Three-Dimensional S-Wave Velocity Structure of the Kinki Region, Southwestern Japan based on Ambient Seismic Noise Tomography using Data from Dense Seismic Array

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Key Points:

- We estimated a high-resolution three-dimensional S-wave velocity model of the Kinki region using ambient noise tomography
- Our velocity model reveals the NE-SW trending Niigata-Kobe Tectonic Zone and the highly-active Biwako-seigan Fault System
- The unidentified probable fault zones were inferred from our fine-scale linear low-velocity anomalies and distribution of earthquakes

Abstract

Research interest in the Kinki region, southwestern Japan, has been aroused by the frequent occurrence of microearthquake activity, particularly on the northern side of the Arima-Takatsuki Tectonic Line. Previous studies in this area focused mainly on deep, large-scale structures and could not efficiently resolve fine-scale (~10 km) shallow crustal structures. Hence, characterization of the upper crustal structure of this region at an improved spatial resolution is required. By cross-correlating the vertical components of the continuous ambient seismic noise data from a dense seismic array, we estimated Rayleigh wave phase velocities using a frequency domain method. The 3D S-wave velocity structure of the Kinki region was then obtained by applying a direct surface wave tomographic method for the phase velocity dispersion data. The estimated velocity model reveals a NE-SW trending low-velocity structure coinciding with the Niigata-Kobe Tectonic Zone (NKTZ) and the active Biwako-seigan Fault Zone (BSFZ). Also, we identified fine-scale low-velocity structures coinciding with known active faults on the eastern side of the NKTZ, as well as sets of low-velocity structures across the Tanba region, that may be attributable to the weathering effects or activity of unidentified concealed fault zones. Furthermore, sedimentary basins manifest as low-velocity zones extending to depths ranging from ~1.5 to 2 km, correlating with those reported in previous studies. Our results therefore contribute towards fundamental understanding of earthquake faulting as well as tectonic boundary and will be useful for hazard assessment and disaster mitigation.

Plain Language Summary

Due to the frequent occurrence of low-intensity earthquakes in the Kinki region, southwestern Japan, there has been a keen research interest aimed at understanding the Earth's internal structure in this region. Geophysical methods based on the speed of seismic waves have been employed by other researchers to examine the structural make-up of the Earth's interior in a wide area. In the Kinki region, previous studies focused on deep, large-scale features of the earth and could not sufficiently map shallow, small-sized (~10 km) structures. In this study, we used ambient vibrations of the earth to extract seismic wave speeds information. Then, we constructed a high-resolution three-dimensional geological model based on seismic wave speeds using specialized geophysical method. Our results reveal linear low-speed zones, interpreted as documented active fault zones and undocumented probable fault zones. These features could be linked to the frequent occurrence of earthquakes in the Kinki region. Our results contribute towards improved understanding of the shallow crustal structure in the Kinki region and can be used to identify earthquake-prone zones, thus facilitating disaster risk reduction. Furthermore, we can use the information of seismic wave speed for accurate earthquake hypocenter estimation.

1 Introduction

To unravel heterogeneities within the crustal structure and upper mantle over a wide area, very few geophysical techniques with proven efficacy are available (Suemoto et al., 2020). Ground-based geophysical methods such as active-source seismic reflection and refraction techniques can be used to map the distribution of faults and geologic boundaries at high resolution. A striking example is a study by Sato et al. (2009), in which deep seismic reflection profiling was employed along a 135 km long seismic line from Metropolitan Osaka to the Ise basin (Osaka-Suzuka seismic survey) across several active reverse faults. Likewise, Ito et al. (2006) conducted a similar survey along the N–S-trending Shingu-Maizuru line. However, this approach only provides details about fault locations and geologic boundaries along the profiles, and heterogeneities across the profiles can only be established from multiple profiles. Therefore, this approach is not well suited to constructing large-scale geological models for areas as large as the Kinki region.

Conversely, P- and S-wave travel-time tomography using local earthquake data over a wide area has provided significant results, resolving major structures such as faults and geologic boundaries (Matsubara et al., 2008; Nakajima et al., 2009; Yolsal-Cevikbilen et al., 2012). Even so, the downside of this approach is that the resolution of geological structures depends on the distribution of natural earthquakes (Suemoto et al., 2020). Using teleseismic data, surface wave tomography can also be applied but teleseismic propagation paths complicate short period (<20 s) measurements due to the scattering and attenuation that occur as distant waves propagate (Bensen et al., 2007; Yang, 2014). Such short-period measurements are the core of our objectives in this study as we seek to resolve shallow crustal features within the Kinki region.

The emergence of ambient noise tomography (ANT) in the last decade has revolutionized seismic tomography because it can circumvent the shortcomings of traditional earthquake surface wave tomography (Sabra et al., 2005; Shapiro et al., 2005). This method utilizes diffuse background ground motion (ambient noise) to extract surface wave empirical Green's functions between station pairs by cross-correlating continuously recorded seismic waveforms (Yang, 2014). It is therefore convenient because surface wave dispersion curves between station pairs can be estimated without requiring the occurrence of earthquakes, as each station can serve as a virtual source and a receiver (Yang, 2014). Since the inception and further developments of permanent and temporary high-quality seismic networks, ANT has been applied to estimate subsurface structures in various geological settings (Chen et al., 2018; Lin et al., 2008; Nishida et al., 2008; Shapiro et al., 2005). Suemoto et al. (2020) constructed a high-resolution shallow-crustal S-wave velocity structure by ambient noise surface wave tomography in the San-in area using data recorded by several seismic arrays, comprising Hi-net stations (Obara et al., 2005) and the dense Manten project array (Iio et al., 2018). Similarly, Nimiya et al. (2020) successfully utilized surface wave data extracted from ambient noise recorded by Hi-net stations to estimate the 3D S-wave velocity structure of central Japan. In contrast, there is still no detailed information about the 3D S-wave velocity structure of the Kinki region.

To provide an improved constraint on fine-scale (~10 km) shallow-crustal structures and geological boundaries in the Kinki region, we performed ambient noise tomographic inversion. We constructed a high-resolution shallow S-wave velocity model using data recorded by the widely distributed Hi-net stations and Japan Meteorological Agency (JMA) stations integrated with data recorded by densely distributed Manten project seismic stations (Iio et al., 2018; Katoh et al., 2019). From our model, the geometry and spatial distribution of major faults and geological boundaries are estimated.

