Title page:

Title: Inversion of Love Waves in Earthquake Ground Motion Records for Two-Dimensional S-wave Velocity Model of Deep Sedimentary Layers

Author #1: Kentaro Kasamatsu, Kajima Corporation (Former Tokyo Institute of Technology), 2-19-1 Tobitakyu, Chofu-shi, Tokyo, Japan, kasmake@kajima.com

Author #2: Hiroaki Yamanaka, Tokyo Institute of Technology, 4259 Nagatsuta, Midori-ku, Yokohama, Kanagawa, Japan, yamanaka.h.aa@m.titech.ac.jp

Author #3: Shin’ichi Sakai, Earthquake Research Institute, The University of Tokyo, 1-1-1 Yayoi, Bunkyo-ku, Tokyo, Japan, coco@eri.u-tokyo.ac.jp

Corresponding author: #1
Abstract

We have proposed a new waveform inversion method to estimate a 2D S-wave velocity structure of deep sedimentary layers using broadband Love waves. As a preprocessing operation in our inversion scheme, we decompose earthquake observation records into velocity waveforms at periods of 1 s interval. Then, we verify an assumption of 2D propagations of Love waves with polarization features based on a principal component analysis to select the segments applied for the inversion. A linearized iterative inversion analysis for the selected Love wave segments filtered at period of every 1 s allows a detailed estimation of boundary shapes of interfaces over the seismic bedrock with an S-wave velocity of approximately 3 km/s. We demonstrate the technique’s effectiveness with applications to observed seismograms in the Kanto plain, Japan. Differences between the estimated and existing structural models are remarkable at basin edges. A regional variation of the near-surface S-wave velocities in our model is similar to a distribution of surface geological classifications. Since a subsurface structure at a basin edge strongly affects earthquake ground motions in a basin with generations of
surface waves, our method can provide a detail model of a complex S-wave velocity
structure at an edge part for a strong ground motion prediction.

Keywords

Waveform inversion, Long-period strong motion, Love wave, Two-dimensional
assumption, Principal component analysis, S-wave velocity, Deep sedimentary layers,
Basin edge, Kanto plain

Introduction

A prediction of an earthquake ground motion using a theoretical approach such as a FD
simulation requires an appropriate model of subsurface S-wave and P-wave velocities,
density, and attenuation factors. Specially, S-wave velocities of deep sedimentary layers
over a seismic bedrock with an S-wave velocity of approximately 3 km/s must be
accurately modeled for site-specific long-period ground motion estimations in a large
Accordingly, several 3D velocity models have been constructed for large basins in Japan (e.g., Sato et al., 1999; Yamanaka and Yamada, 2006; HERP, 2009; HERP, 2012; Yoshimoto and Takemura, 2014; HERP, 2017). Hereinafter, we refer the models by HERP (2009), HERP (2012), and HERP (2017) as the HERP2019, HERP2012, and HERP2017 models, respectively. In particular, the HERP2009 and HERP2012 models cover the whole of Japan. The latter model was constructed from procedures suggested by Koketsu et al. (2012) as an updated version of the HERP2009 model with improvements with additional data such as results of earthquake ground motion simulations and geophysical surveys. The HERP2017 model was constructed only for the Kanto plain based on a model by Senna et al. (2013) who combined a shallow 3D subsurface structural model with deep one which had ever been modeled separately. Senna et al. (2013) established their model with reference to procedures for constructing a 3D structural model indicated by HERP (2017). Digital data of the HERP2009, HERP2012, and HERP2017 models are available for uses in earthquake ground motion evaluations (HERP, 2009; HERP, 2012; HERP, 2017). These velocity models have been validated by comparing simulated earthquake ground motions with observed ones for moderate seismic events. For example,
the HERP2017 model was examined for its appropriateness from estimations of goodness-of-fit analyses between synthetic and observed waveforms for several earthquakes.

It is well known that details of a velocity structure at a basin edge affect ground motion characteristics in an entire basin due to developments of surface waves. For example, Kawase (1996) showed that a large amplitude area was caused by a constructive interference of direct S-waves and basin-induced Rayleigh waves generated at a basin edge during the 1995 Hyogo-ken Nanbu Earthquake though strong ground motion simulations. The most of the existing 3D models were established by interpolating accumulations of 1D and 2D profiles with several velocity discontinuities from geophysical explorations. These 3D models may lack for detailed features in the edge parts because of difficulties to model their complicated shapes.

As one of the alternatives for modeling deep sedimentary layers including a detailed basin-edge structure, inversion analyses using seismograms from moderate events have been performed for 3D basin profiles (e.g., Aoi, 2002; Iwaki and Iwata, 2011). However, such 3D inversions rely on iterative forward calculations for partial derivatives using a
3D FD scheme with numerous computational costs. Thus, it takes a long computational time to conduct many inversion analyses using various conditions and initial models. Therefore, a waveform inversion method for retrieving a 2D profile of the deep sedimentary layers from earthquake records can be practically effective because of its low computational costs. As examples of previous studies on the 2D waveform inversions, Aoi et al. (1995) proposed an inversion method with an assumption of a plane incident wave employing a 2D boundary-element method. Ji et al. (2000) set a 1D structural model outside a basin, and performed an inversion analysis to estimate a 2D basin structure by perturbing depths of several control points in numerical experiments. Most of the studies on the 2D waveform inversions examined performances of their methods through the numerical experiments with synthetic ground motions. There are few case studies applied the 2D inversion techniques to observation records (e.g., Zhao et al., 2004; Hikima and Koketsu, 2010). One of the reasons for the difficulty in the application to actual data is approximating the actual 3D wave propagation to the 2D one. For example, Aoi et al. (1997) pointed out that a simple vertically impinged plane-wave might not sufficiently represent the real complex incident wave and causes errors in the 2D inversion results.
Hikima and Koketsu (2010) assumed that body and surface wave propagations can be fully reproduced by a 2D modeling. However, they could not adequately verify the assumption of the 2D approximations of the body and surface wave propagations. The violation of the 2D assumption might distort an estimated 2D model from a true velocity structure.

In this study, we propose a new waveform inversion method to estimate a 2D S-wave velocity structure of deep sedimentary layers using Love waves from a moderate earthquake. As a preprocessing operation in our method, we first analyze the observation records for the validation of the 2D Love wave propagation assumption to select their segments used in the inversion. We demonstrate our method’s applicability to actual data recorded in the Kanto plain, Japan from comparisons of estimated 2D structures with existing models.

