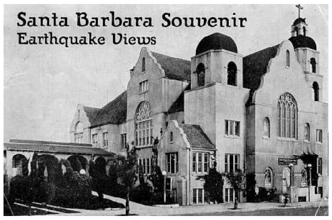

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STUDY ON THE APPLICABILITY OF NON-JACOBIAN ITERATION METHOD FOR MODELING OF THE UPPERMOST CRUSTAL LAYER

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ABSTRACT

Full-waveform inversion method is important for accurate tuning of 3D velocity structure models. In the straightforward linearized inversion method numerical calculation of sensitivity matrix (Jacobian) is most time-consuming step. For the case of a model having thin surface layer, we propose numerically effective inversion method without calculation of the Jacobian. We made the following assumptions. (1) For velocity model having thin surface layer over half-space or over a thick layer, we assume that increasing depth of the thin layer under a site increases amplitude of waveform at this site, and oppositely, decreasing of depth decreases it, while variations of layer depth at another sites have no effect. This is realistic assumption that helps us to skip time-consuming calculation of Jacobian. (2) The degree of layer depth correction is proportional to the misfit between observed and simulated waveforms. (3) Simple pulse-like waveforms from small earthquakes, originated at the rock site, allow us to use ratio of maximum amplitudes *Asim*/*Aobs* as the misfit function. These assumptions lead to simple iteration scheme. Depths of the thin layer under sites are inversion parameters. We applied this method to the estimation of depth of the Uppermost Crustal Layer (*Vs* = 2.4km/s) in the Kinki region, Japan. Results of the 3rd iteration already show negligible difference between amplitudes of simulated and observed waveforms and fit results of the Shingu-Maizuru seismic reflection profile in the same area.

INTRODUCTION

Long-period ground motions from *M*8-class events have often rocked facilities on basin or sedimentary sites located 200–300 km away from the source region, e.g., the well-known damage in Mexico City during the 1985 Michoacan earthquake. In the case of Tomakomai, , Hokkaido, Japan, the long-period ground motions, generated from the source of the 2003 Tokachi-Oki earthquake and then amplified and elongated in the Yufutsu basin, damaged oil storage tanks [*Koketsu et al*., 2005]. The Osaka sedimentary basin is located approximately 150 km away from the source regions of the hypothetical Tonankai and Nankai earthquakes, and the same situation would occur in that the modern megacities of Osaka and Kobe, located in this sedimentary basin, would definitely be shaken by disastrous long-period ground motions from *M*8-class subduction earthquakes. At the moment of the last Tonankai and Nankai events (the 1944 Tonankai and 1946 Nankai earthquakes), there were very few long-period structures in the sedimentary basin, whereas now, the megacities comprise a large number of skyscrapers, oil storage tanks, long-span bridges and so on. For earthquake disaster reduction, reliable strong motion predictions are required at the sites.

Long-period ground motions (2 - 20 s) can be reliably simulated by accurate numerical method; for example, finite-difference method

(FDM). Accurate basin and crustal velocity structure model is necessary in this case. Basin models were intensively generated by many researchers during last decades: [*Koketsu et al*., 2009] for the Kanto basin, [*Kagawa et al.*, 2004] for the Osaka basin and surrounding basins, [*Yoshida et al*., 2007] for the Yufutsu basin, [*Aichi pref.*, 2002] for the Nobi basin. These and other results are compiled into the integrated velocity structure model for the whole Japan [*Koketsu et al..*, 2008].

Crustal velocity model is an important link between earthquake source model and sedimentary basin model. Recently, with developing of dense seismic observational networks like K-NET, KiK-net and Hi-net, new receiver function and seismic tomography results become available for construction of the crustal model, in addition to traditional seismic reflection and refraction exploration methods and seismicity analysis method. *Iwata et al*. [2008] compiled many of these data into a detailed 3D velocity crustal structure model for the western Japan. Uppermost part of this model in inland area, although being important for simulation of long period ground motions (both for body wave amplification and surface wave propagation), rely on a scarce seismic reflection/refraction results mostly. In the areas not covered by data, various kinds of extrapolations, interpolations and generalizations of results of previous studies are used to construct model interfaces. Specifically, 3km depth of the seismic basement (see Fig. 4 below) is assumed for large regions in Shikoku, western Kyushu and Chugoku regions, based on the 1-D velocity model that is used in routine hypocenter determination. In order to improve accuracy of the upper crustal layers (seismological basement layer and layers above it) HF receiver function inversion method [e.g., *Kobayashi et al*., 2000; *Petukhin and Miyakoshi*, 2006] could be applied site-by-site. Considering the self-consistency concept of velocity modeling, i.e. that the model that will be used for simulation of the amplitude waveforms should be estimated from the same kind of data, in this study we propose alternative method that uses large amplitude direct *S*-wave and surface wave data and an iterative full-waveform inversion method. The method is based on the waveform simulation using FDM method and may be effective numerically in case of a wide target area, like western Japan. Simultaneously it also verifies the resultant crustal model by seismic waveform simulation [*Koketsu et al*, 2009].