2 Geologic setting

In the Kinki region, southwestern Japan, the Philippine Sea (PHS) oceanic plate subducts beneath the overriding Eurasian (EUR) plate (Aoki et al., 2016). The southeastward movement of the incipient Amurian plate (Amur Plate) with respect to the EUR plate and a shift in the direction of subduction of the PHS plate (Taira, 2001) has generated relatively new, large fault zones or continually reactivates the old ones, a process referred to as neotectonics (Barnes, 2008).

The major contributors in neotectonics faulting in the Kinki region comprise, among others, the Median Tectonic Line (MTL; black line in Figure 1a), which has been reactivated but now in the right-lateral direction, opposing the Cretaceous left-lateral strike-slip fault action (Barnes, 2008). The MTL divides the Kinki region into outer zone and inner zone (Matsushita, 1963). On the one hand, the outer zone is characterized by four zonally arranged terrains from north to south: namely, the Sanbagawa metamorphic terrain, Chichibu terrain, Hidaka terrain, and Muro terrain (SMZ, CT, HT, and MT; Figure 1a). On the other hand, the zonal arrangement of geologic formations in the inner zone is not prominent, it is divided from north to south into the Tango–Tajima terrain (composed largely of Neogene volcanic and sedimentary series), the Maizuru zone (consisting of Permian and Triassic marine formations and Yakuno intrusive rocks), the Mino–Tanba terrain (including Cretaceous granite), and the Ryoke terrain consisting primarily of Ryoke metamorphic rocks and granites (TTT, MZ, MTT, and RT; Figure 1a) (Matsushita, 1963).

Huzita (1980) reported a triangular neotectonics province (the Kinki triangle, gray-shaded area in Figure 1a) with the E–W compressional stress state and the alternating basins and mountain topography from east to west, which provided central Japan with its civilizational homelands, including the Nara, Osaka, Kyoto and Ise basins (Barnes, 2008). The Kinki triangle is characterized by a dense distribution of predominantly NS-trending Quaternary active faults and some NE-SW or NW-SE strike-slip active faults (Research Group for Active Faults of Japan, 1991). The predominantly E–W compressional movement in the upper crust has engendered the uplifted mountains between the Osaka and Ise basins as a pop-up structure (Sato et al., 2009). A plethora of historical large and destructive earthquakes have occurred in the Kinki region and surrounding areas (Hyodo & Hirahara, 2003; Usami, 2003), mainly in the Jurassic Tanba Accretionary Complex (Wakita, 2013), sharply bounded by the ENE-WSW dextral strike-slip Arima-Takatsuki Tectonic Line (ATTL) to the south (Hallo et al., 2019; Iio, 1996; Katao et al., 1997; Matsushita & Imanishi, 2015) and the reactivated Niigata-Kobe Tectonic Zone (NKTZ) to the east, which is characterized by high strain rates (Sagiya et al., 2000). In recent times, low magnitude earthquakes have been recorded across the entire Tanba region, but their locations do not always coincide with known faults (Kato & Ueda, 2019). However, in such cases, the distribution of their hypocenters makes a chain of seismic alignment between pairs of faults (Oike, 1976). This alignment suggests the possible existence of concealed active faults in those areas, or continuity of known fault systems. Therefore, the possible existence of concealed faults in those areas needs to be investigated in high resolution.

Around the Osaka area, numerous active faults exist, including the E-W to NE-SW oriented dextral strike-slip faults and the N-S striking reverse faults (Research Group for Active Faults of Japan, 1991). The significance of these faults was highlighted by the highly catastrophic 1995 M_w 7.2 Kobe earthquake, which resulted from dextral strike-slip displacements on the NE–SW striking Rokko-Active Fault Zone (Kanamori, 1995; Katao et al., 1997). In addition, a shallow crustal earthquake of magnitude M_w 5.6 occurred in 2018, proximal to the junction of the eastern part of the ATTL and the east-dipping reverse active fault systems, the Uemachi and Ikoma fault zones (Kato & Ueda, 2019; Sato et al., 2009). These earthquakes are a testament to how susceptible life is to tectonic movement along these fault lines and highlight the need to identify zones prone to strong crustal movement in a quest to minimize the effects of destructive earthquakes. Such zones include concealed fault zones, which are difficult to ascertain from surficial evidence, as well as active and new fault systems, which are likely to be the locus of future events.

