Modelling high-frequency seismograms at ocean bottom seismometers: effects of heterogeneous structures on source parameter estimation for small offshore earthquakes and shallow low-frequency tremors

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Running Title

Modelling HF seismograms at OBSs

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Summary

The source characteristics of offshore seismic events, especially regular (or fast) and slow earthquakes, can provide key information on their source physics and frictional conditions at the plate boundary. Due to strong three-dimensional heterogeneities in offshore regions, such as those relating to seawater, accretionary prism, and small-scale velocity heterogeneity, conventional methods using a one-dimensional Earth model may misestimate source parameters such as the duration and radiation energy. Estimations could become severe inaccuracies for small offshore seismic events because high-frequency (> 1 Hz) seismograms, which are strongly affected by three-dimensional heterogeneities, are only available for analysis because of their signal-to-noise ratio. To investigate the effects of offshore heterogeneities on source parameter estimation for small seismic events, we analysed both observed and simulated high-frequency seismograms southeast off the Kii Peninsula, Japan, in the Nankai subduction zone. Numerical simulations of seismic wave propagation using a three-dimensional velocity structure model clarified the effects of each heterogeneity. Comparisons between observations and model simulations demonstrated that the thick low-velocity accretionary prism has significant effects on high-frequency seismic wave propagation. Especially for shallow low-frequency tremors occurring at depths just below the accretionary prism toe, seismogram durations are significantly broader than an assumed source duration, even for stations with epicentral distances of approximately 10 km. Spindle-shape seismogram envelopes were observed even at such close stations. Our results suggest that incorporating three-dimensional heterogeneities is necessary for practical estimation of source parameters for small offshore events.

Keywords:

Computational seismology, earthquake ground motions, earthquake source observations, seismicity and tectonics, wave propagation, wave scattering and diffraction

1. Introduction

To understand the physics and characteristics of earthquake sources, seismologists analyse observed seismograms around the world. Following the correction of the frequency responses of seismometers and local site effects at each station, the observed seismograms provide information on the source rupture complexity and heterogeneous structures along each propagation path. For centroid moment tensor (CMT) inversion for teleseismic events or local crustal earthquakes using low-frequency (< 0.1 Hz) seismograms, path effects are typically evaluated by assuming a simple one-dimensional (1D) Earth model (e.g., Ekström et al., 2012; Kanamori & Rivera, 2008; Nakano et al., 2008) because low-frequency seismic waves are usually considered insensitive to structural heterogeneities. Given the recent developments in computer resources and three-dimensional (3D) subsurface structure models, CMT inversions using 3D velocity models have been practically conducted for not only crustal, but also offshore earthquakes (e.g., Hejrani et al. 2017; Takemura, Okuwaki, et al. 2020; Wang & Zhan 2020). The 3D heterogeneities in the offshore regions, such as seawater, accretionary prism, and small-scale velocity heterogeneity, are developed just beneath the epicentre regions of offshore subduction zone earthquakes. Thus, the 3D CMT method better constrains earthquake source characteristics compared to the 1D method, especially for offshore earthquakes.

Moment tensor inversion is only applicable to moderate-to-large earthquakes because of the signal-to-noise ratios of coherent low-frequency seismic signals. Thus, small events, such as microearthquakes or tectonic and volcanic tremors, are usually evaluated via methods based on the amplitudes or energies of high-frequency (> 1 Hz) seismic waves (e.g., Fletcher & McGarr, 2011; Gusev & Pavlov, 1991; Maeda & Obara, 2009; Nakahara, 2008; Poiata et al., 2018; Sawazaki & Enescu, 2014; Staudenmaier et al., 2016). The precise evaluation of small events is important for not only understanding source physics (e.g., Ellsworth & Bulut, 2018; Gomberg et al., 2016; Hawthorne et al., 2019; Ide et al., 2003; Supino et al., 2020; Thomas et al., 2016; Uchide & Ide, 2010), but also for monitoring crustal and volcanic activities (e.g., Battaglia & Aki, 2003; Kato et al., 2012; Kumagai et al., 2013; Kurokawa et al., 2016; Peng & Zhao, 2009; Yabe & Ide, 2014). The characteristics of high-frequency seismic waves, and related small-scale structural properties in onshore regions, have been widely investigated by numerical, theoretical, and observational approaches (e.g., Carcolé & Sato 2010; Chaput et al. 2015; Margerin 2005; Morioka et al. 2017; Saito et al. 2002; Takahashi et al. 2009; Takemura et al. 2016, 2017; Wegler et al. 2006). Consequently, the estimated location and energy for onshore small seismic events are generally robust even with 1D conventional methods.

Due to long distances between onshore stations and offshore seismic sources, the source parameters of small offshore seismic events cannot be precisely determined from onshore seismic stations. Thus, temporal observations of ocean bottom seismometers (OBSs) have been extensively carried out. Around the Japanese Islands, permanent OBS networks of DONET and S-net (National Research Institute for Earth Science and Disaster Resilience, 2019c, 2019a) have been recently deployed. Several studies have used temporal OBS data or such permanent networks to estimate the source characteristics of small earthquakes and shallow low-frequency tremors (LFTs) (e.g., Nakano et al., 2015, 2019; Nishikawa et al., 2019; Tamaribuchi et al., 2019; Tanaka et al., 2019; Yabe et al., 2019). Offshore 3D heterogeneities, seawater, and bathymetry significantly affect seismic wave propagation even for low-frequency ground motions (e.g., Gomberg 2018; Guo et al. 2016; Nakamura et al. 2015; Shapiro et al. 1998, Takemura, Kubo, et al. 2019; Volk et al. 2017), complicated seismic waves such as T phase, Scholte, acoustic-coupled, and ocean-mode Rayleigh waves, as has been observed in previous studies (e.g., Haney & Tsai, 2017; Muyzert, 2007; Noguchi et al., 2016; Obara & Maeda, 2009; Okal, 2008; Takeo et al., 2014; Tonegawa, Fukao, Takahashi, et al., 2015). However, despite such complexities, 1D velocity structure models have been widely used in seismic source analyses, where incorrect estimations of source parameters may have been provided, especially for the duration and source energy.

Figure 1 shows examples of NS-component filtered velocity seismograms of onshore F-net and offshore DONET stations (National Research Institute for Earth Science and Disaster Resilience, 2019b, 2019a). The sensor orientations of X and Y at the DONET station (M.KMD14) were rotated to North (NS) and East (EW) by using estimations of sensor orientations of Nakano et al. (2012). The estimated event sizes of regular earthquakes in Figure 1 were similar. The event size of a shallow LFT, estimated by using an accompanied shallow very low-frequency earthquake (VLFE), was also similar. However, the observed waveforms of these events were very different. Although an F-net seismogram of an onshore regular earthquake showed a short-duration pulse-like *S*-wave envelope, a DONET seismogram of an offshore regular earthquake was broadened as result of the multiple wave packets. The seismogram of a shallow LFT was more complex, in which a spindle-shaped seismogram envelope was detected. In order to understand the differences in source processes among these seismic events, it is necessary to evaluate the effects of seismic wave propagation through 3D strong heterogeneities in offshore regions, including the seawater, low-velocity accretionary prism, and small-scale velocity heterogeneity.

In this study, to investigate the effects of offshore heterogeneities on high-frequency seismic waves and source parameter estimations, we analysed simulated and observed high-frequency seismograms at offshore seismic stations. Using a realistic 3D offshore heterogeneous model, we conducted 3D numerical simulations of seismic wave propagation via open-source finite-difference method (FDM) software (OpenSWPC; Maeda et al., 2017). Simulated seismograms enable analysis of the effects of each heterogeneity on offshore high-frequency seismograms. By combining various structural model simulations and observed seismograms, we can evaluate the effects of the seawater, accretionary prism, and small-scale velocity heterogeneity on high-frequency seismograms at offshore stations. We also estimated a source time function (STF) of a shallow LFT via the conventional method, and conducted a numerical simulation of seismic wave propagation by using the estimated STF. By comparing simulated and observed seismogram envelope shapes, we also discuss reliable settings for source parameter estimations of offshore small seismic events via conventional methods.

2. Data and methods

The target region, southeast off the Kii Peninsula, southwest Japan (Figures 1 and 2a), has repeatedly experienced large megathrust earthquakes (e.g., Ando, 1975). To understand stress and frictional conditions of the megathrust zone along the Nankai Trough, regular and slow earthquakes around this region have been monitored by DONET stations (locations represented by diamond symbols in Figures 1 and 2a), operated jointly by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC) and the National Research Institute for Earth Science and Disaster Resilience (NIED). Various small seismic phenomena, including small interplate and intraplate regular earthquakes and shallow slow earthquakes, have been observed by broadband seismometers installed at each DONET station. Interplate earthquake activity is quite low, except for activity of the 2016 southeast off Kii Peninsula earthquake (Nakano, Hyodo, et al., 2018; Wallace et al., 2016), but shallow slow earthquakes are often observed (e.g., Annoura et al., 2017; Araki et al., 2017; Nakano, Hori, et al., 2018; Toh et al., 2018). Precise source parameter estimation of such small phenomena is an important issue for understanding earthquake source physics and seismic activity in the Nankai subduction zone.

The effects of offshore heterogeneities on seismic wave propagation were investigated by using both simulated seismograms and three-component broadband velocity seismograms of DONET stations. Simulated seismograms were evaluated at the same locations as the observational stations and virtual seafloor seismic stations, which were uniformly distributed at a horizontal interval of 0.05° in the model region. We used a parallel staggered-grid FDM code of OpenSWPC software (Maeda et al., 2017) to simulate seismic wave propagation of a shallow LFT and small regular earthquakes in the 3D viscoelastic medium. We employed the Cartesian coordinate system of Aki & Richards (2002), where x, y, and z are taken as north, east, and down, respectively. To avoid artificial reflections from model boundaries, perfectly matched layer boundary conditions (e.g., Zhang & Shen, 2010) were assumed at each model boundary.

The 3D velocity structure model developed by Koketsu *et al.* (2012) was used as the model beneath the bedrock. This model has been widely used for various applications across Japan (e.g., Furumura & Kennett 2018; Iwaki et al. 2018; Petukhin et al. 2016; Takemura et al. 2017, 2020). The ETOPO1 model (Amante & Eakins, 2009) was used as the topographic model in simulations. The 1D *S*-wave velocity structures beneath the DONET stations (Tonegawa et al., 2017) were used to model a 3D velocity structure within the accretionary prism. We converted the depth-varying velocity structure model of Tonegawa *et al.* (2017) to a 5-layer model beneath each DONET station. Extrapolation and interoperation of depths of the bottom of each layer via the '*Surface*' gridding algorithm of Generic Mapping Tools (Wessel et al., 2013) were applied to construct the 3D layered accretionary prism model (Figure S1). Details of the model construction are described in the Supporting Material (Text S1, Figures S1-S4). The *P*- and *S*-wave velocities, density (V_P, V_S, and ρ), and attenuations (Q_P and Q_S) for each layer are listed in Table 1. The minimum V_S of 0.5 km/s was assumed in the solid column. The 3D simulation model for shallow LFT and small earthquakes covered an area of 120 × 82.5 × 45 km^3 (delineated by the dashed-blue rectangle in Figure 2a), which was discretised by a grid interval of 0.015 km. Cross-sections of the constructed layered structure model are illustrated in Figures 2b and 2c. This model is the reference model. Simulations of high-frequency seismic wave propagation were conducted using calculation resources of the Earth Simulator at JAMSTEC. Each simulation required 24 TB of computer memory and a wall-clock time of 6.3 h by parallel computing using 1,280 nodes (5,120 cores) of the Earth Simulator to evaluate a seismic wave propagation of 60 s.

According to the observed seismic activity in this region, many intraslab earthquakes occurred at depths of 10–15 km and 20–30 km (e.g., Nakano et al., 2015; Sakai et al., 2005;

Takemura et al., 2020). Interplate earthquakes (~10 km), related with the 2016 southeast off Kii Peninsula earthquake, and shallow (< 7 km) slow earthquakes were also observed. Thus, we investigated the effects of 3D heterogeneities on seismic wave propagations from those three different hypocentres in our simulations (Table 2). Simulation results for earthquakes at different depths allowed us to investigate depth-dependent structural responses. Events A, B, and C occurred on 16 April 2016, 19 April 2016, and 4 December 2014, respectively. Because source parameters for events A–C were not precisely estimated, we assumed double-couple point sources of CMT solutions for the nearest shallow VLFE or moderate size earthquakes (Takemura et al. 2019, 2020). Although source durations of LFTs estimated in previous studies ranged from 10 to 30 s (e.g., Gomberg et al., 2016; Nakano et al., 2019; Yabe et al., 2019), the present source models all employed a simple triangle function duration of 0.2 s, which roughly corresponds to a duration of earthquakes with magnitudes of 3.

Before the analysis of simulated envelopes, we calculated the Fourier amplitude spectrums for both observed and simulated DONET (M.KMB07) seismograms for the 50 s time window that began at the origin time (Figure 3). The multi-radix fast Fourier transformation algorithm (Takahashi, 2013) was employed. Amplification due to the accretionary prism commonly appeared in low-frequency (0.1–0.2 Hz) components for a shallow LFT. Noise levels of observed seismograms around frequencies of 0.1–0.5 Hz were high due to secondary microseism (e.g., Nishida, 2017). The spectrum of a shallow LFT (red line) showed a weaker high frequency (> 2 Hz) component compared with regular earthquakes. However, spectrums of simulated seismograms showed similar levels for the high frequency (> 2 Hz) range. According to the comparison of Fourier spectrums, we note that a 0.1 s triangle function is unrealistic, especially for a shallow LFT case. However, simulation results of such an unrealistic and short duration STF provide opportunities for investigating the effects of structural heterogeneities on high-frequency seismic wave propagation in offshore regions. In other words, the effects of structural heterogeneities are important for source parameter estimation.