With increasing of computer power nonlinear inversion methods become popular for estimation of velocity structures. Fully nonlinear inversion methods, i.e. Genetic Algorithm or Simulating Annealing, can be used if there is no priori information or initial model, simply by running forward simulation many times and selecting proper models. It is not to say that in case 3D FDM waveform simulation these methods become practically impossible. Contrary, having a reliable initial model at hand we can run the Linearised Inversion method or simple and elegant Simplex method [e.g., *Lagarias et al*., 1998] valid for discontinuous models. Second one still need many forward simulations, while first one need calculation of the sensitivity matrix, Jacobian.

Aoi et al. [1995] and Aoi [2002] proposed and tested the Linearized full waveform inversion method for the modeling of a basin structure bottom boundary. *Iwaki and Iwata* [2011] successfully applied this method for the modeling of the Osaka basin in Japan. In their method, FDM numerical calculation of the Jacobian is the most time-consuming step. In this study, for the case of the crustal model having thin surface layer, we propose numerically effective inversion method without calculation of the Jacobian, and apply it to the estimation of 3D depth variations of the Uppermost Crustal Layer (UMC) in the Kinki region.

NON-JACOBIAN METHOD OF VELOCITY STRUCTURE INVERSION

First, we made next assumptions. (1) For velocity model having a thin surface low-velocity layer over half-space or over a thick layer, we assume that increasing depth of the thin layer under a site increases amplitude of waveform at this site, and oppositely, decreasing of depth decreases it, while variations of layer depth at another sites have no effect. This is realistic assumption that helps us to skip time-consuming calculation of Jacobian. (2) The degree of the layer depth correction, necessary to fit waveform, is proportional to the misfit between observed and simulated waveforms. (3) Waveforms of a small-to-medium size earthquake at hard rock site have simple pulse shape, which allows us to use ratio of maximum amplitudes *Asim*/*Aobs* as the misfit function. These assumptions lead to a simple iteration scheme shown in Fig. 1. Sites are control points for the layer interface interpolation. Depths of the thin layer under the sites are inversion parameters.

Similarly to [*Aoi*, 2002] for example, neglecting higher-order terms, the linearized observation equation for the l^{th} iteration is

$$v_{i\_syn}(x_m,t_n\,|\,\mathbf{p}^l)+\sum_{k=1}^{K}\left.\frac{\partial v_{i\_syn}}{\partial p_k}\right|_{\mathbf{p}=\mathbf{p}^l}\cdot\delta p_k^l\cong v_{imn\_obs} \tag{1}$$

where v_{i_syn} $(x_m,t_n|\mathbf{p})$ is the ith component of the synthetic waveform, under model parameter $\mathbf{p}$; v_{imn_obs} is the ith component of the observed waveform at x_m and t_n; x_m is the mth position; t_n is the nth time step; and $\mathbf{p}$ is the model parameter (vector) $(p_1, p_2, \ldots p_k, \ldots, p_K)^T$. $\mathbf{p}^l$ is the model parameter estimated in the (l-1)th iteration ($\boldsymbol{p}^0$ is the initial model). In order to reduce problem to the linear inversion problem we can rewrite:

$$\sum_{k=1}^{K} \frac{\partial v_{i\_syn}}{\partial p_k}\bigg|_{\mathbf{p}-\mathbf{p}^l} \cdot \delta p_k^l \cong v_{imn\_obs} - v_{i\_syn}(x_m, t_n \mid \mathbf{p}^l) \tag{2}$$

or in the matrix form:

$$\mathbf{J} \cdot \delta\mathbf{p} = \mathbf{d} \tag{3}$$

where **d** is the data vector, δ**p** is unknown vector of the model parameters variation, and **J** is sensitivity matrix, Jacobian. [*Aoi*, 2002] and [*Iwaki and Iwata*, 2011] used painfully time-consuming finite-difference method to calculate matrix **J**.

Using advantage of a dense observation network, we will use observation site locations as the control points for model parameters p_k. Next, for velocity model having a thin surface low-velocity layer over half-space or over a thick layer, variations of layer depth at sites k ≠ m have almost no effect. In this case, for k ≠ m components of Jacobian $J_{km} \approx 0$ and inversion become ill-posed. We will assume $J_{km} = 0$ for k ≠ m:

$$\frac{\partial v_{i\_syn}(x_m, t_n)}{\partial p_m}\bigg|_{\mathbf{p}-\mathbf{p}^l} \delta p_m^l \cong v_{imn\_obs} - v_{i\_syn}(x_m, t_n \mid \mathbf{p}^l) \tag{4}$$

We assume that most difference between observed and synthetic waveforms is due to the site effect (effect of velocity structure under site), but source and path effects are accurately simulated. In this case it is convenient to use ratio $v_r = v_{syn}/v_{obs}$ as the inversion variable. This value reflects variation of site effect due to inaccuracy of the velocity model, which is expected to be similar for all components. Effects of non-accounted variations of source and path effects can be reduced by simple averaging over several earthquakes data. Waveforms of a small-to-medium size earthquake at hard rock site have simple pulse shape (see Fig. 2), which allows us to use normalized amplitude parameter, i.e maximum value $v_{i\_r\_max}$, instead of the whole waveform ratio $v_{i\_r}(t_n)$. In this case:

$$mean\left\{\frac{\partial v_{i\_r\_\boldsymbol{max}}(x_m)}{\partial p_m}\right\}_{\mathbf{p}-\mathbf{p}^l} \delta p_m^l \cong mean\{1 - v_{i\_r\_\boldsymbol{max}}(x_m \mid \mathbf{p}^l)\} \tag{5}$$

or

$$s_m \cdot \delta p_m^l \cong mean\left\{\log\left(v_{i\_r\_\boldsymbol{max}}(x_m \mid \mathbf{p}^l)\right)\right\} \tag{6}$$

where s_m is a sensitivity coefficient. We can calculate sensitivity coefficient at each control point (site) in every iteration, or assume the same sensitivity coefficient for all control points and gradually increase it with iterations in order to stabilize inversion, similarly to the Simplex method. Both approaches have similar computational time; we choose second one for this study.

Figure 1 show diagram of the iteration inversion method. We apply "geological" constrain on the inversion results; that mean that layer depth should fit geological, seismic exploration, gravity anomaly and etc. data. Method can be applied hierarchically to the model with several surface layers: first the deepest layer is estimated assuming zero or a fixed depth of the upper layers and using longer-period waveforms, and then upper layers can be calibrated successively using shorter and shorter-period waveforms.

NUMERICAL TESTING OF THE METHOD THROUGH INVERSION OF SIMULATED WAVEFORMS

To show the validity of the proposed method, a numerical experiment was carried out where waveforms generated with an assumed true model (the target model) were used as the input data. Fig. 2a shows the UMC layer topography of the target model, which is a simplified version of the inverted UMC layer for the Kinki region (see below). The sedimentary layers are modeled by a single soft layer, and crustal layers below UMC layer are exactly the same as in the crustal model for the Kinki region (see below). Their physical

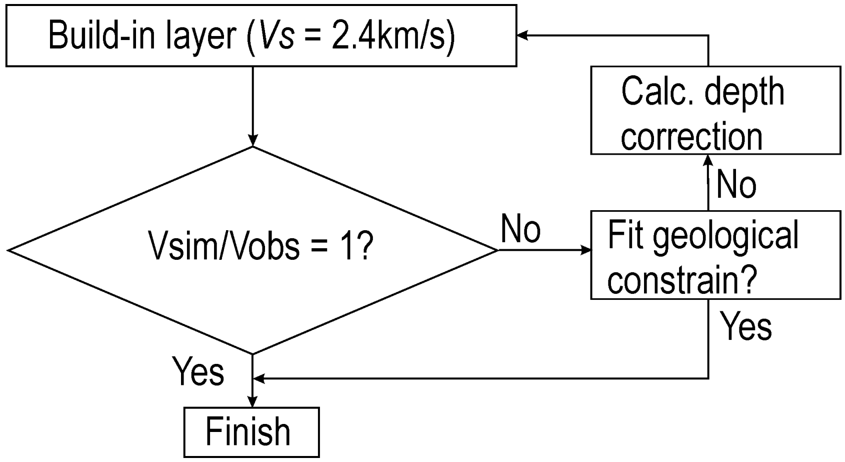

Fig. 1. Iteration method. Depth corrections are calculated for each observation site proportionally to the v_{syn}/v_{obs} ratio at this site. For inversion stability, coefficient proportionality is decreased gradually. Corrected depths are then interpolated between control points at the sites and new layer depth distribution is used for the next iteration.

parameters are shown in Table 2. We also used the same set of point seismic sources as for inversion below (Table 1). Source time function in the bell-shape is employed. Waveforms generated by the FDM at the hard rock KiK-net and K-NET stations shown by squares in Figs. 2b and c, are used as the data for this numerical inversion experiment. Fig. 2b shows the initial model for the inversion, which is a constant depth layer, h = 1.0 km, with the sedimentary basins filed by uniform material having V_p = 2.4 km/s and V_s = 1.0 km/s.

Maximum amplitude ratios v_r generated by this initial model are plotted in Fig. 2b using blue and red squares for overestimated and underestimated amplitudes respectively. The difference of v_r value from 1.0 comes from the disagreement of the initial model and the target. In areas A, B, and C, the depth of the target is larger than initial model and $v_r < 1.0$, and contrary in area D the depth of the target is smaller than initial model and $v_r > 1.0$ respectively.

Inverted model is shown in Fig. 2c. Since the maximum amplitude ratio is not sensitive to the layer depth changes in cases when a site is close to source, detailed features of the estimated model have not been recovered. Major characteristics, however, have been recovered well.