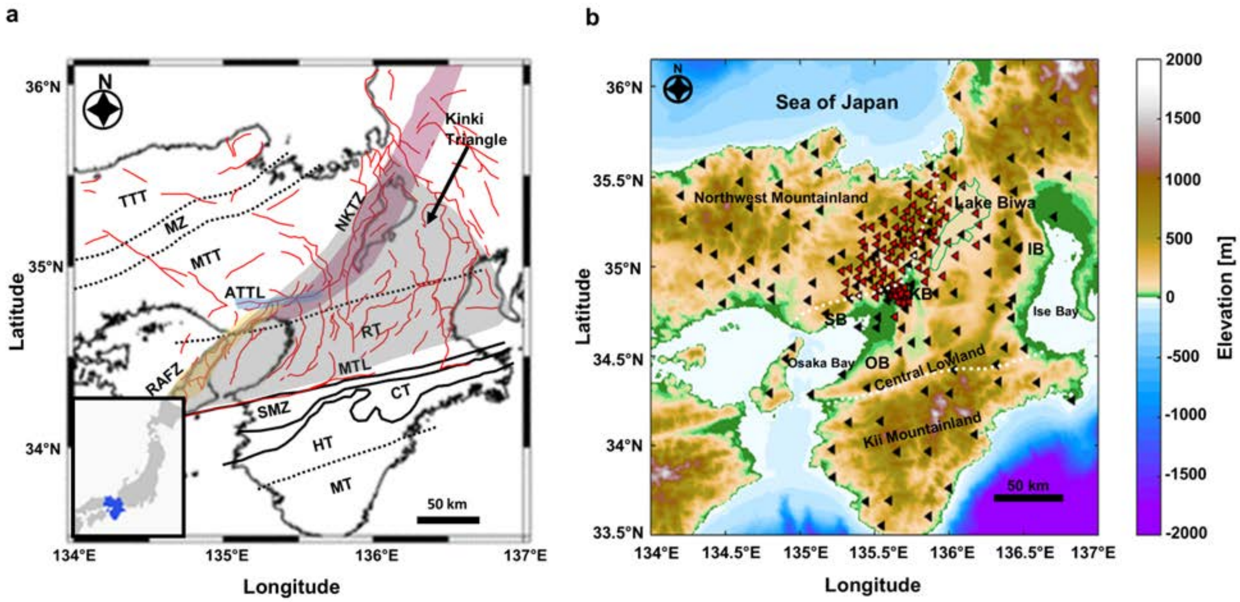


Figure 1. (a) Map of the Kinki region showing the spatial distribution of tectonic structures. Red lines represent active faults (retrieved on 19 November 2021 from https://gbank.gsj.jp/subsurface/english/ondemand.php), thick purple, yellow, and light blue lines represent the Niigata-Kobe Tectonic Zone (NKTZ), Rokko Active Fault Zone (RAFZ), and the Arima-Takatsuki Tectonic Line (ATTL), respectively. Also shown are the locations of the Median Tectonic Line (MTL), and tectonic divisions of the Kinki region, comprising Tango-Tajima Terrain (TTT), Maizuru Zone (MZ), Mino-Tamba Terrain (MTT), Ryoke Terrain (RT), Sanbagawa Metamorphic Zone (SMZ), Chichibu Terrain (CT), Hidaka Terrain (HT), and Muro Terrain (MT). The insert shows the location of the Kinki region within Japan. (b) Topographic map of the Kinki region. Red and black triangles represent the locations of temporary and permanent stations, respectively. White, broken lines indicate the boundaries between the Northwestern Mountainland, Central Lowland, and Kii Mountainland.

3 Data and Methods

We used the vertical component of continuous seismic waveforms recorded by permanent and temporary stations from April 1 to September 30 during the year 2019. The permanent stations included 78 Hi-net stations, 1 Kyushu University station, 1 Tokyo University station, 2 Nagoya University stations, 10 AIST stations, 16 Kyoto University stations and 9 JMA stations, and temporary stations comprised 104 Kyoto University Manten project stations (Iio et al., 2018; Katoh et al., 2019), that are distributed around the central part of the Kinki region. Combining these set of stations enabled us to obtain a dataset with very dense ray paths coverage of short-period surface waves and a subsequent high-resolution 3D S-wave velocity structure. To construct the 3D S-wave velocity structure, we first extracted surface waves propagating between station pairs by computing the cross-correlation of ambient noise. We then employed the zero-crossing method (Ekström et al., 2009) to estimate Rayleigh wave phase velocities between station pairs. Finally, we used the direct surface wave inversion (Fang et al., 2015) to obtain the 3D S-wave velocity model.

3.1 Preprocessing and cross-correlation

After partitioning continuous ambient noise data for each day into 30 minute segments with a 50% overlap, the instrumental response of each dataset was corrected. Next, cross-correlation spectra for all possible combinations of station pairs were computed from the resulting seismograms (Ekström, 2014). Then, the daily cross-correlation spectra were stacked over a six-month-long time series. Rayleigh wave propagation can be clearly observed in the time-domain cross-correlation computed from the stacked cross-correlation spectra (Figure 2).

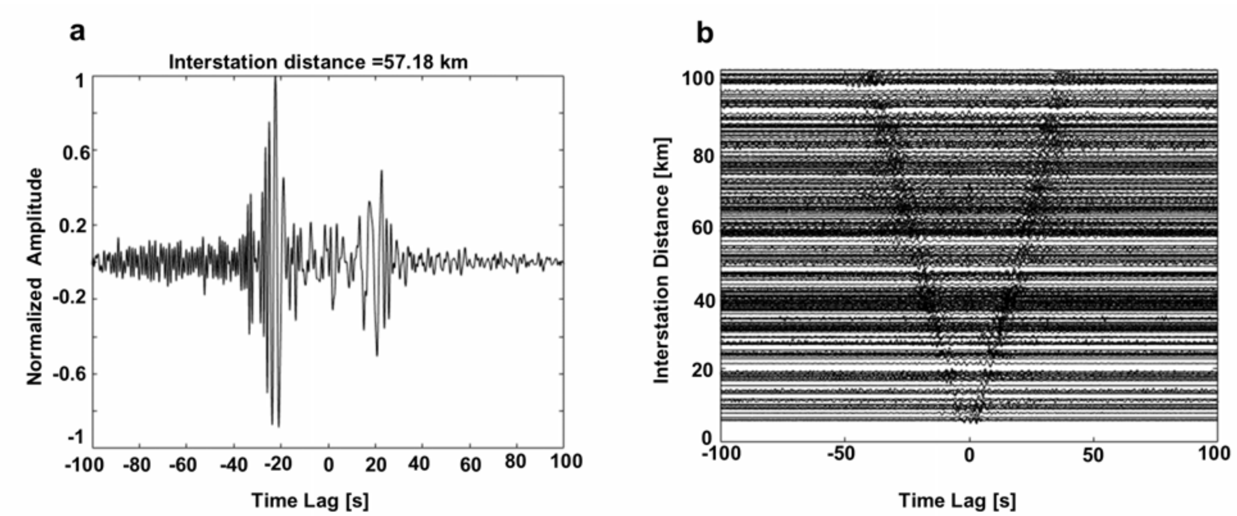


Figure 2. Cross-correlation functions of the vertical component in the 0.05-0.95 Hz frequency band showing the empirical Green's functions between station pairs. (a) Cross-correlation function for a station pair with an interstation distance of 57.18 km (shown in Figure 3b), and (b) stacked cross-correlation functions from randomly selected station pairs, exhibiting surface wave propagation between station pairs.

3.2 Surface wave phase velocity measurement

Phase velocity measurements can be performed in either the frequency domain or time domain. In the time domain, only those interstation distances exceeding three wavelengths (λ) are typically considered due to the high-frequency approximation required in time domain analysis (Bensen et al., 2007; Lin et al., 2008; Yao et al., 2006). In contrast, there is no theoretical limitation for interstation distances in the frequency domain method (i.e., zero-crossing method; Ekström et al., 2009), and interstation distances up to approximately one wavelength can be practically used (Ekström et al., 2009; Tsai & Moschetti, 2010). We applied the zero-crossing method to estimate phase velocities between station pairs. The zero-crossing method is based on modeling cross-correlation spectra by the spatial autocorrelation (SPAC) method (Aki, 1957; Asten, 2006) and uses the zero-crossing frequencies of the real part of the cross-correlation spectra. The SPAC method is premised on the assumption that ambient noise is dominated by surface waves and that ambient noise sources are distributed uniformly (Aki, 1957). Under this assumption, the real part of the vertical cross-correlation spectra can be modeled using a Bessel function of the first kind and zeroth order as follows:

$$Real\big(\rho(f,x)\big) = J_0\left(\frac{2\pi f x}{C_R(f)}\right), \tag{1}$$

where ρ is the vertical cross-spectrum, x is the interstation distance, J_0 is the Bessel function of the first kind and zeroth order, and $C_R(f)$ is the Rayleigh wave phase velocity. In the zero-crossing method, we focus only on the zero crossings where both sides of equation (1) should be zero. The zero-crossing points are insensitive to fluctuations in the spectral power of the background noise and non-linear filtering in the data processing (Ekström et al., 2009). Using zero crossings simplifies phase velocity measurements and stabilizes the estimation of phase velocities because phase velocity estimation is not affected by incoherent noise (Cho et al., 2021).