**Earthquake Records**

Here, we describe characteristics of long-period earthquake ground motion records
acquired during a moderate $M_J$ 6.4 (JMA magnitude) earthquake in the Kanto plain, Japan for the use of the later waveform inversion. Figure 1 shows locations of the earthquake epicenter estimated by JMA and seismic observatories whose records are used in this study. The top depths of the seismic bedrock with an S-wave velocity of 3.2 km/s from the HERP2009 model are also illustrated by red contours in the figure. The event is a strike-slip crustal earthquake with a focal depth of 14 km, and has a fault mechanism with a compression axis in the northwest–southeast direction (Fukuyama et al., 1998). Densely distributed strong motion stations have been installed as the parts of strong motion networks such as K-NET, KiK-net (NIED, 2019), and MeSO-net (Hirata et al., 2009) in the study area. The K-NET stations have accelerometers installed on the surface, while there are two accelerometers on the surface and at a depth more than 100 m at the KiK-net stations. Accelerometers at the MeSO-net stations have been installed at a depth of 20 m. We used the surface records at the K-NET and KiK-net stations and the borehole records at the MeSO-net stations regarding as the surface motions. The velocity Fourier amplitude spectra at stations of SIT003, TKY007, and E.KNJM are shown in Figure 2 as examples of the observed ground motion records. These stations
are located in the areas with the thick sedimentary layers (e.g., Shima, 1977) which can amplify ground motions with a dominant period of approximately 8 s. It is well known that the sediments in the central part of the Kanto plain has a predominant period of 8 s of motions in the previous studies with a focus on long-period surface wave propagations (e.g., Tanaka et al., 1980; Yamanaka et al., 1989; Kinoshita et al., 1992).

Figure 3 shows velocity waveforms at eight stations in the northwest–southeast direction, which is almost equal to the transverse direction, along Line-1 from TKY003 to E.MZUM. The vertical axis of each panel denotes a distance from TKY003 at the edge of the basin. This distance is hereinafter referred as a basin-edge distance. The velocity waveforms were calculated from the acceleration records by integrating with the fast Fourier transform. A Gaussian band-pass filter represented by Eq. (1) was also applied to the records in the integration at central periods, $T^m$, of 1, 3, 5, and 9 s for the filtered velocities. This filter shape is expressed as:

$$W(T^n) = \exp\left[-\gamma \left(\frac{(T^n - T^j)}{T^m}\right)^2\right], \quad (1)$$

where $T^j$ represents a discrete period calculated by the Fourier transform. $\gamma$ is a coefficient for determining the filter width, which is set to 50 with reference to Uetake
and Kudo (2001) who investigated characteristics of long-period Love wave propagations at the Kanto area. Figure 3 shows that large amplitude surface waves propagate to the northeast direction, which is almost equal to the radial direction, after arrivals of S-waves. For example, the record at E.IMIM at a period of 3 s has the surface wave arriving approximately at a time of 70 s with much larger amplitude than that of the S-wave. We can also see the surface waves with amplitudes larger than those of the S-waves at the other periods.

**Particle motions**

Horizontal orbits of the velocity motions at the eight stations along Line-1 are shown in Figure 4a. These orbits are calculated for the segments including the surface waves shown in Figure 3. We use the motions at periods from 1 to 10 s because the surface wave motions are predominant especially in this period range. The particle motions at periods of 1 s interval are arranged from the top to the bottom in Figure 4a. The horizontal axes in the figures denote the basin-edge distances of each station. Principal axis directions evaluated by a principal component analysis (Montalbetti and Kanasewich, 1970) are superimposed with red lines. The motions are polarized in the transverse direction
especially at the long-periods such as a period of 10 s. This means that transversely
oriented components of the ground motions are mainly Love waves propagating to the
radial direction.

**Methodology for Estimating 2D S-wave Velocity Structure**

**Preprocessing**

We select the segments used for the waveform inversion by inspecting polarization
features of the horizontal particle motions described in the previous section. The
amplitudes of the motions in the radial direction become large as the basin-edge distances
increase such as in the horizontal motions at SIT003 and E.MZUM at a period of 7 s in
Figure 4a. Accordingly, we observe deviations of the principal directions from the
transverse direction and increases of the ellipticities of the horizontal particle motions.

We also see that the principal axis directions of the particle motions tend to largely differ
from the transverse direction at the short-periods. The short-period ground motions may
be much influenced by scattering waves due to 3D irregularities. These biases of the
principal axis directions of the motions clearly suggest a difficulty in the assumption of
the 2D approximation of the Love wave propagations especially at the short-periods. It is therefore crucial to select band-pass filtered segments of Love wave motions for our inversion to estimate a 2D S-wave velocity structure. Hereinafter, the assumption of the 2D approximation in the Love wave propagations is referred to two-dimensionality. From the above analyses of the observed Love waves at each period, we can suppose that the two-dimensionality is satisfied in the propagation of the long-period Love wave motion at the stations in the long basin-edge distances.

The band-pass filtered segments to be used in the 2D inversion are selected from the polarization features (e.g., Vidale, 1986). As described above, the propagations of the short-period Love waves contain non-two-dimensional waves due to the 3D irregularities. We thus define an E-value to examine the two-dimensionality quantitatively as follows:

\[ E_i^m = \frac{\left| \theta_i^m - \theta_{i\text{ref}}^m \right|}{90} \cdot \left| \phi_i^m - \phi_{i\text{ref}}^m \right|, \] (2)

where \( \theta_i^m \) and \( \phi_i^m \) represent the principal axis direction and the ellipticity of the motions calculated by the principal component analyses using the segments in the transverse and radial directions at the \( i \)-th station and the \( m \)-th period, respectively. Here, \( \theta_i^m \) is measured from \(-90^\circ\) to \(90^\circ\). We suppose that the two-dimensionality of the Love
wave propagation is sufficiently satisfied at a period of 10 s because this period is longer than the predominant one of the Love wave amplification expected from the S-wave velocity structure of the deep sedimentary layers as explained earlier. Thus, the principal axis direction $\theta_i^{ref}$ and the ellipticity $\phi_i^{ref}$ of the motions at a period of 10 s are regarded as reference values. The biases of $\theta_i^m$ and $\phi_i^m$ from $\theta_i^{ref}$ and $\phi_i^{ref}$ are used for the E-value $E_i^m$ as their product. The reference period can be set to the longest one considering ratios of the Fourier amplitude spectra between pre-trigger and principal motions of the observation records. We set the reference period to 10 or 9 s in this study.