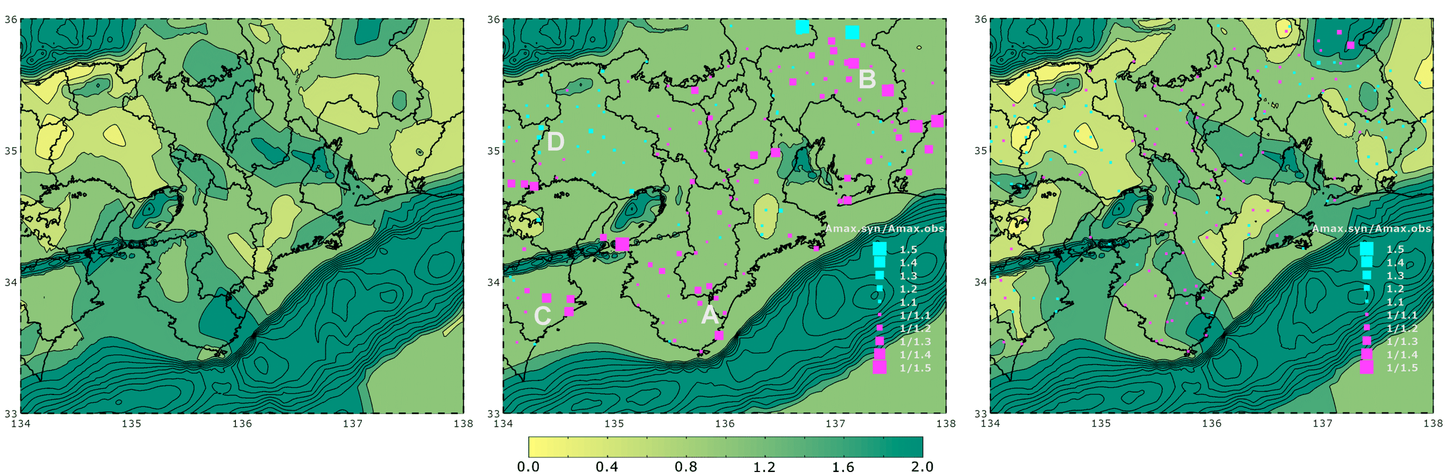

Fig. 2. Results of inversion test: UMC layer depth target model used for calculation of "observed" waveforms (left), initial model (center) and inverted model (right).

DATA PREPARATION AND FINITE-DIFFERENCE CALCULATIONS

For this study we choose 7 medium size earthquakes (*M*5-6). Source parameters are listed in Table 1 and source locations are shown in Fig. 3. Source mechanism, depth and rise time T_r are critical parameters for getting good waveform fit. Strictly speaking, for waveform simulation in the 3D velocity model, which is different from the 1D velocity model used by the JMA for hypocenter determination and the F-net for source mechanism and source depth estimation, we need to relocate earthquake (including source time) and reevaluate source mechanism and depth using our 3D velocity model. Here we used simplified approach. In the preliminary run of simulations with source parameters evaluated by two agencies, JMA and F-net, we choose those source parameter determinations that results in a better waveform fit; relevant agencies are shown in parenthesis in Table 1. Reasonable difference is possible due to difference of period range: short period body wave arrivals are used by JMA for hypocenter location, long period waves (around 3-10 s, mostly body waves) are used by JMA for the CMT solution, and very long period 20-40 s mostly surface waves are used by F-net for the CMT solution.

For determination of UMC structure we selected hard rock KiK-net and K-NET sites. For 100-200m borehole KiK-net sites it is easy to separate hard rock sites simply by referring *PS*-logging results at the bottom of borehole. We assumed that sites having $V_s >$ 1000m/s at 100m depth are the hard rock sites. For K-NET sites having *PS*-logging results only to 20m depth, such approach is impossible. Instead of *PS*-logging data for K-NET sites we used deep and shallow velocity structure model for the whole Japan [*Koketsu et al.*., 2008] and applied the same criterion. In total 557 sites were selected for the western Japan.

3-D crustal structure model for this study were combined from the results of *Iwata et al*., [2008] for the Moho and Conrad and Seismic basement, *Baba et al*. [2006] for the oceanic sediments (accretion prism) and the subduction plate interfaces. Schematically this model is shown in Fig. 4 and Table 2. General principle of the crustal velocity modeling is to build model from several layers with constant values of velocity, density and quality factor Q of the each layer. This traditional approach allows us directly incorporate into the model results of many previous studies, e.g. seismic profiling and receiver function inversion, which employ the same approach with constant velocity layers. Model interfaces between layers are three-dimensional. They are constructed using spline interpolation method. Oceanic water layer was reduced in this study, simply by subtracting depth of the oceanic layer from the depth of all velocity interfaces under the ocean (squashing). This procedure keeps thickness of layers (oceanic sediments, oceanic upper crust and lower crust) the same as in the model with oceanic layer, and reduces simulation errors.

For simulations, we employed the 3D staggered grid finite difference method [*Graves*, 1996]. The shortest target period in the simulations was 2.8 s. The smallest shear-wave was assumed 700 m/s. Finite difference grid size in horizontal directions was 400 m (5 grids for the shortest wavelength). For depth direction, non-uniform grid size was used [*Pitarka*, 1999]. For the sedimentary layers and crust shallower than 6800 m (deepest of the oceanic sediments layer 1) grid span was designed to be 200 m. For seismic basement and oceanic sediments layer 2 deeper than 6800 m and shallower than 9200 m, the grid span was 400 m. The rest of the model had grid span 800 m. The depth of calculation volume was 40000 m (deeper than the deepest Moho). For the calculations we used 6250 time steps at 0.016 s time interval.

Table 1. Source parameters of used earthquakes.