If f_n represents the frequency of the nth observed zero crossing of the cross-correlation spectrum, and Z_n denotes the nth zero of the Bessel function, we can match each f_n with the zero crossings of the Bessel function to have all the possible phase velocity dispersion curves according to the following equation:

$$C_m(f_n) = \frac{2\pi f_n x}{Z_{n+2m}}, \tag{2}$$

where m representing the number of missed or extra zero points, takes the values (0, ±1, ±2,…). Applying equation (2) for all observed values of f_n yields numerous possible dispersion curves.

We used the GSpecDisp package (Sadeghisorkhani et al., 2018) to determine Rayleigh-wave phase-velocity dispersion curves uniquely by the zero-crossing method from the stacked cross-correlations. To reduce noise effects in the correlations, we applied a velocity filter of 1–4.5 km/s with a taper interval of ~0.2 km/s. Then, we applied spectral whitening to each correlation for amplitude equalization (Sadeghisorkhani et al., 2018). With many possible phase velocities occurring at each frequency with regard to equation (2) (colored dots; Figure 3a), it is difficult to uniquely determine the phase velocity dispersion curves without using a reference velocity dispersion curve as a guide. To circumvent this, we manually picked the dispersion curve appearing closest to the reference dispersion curve. In the GSpecDisp, average velocities can be estimated by combining all cross-correlation spectra (average velocity module). We estimated average velocities in the period range from 2 to 8 s and used the result as a reference velocity for dispersion curve estimation in single station-pair phase-velocity picking mode in GSpecDisp (dashed black dots; Figure 3a). Finally, we estimated phase-velocity dispersion curves between each station pair (red circles in Figure 3a).

For our dataset, the maximum measurable period required an interstation distance (x, in km) of at least three wavelengths (λ), defined as the r/λ ratio in GSpecDisp ($x/\lambda \geq 3$). For each cross-correlation function, the signal-to-noise ratio (SNR) was defined as the ratio between maximum absolute amplitude in the signal window (between arrival times corresponding to waves with 1 and 4.5 km/s) and the root mean square amplitude in the noise time window (between 500 and 700 s). We used an SNR threshold of 10 to reject correlations with low signal. Finally, we obtained a total of 23,647 dispersion curves (Figure 3c).

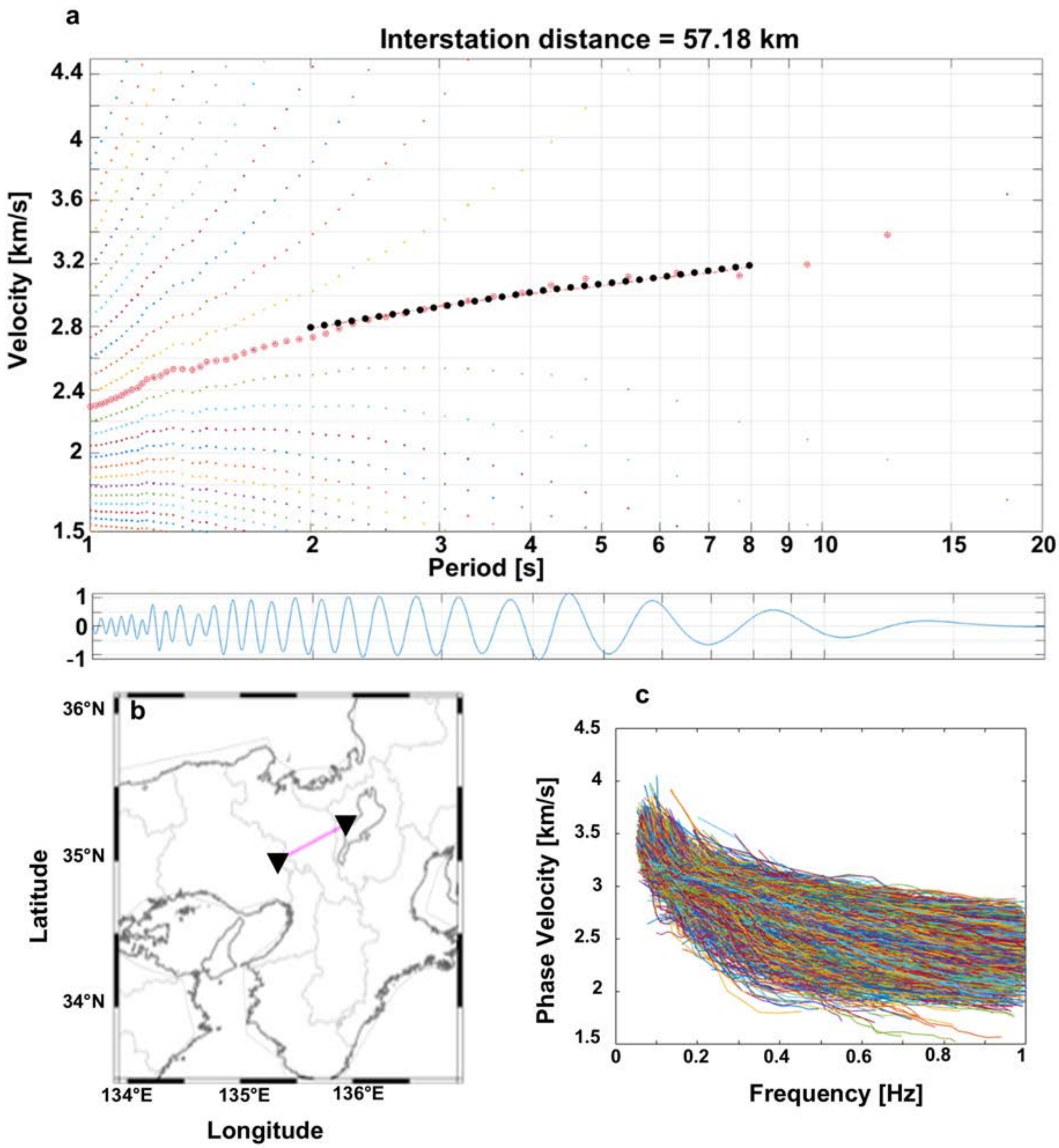


Figure 3. (a) Observed phase velocity dispersion curves (upper panel) and the real part of the cross-correlation spectrum (lower panel). Red and black circles in the upper panel represent the selected points of the dispersion curve and the average phase-velocity dispersion curve for the region, respectively. (b) Location of the station pair for which dispersion data are displayed in (a). (c) Phase-velocity–frequency plot showing the 23,647 selected dispersion curves for all the station pairs used.