3. Simulation results

Figure 4 shows snapshots of the simulated *P*- and *S*-wave propagations of a shallow LFT. The *P*- and *S*-wavefields were evaluated by calculating the divergence and rotation, respectively, of the simulated velocity wavefield of a shallow LFT source. The *P* and *S* waves, which radiated from a shallow LFT on the plate boundary, showed complex propagation through heterogeneous structures and repeating scattering and reflection from each layer boundary. The intensities of *P* and *S* waves at depths below the bedrock were very weak, but large seismic energies were trapped within the low-velocity accretionary prism. The *P* waves were also trapped within the seawater layer as ocean acoustic waves. According to these simulated *P* and *S* wavefields, long-duration complicated simulated seismograms can be expected.

The root-mean-square (RMS) envelopes for the shallow LFT simulation are shown in Figure 5a. After synthesizing three-component MS envelopes for the frequencies of 1–5 Hz, the RMS envelopes were evaluated by taking the mean square for the sum of the three-component MS envelopes. The RMS envelope was smoothed by taking the 1 s moving average. Because the FDM can evaluate the seismic wave propagation for frequencies less than 5 Hz due to our simulation settings, we investigated simulated seismograms for the 1–5 Hz frequency band, which is lower than the typical dominant frequencies of LFTs (2–8 Hz). After calculating the vector sum of three-component envelopes, a moving average with a time window of 1 s was applied. Each trace was normalized by each maximum amplitude. The envelopes at stations around the epicentre (< 10 km) clearly showed a short-duration *S*-wave pulse and weak later phases. However, as distance increased, RMS envelopes broadened rapidly. At distances greater than 30 km, the onsets of *S* waves were not clear and RMS envelopes were characterized by spindle shapes with durations of 20–30 s. These *S*-wave durations were significantly longer than the assumed source pulse (0.2 s). This broadening and delayed peak of simulated seismogram envelopes was much larger than for seismic waves propagating through the typical lithosphere (e.g., Saito et al., 2005; Takahashi et al., 2007; Tripathi et al., 2010).

On the other hand, the simulated RMS envelopes for regular earthquakes were composed of pulse-like *S* waves and multiple distinct later packets (Figures 5b and 5c). Multiple later packets may be developed by the seawater layer, accretionary prism, or subducting oceanic plate. The simulation of the deepest source (event C; Figure 5c) showed simpler RMS envelopes. Simulations of all events were conducted by using the same source time function (0.2 s triangle function), but the simulated envelopes for event A were very different from those for events B and C. Significant differences were found in simulated RMS envelopes between events A and B, despite similar source mechanisms, both of which were low-angle thrust faulting on the plate boundary. Thus, we considered that differences in simulated RMS envelopes between a shallow LFT and regular earthquakes could be caused by shallower heterogeneous structures.

Figure 6 shows observed RMS envelopes of a shallow LFT on 16 April 2016, an interplate earthquake on 19 April 2016, and an intraplate earthquake on 4 December 2014. Because we did not know the precise source parameters of these small events and incorporated the effects of site amplification at each station, we compared the shapes of seismogram envelopes between simulations (Figure 5) and observations (Figure 6). Similar characteristics of simulated RMS envelopes were found in observations. Although the durations of observed RMS envelopes for a shallow LFT were longer than the simulation with a 0.2 s triangle function (Figure 5 a), the observed RMS envelopes also broadened with increasing distance (Figure 6a). Pulse-like wave packets of *S* waves and later phases appeared in observed RMS envelopes for regular earthquakes (Figures 6b and 6c). To quantify the similarity between observed and simulated envelopes, we calculated cross-correlation coefficients (CCs) between both envelopes. In calculations of CCs, we allowed time shifts within ±3 s because we did not know the precise source parameters for observed seismograms. Figure 7 shows spatial variations of cross-correlation coefficients for Events A–C. Due to sensor conditions, we discarded DONET E-node (northwest four stations) in the comparison for Event C. We confirmed high (> 0.7) correlation coefficients except for Event A. The cause of CCs for Event A can be explained by an assumed STF. The comparison of simulation with a complex STF is discussed in the later section. The CCs of RMS envelopes between simulations and observations suggest that our local 3D model reliably characterizes seismic wave propagation southeast off the Kii Peninsula, Japan.

Figure 8 shows the spatial variations of maximum RMS amplitudes and envelope half-value widths derived from ground motion simulation for a shallow LFT. The envelope half-value width is the period for which the amplitude is greater than half of the maximum amplitude (Figure 8a) and has been used to represent the event duration of LFTs (e.g., Ide, 2010; Yabe et al., 2019). We used simulated RMS envelopes at both DONET and virtual seismic stations for detailed evaluations of spatial variations in amplitude and duration. The simulated amplitude distribution did not show a simple two-lobe pattern, which was expected from the assumed source mechanism. Generally, the high-frequency maximum amplitude distribution is distorted from the expected source radiation pattern by seismic wave scattering and diffraction due to small-scale subsurface heterogeneities (e.g., Imperatori & Mai 2013; Morioka et al. 2017; Takemura et al. 2009, 2016). Because we did not introduce a small-scale velocity heterogeneity into the reference model in this simulation, we confirmed that the shallower heterogeneities, such as seawater and the thick low-velocity accretionary prism, also distorted the maximum amplitude distribution. Thus, the assumption of isotropic radiation for energy estimations of small offshore seismic events in many studies could be suitable (e.g., Tamaribuchi et al., 2019; Yabe et al., 2019). Because our simulations did not include site amplifications caused by structures with V_S < 0.5 km/s, the observed maximum amplitude distribution could be more complicated because of site amplification factors at DONET stations for high-frequency seismic waves (e.g., Kubo et al., 2018; Yabe et al., 2019).

Figures 8c and 8d show spatial variations and distance-change properties of half-duration widths from the simulation results for a shallow LFT source. We also plotted the theoretical values of t_q via the method of Sato & Emoto (2018) as a typical envelope broadening for onshore earthquakes. Parameter t_q is defined as the time between *S*-wave onset and the time when the RMS envelope decays to half the maximum amplitude. This t_q value only reflect the effect of envelope broadening due to small-scale velocity heterogeneity. Typical t_q values in onshore regions gradually increase from 2 to 10 s at distances of 20–200 km (e.g., Tripathi et al., 2010). Although the half-value widths exhibited heterogeneous distribution and were widely scattered, these values increased with increasing distance and were of much longer duration than values of t_q (red lines). These longer and distance-dependent *S*-wave durations could be caused by shallower heterogeneous structures, such as seawater and the accretionary prism. Thus, to reliably estimate source duration, the effects of elongation due to shallower structures should be incorporated.

On the other hand, the half-value widths for a simulated interplate earthquake did not show clear distance-dependent properties (Figure 9). Some stations showed distinct reflection phases from the sea surface or bedrock (Figures 5b, 6b, and 9b). Similar features appeared in cases of an intraslab earthquake (Figure S5). Weak distance-dependent features were confirmed in half-value widths for later portions. Step-like elongation of half-value widths could be explained by multiple reflections from sea surface and bedrock boundary. Source durations of interplate earthquakes could be overestimated if such multiple reflections are counted as the durations of *S* waves. Conversely, by excluding such reflection phases and site amplifications from the analysis, the source parameters of offshore regular earthquakes, which occurred at sufficient depths below the accretionary prism, could be robustly estimated.

4. Effects of heterogeneous structures on source parameter estimation for small seismic events

4.1. Effects of heterogeneous structures on offshore seismograms

We demonstrated that high-frequency seismic waves show complicated propagation in offshore regions due to the influence of 3D heterogeneous structures. Our model in previous 3D simulations contains seawater, low-velocity accretionary prism, crust, and the subducting Philippine Sea plate. According to the snapshots of the simulated wavefield and differences in simulated envelopes due to source locations, shallower heterogeneities could strongly affect high-frequency seismic wave propagation. Shallower heterogeneities affect the maximum amplitude distribution and envelope broadening of high-frequency seismic waves. At frequencies greater than 1 Hz, characteristics of seismic wave propagation were usually affected not only by deterministic layered structures but also small-scale velocity heterogeneities along the propagation path (summarized in Ch. 2 of Sato et al., 2012). In this sub-section, we discuss the effects of a small-scale velocity heterogeneity, seawater, and low-velocity accretionary prism on high-frequency seismic wave propagation through offshore regions.

Small-scale velocity heterogeneities were modelled by using stochastic random velocity fluctuations characterized by an exponential autocorrelation function (ACF). The parameters of small-scale velocity heterogeneity (correlation length and RMS value) within the slab mantle and others were derived from Furumura & Kennett (2005) and Takemura et al. (2017), respectively (see Table 3). We assumed the small-scale velocity heterogeneity within the accretionary prism as the same as that within the crust. Similar small-scale heterogeneity models were previously used for waveform modelling in the Kanto sedimentary basin (Takemura & Yoshimoto 2014).

We employed the following four heterogeneous models to evaluate the effects of each heterogeneity on offshore seismograms: model with a small-scale velocity heterogeneity; model without a seawater layer; model without an accretionary prism; and model without a seawater layer and accretionary prism. The model with a small-scale velocity heterogeneity was constructed by superposing small-scale velocity heterogeneities (in Table 3) on the reference model. In the model without the seawater layer, the physical parameters within the seawater layer were replaced with those within the air column, and the seafloor was treated as the free surface. In the model without the low-velocity accretionary prism, physical parameters were replaced with those of the upper crust (Table 1). All models included bathymetry, but we do not discuss its effects because previous studies have reported limited effects of topographic scattering on body wave propagation (e.g., Imperatori & Mai, 2015; Takemura et al., 2015).

Figure 8 shows the simulated RMS envelopes at DONET stations using the four different heterogeneous models. To visualize the effects of heterogeneities, RMS envelopes of the reference model were also plotted by grey lines. Although maximum amplitudes and later phases were slightly changed by introducing small-scale velocity heterogeneities, these effects were not significant (Figure 10a). This tendency remained unchanged when using stronger models or those with different random seeds (Figure S6). The durations of RMS envelopes also showed slight differences between models with/without small-scale velocity heterogeneities (Figure S7). In typical crustal earthquakes, the effects of small-scale velocity heterogeneity appear as amplitude fluctuations of high-frequency seismic waves even for stations at shorter distances (e.g., Yoshimoto et al., 2015). Numerical simulations by Iwaki et al. (2018) showed that crustal small-scale heterogeneity has limited effects on ground motions for frequencies lower than 1 Hz in the Kanto sedimentary basin. Our large-scale simulations demonstrate that the effects of the sedimentary layers (accretionary prism) on high-frequency seismic waves are stronger than those of small-scale velocity heterogeneities.

In the model without seawater, the seafloor was treated as the free surface, and consequently, RMS amplitudes were amplified (Figure 10b). The low-velocity accretionary prism has dominant effects on high-frequency seismic waves in this region. The model without the accretionary prism exhibited simple pulse-like RMS envelopes, which were very similar to the typical envelopes of onshore small earthquakes (Figure 10c). Pulse-like *S* waves and multiple small pulses after *S* waves were found in simulated RMS envelopes. The differences between the RMS envelopes of the reference model and the model without the accretionary prism indicate that elongations of *S* waves at offshore stations were mostly caused by the low-velocity accretionary prism.

Multiple later packets after *S* waves were weak but also present in the RMS envelopes of the model without both seawater and an accretionary prism (Figure 10d). These phases could be interpreted as reflections from the subducting Philippine Sea slab. It was difficult to find the phases reflected from the subducting Philippine Sea slab in the RMS envelopes of the model that included the accretionary prism, as the elongation and amplification effects of the accretionary prism masked reflections from the Philippine Sea plate. Tonegawa et al. (2015) also reported that reflections from the bedrock (bottom of oceanic sediments) are also dominant in ambient noise wavefields, and it can be difficult to identify reflections from the boundaries at depths below the bedrock.

According to the above results, we conclude that the low-velocity accretionary prism is the dominant influence on high-frequency seismic wave propagation through the offshore region, whereas other heterogeneities, such as seawater and small-scale velocity heterogeneities, have minor effects. Similar tendencies were also found in simulations of interplate and intraslab earthquakes (Figures S8 and S9). The low-velocity accretionary prism is also important for low-frequency surface wave propagation (e.g., Takemura, Kubo, et al. 2019; Volk et al. 2017). Strong amplification or waveguide effects for both low- and high-frequency seismic waves due to the accretionary prism were also reported in the Cascadia, Mexico, and Hikurangi subduction zones (e.g., Gomberg, 2018; Kaneko et al., 2019; Shapiro et al., 2000). To achieve reliable modelling of broadband seismic wave propagation through offshore regions, and precise source-parameter estimations for offshore seismic events, a detailed model of the low-velocity accretionary prism should be considered.

4.2. Source parameter estimation for a shallow LFT via the conventional method

For small events, especially LFTs, the durations, source energies, or band-limited moments have been estimated by using stacked coherent RMS envelopes or average values for the used stations after correction of site amplification and attenuations (e.g., Annoura et al., 2016; Ghosh et al., 2009; Kao et al., 2010; Maeda & Obara, 2009; Yabe et al., 2019; Yabe & Ide, 2014). However, our simulations demonstrated that elongation of RMS envelopes for a shallow LFT, due to the low-velocity accretionary prism, occurs even for stations as close as 10 km from the epicentre (Figures 5a and 8). Thus, incorrect estimation of source parameters via the conventional method can be expected. In this sub-section, we conducted an FDM simulation of seismic wave propagation and compared it with observed RMS envelopes to investigate the validity and limitations of STFs estimation by the conventional method.