Figure 4b shows the E-values of the Love wave segments at the eight stations along Line-1. The horizontal axes in the figure indicate the basin-edge distances. The E-values at periods from 1 to 9 s are arranged from the top to the bottom. The E-values of the stations at a period of every 1 s are plotted with circles in Figure 4b. The E-values have a clear dependency of the periods. Namely, we see that the long-period segments have the small E-values at the stations far from TKY003. For example, the principal axis direction at a period of 8 s deviates from the transverse direction by approximately $45^\circ$ at distances more than 50 km (as shown in Figure 4a) with the E-value of approximately
0.2. Furthermore, the biases of the principal axis directions of the motions at a period of 5 s are also large at distances more than 15 km (as shown in Figure 4a) with E-values of more than 0.2. We therefore considered that the small E-values less than approximately 0.2 indicate the satisfaction of the two-dimensionality assumption for the 2D waveform inversion. The maximum basin-edge distances of the stations with E-values less than 0.2 at the individual periods are shown by arrows in Figures 4a, 4b, and 4c. In Figure 4c, the horizontal and vertical axes denote the basin-edge distance and the period of the motion, respectively. We also illustrate a range of the basin-edge distances and the periods where the two-dimensionality can be satisfied by a gray-shade area. We use the band-pass filtered segments at periods and distances inside this range for the inversion analysis.

The important point of the above preprocessing is that the number of the segments used in the 2D waveform inversion differs depending on the periods of the motions. Table 1 shows the numbers of the stations whose E-values are less than 0.2 and the maximum basin-edge distances used in later inversion analyses at each period. The shorter the periods of the Love wave motions are, the smaller the numbers of the segments are used in the inversion. The short-period Love waves at the stations only around the edge part
can be used in the 2D inversion, while the long-period Love waves at the stations at distances far from the edge can be used.

Model Parameterization

In the waveform inversion, we assume a homogeneous layered model with irregular interfaces. Boundary shapes of layers in the 2D model are deduced in the inversion of the band-pass filtered segments of the Love wave motions as selected in the preprocessing. Physical parameters such as S-wave and P-wave velocities, density, and attenuation factors are given in advance.

We employed a boundary shape parameterization suggested by Aoi et al. (1997). The difference of discretized boundary shapes between present and initial 2D models, \( \zeta(x) - \zeta_0(x) \), is described by a linear combination of coefficients \( p_{n,k} \) and basis functions \( c_k \) as in Eq. (3):

\[
\zeta(x) - \zeta_0(x) = \sum_{k=0}^{K+1} p_{n,k} c_k(x), \quad (3)
\]

where \( x \) denotes a horizontal distance from the left end of the model. Suffixes \( n \) and \( k \) represent the \( n \)-th boundary interface and the \( k \)-th basis function, respectively. \( c_k \) is common to all the interfaces and represented as follows:
\begin{align}
  c_k(x) &= \begin{cases} 
  \frac{1}{2} \left[ 1 + \cos \frac{\pi}{\Delta} (x - x_k) \right], & \text{if } (x_{k-1} \leq x \leq x_{k+1}) \\
  0, & \text{otherwise} 
  \end{cases}, \\
  (k = 1, 2, \ldots, K) \\
  \sin \frac{\pi}{\Delta} (L + x), & \text{if } (-L \leq x \leq -L + \Delta/2) \\
  \frac{1}{2} \left[ 1 + \cos \frac{\pi}{1-\Delta/2} (L - \Delta/2 + x) \right], & \text{if } (-L + \Delta/2 \leq x \leq -L + \Delta) \\
  0, & \text{otherwise} 
\end{align}

\begin{equation}
  c_n(x) = \frac{1}{2} \left[ 1 + \cos \frac{\pi}{1-\Delta/2} (L - \Delta/2 + x) \right], \quad \text{if } (-L + \Delta/2 \leq x \leq -L + \Delta) \quad (5)
\end{equation}

\begin{equation}
  c_{K+1}(x) = c_n(-x), \quad (6)
\end{equation}

The model to be estimated has a length of 2L in the range of \(-L \leq x \leq L\). The model in the horizontal direction is equally divided into \(K+1\) pieces with \(K+2\) node points that are numbered from 0 to \(K+1\). \(\Delta\) indicates the horizontal interval of the basis functions, which is equal to \(2L/K+1\). \(x_k\) denotes the horizontal distance of the \(k\)-th basis function contributing to the depths at distances from \(x_{k-1}\) to \(x_{k+1}\). The \(p_{n,k}\)'s are unknown parameters in the inversion to minimize residuals between the observed and synthetic Love waves as described in the next subsection.

**Inversion Method**

We employed a linearized iterative inversion algorithm using the Gauss-Newton method to obtain an optimized solution. The linearized system to be solved can be...
expressed as:

\[
\begin{bmatrix}
A \\
G
\end{bmatrix}
\begin{bmatrix}
\delta p \\
\end{bmatrix} =
\begin{bmatrix}
e \\
0
\end{bmatrix}, \quad (7)
\]

where \( \delta p \), \( A \), \( G \), and \( e \) are the correction matrix, the Jacobian matrix, the smoothing matrix, and the residual matrix, respectively. The partial derivatives of the Jacobian matrix \( A \) can be numerically computed with model perturbations, \( dp \), using the backward FD scheme. \( dp \) is set to 2.5 times the vertical FD grid spacing in the 2.5D forward calculation, which is described in the next subsection.