Earthquake name	M_{JMA}	Epicenter	Source mechanism	Depth, km	Rise time, T_r (s)
1998 Northern Mie pref.	5.4	35.17N/136.57E	355/40/63 (JMA)	10.5 (JMA)	1.0
1999 NW Shiga pref.	4.9	35.27N/135.94E	17/66/101 (F-net)	10.0 (JMA F-net)	1.0
1999 Central Wakayama pre.	5.4	34.04N/135.47E	339/82/-58 (JMA)	69.58 (JMA)	1.0
2000 Shima peninsula region	5.5	34.28N/136.35E	306/72/130 (F-net)	43.1 (JMA)	1.1
2001 Mid Kyoto pref.	5.1	35.15N/135.66E	353/18/71 (JMA)	9.5 (JMA)	1.3
2001 Northern Hyogo pref.	5.6	35.47N/134.49E	90/89/172 (F-net)	10.6 (JMA)	2.4
2007 Northern Mie pref.	5.4	34.79N/136.41E	347/46/103 (F-net)	16.0 (JMA)	2.4

Example of simulated waveforms is shown on Fig. 5. The wave fit is good except of a general tendency for underestimation of amplitudes, which is critical point for the strong ground motions prediction, and tendency of time delay of simulated waveforms in comparison with observed waveforms.

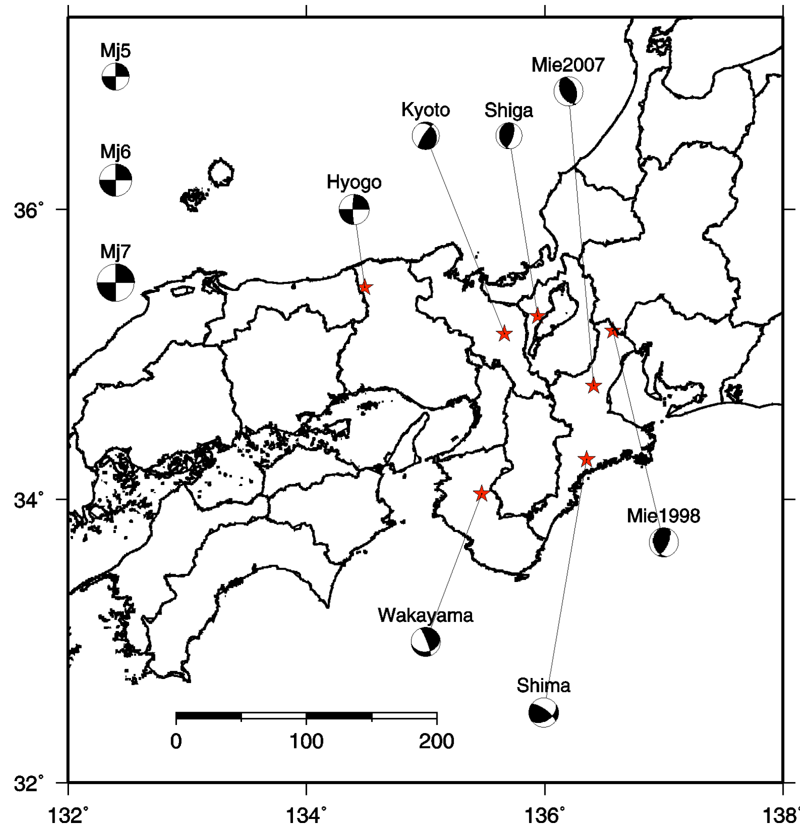

Fig. 3. Location and source mechanisms of used earthquakes.

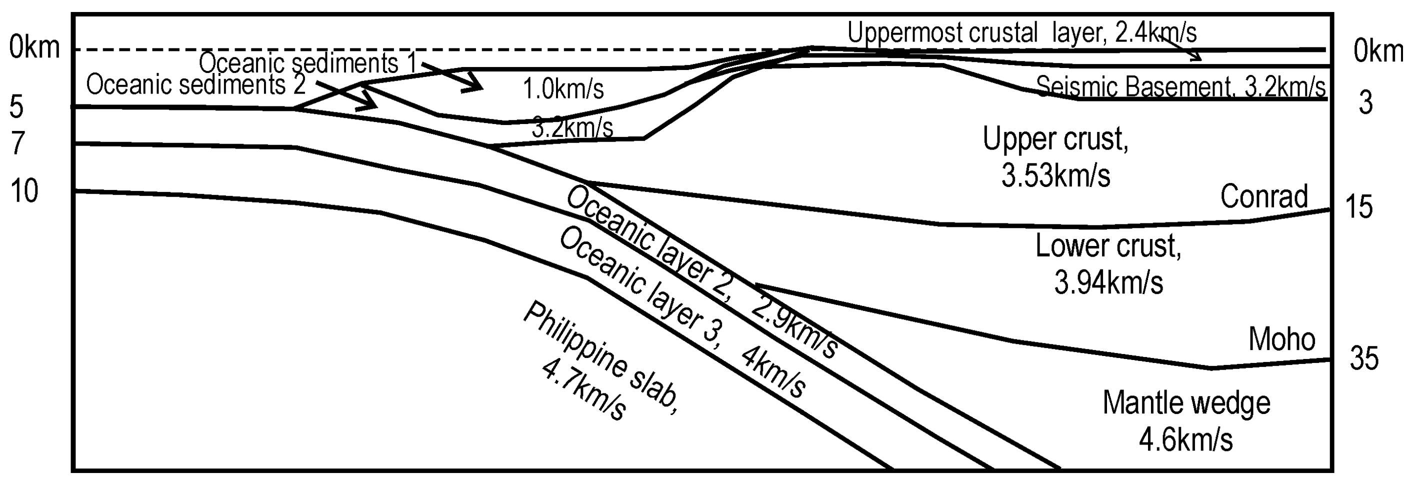

Fig. 4. 3D crustal velocity structure model (schematic). S-wave velocity values and approximate layer depth are given for reference.