We used observed DONET seismograms for a shallow LFT that occurred at 11:18 on 24 October 2015 (JST), located at 136.91°E, 33.35°N. For deep slow earthquakes, the proportionality between seismic energy rate functions estimated from high-frequency seismograms and seismic moment rate functions estimated from low-frequency seismograms has been confirmed (Ide et al., 2008). Yabe et al. (2019) showed that the same proportionality holds true for shallow slow earthquakes. Therefore, seismic energy rate functions estimated from high-frequency seismograms can be converted into STFs by dividing them by the value of scaled energy. The seismic energy rate function of the event observed at each station is conventionally calculated from squared seismograms at each station after correction of site and attenuation effects within time windows of half-value width measured for the stacked RMS envelope. Seismograms were excluded from the analysis if the cross-correlation coefficient with the same event envelopes at any other stations did not exceed 0.6. This procedure removed stations lacking tremor signals. Modelling for direct-*S* wave energies using the uniform background velocity structure (V_S = 3.5 km/s and ρ = 2700 kg/m^3) and distance-dependent attenuation model (Figure 3 of Yabe et al., 2019) was employed. The seismic energy rate functions were converted into STFs by assuming a scaled energy value of 3.0×10^{-9} (Ide & Maury, 2018; Ide & Yabe, 2014; Yabe et al., 2019). The STF of the event was calculated by stacking STFs at each station after applying a 10 s low-pass filter.

Figure 11 shows the STFs of a shallow LFT that occurred on 24 October 2015. The conventional method typically provided STFs represented by blue dashed or red lines, which were estimated by using stacking seismic energy rate functions of B-node stations and B- and C-node stations, respectively. The moment releases of these STFs continued 50–60 s from the origin and peaks of moment rate appeared at around 25 s. The STF described by the solid-blue line was constructed by using the seismic energy rate function at M.KMB06 only, which was located just above the shallow LFT hypocentre. This STF was completely different from the others. The peak of this STF appeared within 3.2 s of the origin, and the moment release continued 30–40 s from the origin. By using these STFs, we synthesized seismograms of the shallow LFT. Hypocentre depth was fixed on the plate boundary, and the focal mechanism was the same as in previous simulations of event A (Table 2). After FDM simulations, we convolved the estimated STFs to simulated envelopes. To achieve suitable STF convolutions, we assumed a simple 0.1 s triangle function, which was sufficiently shorter than the STFs in Figure 11. After FDM simulations, we convolved the estimated STFs to simulated envelopes.

Figure 12 compares 1–5 Hz RMS envelopes between simulations and observations. In this comparison, the amplitudes of each trace were normalized by each maximum amplitude because the precise seismic moment and site amplification factors from the V_S = 0.5 km/s layer were not well known. The analysis focused on differences in envelope shapes between simulations and observations. To quantify fitness of envelope shapes, we also calculated CCs between observed and simulate envelopes (blue numbers in each trace). The simulation results with the STF from M.KMB06 reproduced the observed envelope shapes and durations at all stations. Almost all of the simulated envelopes had CCs larger than 0.7. This STF also reproduced the seismograms observed for a shallow VLFE (Yabe et al., 2019), which occurred in the same time window as the target shallow LFT but appeared for frequencies of 0.03–0.05 Hz.

The simulations with the other STFs were much longer than the observations. Two of the simulated STFs also showed delayed peak amplitudes compared to the observations. The epicentral distances of B-node stations except for M.KMB06 ranged from 10 to 16 km. The simulation with a simple STF (Figures 5a and 8) demonstrated that the durations of *S* waves increased rapidly with increasing distance, and values reached to 5–15 s at stations located at distances of 10–15 km. Thus, STFs from stacked seismic energy rate functions via the conventional method were delayed and overestimated. The simulation with a simple STF of duration 0.2 s (see Figure 5a) shows that the RMS envelopes at very close (< 10 km) stations are characterized by a pulse-like *S* wave packet, and consequently, seismic energy rate functions at close stations could preserve source information. Thus, the shapes and durations of observed RMS envelopes were only well reproduced in the STF estimated from the nearest station (~1.4 km, M.KMB06).

Our numerical tests in this sub-section revealed that STFs from stations very close to the source reproduced the observed seismogram envelops of the target shallow LFT. In other words, the effects of offshore heterogeneities were limited at stations with epicentral distances less than 10 km, even for shallower (~5 km) seismic events (see Figure 5), and, consequently, the source parameter information could be preserved. At stations with epicentral distances greater than approximately 10 km, RMS envelopes were strongly elongated due to the effects of the accretionary prism (Figure 5a). At stations with epicentral distances greater than approximately 10 km, durations of the RMS envelope were controlled by the heterogeneous structure of the accretionary prism along the propagation path. Thus, STFs calculated by using such stations via the conventional method (e.g., Yabe et al., 2019) were overestimated and could not reproduce the observed RMS envelopes. When the conventional method is applied to OBS data, seismograms should be selected from stations located closer than approximately 10 km. If stations further than 10 km from the source are used, the effects of 3D offshore heterogeneities should be included in the method of source parameter estimation.

5. Conclusions

We investigated the effects of offshore heterogeneities, such as a small-scale velocity heterogeneity, seawater, and the accretionary prism, on high-frequency seismic propagation southeast off the Kii Peninsula, southwest Japan. Our simulations demonstrated that the low-velocity accretionary prism affects the shapes of seismogram envelopes. A thick low-velocity accretionary prism is also developed in the Cascadia, Mexico, and Hikurangi subduction zones, and this affects not only low-frequency surface waves, but also high-frequency seismic waves. The effects of the accretionary prism are significant for shallow-depth events along the plate boundary, which correspond to shallow LFTs along the Nankai Trough, as these sources are typically located just beneath the accretionary prism toe at depths of 5–8 km near the trench axis. On the other hand, seismogram envelopes for sufficiently deep-depth events, which correspond to interplate and intraslab regular earthquakes along the Nankai Trough, comprise multiple pulse-like wave packets that consist of direct-*S* waves and multiple reflections from the bedrock.

The seismogram envelopes of a source at shallow LFT depths (5–7 km) were broadened due to the low-velocity accretionary prism, even when assuming a simple STF. The durations of RMS envelopes exceeded 10 s, even at stations located at distances greater than 10 km. Multiple reflected phases found in seismograms of regular earthquakes also can affect half-value widths. Such elongation of envelopes will cause incorrect estimations of event durations. Event durations are important for discussing the scaling law of regular and slow earthquakes. After removing such reflection phases and site amplification from the analysis, the source parameters of offshore regular earthquakes, which occurred at sufficient depths below the accretionary prism, could be reliably estimated.

At stations very close (< 10 km) to the source, pulse-like *S* wave envelopes, reflecting an assumed STF, were preserved for shallow (5–30 km) small seismic events. If stations at distances greater than 10 km are used, the STF could be overestimated in terms of the duration and peak time of the moment rate. By estimating the STF by using only the nearest stations, the FDM simulations reproduced the shapes and durations of observed RMS envelopes. Thus, conventional approaches might work well by using stations at distances less than 10 km.

High-frequency seismogram envelopes are useful for both studies of structural property (e.g., Carcolé & Sato, 2010; Przybilla et al., 2009; Saito et al., 2005; Takemura et al., 2015) and source parameter estimations (e.g., Battaglia & Aki, 2003; Gusev & Pavlov, 1991; Ide et al., 2008; Yabe et al., 2019). Envelope based source inversions have been conducted by using theoretical Green's function envelopes (e.g., Nakahara, 2013; Petukhin et al., 2004; Sawazaki & Enescu, 2014).Theoretical envelopes can be easily evaluated, but 3D heterogeneous structures cannot be incorporated. Combined use of the Green's functions of numerical simulations in a 3D local model and site amplifications due to lower (V_S < 0.5 km/s) velocity sediments at each station can potentially enable reliable source parameter estimations even for small offshore events. Green's functions within the local layered model are also useful for estimations of small-scale heterogeneities within oceanic sediments, which may reflect physical properties or clack distributions of the accretionary prism. Such challenging work remains for future studies.

Acknowledgements

F-net and DONET waveform data are available via the NIED website (https://doi.org/10.17598/NIED.0005, https://doi.org/10.17598/NIED.0008). Bathymetric depth data were obtained from ETOPO1 (Amante & Eakins, 2009). OpenSWPC software (Maeda et al., 2017) and the 3D model of Koketsu et al. (2012) were obtained from https://doi.org/10.5281/zenodo.3712650 and https://www.jishin.go.jp/evaluation/seismic_hazard_map/lpshm/12_choshuki_dat/, respectively. Generic Mapping Tools (GMT; Wessel et al., 2013) and Seismic Analysis Code (SAC; Helffrich *et al.* 2013) were used to produce the figures and in signal processing, respectively. The multi-radix FFT package (ffte; Takahashi, 2013) is available at http://www.ffte.jp/. The catalogues of slow earthquakes (Nakano, Hori, *et al.* 2018, Takemura *et al.* 2019) were downloaded from the Slow Earthquake Database website (Kano et al., 2018; http://www-solid.eps.s.u-tokyo.ac.jp/~sloweq/). The CMT results of the 2016 southeast off Kii Peninsula earthquake, Japan, and an intraslab earthquake used in regular earthquake simulations are available from https://doi.org/10.5281/zenodo.3523583. We also used the unified hypocentre catalogue of the Japan Meteorological Agency (https://www.data.jma.go.jp/svd/eqev/data/bulletin/index.html). The FDM simulations of seismic wave propagation were conducted by using the Earth Simulator of JAMSTEC. This study was supported by the Japan Society for the Promotion of Science (JSPS) KAKENHI Grant Numbers 17K14382, 18K13639 in the Grant-in-aid for Young Scientists and Grant Number 19H04626 in the Scientific Research under Innovative Areas 'Science of Slow Earthquakes'. We would like to thank Editage (www.editage.com) for English language editing. We also thank three anonymous reviewers and Editor Prof. J. Virieux for carefully reviewing the study and providing constructive comments, which have helped to improve the manuscript.

Tables

Table 1. Physical parameters of each layer in the 3D velocity structure model. The parameters were obtained from the Japan Integrated Velocity Structure Model (JIVSM) (Koketsu et al., 2012). The air and seawater layers were treated as being the same, following Maeda *et al.* (2017).

	V_P [km/s]	V_S [km/s]	ρ [kg/m^3]	Q_P	Q_S
Air	0.0	0.0	0.001	10^{10}	10^{10}
Seawater	1.5	0.0	1.04	10^{6}	10^{6}
Sedimentary layer 1	1.8	0.5	1.95	170	100
Sedimentary layer 2	2.2	0.8	2.07	272	160
Sedimentary layer 3	2.4	1.0	2.15	340	200
Sedimentary layer 4	3.0	1.5	2.25	510	300
Sedimentary layer 5	3.5	2.0	2.35	680	400
Basement	5.5	3.2	2.65	680	400
Upper crust	5.8	3.4	2.70	680	400
Lower crust	6.4	3.8	2.80	680	400
Upper mantle	7.5	4.5	3.20	850	500
Philippine Sea plate					
Oceanic crust layer 2	5.0	2.9	2.40	340	200
Oceanic crust layer 3	6.8	4.0	2.90	510	300
Oceanic mantle	8.0	4.7	3.20	850	500

Table 2. Source parameters used in FDM simulations. A shallow LFT and intraslab earthquakes were referred from the nearest shallow VLFE and intraslab earthquakes of CMT solutions in our previous studies (Takemura, Okuwaki, *et al.* 2019, Takemura *et al.* 2019). Double-couple point sources were assumed. The value of T_0 represents the STF duration.

	Type	Lon. [E°]	Lat. [°N]	Depth [km]	Strike [°]	Dip [°]	Rake [°]	*Mw*	T_0 [s]
A	LFT	136.90	33.20	6.20	255.0	7.0	116.0	3	0.2
B	Interplate earthquake	136.34	33.40	11.15	243.3	10.1	114.4	3	0.2
C	Intraslab earthquake	137.11	33.28	28.7	264.3	37.9	37.9	3	0.2

Table 3. Parameters of small-scale velocity heterogeneities in each layer. The models of the crust and oceanic crust were assumed to be the same as the model of crustal heterogeneity (Takemura *et al.* 2017). The model of the oceanic mantle was derived from Furumura & Kennett (2005).

Layer	ACF-type	Correlation length *a*	RMS value ε
Air and seawater	-	-	-
Accretionary prism	Exponential	Isotropic: 1 km	0.03
Crust	Exponential	Isotropic: 1 km	0.03
Mantle	-	-	-
Oceanic crust	Exponential	Isotropic: 1 km	0.03
Oceanic mantle	Exponential	Horizontal: 10 km Vertical: 0.5 km	0.03

Figures

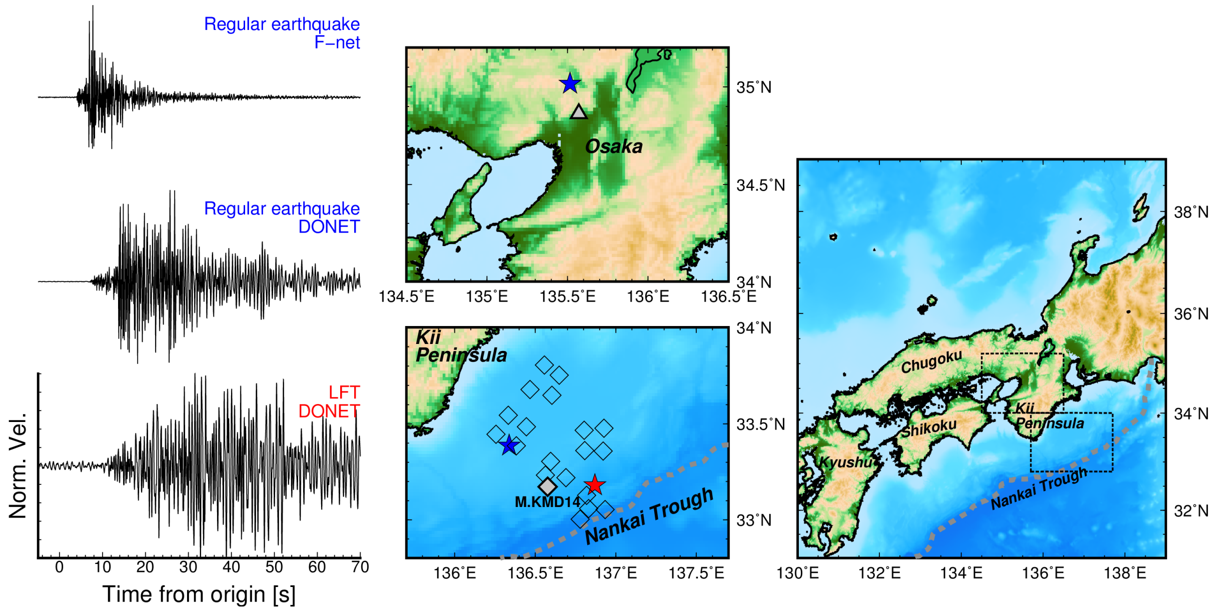

Figure 1. Examples of observed NS-component seismograms for a crustal earthquake, an offshore interplate earthquake, and a shallow LFT. Crustal and offshore interplate earthquakes occurred on 24 January 2015 and 1 April 2016, respectively. A shallow LFT occurred on 3 April 2016. The magnitudes of these events were approximately 3, and epicentral distances were also similar. The blue stars and red star in enlarged maps are the epicentres of regular earthquakes and the shallow LFT, respectively. The triangle and diamond symbols show the locations of the F-net and DONET stations, respectively.