The residual matrix \( e \) is defined from the residuals between the observed and synthetic segments. We evaluate the residuals with the L2 norm as follows:

\[
e = \frac{1}{M} \sum_{m=1}^{M} \left[ \frac{1}{N_i^{m}} \sum_{i=1}^{N_i^{m}} \left\{ \sum_{j=1}^{T(m)} \frac{o_{i}^{m}(t_j) - s_{i}^{m}(t_j)}{o_{i}^{m}(t_{max})} \right\} \right]^2, \quad (8)
\]

where \( o_{i}^{m}(t_j) \) and \( s_{i}^{m}(t_j) \) are the observed and synthetic velocities of the Love wave segments selected with the E-values at periods of 1 s interval. Suffixes \( i \), \( j \), and \( m \) represent the \( i \)-th station, the \( j \)-th time interval of the digitized segment, and the \( m \)-th period, respectively. The residual between \( o_{i}^{m}(t_j) \) and \( s_{i}^{m}(t_j) \) is normalized with the observed maximum amplitude \( o_{i}^{m}(t_{max}) \) of each band-pass filtered segment. \( M \) is the...
number of the periods. \( N^m \) is the number of the stations at the individual periods. \( ts^m_i \) and \( te^m_i \) are the start and end times of the segments to be inverted, respectively. The number of rows of the matrix \( e \) is given by multiplying \( M \) with \( N^m \), and the number of samples of the inverted digitized waveforms.

To make the inversion robust, we include the smoothing matrix \( G \) in which differences between the \( p_{n,k} \) and \( p_{n,k+1} \) at the \( n \)-th boundary interface do not change largely as follows:

\[
g = \frac{\beta^2}{L \cdot (K + 1)} (p_{n,k} - p_{n,k+1}), \quad (9)
\]

where \( L \) denotes the number of the layer interfaces to be deduced. \( \beta \) is a coefficient for determining the strength of the smoothing, which is set to 30 in consideration of the stability of solutions from results of several inversion analyses using different coefficients in this study. The number of rows of the matrix \( G \) is given by multiplying \( L \) with \( K + 1 \).

We solve Eq. (7) using the singular value decomposition method (Lawson and Hanson, 1974) to obtain \( \delta p \) for the optimal \( p_{n,k} \)s.

**Forward FD Modeling**

We use a FD staggered-grid formulation of the 2.5D elastic equation of the \( SH \) motion.
(Liner, 1991) with the fourth-order approximation in space and the second-order approximation in time to simulate the Love wave behaviors in the 2D model. Temporal evolutions of a velocity and a stress are solved using the explicit FD scheme. The A1 absorbing boundary condition suggested by Clayton and Engquist (1977) and the buffer region (Cerjan et al., 1985) are introduced at the right, left, and bottom edges of our numerical model to reduce non-physical reflections. The top surface is implemented using the improved vacuum formulation suggested by Zeng et al. (2012). An attenuation is introduced by employing the method of Graves (1996) with a reference frequency of 0.4 Hz (Kasamatsu and Kato, 2020).

A point load is given at horizontal distances more than 20 km from the basin-edge at the top of the bedrock as an external force to generate the Love waves. One of major difficulties in a waveform inversion is a treatment of effects due to a fault rupture process. We therefore not only give the point load far from the basin part but also convolute the observed and synthetic waveforms to solve this problem (e.g., Ji et al., 2000; Amrouche and Yamanaka, 2015). The convoluted waveform $s_i^m(t)$ can be expressed as:

$$s_i^m(t) = F^{-1}\left[T_i^m \cdot O_r^m\right], \quad (10)$$
\[ T_i^m = \left| \frac{C_i^m}{C_r^m} \right| \left| \frac{C_i^m \cdot C_r^{m*}}{C_i^m \cdot C_r^{m*}} \right|, \] (11)

where \( T_i^m \) is a transfer function at the \( m \)-th period defined from the Fourier spectra of the synthetic waveforms at the \( i \)-th station, \( C_i^m \), and a reference station, \( C_r^m \), in Eq. (11).

An asterisk indicates a complex conjugate of the Fourier spectrum. \( T_i^m \) is multiplied to the observed Fourier spectrum of the reference station, \( O_r^m \), to generate the convoluted waveform, \( s_i^m(t) \), using the inverse Fourier transform, \( F^{-1} \), as shown in Eq. (10).

Result of Line-1

Inversion Result

We applied our method to the records of the Mt. Fuji region earthquake in 2011 (\( M_{J}6.4 \)) to retrieve 2D S-wave velocity profiles of the deep sedimentary layers for three survey lines shown in Figure 1. Here, we explain the inversion results for Line-1 in detail.

We prepared an initial model by HERP (2009) with seven homogenous layers with S-wave velocities of 0.5 to 4.6 km/s. The sediments over the seismic bedrock consist of three layers with S-wave velocities of 0.5, 0.9, and 1.5 km/s. The optimal bottom depth
shapes of these three layers were detected from the inversion of the filtered segments of the Love wave motions selected in the previous section (see Figures 4a, 4b, and 4c). Since the HERP2009 model is an old version of the HERP2012 model, we used the HERP2009 model as the initial model in the inversion analysis. We then compare our 2D models with the HERP2012 model.

We estimated a velocity structure at horizontal distances from 0 to 57 km using the 15 basis functions. The horizontal interval of the basis functions, $\Delta$, represented in Eqs. (4) and (5) was approximately 4 km. A total of 45 $p_{n,ls}$ for the three boundaries of the sediments expressed by Eq. (3) were optimized. We used TKY003 as the reference station for the convolution processing. The grid spacings of the 2.5D FD forward modeling were 50 m in the horizontal direction and 20 m in the vertical direction. The vertical grid spacing was 50 m at depths more than 5 km to reduce the computational costs.

Figure 5 shows the inverted and observed velocity waveforms at periods of 1, 3, 5, and 9 s. The amplitudes of the synthetic motions at a period of 9 s are slightly underestimated at E.MZUM, E.IMIM, and SIT009. However, the overall characteristics at all the periods are well reproduced at every station. Even the observed and inverted
motions at relatively short-periods such as 1 and 3 s show high similarities at SIT012, E.KSRM, and E.OKDM. Synthetic waveforms in the initial 2D model (HERP, 2009) were calculated by the forward simulation as shown in Figure 5. These synthetic waveforms at a period of 9 s contain the Love waves with slightly later arrival times than the observed ones at distances more than 47 km. Furthermore, the amplitudes of the synthetic motions at periods of 3 and 5 s in the initial model are overestimated at E.OKDM and E.KSRM.