3.3 Direct inversion of the surface wave dispersion curves

Ambient noise tomography using phase velocity dispersion curves typically involves a two-step procedure. Firstly, 2D phase velocity maps are constructed by travel-time tomography at discrete frequencies. Secondly, pointwise inversion of dispersion data for 1D profiles of S-wave velocity as a function of depth at each grid point is implemented, and combining multiple 1D profiles subsequently yields the 3D S-wave velocity structure (Shapiro & Ritzwoller, 2002; Yao et al., 2008). Nonetheless, a 3D S-wave velocity structure can equally be estimated by direct inversion of dispersion data without the intermediate step of constructing 2D phase velocity maps (Boschi & Ekström, 2002; Fang et al., 2015; Feng & An, 2010; Pilz et al., 2012). Typically, these direct inversion approaches do not update the ray paths and sensitivity kernels for the newly constructed 3D models (Fang et al., 2015). Also, one-step linearization may produce biased wave velocity estimations in a medium akin to the shallow crustal structure, where S-wave velocity variations can exceed 20% (Lin et al., 2013).

To estimate the 3D S-wave velocity structure from phase velocity dispersion data, we applied a direct surface wave tomography method (DSurfTomo), which is based on frequency-dependent ray-tracing and a wavelet-based sparsity-constrained inversion (Fang et al., 2015). This approach circumvents the intermediate step of constructing 2D phase velocity maps and iteratively updates the sensitivity kernels of period-dependent dispersion data (Fang et al., 2015). Furthermore, it accounts for the ray-bending effects of period-dependent ray paths by using the fast-marching method (Rawlinson & Sambridge, 2004). Accounting for such effects in the inversion is especially useful for short-period surface waves, which are significantly sensitive to the highly complex shallow crustal structure (Fang et al., 2015; Gu et al., 2019). Therefore, this approach is a well-suited tool for determining the shallow-crustal structure of the Kinki region using short-period surface-waves dispersion data.

In tomographic inversion, the objective is to find a model $\mathbf{m}$ that minimizes the differences $\delta t_i(f)$ between the measured travel times $t_i^{obs}(f)$ and the calculated travel times $t_i(f)$ from the model for all frequencies f. The travel time for path i is given as

$$\delta t_i(f) = t_i^{obs}(f) - t_i(f) \approx -\sum_{k=1}^{K} v_{ik} \frac{\delta C_k(f)}{C_k^2(f)}, \tag{3}$$

where $t_i(f)$ represents the computed travel times from a reference model that can be updated in the inversion, v_{ik} denotes the bilinear interpolation coefficients along the ray path associated with the ith travel-time data, $C_k(f)$ is the phase velocity, and $\delta C_k(f)$ denotes its perturbation at the kth 2D surface grid node at frequency f. Surface wave dispersion is sensitive primarily to S-wave velocity (Fang et al., 2015). However, short-period Rayleigh wave dispersion also has a sensitivity to P-wave velocity in the shallow crust (Fang et al., 2015). The P-wave velocity perturbations together with mass density are therefore explicitly included in the calculation of surface wave dispersion, with R'_α and R'_ρ as scaling factors, leading to the following equation:

$$\delta t_i(f) = \sum_{k=1}^{K} \left(-\frac{v_{ik}}{C_k^2}\right) \sum_{j=1}^{J} \left[R'_\alpha(z_j) \frac{\partial C_k}{\partial \alpha_k(z_j)} + R'_\rho(z_j) \frac{\partial C_k}{\partial \rho_k(z_j)} + \frac{\partial C_k}{\partial \beta_k(z_j)} \right] \Bigg|_{\theta_k} \delta\beta_k(z_j) = \sum_{l=1}^{M} G_{il} m_l, \tag{4}$$

where θ_k denotes the 1D reference model at the k^{th} surface grid point, $\alpha_k(z_j)$, $\rho_k(z_j)$, and $\beta_k(z_j)$ represent the P-wave velocity, the mass density, and the S-wave velocity, respectively. J is the number of grid nodes in the depth direction, and $M = KJ$ represents the total number of grid points. Equation (4) can be written in matrix form, as follows:

$$\mathbf{d} = \mathbf{G}\mathbf{m}, \tag{5}$$

where **d** is the surface wave travel-time residual vector for all ray paths and discrete frequencies, **G** represents the data sensitivity matrix, and **m** represents the model parameter vector. Damping and weighting parameters are applied to balance data fitting and smoothing regularization. In addition to the damping and weighting parameters, the sparsity fraction, which is a parameter indicating how sparse the sensitivity matrix is, was selected on a trial-and-error basis for our data considering the diverse patterns in inverted S-wave velocity models (weakly smoothed and strongly smoothed S-wave velocity models are shown in Figures S1 and S2, respectively).

In our inversion, the entire Kinki region was parameterized into 55 by 60 grid points on the horizontal plane with an interval of 0.05° in each horizontal direction (latitude and longitude), as well as 7 grid points along the depth direction (i.e., 0, 0.5, 1.0, 2.2, 4.0, 6.0 and 9.0 km). These parameters along with the large volume of dispersion data were memory intensive, we therefore used dispersion data within a narrow frequency bandwidth of 0.083 - 0.67 Hz to circumvent the computer memory limitations during the inversion. Dispersion measurements within a broad frequency bandwidth of 0.05 - 0.95 Hz was used for the northern part of the Kinki area, which was parameterized into 29 by 96 grid points on the horizontal plane with an interval of 0.02° in the latitude and longitude directions, and 11 grid points along the veritical direction (0, 0.1, 0.3, 0.5, 0.8, 1.4, 2.0, 3.0, 4.0, 5.5 and 7.0 km). Empirically, the fundamental mode Rayleigh wave phase velocity is primarily sensitive to 1.1 × S-wave velocity at depths of about one-third of its corresponding wavelength (Fang et al., 2015; Foti et al., 2014; Hayashi, 2008). Consequently, an initial S-wave velocity model was constructed by multiplying the average of the observed Rayleigh wave phase velocities at a depth of one-third of the wavelength by 1.1 (i.e., a one-third wavelength transformation; Figure 4). Topographic effects on our inverted S-wave velocity models were accounted for by subtracting altitude from the depth value at each grid point because the inversion is based on the assumption of a flat surface. Therefore, the depth shown in our final 3D S-wave velocity models is the depth below sea level (S-wave velocity models before topographic correction are shown in Figures S3, S4 and S5).

