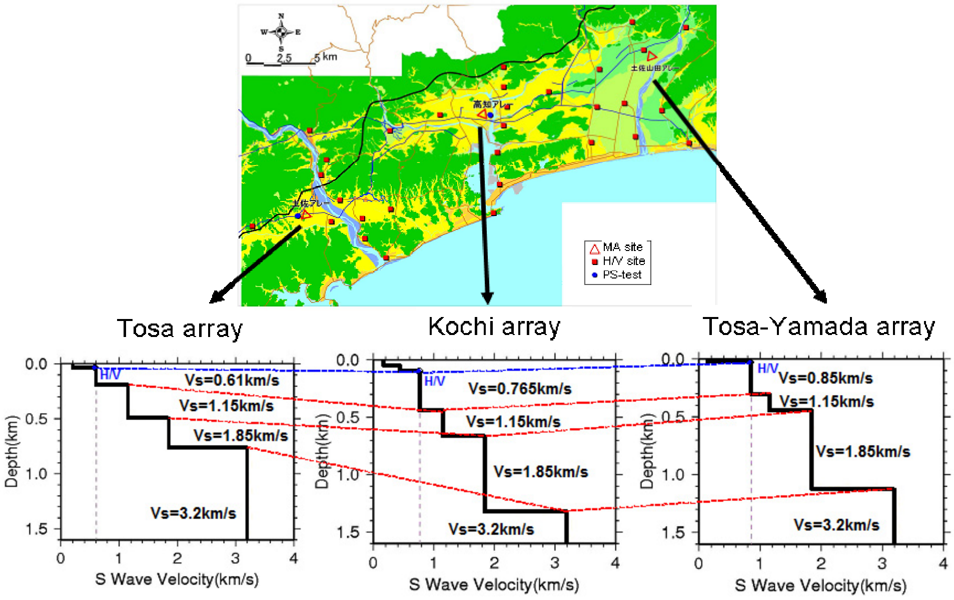


Figure 9. Microtremor array location (red triangle, upper plot) and measurements results (lower plot). Locations of the one site microtremor H/V measurements (red points) and borehole sites (blue points) are also shown. Color on the upper plot show topography: green – bedrock mountains, yellow – sedimentary plains. (From Yamamoto et al., 2010)

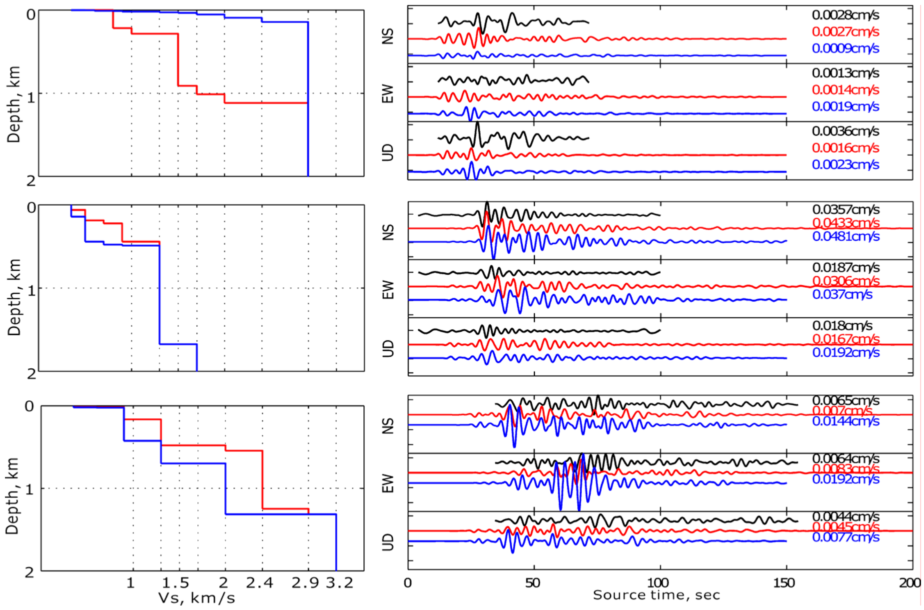


Figure 10. Comparison of observed (black line), simulated for this study model (red line) and the (Koketsu et al., 2008) model (blue line) waveforms. Numbers – peak amplitudes of each component.