Figure 6 shows the estimated 2D structure along Line-1. Physical parameters of each layer are listed in Table 2. The third layer can be regarded as an outcrop at distances from 0 to 5 km. The bottom depth of the first layer becomes deep at distances more than 5 km. Then, it reaches 0.2 km at distances around 10 km. The bottom depth of the first layer becomes shallow at distances from 12 to 25 km, and the second layer exists near the surface at approximately 17 km.

The initial model (HERP, 2009) and the HERP2012 model are also shown in Figure 6. At distances more than 30 km, all of our estimated boundary depths of the sedimentary layers are shallower than those in the HERP2012 model. The differences between the
inverted and HERP2012 models are remarkable in the shallow parts of the basin-edge area. The estimated bottom depth of the second layer becomes shallow at distances less than 10 km and deep at distances more than 10 km. As the results, the estimated bottom interface of the second layer becomes steeper than that in the HERP2012 model at distances from 5 to 15 km. The bottom depth of the first layer becomes shallow slightly around SIT009 in the HERP2012 model, but the top of the second layer does not exist near the surface unlike our results.

As above, our obtained 2D profile shows the more irregular structural features around the basin-edge area than the HERP2012 model. This suggests we could model the detailed velocity structure especially in the edge part due to the introduction of the short-period Love waves in the inversion.

Comparison with HERP2017 Model

In the Kanto plain, the HERP2017 model was constructed as a further updated version of the HERP2012 model. The HERP2017 model contains layers with S-wave velocities less than 0.5 km/s, which are not included in the HERP2009 and HERP2012 models. Our inversion results, therefore, cannot be directly compared with the HERP2017 model.
However, we can still find a similarity between them around the basin-edge part along Line-1. Figure 7 compares their 2D structural models at distances from 0 to 22 km as the edge part. The top depths of the bedrock in the HERP2017 and our models show a high similarity. On the other hand, our estimated upper depths of the second and third layers with S-wave velocities of 0.9 km/s and 1.5 km/s are different from those of the HERP2017 model.

**Comparison with Surface Geological Classifications**

The surface geological classifications (Geological Survey of Japan, 2015) around the basin-edge area along Line-1 are indicated in Figure 7. According to the previous study on relations to S-wave velocities with geological ages (Yamamizu et al., 1981), the layers with S-wave velocities of 1.2 to 1.6 km/s and 0.4 to 0.9 km/s belong to Neogene and Quaternary ages, respectively.

In the estimated structure, the third layer with an S-wave velocity of 1.5 km/s can be regarded as an outcrop at distances from 0 to 5 km because of the thin thicknesses of the overlying layers. The surface geology in this area mostly belongs to Neogene age. At distances from 5 to 13 km, the estimated bottom depth of the first layer with an S-wave
velocity of 0.5 km/s becomes deep gradually away from a station of SIT012. The geological age of the surface formation in this area is categorized as Quaternary with lower terrace sediments. In the estimated structure, the second layer with an S-wave velocity of 0.9 km/s exists near the surface locally at distances around 17 km. Although the surface geology around this part also belongs to Quaternary age, Geological Survey of Japan (2015) reported an existence of middle terrace sediments which are older than the lower terrace ones in this part. Thus, we can confirm qualitatively that the regional variation of the near-surface S-wave velocities in the estimated model is similar to a distribution of the surface geological classifications.

Results of Lines-2 and 3

Line-2

Figure 8a illustrates an inverted 2D velocity structure along Line-2. The horizontal length along Line-2 is approximately 58 km from the basin edge. We used 28 stations along this line in the inversion analysis. Segments of the Love waves used in the inversion were selected at individual periods from 1 to 10 s except for 6 s from the E-values as
shown in Table 1. We excluded the segments at a period of 6 s in the inversion because
the Fourier amplitude spectrum of the motion at each station has a trough at this period.
The horizontal axis in Figure 8a indicates a horizontal distance from the reference station
of OK.NKYM, where the seismic bedrock exists near the surface. Each estimated
boundary depth is shallower than that in the initial model (HERP, 2009) at almost all the
distances. These boundary depths are also shallower than those in the HERP2012 model
at distances more than 25 km. We found remarkable differences between the estimated
and initial structures in the basin-edge part. The estimated bottom depth of the first layer
is shallower than that in the initial model at approximately 10 km, then it gradually
deeps away from OK.NKYM. The shape of the bottom interface of the first layer in
the edge part is similar to that in the HERP2012 model.

Line-3

Figure 8b shows an inverted 2D velocity structure along Line-3. The horizontal length
along Line-3 is approximately 34 km. We used seven records at periods from 1 to 9 s
except for 6 s in the inversion analysis as shown in Table 1. The segments at a period of
6 s were excluded with the same reason in the Line-2 inversion as explained in the
previous subsection. All the boundary depths of the sedimentary layers are significantly shallower than those in the initial model (HERP, 2009) at distances less than 30 km, and they are in good agreement with the HERP2012 model. The second layer with an S-wave velocity of 0.9 km/s exists near the surface at distances from 15 to 20 km. It was reported that the S-wave velocities of the top layers in the deep sediments around this area are 0.7 to 0.8 km/s (Yamanaka and Yamada, 2006). Although we do not consider lateral variations of the S-wave velocities, the existence of the second layer with an S-wave velocity of 0.9 km/s near the surface means that our estimated structure has similar features to the surface S-wave variation of the model by Yamanaka and Yamada (2006).

The estimated bottom depths of the layers 1 to 3 at distances more than 30 km in our model are almost the same as those of the HERP2009 and HERP2012 models.

Effect of Introducing Short-Period Motions

We indicated the importance of the preprocessing operation for confirming the two-dimensionality of the Love wave propagations before the 2D waveform inversion with
its applications to the earthquake records in the previous sections. As the result of introducing the preprocessing step, we were able to obtain the detailed 2D profiles especially at the edge parts with the introduction of the short-period Love wave motions as well as the long-period ones. In this section, a numerical experiment using synthetic waveforms generated by a 3D computation is performed to examine an effectiveness of the introduction of the short-period motions in the waveform inversion.

The synthetic records were generated using the 3D model by HERP (2012) for the numerical test. The ground motions of the Mt. Fuji region earthquake in 2011 ($M_s6.4$) were simulated at the seven stations along Line-3 employing a 3D FD scheme (e.g., Graves, 1996; Pitarka, 1999). We regard them as the synthetic-earthquake ground motion records. These seven stations are located at the same locations as the actual observatories.