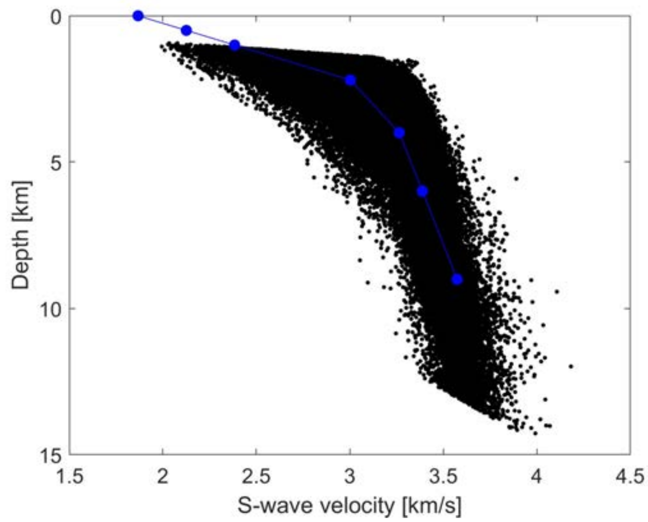


Figure 4. Initial S-wave velocity model. Black dots represent transformed interstation Rayleigh wave phase-velocity dispersion curves measured using the zero-crossing method to a depth–S-wave velocity approximation. The blue line and blue dots represent the average S-wave velocity model used as the initial reference S-wave velocity model in the inversion process.

4 Results

We employed the direct surface wave tomographic inversion method to construct a 3D S-wave velocity model of the Kinki region using Rayleigh wave dispersion curves. To ensure the reliability of our measurements, we plotted the spatial ray paths coverage within the Kinki region (Figure 5). From Figure 5, it is apparent that the ray paths density is sufficient to provide reliable measurements, especially in the most central part of the study area, where seismic stations are densely distributed. At the edges, however, the ray paths coverage is slightly limited. To further corroborate our inverted S-wave velocity model, we conducted a checkerboard resolution test using anomalies of ~0.2° (~22 km; Figure 6a, b) and ~0.1° (~11 km; Figure S6) for the entire Kinki region and the northern part of the Kinki region, respectively, with an amplitude of the velocity anomaly set to ~10%. In Figure 6, we display the results of the checkerboard resolution test for horizontal slices at different depths. Using these parameters, tectonic and geologic features with sizes greater than 10 km could be resolved clearly in the center of the study area using observed data.

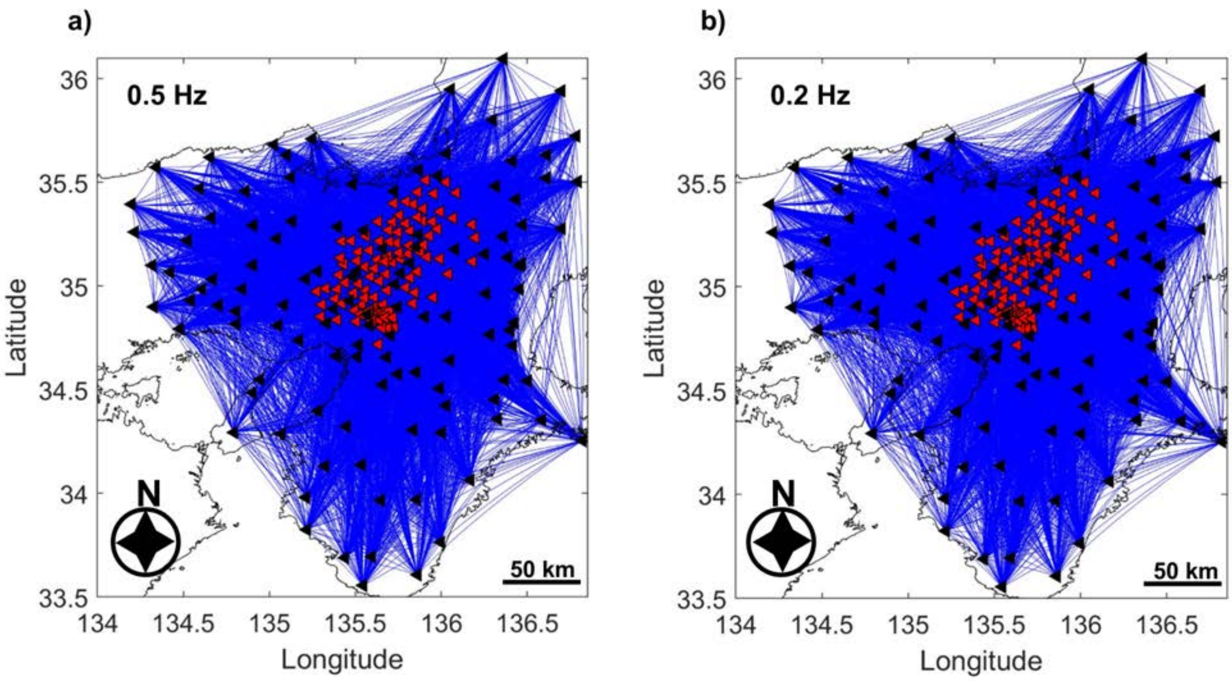


Figure 5. Ray paths obtained from the final 3D inversion model at two selected frequencies by the fast-marching method: (a) 0.5 Hz and (b) 0.2 Hz. Red and black triangles represent the locations of temporary and permanent stations, respectively. Blue lines indicate ray paths.