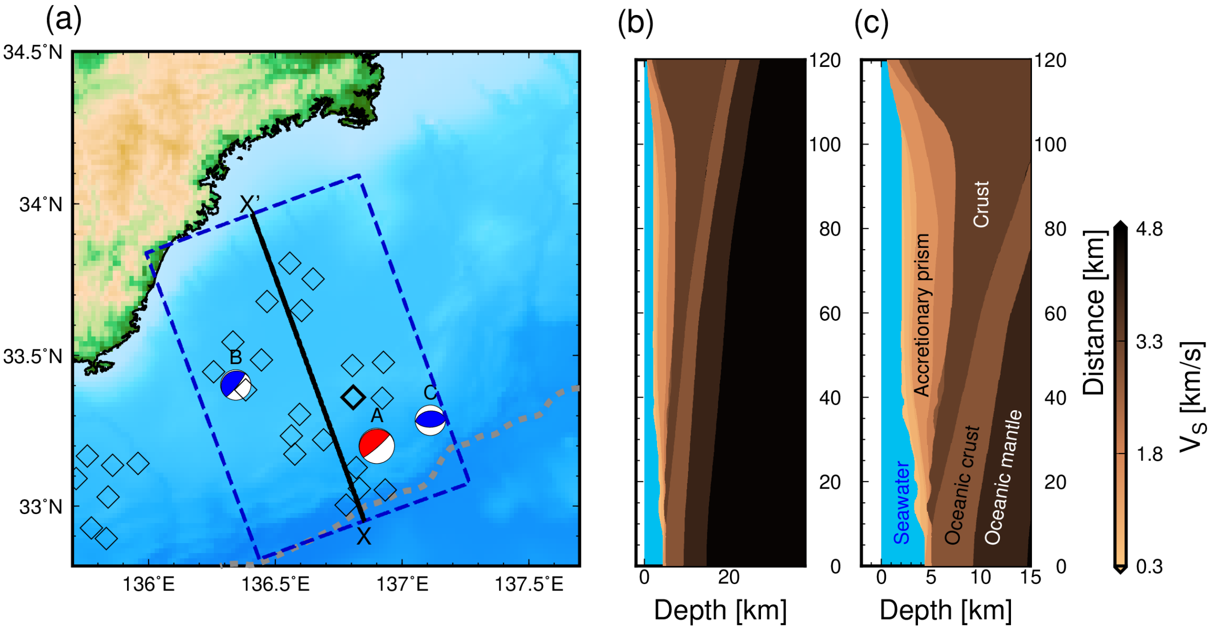

Figure 2. (a) Map of the simulation region, and cross-sections of the (b) *S*-wave velocity model along the X—X′ profile and (c) *S*-wave velocity model at shallower (< 15 km) depths. The red and blue focal mechanisms are source models of a shallow LFT and regular earthquakes, respectively. Details of physical parameters in each layer are listed in Table 1.

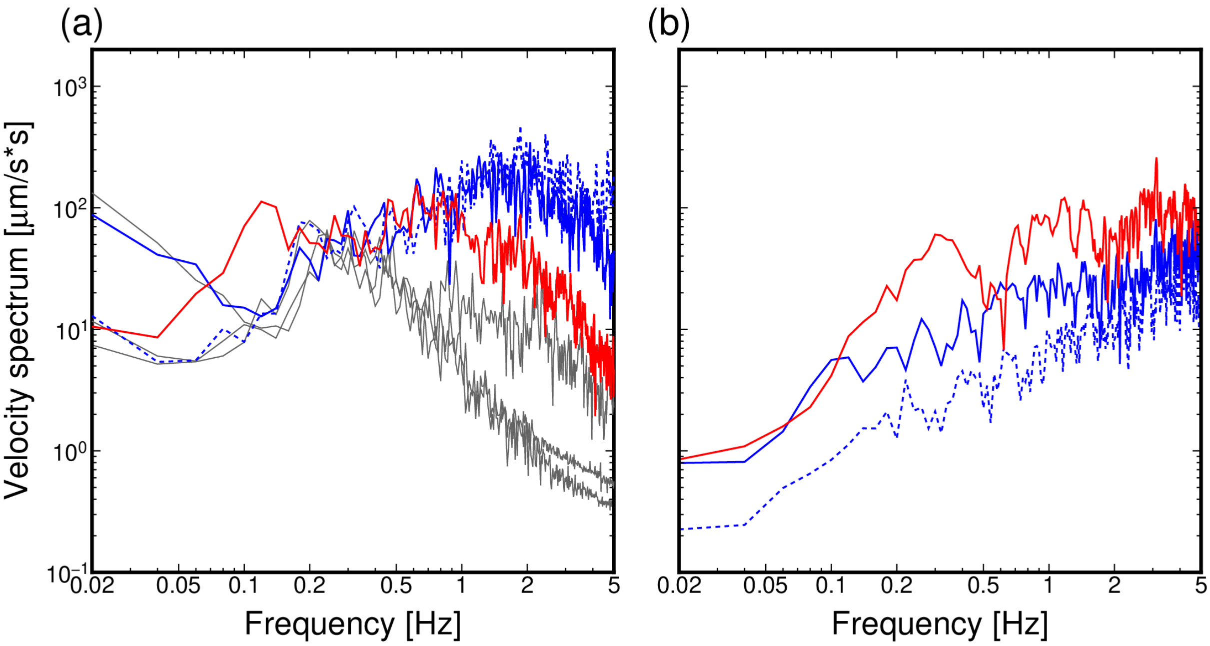

Figure 3. Fourier spectrums of (a) observed and (b) simulated seismograms at M.KMB07. The location of M.KMB07 is shown in Figure 2 (bold diamond). The red, blue solid, blue dashed, and grey lines in (a) are spectrums of a shallow LFT on 16 April 2016, an interplate earthquake on 19 April 2016, an intraplate earthquake on 4 December 2014, and noise parts, respectively. The red, blue, and blue dashed lines in (b) are spectrums of simulated seismograms for Events A–C. The time window for the event signal is from the origin time to the lapse time of 50 s.

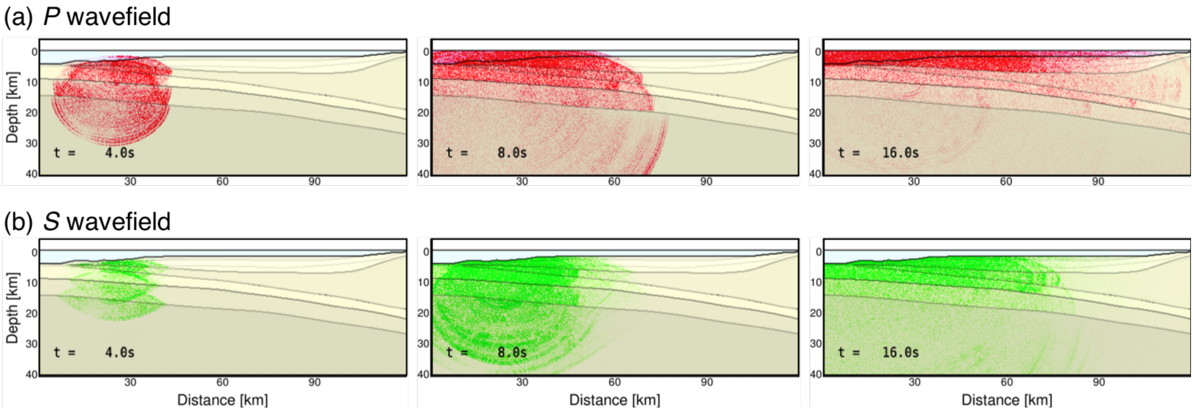

Figure 4. Simulated snapshots of (a) *P* and (b) *S* wavefields along profile X—X′ for the shallow LFT simulation (red focal sphere in Figure 2).

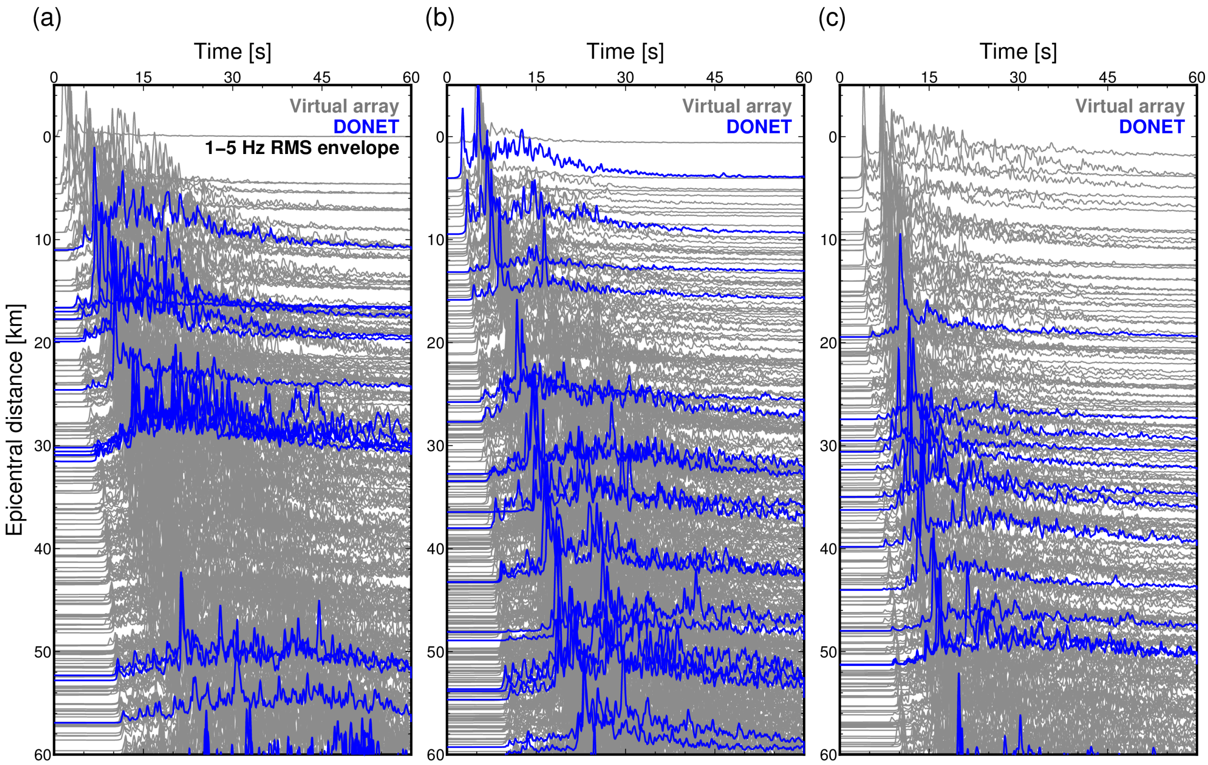

Figure 5. Simulated RMS envelopes for frequencies of 1–5 Hz from simulations of the (a) shallow LFT, (b) interplate earthquake, and (c) intraslab earthquake. Source parameters and locations are shown in Table 2 and Figure 2a. The RMS envelopes at DONET and virtual stations are represented by blue and grey lines, respectively.

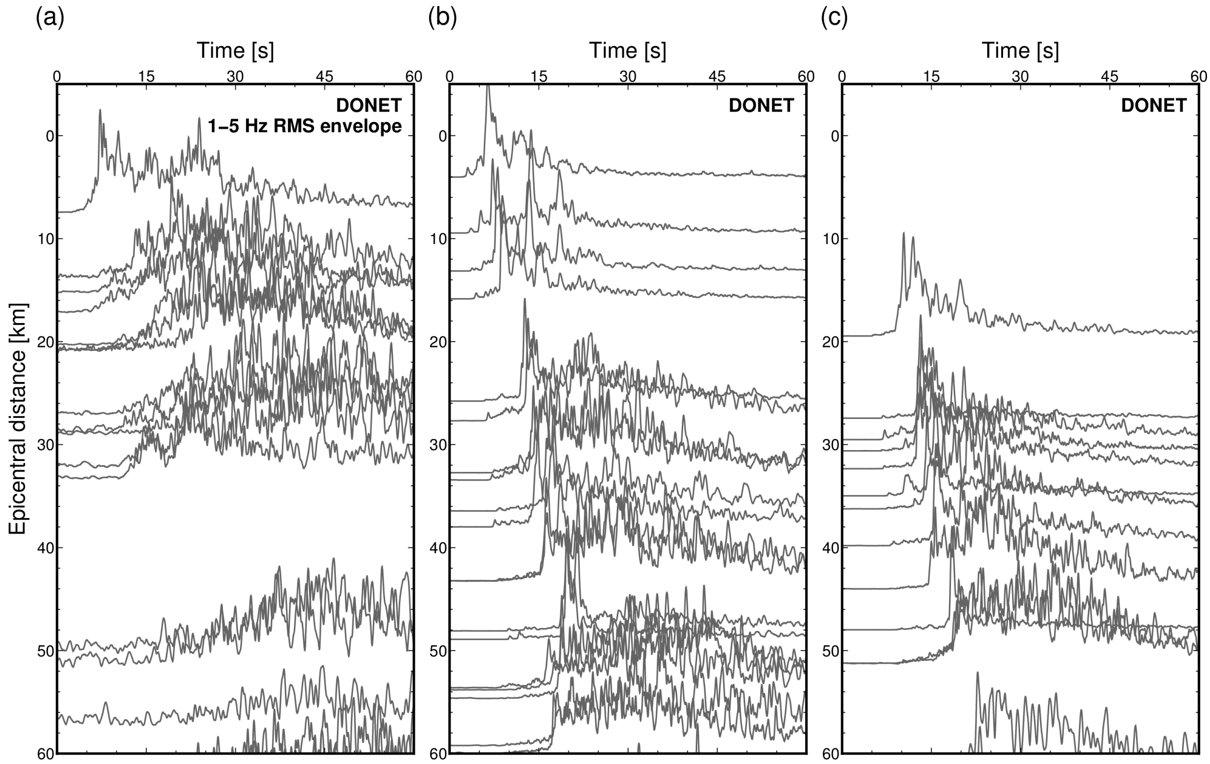

Figure 6. Observed RMS envelopes for frequencies of 1–5 Hz at DONET stations. The RMS envelopes for (a) a shallow LFT on 16 April 2016, (b) an interplate earthquake on 19 April 2016, and (c) an intraplate earthquake on 4 December 2014.