A double couple point force with a source mechanism by F-net (Fukuyama et al., 1998) was given as an external force at the hypocenter location. Fault parameters such as a seismic moment are listed in Figure 1. The HERP2012 model used in the 3D calculation was discretized with a horizontal grid spacing of 100 m and a vertical grid spacing of 10 m.
We performed two inversion analyses to model sedimentary layers from 0 to 34 km along Line-3 using the synthetic records at periods from 1 to 9 s and from 7 to 9 s, respectively. Then, we compare retrieved 2D profiles from the two inversions with the above true one (HERP, 2012). In the former inversion, the segments at periods from 1 to 9 s selected from the E-values were inverted as the same as the inversion using actual observation records in the previous section. In the latter inversion, only the segments at periods of 7 to 9 s were used at entire the distances from 0 to 34 km without consideration of E-values. Other conditions in the two inversions were the same as those in the inversion described in the previous section.

The 2D profile from the inversion using the segments at periods from 1 to 9 s is illustrated in Figure 9 with the true structure (HERP, 2012). The true model is well reconstructed by the inversion. The 2D velocity structure estimated using the segments at periods from 7 to 9 s is also shown in Figure 9. Obvious differences are observed between the estimated and true structures at the basin-edge part. The bottom depth of the first layer estimated using only the long-period segments is shallower than that of the true structure at distances from 0 to 15 km. Furthermore, the upper depths of the second and
third layers in the estimated structure using the segments at periods from 7 to 9 s also become shallow at distances around 5 km. Thus, the inversion using only the long-period Love waves cannot offer the accurate 2D profile at the basin-edge part. The preprocessing operation with the E-values to include the short-period segments in the 2D inversion enhances a reliability of an inverted 2D profile.

Discussion and Conclusion

We have proposed a waveform inversion method to estimate a 2D S-wave velocity structure of deep sedimentary layers using broadband Love waves. As a preprocessing operation in our inversion scheme, we decompose earthquake observation records into velocity waveforms at periods of 1 s interval for a confirmation of an assumption of 2D propagations of the Love waves with a principal component analysis of horizontal ground motions. Our linearized iterative inversion analysis allows an accurate estimation of boundary shapes from a top depth of a seismic bedrock to a bottom depth of a layer with an S-wave velocity of approximately 0.5 km/s from the segments of the Love wave
motions at a period every 1 s which is selected in the preprocessing step.

We have demonstrated the effectiveness of our technique with the applications to the observed seismograms in the Kanto plain, Japan. The remarkable differences were observed between our inverted and the existing models (HERP, 2009; HERP, 2012) especially at the basin-edge area. Our method can offer the detailed 2D velocity profiles at the edges due to the introduction of the short-period segments selected from the E-values for the inversions. A velocity structure at a basin edge strongly affects earthquake ground motions in a basin because of generations of surface waves (e.g. Vidale and Helmberger, 1988; Kawase, 1996; Hartzell et al., 2016). 2D profiles including the details of the basin-edge structure along many lines from our proposed inversions can be used to improve the existing 3D models for accurate earthquake ground motion evaluations.

In an actual work for a prediction of a strong ground motion from a large earthquake, a lot of the inversion analyses using various conditions are required in estimations of S-wave velocity structures. Since a high-performance computer for the inversions is not always available in the actual work, we have to reduce a calculational cost in a forward calculation as possible. In this study, our 2D inversion employing the 2.5D FD
calculations took approximately one day to obtain an optimal solution by performing parallel computing using the open MP algorithm with a conventional PC equipped with four Intel Xeon processors (E7-8891 v4, 2.8GHz). On the other hand, the 3D FD calculation to generate the synthetic records for the numerical test in the previous section took several days with the same conventional PC. This indicates several hundred times differences in the calculation costs between the 2D and 3D inversion analyses. Although the computational costs depend on a PC’s performance used in the calculation, it is still difficult to obtain an optimal solution from a 3D waveform inversion in a reasonable computational cost with a conventional PC.

The above differences in the computational cost can be very critical to conduct many inversions of records at dense seismic stations in a large basin. Recently, many earthquake observatories have been installed in large basins, especially in seismically active areas. Our 2D inversion technique can be effectively applied to earthquake records along many lines in such the areas with reasonable computational costs.

**Declarations**
Ethics approval and consent to participate

Consent for publication

List of abbreviations

HERP: The Headquarters for Earthquake Research Promotion

NIED: National Research Institute for Earth Science and Disaster Resilience

JMA: Japan Meteorological Agency

3D: Three-dimensional

2D: Two-dimensional

2.5D: Two and one-half dimensional

1D: One-dimensional

FD: finite-difference

Availability of data and materials

The K-NET and KiK-net strong motion data are available at the Web site of strong-motion seismograph networks operated by the NIED (http://www.kyoshin.bosai.go.jp/).
The MeSO-net strong motion data are available at the Hi-net web site operated by the NIED (https://hinetwww11.bosai.go.jp/auth/download/cont/?LANG=en). The JMA catalog data are available at the Hi-net web site operated by the NIED (http://www.hinet.bosai.go.jp/?LANG=en).

Competing interests

The authors declare that they have no competing interests.

Funding

This work was supported by JSPS KAKENHI Grant Number JP17K01324

Authors' contributions

KK conducted the analysis. KK and HY drafted the manuscript. All authors read and approved the final manuscript.

Acknowledgements

We would like to thank the NIED for providing earthquake observation record data and the JMA for their catalog data. We used the Generic Mapping Tools (Wessel et al., 2013) for drawing the figures.