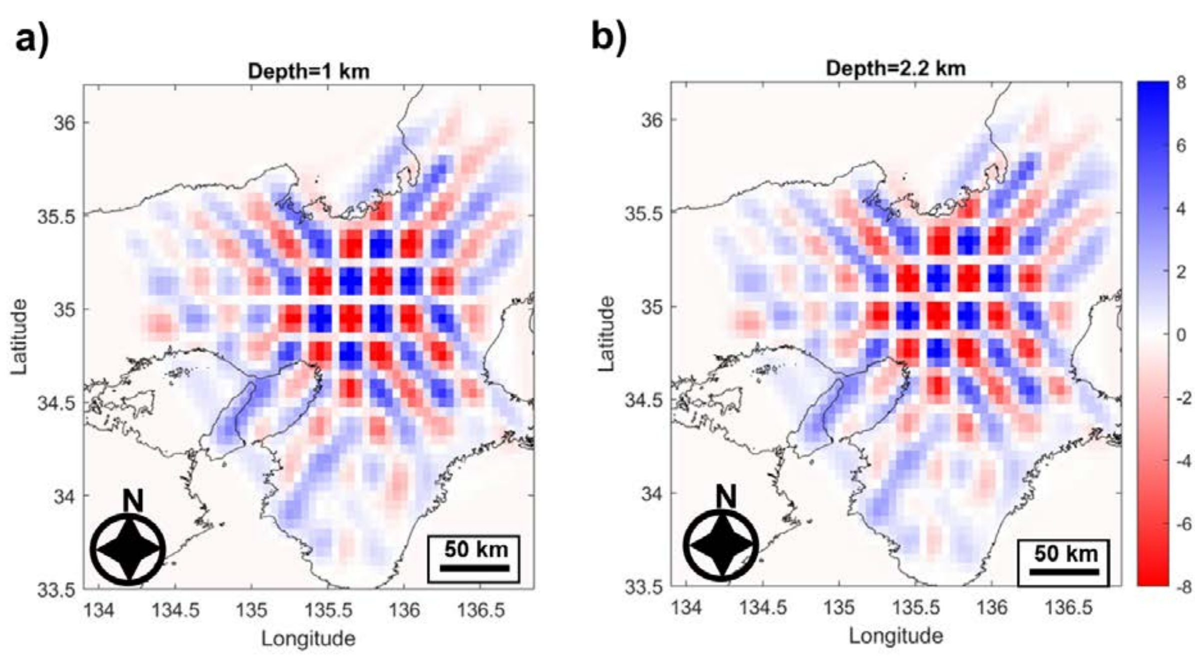
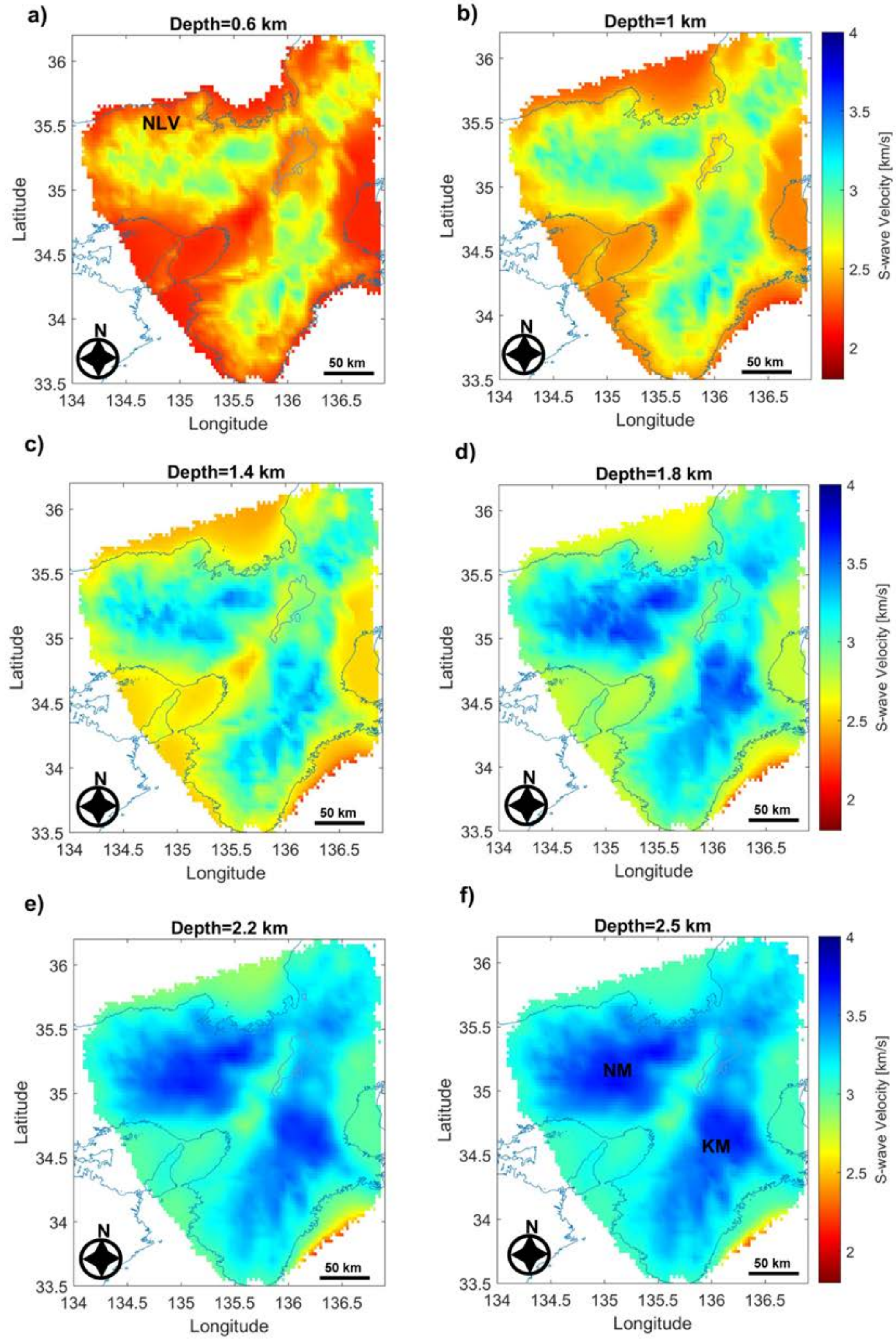


Figure 6. Horizontal velocity perturbation slices of the checkerboard resolution test results at 1 km (a) and 2.2 km (b) depths. The anomaly size was ~22 km (0.2°), and the velocity amplitude was ~10%. Depth is shown above each horizontal slice.

Figure 7 displays selected horizontal slices (map views) at different depths, exhibiting the lateral distribution of S-wave velocities within the study area. The third dimension (depth, in km) is given in the numerical form above each horizontal slice. Significant S-wave velocity heterogeneities are apparent and are discussed in detail in the following sections. These anomalies highlight tectonic and geologic features associated with the study area.

Two broad high-velocity anomalies can be observed in the displayed horizontal slices. The first anomaly (marked NM, Figure 7f) appears to be trending in the E–W direction, whereas the second high-velocity anomaly (marked KM) occurs from the southern side of the study area, trending roughly NE–SW across the MTL. These anomalies agree with the results of Nishida et al. (2008), which indicated comparable S-wave velocities in those areas, particularly at a depth of about 2 km (see Figure 20 in Nishida et al., 2008). Between the two distinct high-velocity zones exhibited in Nishida et al. (2008), an elongated low-velocity anomaly is evident. Likewise, a prominent low-velocity anomaly is apparent in our results, flanked on both sides by high-velocity zones (NM and KM) and trending roughly NE–SW. Although our results and those of Nishida et al. (2008) at a depth of about 2 km are similar, our results show prevalent small-scale (~10 km) low-velocity features at a depth of about 2 km and shallower. In the S-wave velocity model of Nishida et al. (2008), such narrow low-velocity zones could not be revealed. The higher lateral resolution of our velocity model at shallow depths (≤ 2 km) than Nishida et al. (2008) model ascribes to the use of shorter wavelength surface waves and the dense seismic array. Most importantly, prominent anomalies identified in our results correlate well with known geologic features, including fault zones, sedimentary basins, and mountain ranges.