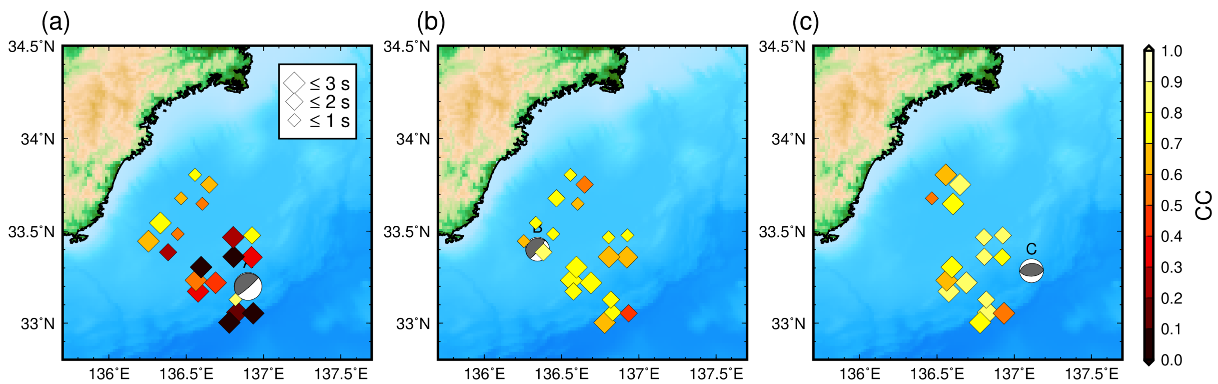

Figure 7. Spatial variations of cross-correlation coefficients between observed and synthesized envelopes for (a) Event A, (b) Event B, and (c) Event C. The sizes of symbols represent the absolute values of time shifts, which provide maximum correlation coefficients at each station.

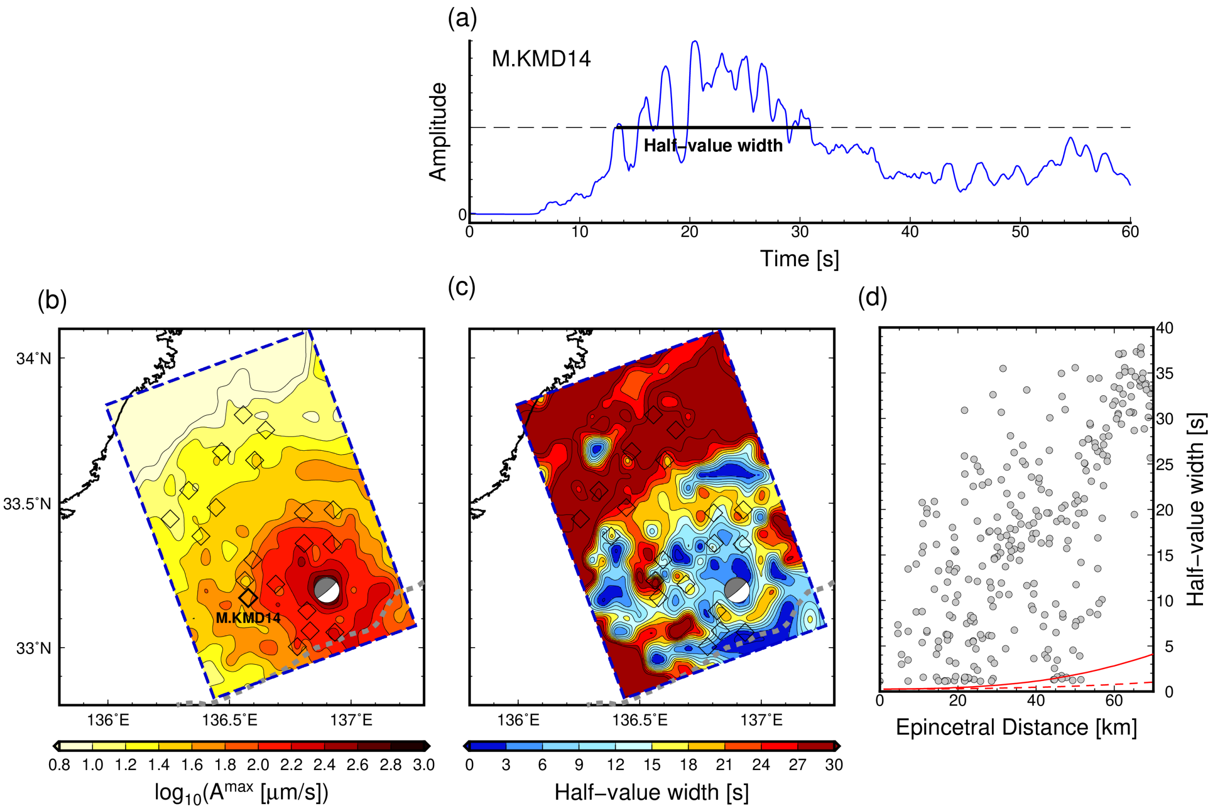

Figure 8. Spatial variations of maximum RMS amplitudes and half-value widths of simulated RMS envelopes for a shallow LFT: (a) an example of half-value width estimations, (b) maximum RMS amplitude distribution, (c) half-value width distribution, and (d) half-value widths as a function of the epicentral distance. The example in (b) is an RMS envelope at M.MRD14 (location shown in (a)). The red dashed and solid lines in (d) are values of t_q via the method of Sato & Emoto (2018). Here, t_q is defined as the time between *S*-wave onset and the time when the RMS envelope decays to half the maximum amplitude. The assumed small-scale random velocity heterogeneity models for dashed and solid lines are characterized by an exponential-type autocorrelation function with correlation length of 1 km and RMS values of 0.03 and 0.05, respectively. The central frequency in theoretical synthetics is 4 Hz.

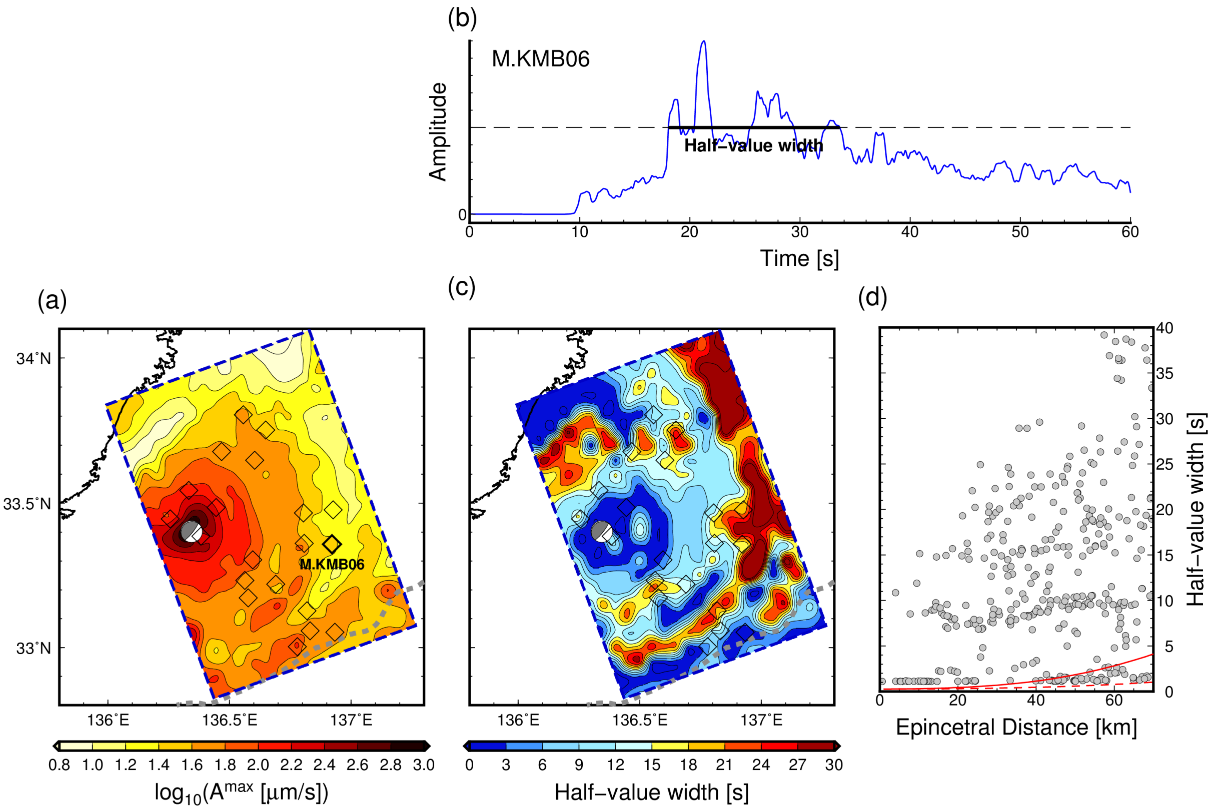

Figure 9. Spatial variations of maximum RMS amplitudes and half-value widths of simulated RMS envelopes for an interplate earthquake: (a) an example of half-value width estimations, (b) maximum RMS amplitude distribution, (c) half-value width distribution, and (d) half-value widths as a function of epicentral distance. The example in (b) is an RMS envelope at M.MRB06 (location shown in (a)). The red dashed and solid lines in (d) are values of t_q via the method of Sato & Emoto (2018). Here, t_q is defined as the time between *S*-wave onset and the time when the RMS envelope decays to half the maximum amplitude. The assumed small-scale random velocity heterogeneity models for dashed and solid lines are characterized by an exponential-type autocorrelation function with correlation length of 1 km and RMS values of 0.03 and 0.05, respectively. The central frequency in theoretical synthetics is 4 Hz.

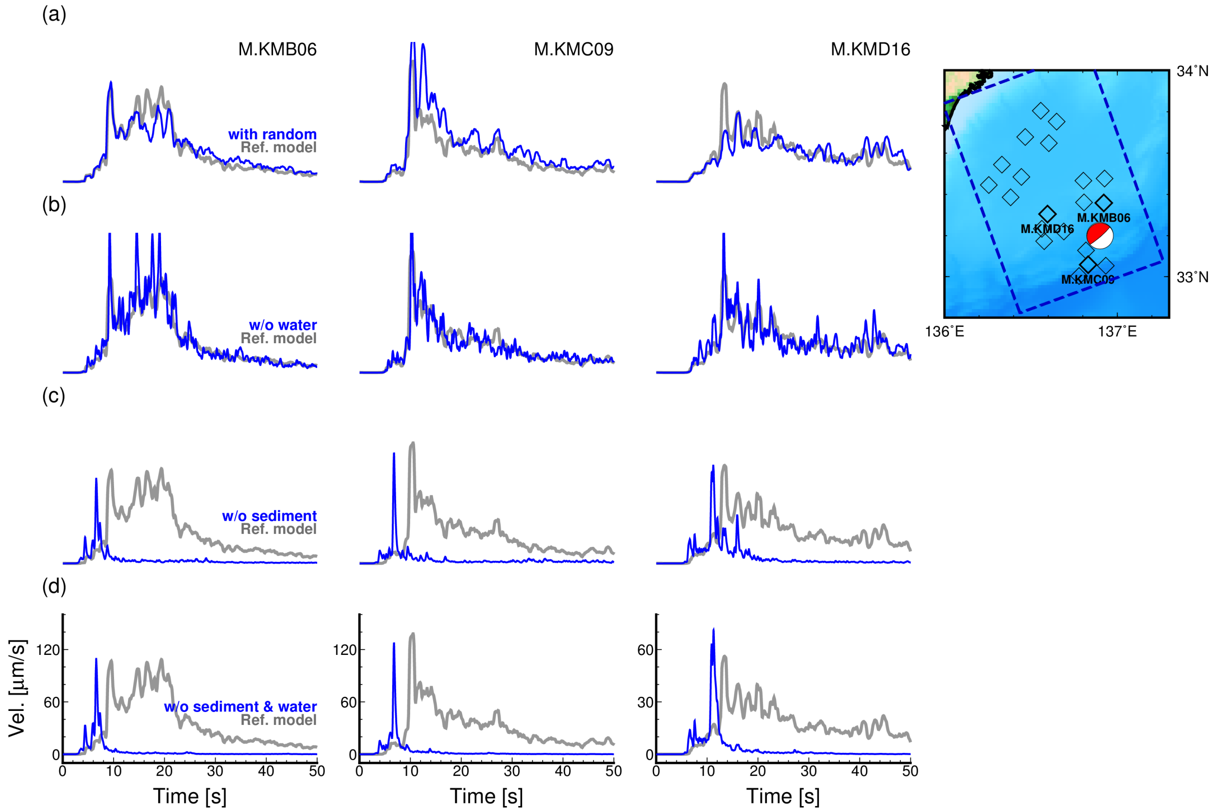

Figure 10. Simulation results for various heterogeneous models of a shallow LFT. The blue lines are RMS envelopes derived from models (a) with small-scale random velocity heterogeneities, (b) without seawater, (c) without oceanic sediments (accretionary prism), and (d) without oceanic sediments and seawater. Grey lines are the simulation results with the reference model.