Authors' information
References

• Amrouche, M. and Yamanaka, H. (2015) Two-dimensional shallow soil profiling using time-domain waveform inversion. Geophysics 80(1):EN27-EN41, doi:10.1190/geo2014-0027.1

• Aoi, S., Iwata, T., Irikura, K., and Sanchez-Sesma, J. (1995) Waveform inversion for determining the boundary shape of a basin structure. Bull. Seism. Soc. Am. 85(5):1445-1455

• Aoi, S., Iwata, T., Fujiwara, H., and Irikura, K. (1997) Boundary shape waveform inversion for two-dimensional basin structure using three-component array data of plane incident wave with an arbitrary azimuth. Bull. Seism. Soc. Am. 87(1):222-233

• Aoi, S. (2002) Boundary shape waveform inversion for estimating the depth of three-
dimensional basin structures. Bull. Seism. Soc. Am. 92(6):2410-2418.

doi:10.1785/0120010245

- Cerjan, C., Kosloff, D., Kosloff, R., and Reshef, M. (1985) A nonreflecting boundary condition for discrete acoustic and elastic wave equations. Geophysics 50(4):705-708, doi:10.1190/1.1441945

- Clayton, R. and Engquist, B. (1977) Absorbing boundary conditions for acoustic and elastic wave equations, Bull. Seism. Soc. Am. 67(6):1529-1540

- Fukuyama, E., Ishida, M., Douglas, S., and Kawai, H. (1998) Automated seismic moment tensor determination by using on-line broadband seismic waveforms. Zisin (Journal of the Seismological Society of Japan. 2nd ser.) 51(1):149-156, doi:10.4294/zisin1948.51.1_149 (in Japanese with English abstract)

- Geological Survey of Japan, AIST (ed.) (2015) Seamless digital geological map of Japan 1: 200,000. May 29, 2015 version. Geological Survey of Japan, National Institute of Advanced Industrial Science and Technology.

- Graves, R. (1996) Simulating seismic wave propagation in 3D elastic media using staggered-grid finite differences. Bull. Seism. Soc. Am. 86(4):1091-1106
Hartzell, S., Leeds, L., Ramirez-Guzman, L., Allen, and P., Schmitt, G. (2016) Seismic site characterization of an urban sedimentary basin, Livermore Valley, California: Site response, basin-edge-induced surface waves, and 3D simulations. Bull. Seism. Soc. Am. 106(2):609-631, doi:10.1785/0120150289

Hikima, K. and Koketsu., K (2010) Waveform inversion for 2-D velocity structures and construction of 3-D velocity structure using its results. Abstracts of the 13th Japan Earthquake Engineering Symposium, International Congress Center Epochal Tsukuba, Ibaraki, 17-20 November 2010 (in Japanese with English abstract)

Hirata, N., Sakai, S., Sato, H., Satake, K., and Koketsu, K. (2009) An outline of the special project for earthquake disaster mitigation in the Tokyo Metropolitan Area - subproject 1: Characterization of the plate structure and source faults in and around the Tokyo Metropolitan area. Bulletin of the Earthquake Research Institute, University of Tokyo 84(2):41-56 (in Japanese with English abstract)

Iwaki, A. and Iwata, T. (2011) Estimation of three-dimensional boundary shape of the Osaka sedimentary basin by waveform inversion. Geophys J Int 186(3):1255-1278, doi:10.1111/j.1365-246X.2011.05102.x
Ji, C., Helmerger, D., and Wald, D. (2000) Basin structure estimation by waveform modeling: Forward and inverse methods. Bull. Seism. Soc. Am. 90(4):964-976, doi:10.1785/0119990080

Kasamatsu, K. and Kato, K. (2020) Long period ground motion simulation of the 2011 off the Pacific coast of Tohoku Earthquake using theoretical Green's function: Examination based on inverted pseudo point source model. Journal of Structural and Construction Engineering 85(769):427-437, doi:10.3130/aijs.85.427 (in Japanese)

Kawase, H. (1996) The cause of the damage belt in Kobe: “The Basin-Edge Effect,” constructive interference of the direct s-wave with the basin-induced diffracted/ Rayleigh waves. Seis. Res. Lett. 67(5):25-34, doi:10.1785/gssrl.67.5.25

Kinoshita, S., Fujiwara, H., Mikoshiba, T., and Hoshino, T. (1992) Secondary love waves observed by a strong- motion array in the Tokyo Lowlands, Japan. Journal Physics Earth 40(1):99-116, doi:10.4294/jpe1952.40.99

Koketsu., K., Miyake, H., and Suzuki, H. (2012) Japan integrated velocity structure model version 1. Abstracts of Proceedings of the fifteenth World Conference on Earthquake Engineering, Lisbon, Portugal, 24-28 September 2012
Lawson, C. and Hanson, R. (1974) Solving least squares problems. Society for Industrial and Applied Mathematics Classics in Applied Mathematics 15, doi: 10.1137/1.9781611971217

Liner, L. (1991) Theory of a 2.5-D acoustic wave equation for constant density media. Geophysics 56(12):2114-2117, doi:10.1190/1.1888858

Montalbetti, F. and Kanasewich, R. (1970) Enhancement of teleseismic body phases with a polarization filter. Geophysical Journal International 21(2):119-129, doi:10.1111/j.1365-246X.1970.tb01771.x

National Research Institute for Earth Science and Disaster Resilience (2019) NIED K-NET, KiK-net. National Research Institute for Earth Science and Disaster Resilience, doi:10.17598/NIED.0004

Pitarka, A. (1999) 3D Elastic finite-difference modeling of seismic motion using staggered grids with nonuniform spacing. Bull. Seism. Soc. Am. 89(1):54-68

Sato, T., Graves, R., and Somerville, G. (1999) Three-dimensional finite-difference simulations of long-period strong motions in the Tokyo metropolitan area during the 1990 Odawara earthquake (Mj 5.1) and the great 1923 Kanto earthquake (Ms 8.2) in
Senna, S., Maeda, T., Inagaki, Y., Suzuki, H., Matsuyama, H., and Fujiwara, H. (2013) Modeling of the subsurface structure from the seismic bedrock to the ground surface for a broadband strong motion evaluation. Journal of Disaster Research 8(5):889-903, doi:10.20965/jdr.2013.p0889

Shima, E. (1977) On the deep underground structure of Tokyo Metropolitan Area. Abstracts of Proceedings of the sixth World Conference on Earthquake Engineering, New Delhi, India, 10-14 January 1977

Tanaka, T., Yoshizawa, S., and Osawa, Y. (1980) Characteristics of strong earthquake ground motion in the period range from 1 to 15 seconds, Abstract of Proceedings of the seventh World Conference on Earthquake Engineering, Istanbul, Turkey, 8-13 September 1980

The Headquarters for Earthquake Research Promotion (2009) Seismic Hazard Map for the Long-period Ground Motion, the Trial Version in 2009 (Choshuki Jishindo Yosoku Chizu 2009 Shisakuban) (in Japanese). https://www.jishin.go.jp/evaluation/seismic_hazard_map/lpshm/09_choshuki/.
Accessed 11 Apr 2016

- The Headquarters for Earthquake Research Promotion (2012) Seismic Hazard Map for the Long-period Ground Motion, the Trial Version in 2012 (Choshuki Jishindo Yosoku Chizu 2012 Shisakuban) (in Japanese).

https://www.jishin.go.jp/evaluation/seismic_hazard_map/lpshm/12_choshuki/.