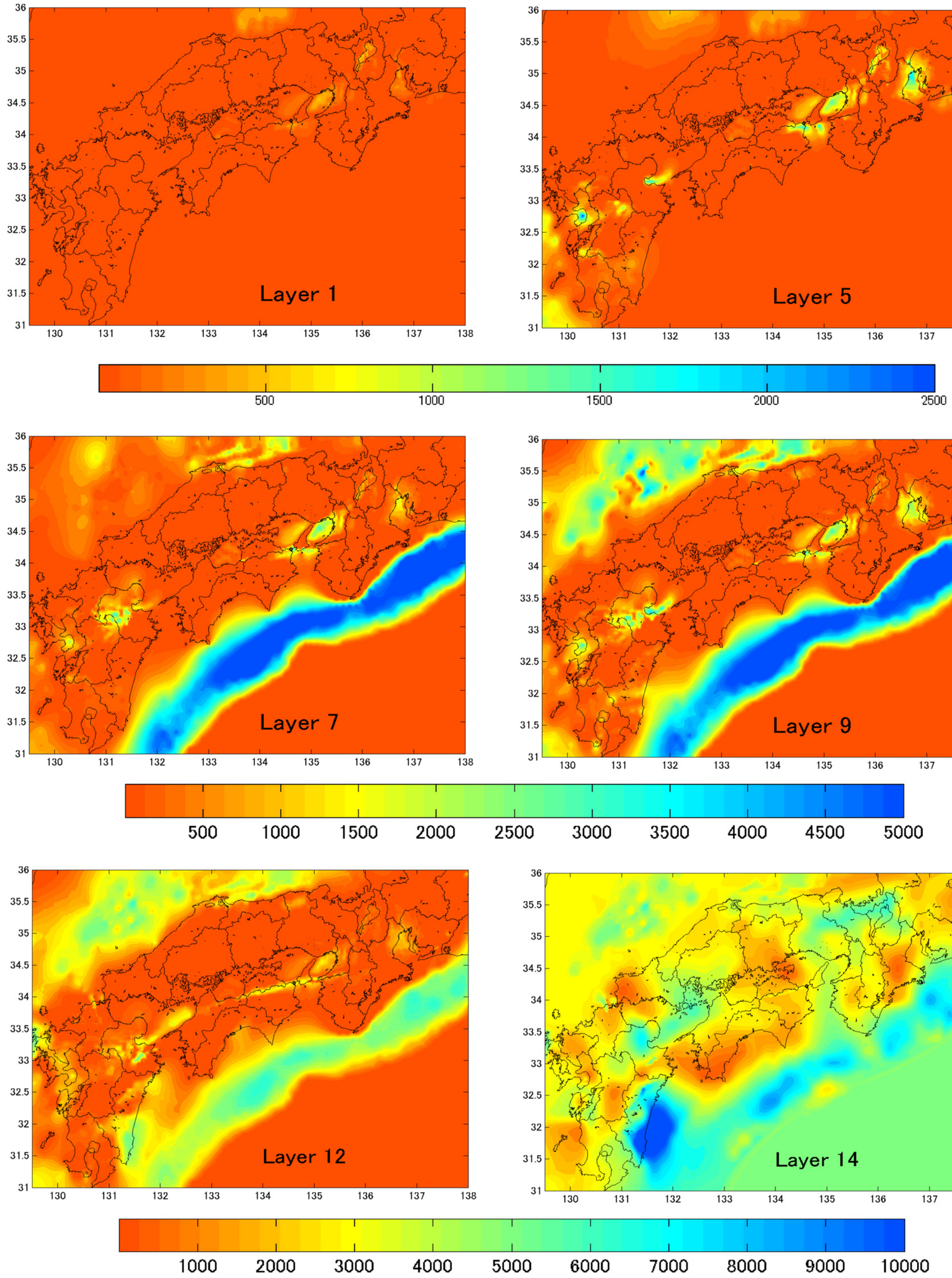
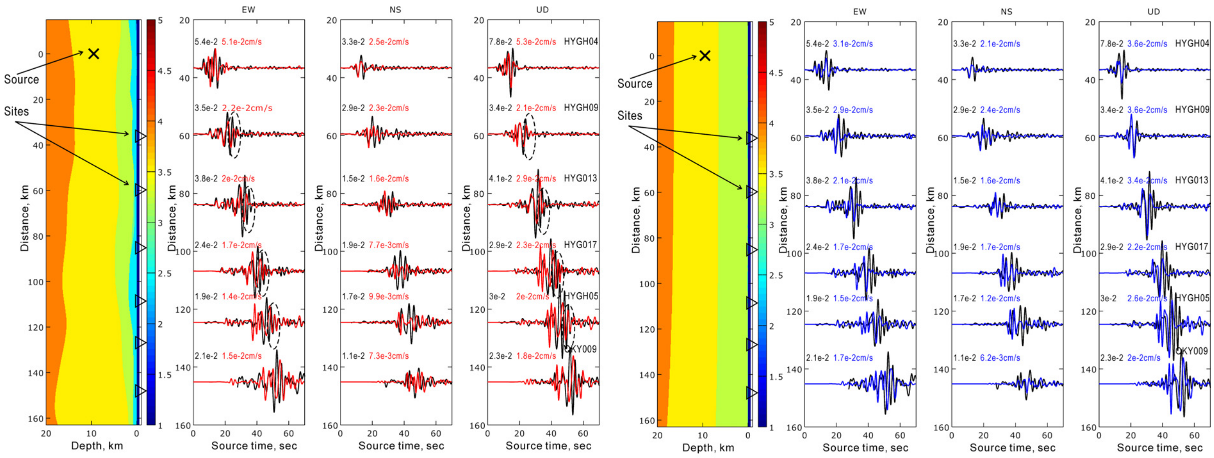