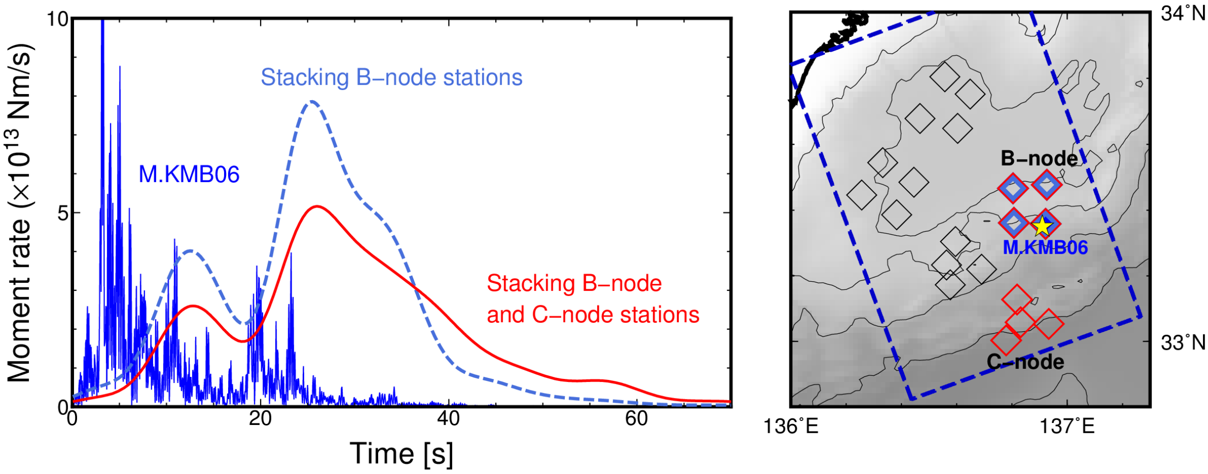

Figure 11. Source time functions (STFs) estimated from observed RMS envelopes of a shallow LFT at 11:18 on 24 October 2015 (JST). The blue dashed, blue, and red lines are source time functions estimated from the RMS envelope of M.KMB06, envelopes of B-node stations, and envelopes of B- and C-node stations, respectively. The yellow star on the map is the epicentre. The epicentral distance of M.KMB06 was 1.4 km. The stations represented by coloured symbols were used in the STF estimation for corresponding coloured lines.

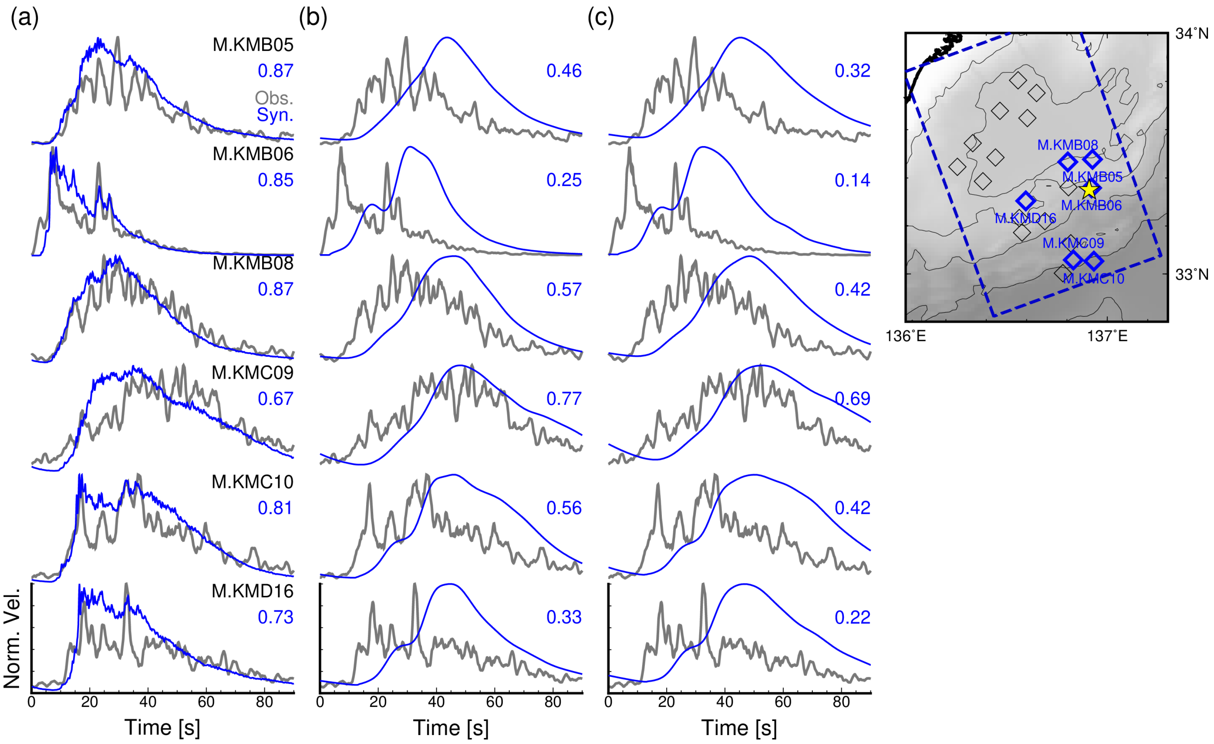

Figure 12. Comparison between simulated RMS envelopes with observations. The RMS envelopes were synthesized from simulation results of source time functions estimated by using (a) M.KMB06, (b) B-node stations, and (c) B- and C-node stations. Because the precise seismic moment for the target shallow LFT and site amplification factors were not well known, the amplitudes of each trace were normalized by each maximum amplitude. The numbers written by blue characters are cross-correlation coefficients between observed and synthesized envelopes at each station. The upper-right panel shows the stations used (blue diamonds) and the epicentre location (yellow star).

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Supporting information:

Modelling high-frequency seismograms at ocean bottom seismometers: effects of heterogeneous structures on source parameter estimation of small offshore earthquakes and shallow low-frequency tremors

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Text S1

The 3D model used in the simulations was basically constructed from the Japan Integrated Velocity Structure Model (JIVSM) (Koketsu et al., 2012; Table 1 and Figure S1b). Because the accretionary prism of the JIVSM is modelled by a V_S = 1 km/s constant layer and is too simple for realistic ground motion simulation, we introduced estimations of 1D *S*-wave velocity structures beneath DONET stations by Tonegawa *et al.* (2017). We converted the depth-varying velocity structure model of Tonegawa *et al.* (2017) to a 5-layer model beneath each DONET station. The physical parameters of each layer are listed in Table 1 of the main text. Thicknesses of each layer were determined by fitting the depth-average *S*-wave velocities derived by Tonegawa *et al.* (2017). For example, if the depth-average V_S of Tonegawa *et al.* (2017) become the V_S of layer 1 at a certain depth, this depth is considered as the bottom of layer 1. After obtaining the bottom depths of layers 1–4 beneath DONET stations, the bottom depths of each station were interpolated and extrapolated via the '*Surface*' gridding algorithm in Generic Mapping Tools software (GMT; Wessel *et al.* 2013). Interpolation and extrapolation were only applied within the region of the accretionary prism (Figure S1). In other words, the JIVSM onshore and outer-rise sedimentary structures, and structures beneath bedrock, were fixed.

This construction method was basically the same as in Takemura, Kubo, *et al.* (2019). However, in the present model, a layered structure was employed rather than a depth velocity-gradient function. The simulation results of both the smooth-gradient and layered models were almost the same (Figures S2–S4).

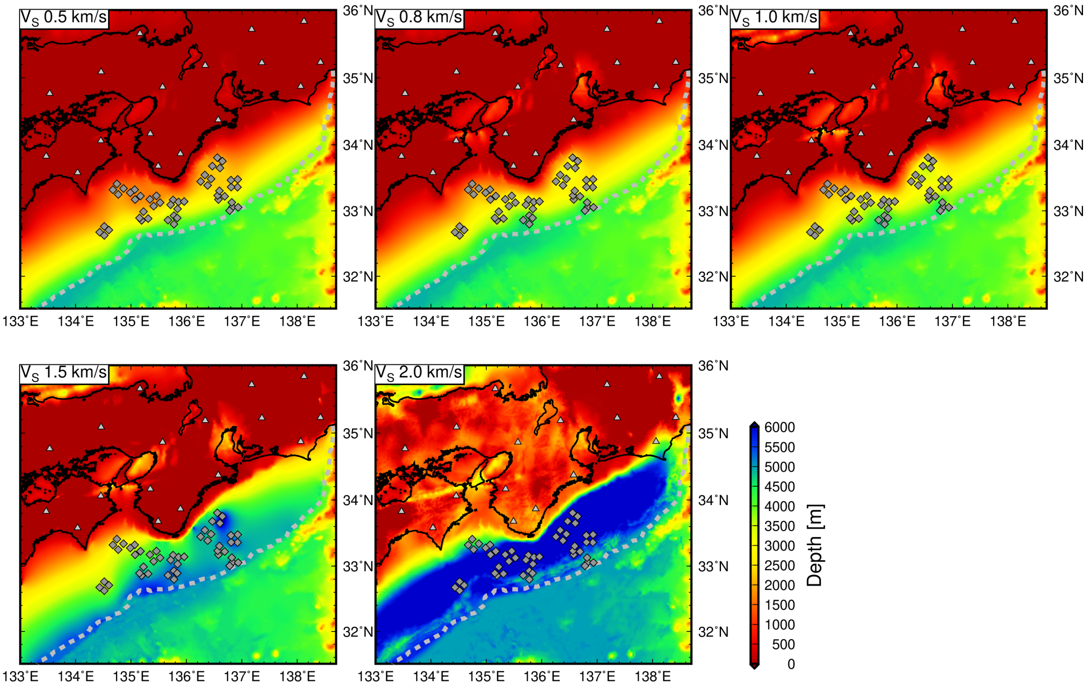

Figure S1a. The 3D layered structure of the accretionary prism. Background colours represent depths of the bottom of each layer (see Table 1). The bottom of layer with V_S of 2. 0 km/s is corresponding to the bedrock depths of the JIVSM (Koketsu et al., 2012).

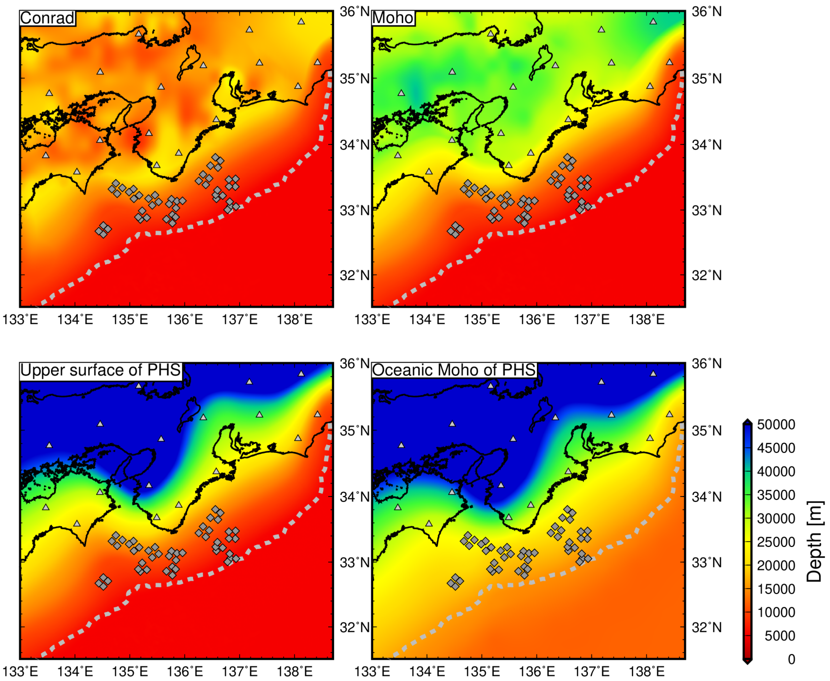

Figure S1b. Spatial variations of depths for Conrad, Moho, upper surface and oceanic Moho of the Philippine Sea plate in the assumed model, JIVSM (Koketsu et al., 2012).