Table 2. Crustal velocity structure parameters.

Layer Name	*Vp*, km/s	*Vs*, km/s	Density g/cm^3	*Qp*	*Qs*
Oceanic sediments 1	2.4	1.0	2.15	340	200
Oceanic sediments 2	5.5	3.2	2.65	680	400
Uppermost crustal layer	4.2	2.4	2.45	680	400
Seismic basement	5.5	3.2	2.65	680	400
Upper crust	6.0	3.53	2.7	680	400
Lower crust	6.7	3.94	2.8	680	400
Mantle wedge	7.8	4.6	3.2	850	500
Oceanic layer 2	5.0	2.9	2.4	340	200
Oceanic layer 3	6.8	4.0	2.9	510	300
Philippine slab	8.0	4.7	3.2	850	500

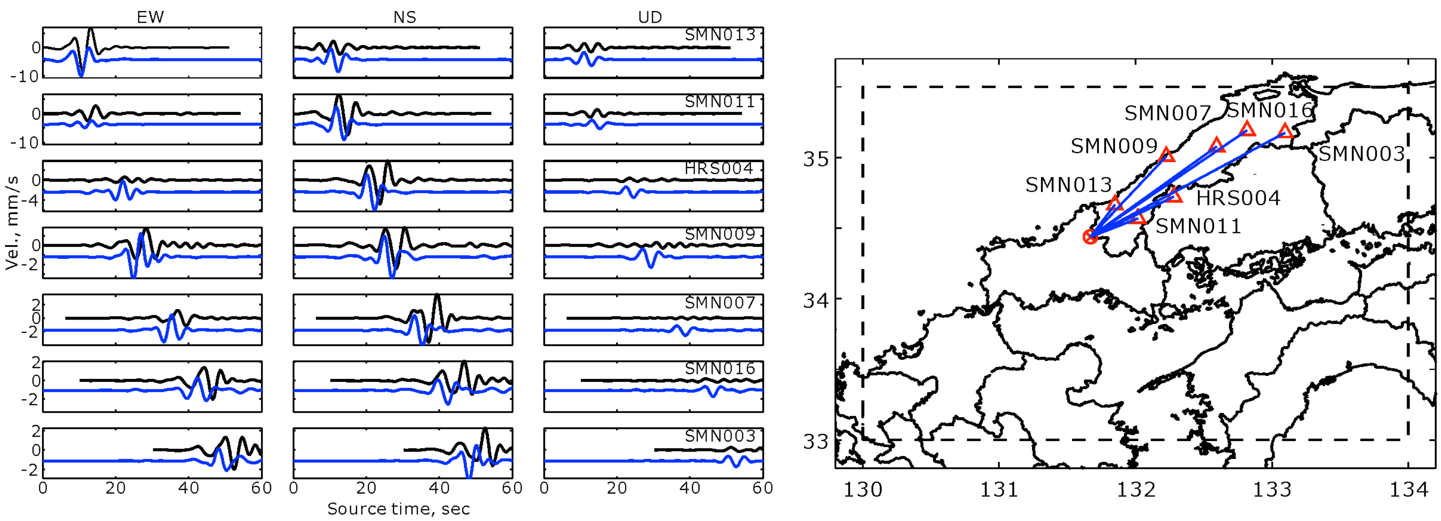

Fig. 5. Left: comparison of synthetic (blue) and observed (black) waveforms for the model without UMC layer. Right: location of the source and sites.

INVERSION RESULTS

In order to get larger amplitudes of the synthetic waveforms and better fit with the observed waveforms respectively, we introduced the UMC layer into our velocity structure. In the target area such a layer was observed on the Shingu-Maizuru seismic reflection/refraction profile for example [*Ito et al.*, 2005]. Velocity of the UMC layer was estimated using borehole *PS*-logging results in the studied area. They show existence of V_s = 2.4 km/s layer in the plutonic rock (e.g. granite) areas and V_s = 2.0 km/s layer in other areas. We assumed two-layer model for the UMC layer: upper V_s = 2.0 km/s layer with fixed thickness proportional to the waveform misfit, 0.5 km maximum, and V_s = 2.4 km/s with initial thickness 0.5 km. This was our initial model $\mathbf{p}^0$ for inversion.

Some areas had high density of observation sites with a week consistency of amplitudes between sites. In order to stabilize results we used additional averaging over 0.25^{O} x 0.25^{O} grid. A few grids had no sites; we manually adjusted layer depth for such grids. Results for some iterations are shown in Fig. 6. Layer depth for Iteration 3 already is consistent with the results of the Shingu-Maizuru reflection profile (geological constrain). Depths of layer for Iteration 4 and Iteration 5 are too large. We assumed Iteration 3 as the result of the iteration inversion process.