Figure 11. Examples of lower interfaces of sedimentary layers: layers 7 to 10 are smoothly combained with the Oceanic sediment 1 layer of the crustal model (large off-shore area painted in blue), layer 14 is also Seismic basement/Oceanic sediments 2 layer.

4. VERIFICATION OF THE VELOCITY STRUCTURE MODEL BY WAVEFORM SIMULATION

For the resulting velocity structure model validation we use *M*5.1 2001 Mid Kyoto earthquake, located in center of the most populated Kinki region. Because target of velocity modeling is the simulation of 2 sec and more long-period motions (see companion paper of Kagawa et al. (2012)) shortest period of calculated waveforms is 2 sec that is smaller than 2.8 sec used for velocity model calibration. Fig. 12 compares observed and simulated waveforms. Introducing of the UMC improves waveform fit by increasing of later phases, circled by dashed line.



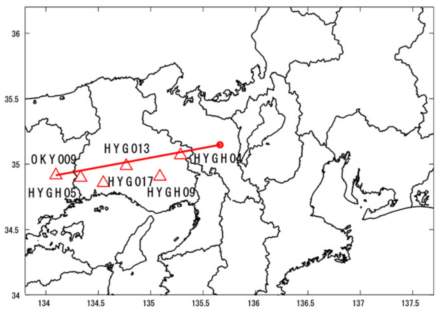


Figure 12. Comparison of observed (black line) and simulated waveforms of 2001 Mid Kyoto earthquake for final velocity model having UMC layer (red) and model without UMC layer (blue). Numbers – peak amplitudes, left subplots show velocity model cross-section (*S*-wave velocity), right lower plot shows location of source, sites and velocity cross-section.

5. DISCUSSION AND CONCLUSIONS

We constructed, calibrated and verified crustal and basin velocity structure models for the Southwest Japan. Comparison of observed and simulated waveforms demonstrates good fit both in waveform shape, time delay and amplitude. With a few exceptions, amplitude fit of observed and simulated maximum amplitudes for hard-rock sites is within factor of 1.3.

Uppermost Crustal layer was introduced and calibrated using the Non-Jacobian waveform inversion method. We proved that the Uppermost Crustal layer is necessary for good waveform fitting. *S*-wave velocity of the layer is around 2.0-2.4 km/s; its thickness is around 2 km in average. Introducing of additional layer strongly improve both fitting of amplitudes and similarity of waveforms themselves due to increasing of amplitudes of later phases. In addition to seismic exploration results, this effect is a proof of the necessity of thin low-velocity layer above of the seismic basement layer, even in the mountain areas having hard surface rocks.

We found some correlation of the layer thickness with the seismic exploration results, but didn't find stable correlation with geological features of studied region. As an alternative approach we collected velocity measurements in boreholes and studied their distributions separately for each geological type of rocks. We found stable differences of *S*-wave velocities between dense/monolithic types of rocks (i.e. granite) and cracked/heterogeneous types of rocks (i.e. conglomerates): higher velocities for the first group and lower velocities for the second group. Within each group of rocks, e.g. between granites and schist's, or between conglomerates and breccia, difference of velocities was unconfident, although standard deviation of velocity samples is large. This is a probable reason of the absence of correlation of depth of UMC layer and geology.

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