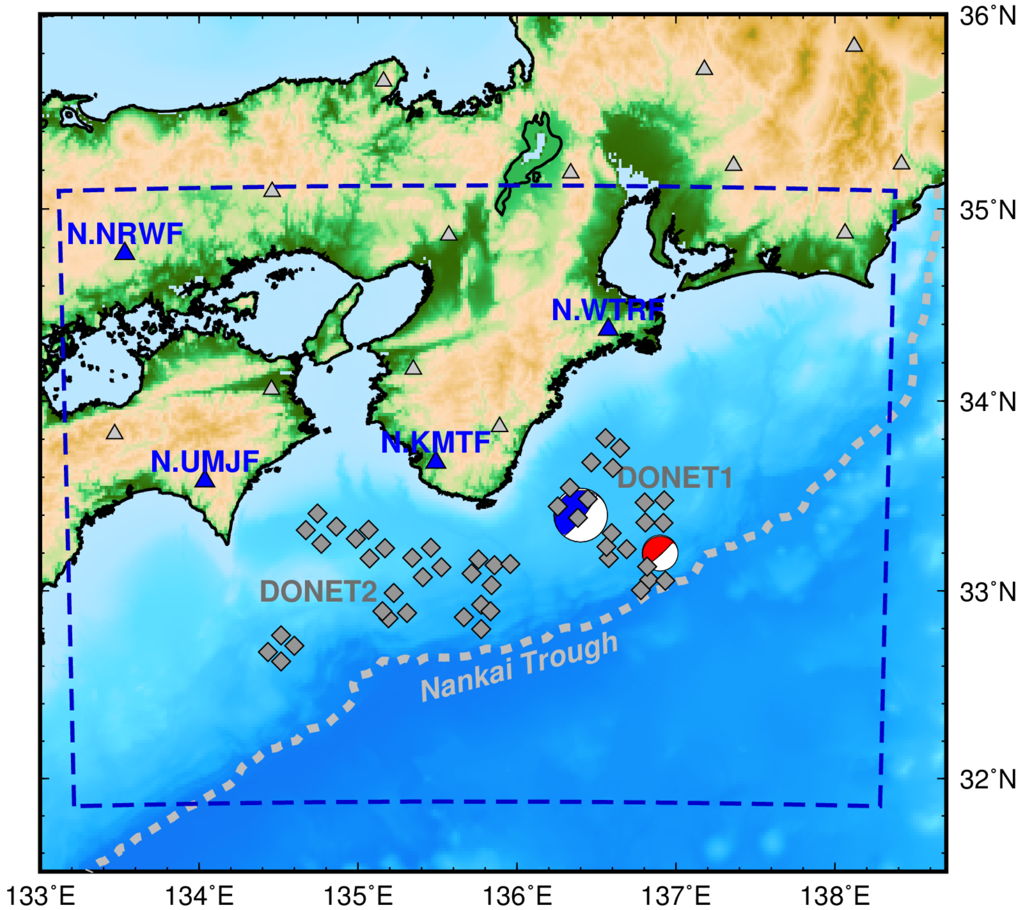

Figure S2. Map for validation of simulations. The dashed-blue rectangle represents the horizontal coverage of the calculation region, which was discretised at intervals of 0.2 km in horizontal directions and 0.1 km in the vertical direction. Simulations were conducted using the computer system at the Earthquake and Volcano Information Center, Earthquake Research Institute, University of Tokyo. The blue and red focal spheres are CMT solutions of the 2016 *Mw* 5.9 earthquake southeast off Kii Peninsula, Japan (Takemura et al., 2020), and a shallow very low-frequency earthquake on 10 April 2016 (Takemura et al., 2018, 2019), respectively. The triangles and diamonds represent F-net and DONET stations, respectively.

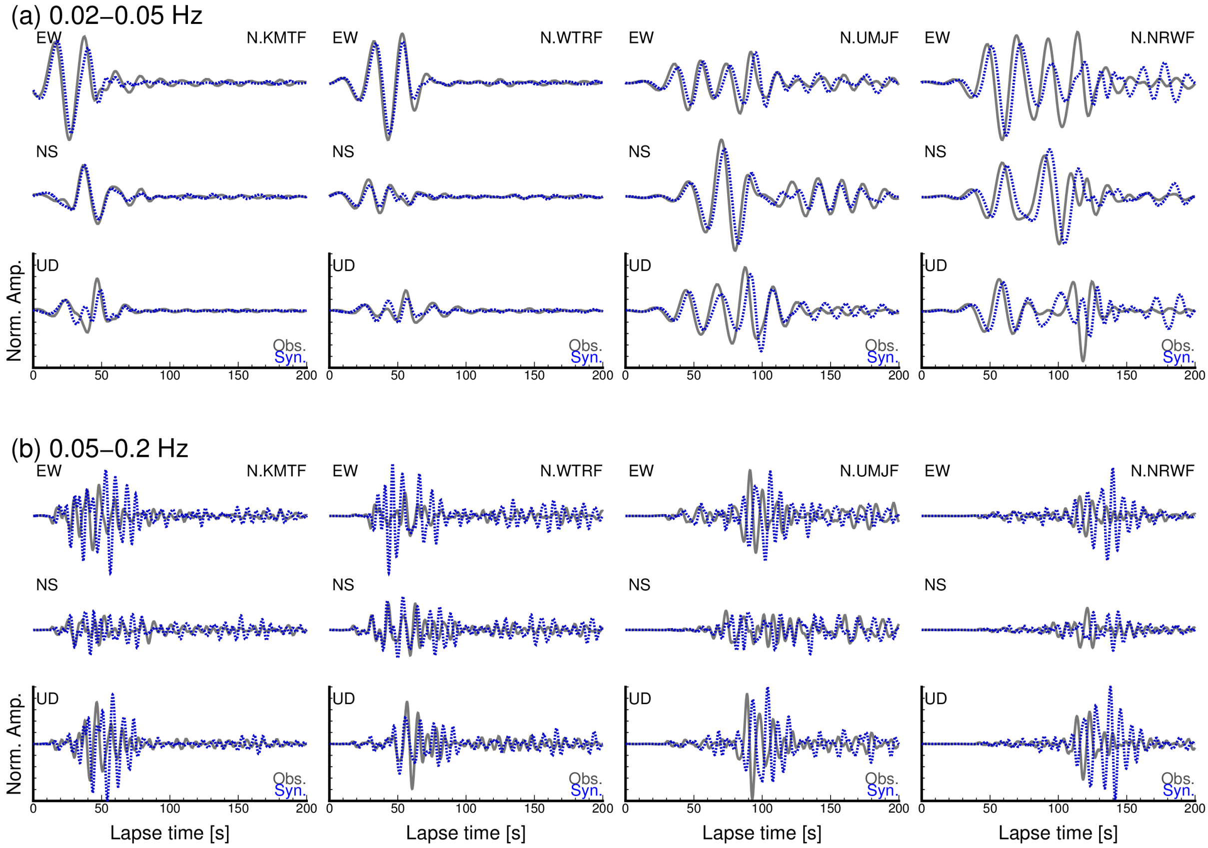

Figure S3. Simulation of the 2016 *Mw* 5.9 earthquake southeast off Kii Peninsula, Japan (Shunsuke Takemura et al., 2020): (a) 0.02–0.05 Hz, (b) 0.05–0.2 Hz. Seismic wave propagation in our model was evaluated via OpenSWPC software (Maeda et al., 2017). The simulation model covered the area within the blue rectangle shown in Figure S1, which was discretised by grid intervals of 0.2 km in horizontal directions and 0.1 km in the vertical direction. Observed and synthetic seismograms at F-net stations are represented by grey and blue dotted lines, respectively. Amplitudes at each station were normalized by the maximum amplitude of six-component filtered seismograms.

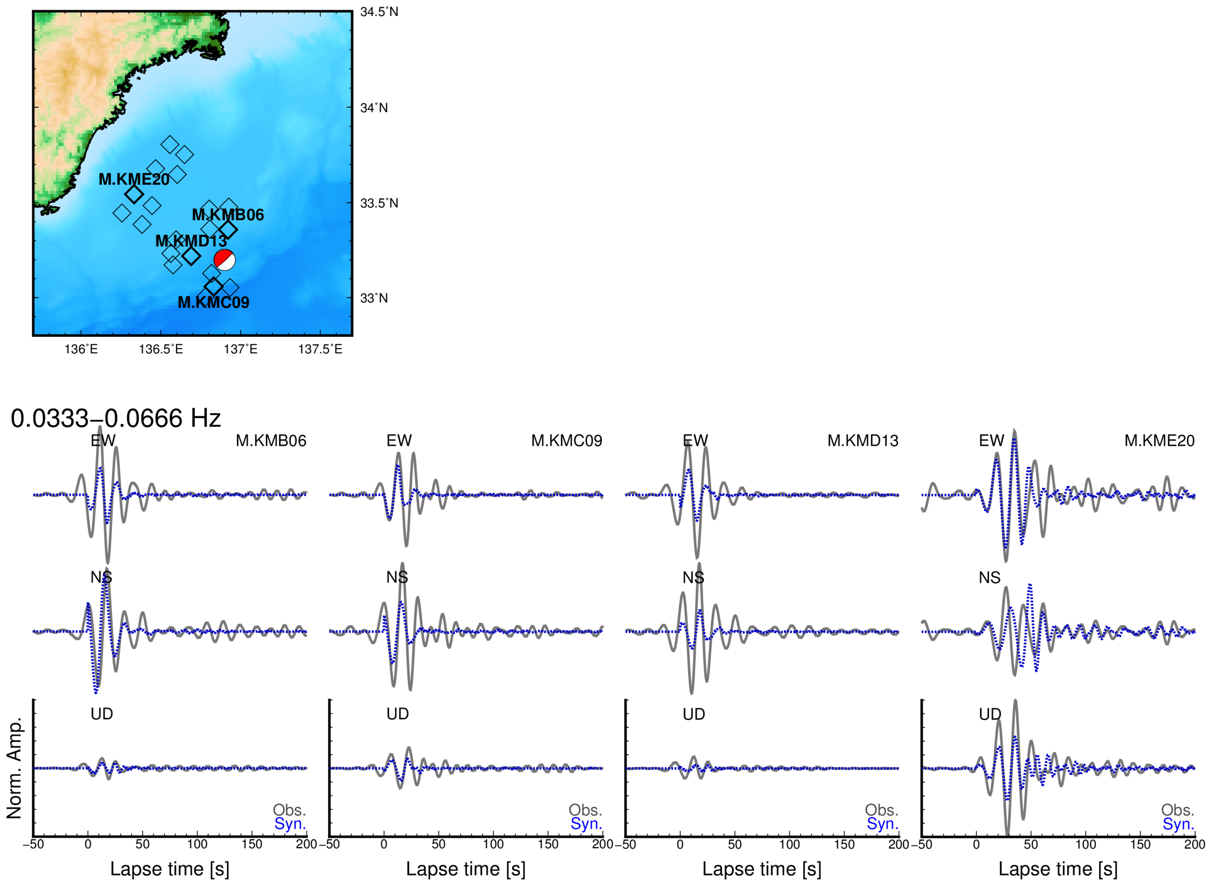

Figure S4. Simulation result of a shallow very low-frequency earthquake on 10 April 2016 (S. Takemura et al., 2018; Shunsuke Takemura et al., 2019). Observed and synthetic seismograms at DONET stations are represented by grey and blue dotted lines, respectively. Amplitudes at each station were normalized by the maximum amplitude of six-component filtered seismograms. Locations of stations and the source are shown in the upper-left panel. The simulation model and settings are similar to those in Figure S2.

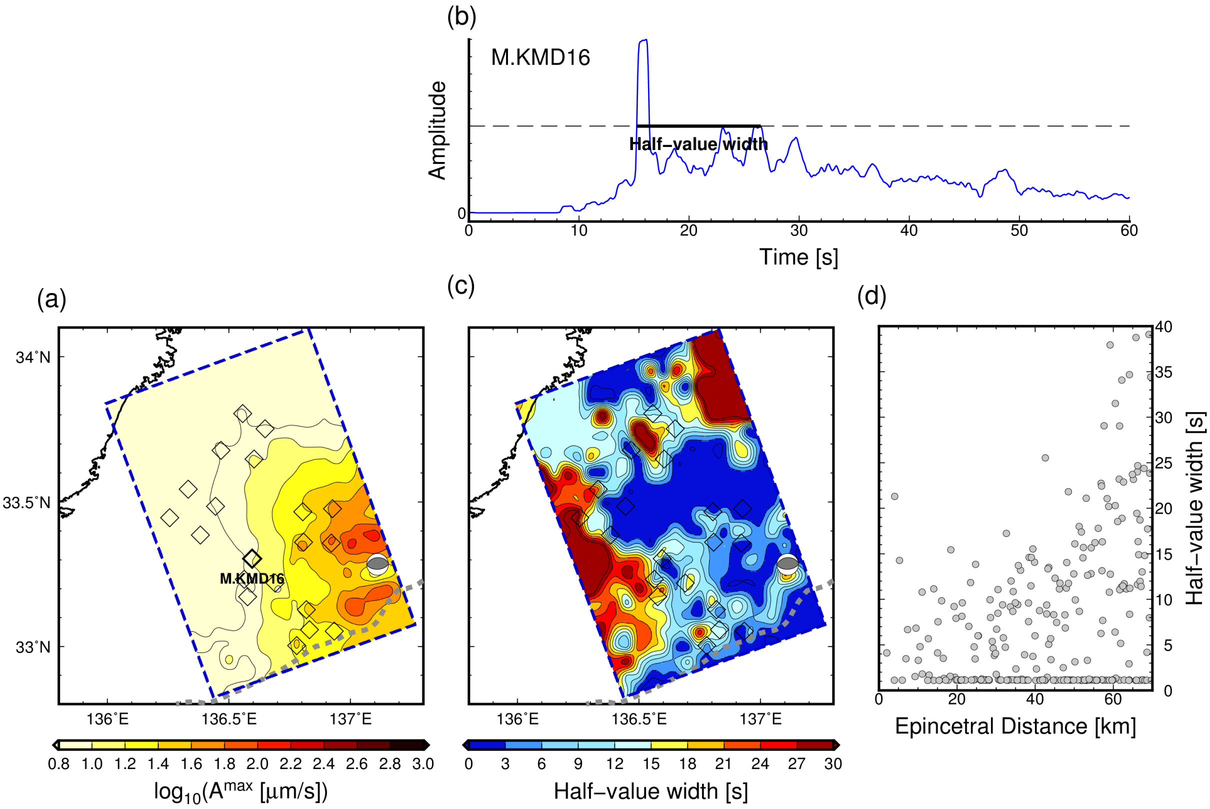

Figure S5. Spatial variations of maximum RMS amplitudes and half-value widths of simulated RMS envelopes for an intraslab earthquake: (a) Maximum RMS amplitude distribution, (b) example half-value width estimations, (c) half-value width distribution, and (d) half-value widths as a function of epicentral distance. An example in (b) is an RMS envelope at M.MRD16 (location shown in (a)).