Although gradual reduction of the v_{r_max} values to the target v_{r_max} = 1.0 is obvious for the most area, some sites still have underestimated maximum amplitudes. Most prominent underestimation is observed in southern Kii peninsula area. For this area we applied the receiver function analysis. Results show that even for Iteration 3 depth of the UMC layer is overestimated and reason of the underestimation of amplitudes is other than the shallow crustal structure. This is the area with a complex tectonic structure [e.g., *Kodaira et al.*, 2006] having strong gravity and anisotropy [*Saiga et al.*, 2011] anomalies. More complicated modeling, other than simplified constant velocity layering, is necessary for this area.

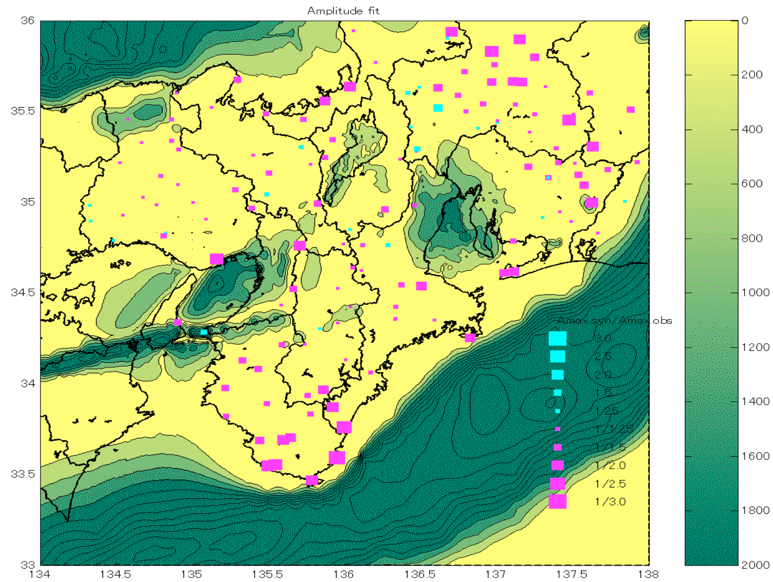

Fig. 6a. Distribution of the v_r values (squares) and depth of the UMC layer (contour) for model without the UMC layer [Koketsu et al.., 2008].

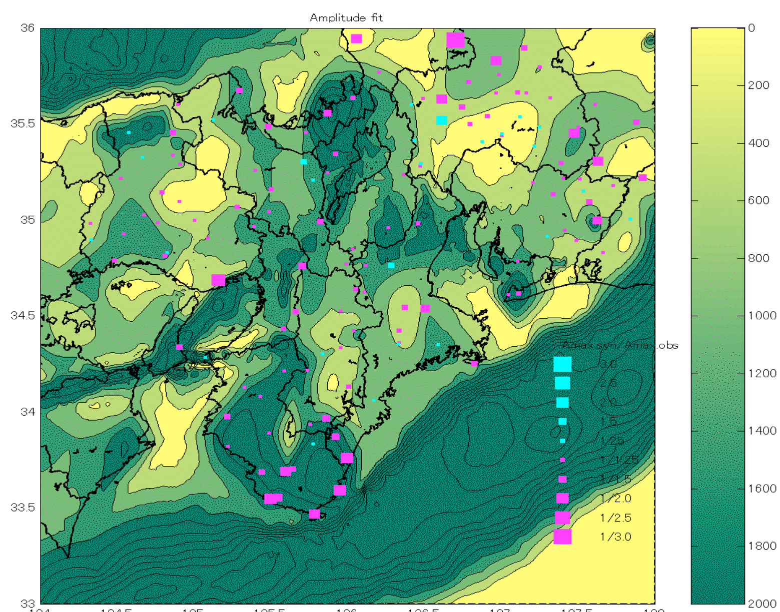

Fig. 6b. Same for iteration 3 model.

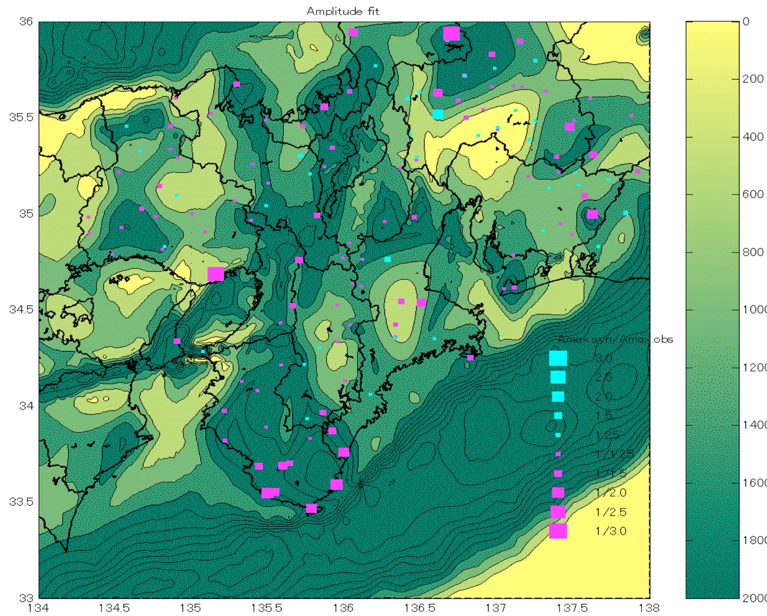

Fig. 6c. Same for iteration 5 model.