Accessed 2 May 2016

- The Headquarters for Earthquake Research Promotion (2017) Kantou Chihouno Senbu Sinbu Tougou Jibankouzou Moderu (in Japanese).

https://www.jishin.go.jp/evaluation/strong_motion/underground_model/integration_model_kanto/.

Accessed 6 Apr 2020

- Uetake, T. and Kudo, K. (2001) Three dimensional S-wave velocity structure in and around Ashigara valley, west of Kanagawa prefecture, Japan, evaluated from love wave dispersion data. Zisin (Journal of the Seismological Society of Japan. 2nd ser.) 54(2):281-297, doi:10.4294/zisin1948.54.2_281 (in Japanese with English abstract)

- Vidale, E. (1986) Complex polarization analysis of particle motion. Bull. Seism. Soc.
Vidale, E. and Helmberger, V. (1988) Elastic finite-difference modeling of the 1971 San Fernando, California earthquake. Bull. Seism. Soc. Am. 78(1):122-141

Wessel P, Smith WHF, Scharroo R, Luis J, and Wobbe F (2013) Generic mapping tools: improved version released. EOS Trans AGU 94:409, doi:10.1002/2013EO450001

Yamamizu, F., Takahashi, H., Goto, N., and Ohta, Y. (1977) Shear wave velocities in deep soil deposits Part III: Measurements in the borehole of the Fuchu observatory to the depth of 2,750 m and a summary of the results. Zisin (Journal of the Seismological Society of Japan. 2nd ser.) 34(4):465-479, doi:10.4294/zisin1948.30.4_415 (in Japanese with English abstract)

Yamanaka, H., Seo, K., and Samano, T. (1989) Effects of sedimentary layers on surface-wave propagation. Bull. Seism. Soc. Am. 79(3):631-644

Yamanaka, H. and Yamada, N. (2006) Modeling 3D S-wave velocity structure of Kanto basin for estimation of earthquake ground motion. Butsuri-Tansa 59(6):549-560, doi:10.3124/segj.59.549 (in Japanese with English abstract)
Yoshimoto, K. and Takemura, S. (2014) Surface wave excitation at the northern edge of the Kanto Basin, Japan. Earth, Planets and Space 66(16), doi:10.1186/1880-5981-66-16.

Zeng, C., Xia, J., Miller, R., and Tsoflias, G. (2012) An improved vacuum formulation for 2D finite-difference modeling of Rayleigh waves including surface topography and internal discontinuities. GEOPHYSICS 77(1):T1-T9, doi:10.1190/geo2011-0067.1.

Zhao, L., Zheng, T., and Xu, W. (2004) Modeling the Jiyang depression, northern China, using a wave-field extrapolation finite-difference method and waveform inversion. Bull. Seism. Soc. Am. 94(3):988-1001, doi:10.1785/0120030167.

**Figure Legends**

**Fig. 1.** Locations of epicenter and strong ground motion observatories used in this study indicated by focal sphere and squares, respectively. Red contours denote top depths of seismic bedrock modeled by HERP (2009). 2D S-wave velocity structures...
of deep sedimentary layers were estimated along Line-1 to Line-3 as represented
with black thick lines.

- **Fig. 2.** Fourier amplitude spectra observed at three stations. Each line shows a
  spectrum in the normal direction along Line-1 to Line-3 in Figure 1. These spectra
  are smoothed using the Parzen window with a bandwidth of 0.02 Hz.

- **Fig. 3.** Observed velocity waveforms in transverse direction. These waveforms are
  band-pass filtered at periods of 1, 3, 5, and 9 s. Solid triangles and circles indicate
  arrival times of P-waves and S-waves, respectively. Vertical axes denote basin-edge
  distances.

- **Fig. 4.** Results of principal component analyses for records along Line-1. (a)
  Horizontal particle motions and principal axis directions indicated with black and
  red lines, respectively. The horizontal axis indicates a basin-edge distance. (b) E-
  values for observatories at individual periods. Waveforms at stations located within
  distances indicated by arrows are used in 2D inversion analysis. (c) Relationship
  between periods of ground motions and basin-edge distances for a confirmation of a
  two-dimensionality in 2D inversion.
Fig. 5. Comparison between observed and calculated velocity waveforms at periods of 1, 3, 5, and 9 s along Line-1. Black, red, and blue dotted lines represent observed, inverted, and synthetic waveforms, respectively. The synthetic waveforms were generated from initial model by HERP (2009).

Fig. 6. Comparison of inverted 2D velocity structure with previous models. Red, gray dotted, and black lines represent the estimated structure and structural models by HERP (2009, 2012) along Line-1, respectively. The horizontal axis indicates a basin-edge distance. Triangles denote observatory locations.

Fig. 7 Comparison of 2D velocity structural models around a basin-edge area along Line-1. Red and black lines denote the inverted model and the model by HERP (2017), respectively. S-wave velocities (Vs) denoted by black lines represent those of the model by HERP (2017). Surface geological classifications (Geological Survey of Japan, 2015) are also indicated with gray, black broken, and black solid lines on the top of the models.

Fig. 8. Inverted 2D velocity structures along (a) Line-2 and (b) Line-3. Legends of the figures are the same as Figure 6.
Fig. 9. Comparison of 2D velocity structures in a numerical test. Gray line represents true structure (HERP, 2012). Black solid and dotted lines indicate inverted structures from band-pass filtered segments at periods from 1 to 9 s and from 7 to 9 s, respectively.

Table Legends

Table 1. Number of stations and maximum basin-edge distances used in inversion analyses along Lines-1 to 3 at each period.

Table 2. Physical parameters of layers 1–4 in Figures 6, 7, 8a, 8b, and 9.