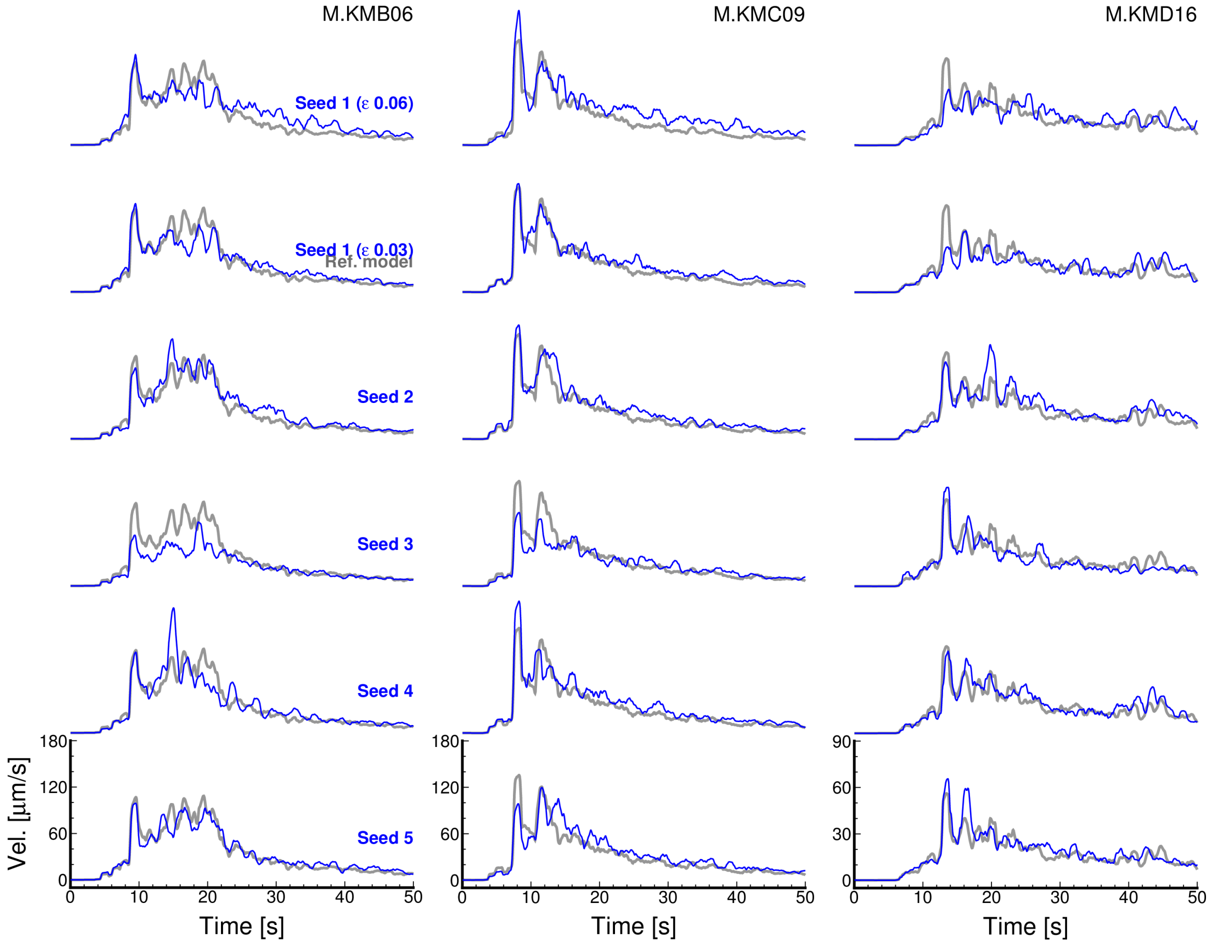

Figure S6. Comparisons of simulated RMS envelopes with different small-scale random velocity heterogeneity models. The PSDF type is an exponential type, and the parameters for small-scale random velocity heterogeneity models within the crust and accretionary prism are correlation length of a = 1 km and RMS values of ε = 0.03, 0.06. The models with a = 1 km and ε = 0.03 were conducted using different random seeds (numbered 1–5). Details of small-scale velocity heterogeneity models are described in Table 3. Locations of DONET stations are shown in Figure 8.

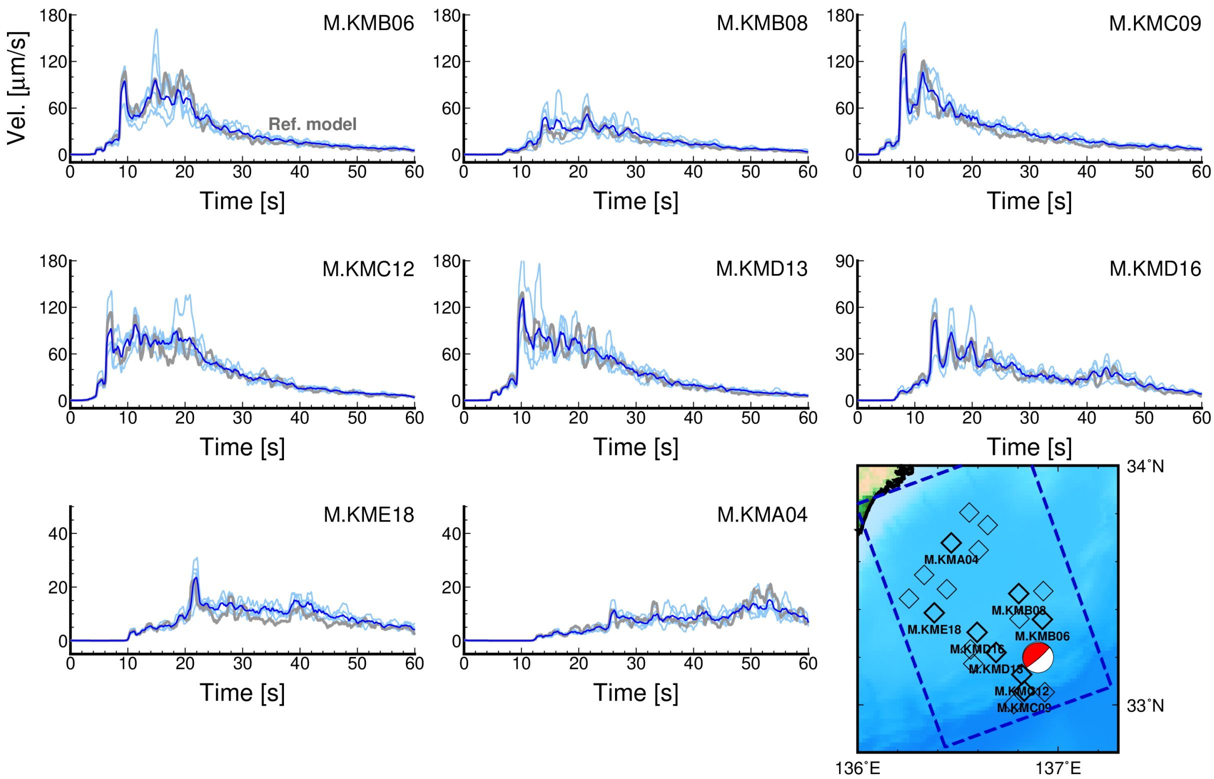

Figure S7. Comparisons between simulated envelopes with/without small-scale velocity heterogeneities. Grey lines are envelopes from the reference model (without small-scale velocity heterogeneities). Blue and light blue lines are stacked and individual envelopes in the model with small-scale velocity heterogeneities. Details of small-scale velocity heterogeneity models are described in Table 3.

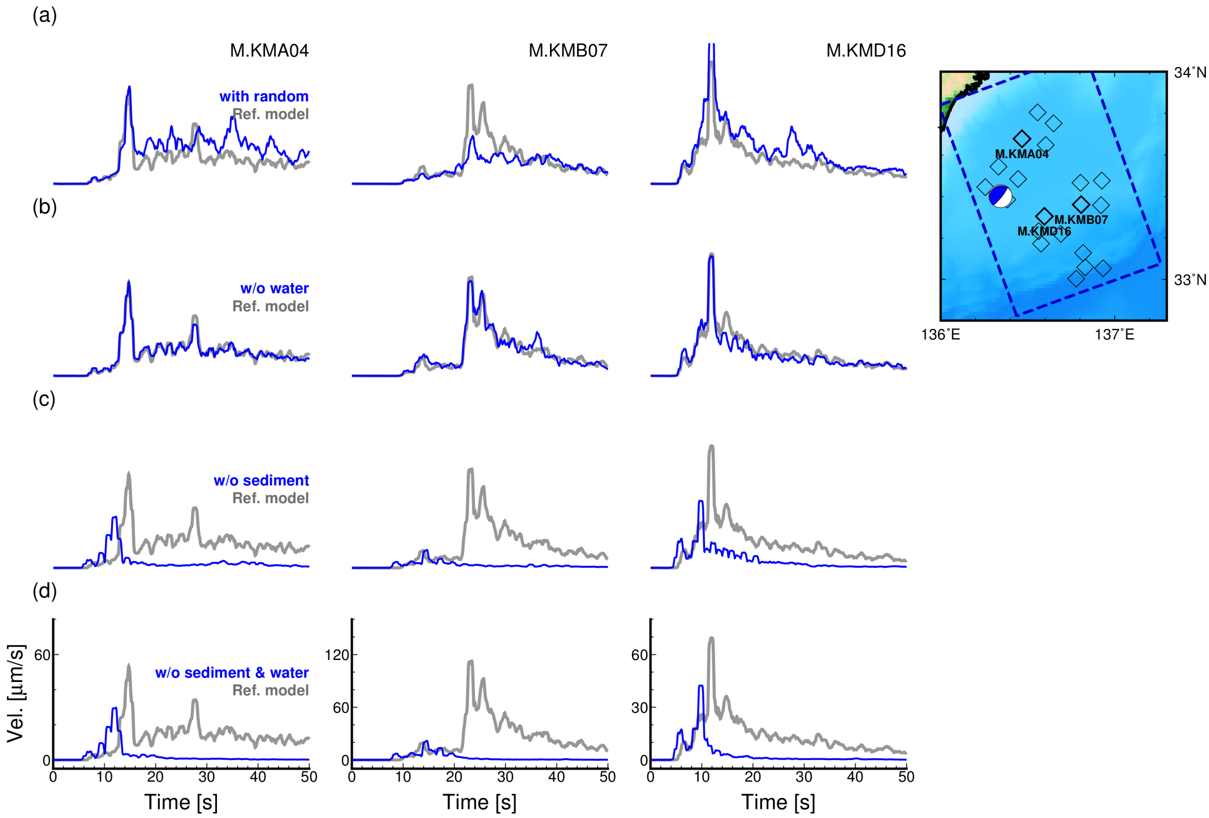

Figure S8. Simulation results of various heterogeneous models for an interplate earthquake. The blue lines are RMS envelopes derived from models: (a) with small-scale random velocity heterogeneities, (b) without seawater, (c) without oceanic sediments (accretionary prism), and (d) without oceanic sediments and seawater. Grey lines are the simulation results with the reference model.

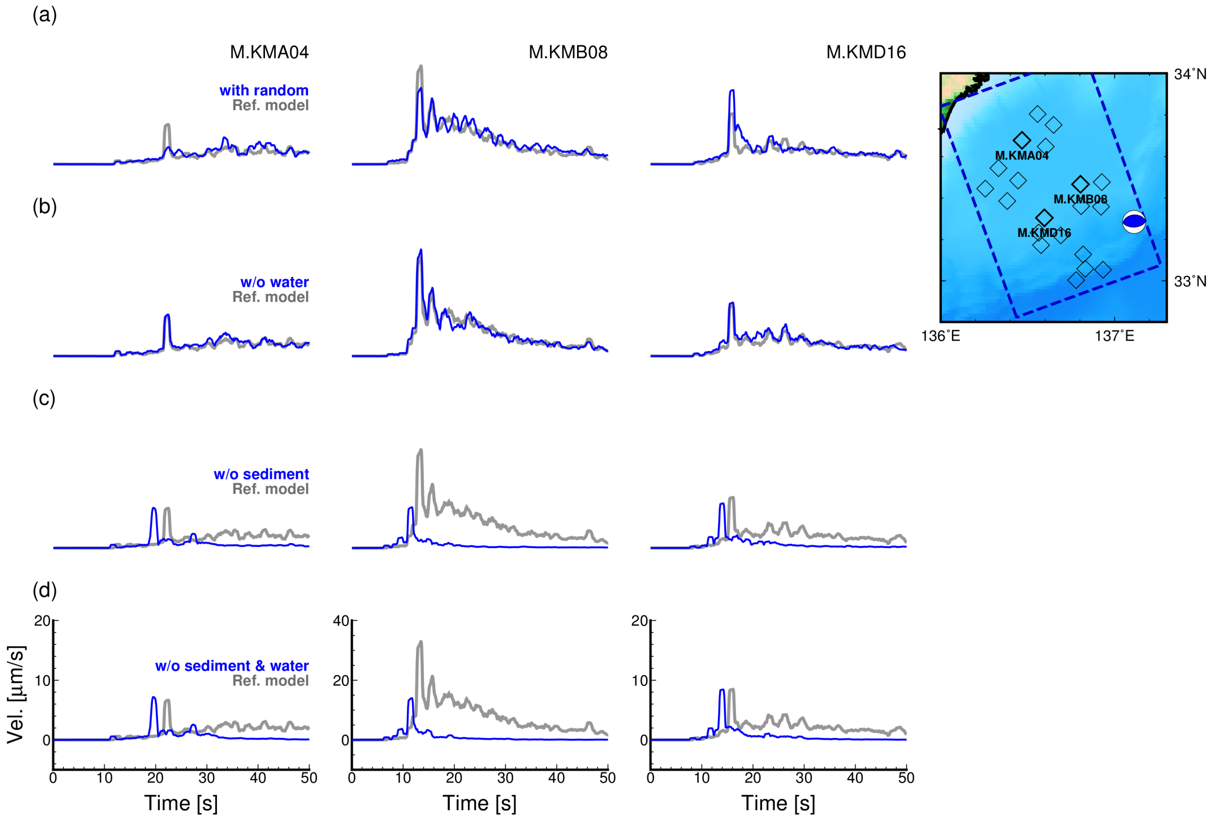

Figure S9. Simulation results of various heterogeneous models for an intraslab earthquake. The blue lines are RMS envelopes derived from models: (a) with small-scale random velocity heterogeneities, (b) without seawater, (c) without oceanic sediments (accretionary prism), and (d) without oceanic sediments and seawater. Grey lines are the simulation results with the reference model.

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