TESTING OF THE UMC MODEL

For testing of estimated velocity structure we calculated gravity anomalies and compared them with observed Bourge anomalies. Effects of shallow velocity and density structure may be masked by effects of strongly non-uniform deep (Moho and deeper) mass distributions in subduction zone. Deep distributions of densities were estimated and their effects were removed using next approach. Density distribution at depth 30km and deeper was estimated from seismic tomography results of *Matsubara et al*. [2008] using empirical relation of *Ludwig et al.* [1970] between V_p and density. Results are shown in Fig.7. Simulated and observed gravity anomalies have good correlation for final velocity model in the Kii peninsula area (circled by the dashed line).

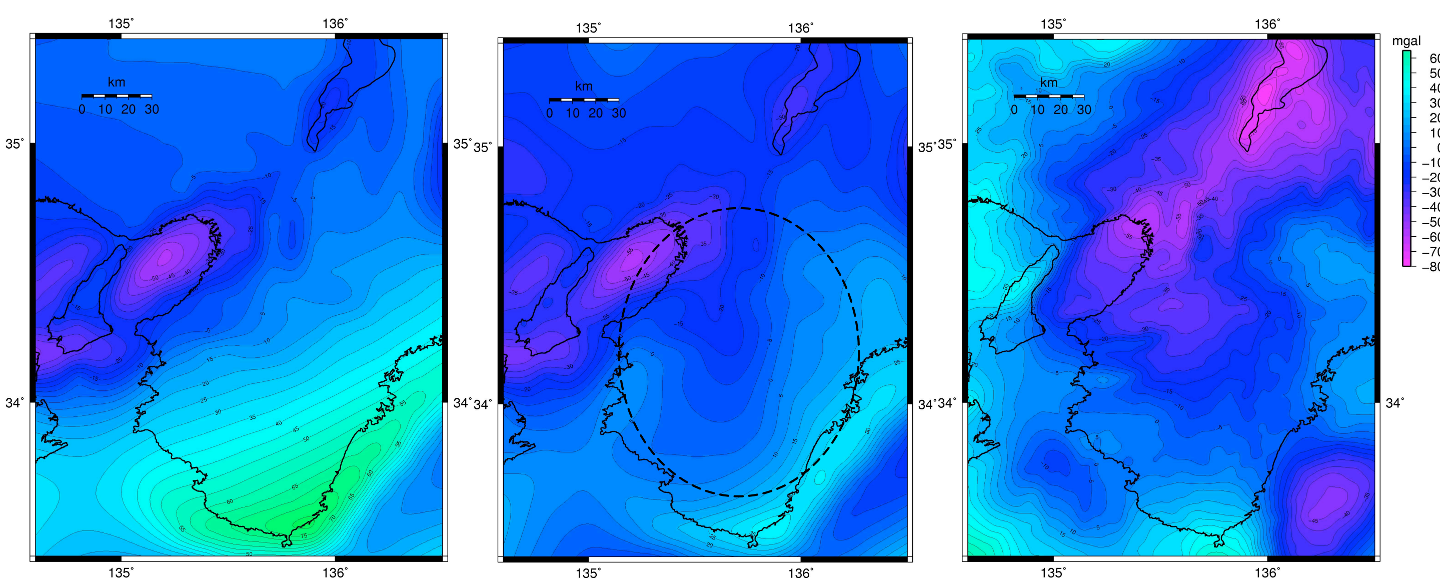

Fig. 7. Gravity distributions calculated for model without UMC layer (left), for final model (center) and observed gravity anomaly (right). Dashed line indicate strongly improved area.

DISSCUSION AND CONCLUSIONS

Non Jacobian waveform inversion method for velocity structure is proposed and tested on real data. Similar to seismic exploration results, we confirmed that the UMC layer is necessary for good waveform fitting. *S*-wave velocity of the layer is around 2.0-2.4 km/s, where this thickness is around 1.5 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. This effect is an additional proof of the necessity of thin low-velocity layer above of the seismic basement layer, even in the mountain areas having hard surface rocks.

In this study we used period range 3-10 s. With a few exceptions, for hard rock sites in the studied region *S*-waves have pure pulse-like shape in this period range. For this reason we used pure synthetic-to-observed peak amplitude ratio as the target function for inversion. Decreasing of shortest period to 1-2 s result in increasing of later phase content of waveforms, the deeper surface layer the longer later phase. It is expected that later phase content is less affected by source and path effects and can be used for waveform inversion as well [Aoi 2002].

We found some correlation of the layer thickness with the seismic exploration results, but didn't find stable correlation with geological futures 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 (e.g., granite) and cracked/non-uniform types of rocks (e.g., conglomerates), higher velocities for the first group and lower velocities for the second group. Within each group of rocks, i.e. between granite and schist, or between conglomerate and breccia, difference of velocities is unconfident, although standard deviation of velocity samples is large.

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