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375 **Figure 7**. S-wave velocity models at different depths below sea level, given above each panel. (a–f) S-
376 wave velocity models without showing the active faults (S-wave velocity models before topographic
377 correction are shown in Figure S4). NM and KM represent the prominent high-velocity anomalies.

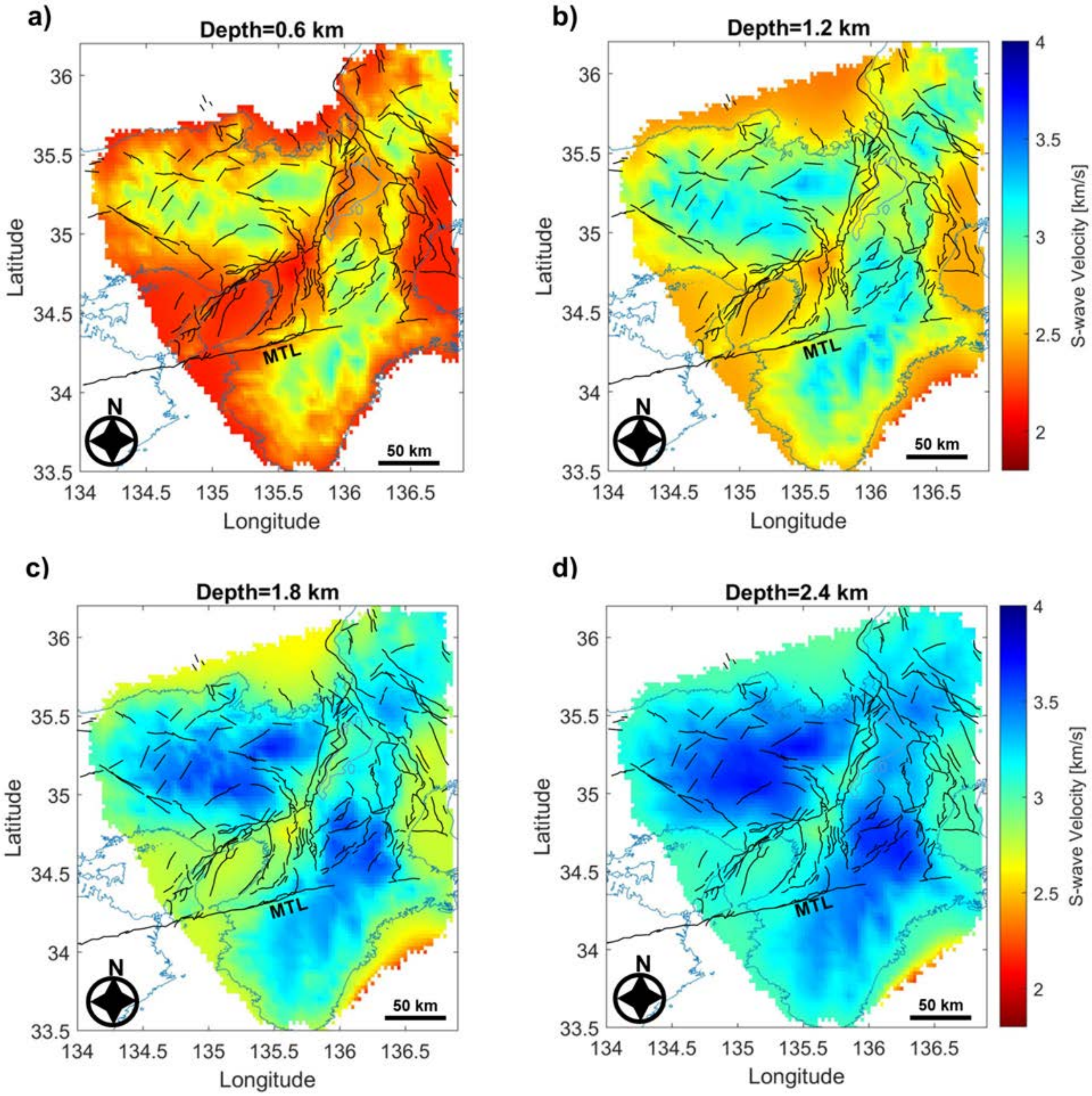

Figure 8. S-wave velocity models at different depths below sea level, given above each panel. (a–d) S-wave velocity models overlaid with active faults (black lines). Also shown is the location of the Median Tectonic Line (MTL).

5 Interpretations

We attribute the high S-wave velocity anomaly observed in the northwestern part of the study area (marked NM in Figures 7f and 9b) to the presence of the Yakuno intrusive rocks and the Mino/Tamba belts (Figure 1a). The Yakuno intrusive rocks constitute the Maizuru zone, and the Mino/Tamba belts are Jurassic accretionary complexes composed of non-marine sediments, and the extensively distributed granite batholith (Matsushita, 1963; Nakae, 1993; Nakajima, 1994).

Towards the edges of the study area, the resolution of our S-wave velocity model is compromised by the moderate-low ray paths coverage. However, an extensive low-velocity anomaly proximal to the Sea of Japan (NLV in Figure 7a) is evident. This low-velocity anomaly is attributable to the presence of the Tango-Tajima terrain, which is composed of the Neogene volcanic and sedimentary series (Matsushita, 1963). The high velocities on the southeastern side of the study area (around the MTL) may be indicating the presence of the zonally arranged Sanbagawa metamorphic terrain, which consists of the dynamo-thermally metamorphosed Paleozoic, the Chichibu terrain, which is composed chiefly of Paleozoic and fossiliferous Mesozoic, and the Hidaka terrain with undivided Mesozoic and scanty fossils (Figure 1a).

A prominent elongated NE–SW trending low-velocity anomaly occurring at the boundary between the high-velocity anomalies denoted by NM and KM (Figures 7f and 9b) is observed. This low-velocity anomaly coincides with the Niigata–Kobe Tectonic Zone (NKTZ, Figure 1a) and the Biwako-seigan Fault Zone on the western shoreline of Lake Biwa (BSFZ; Figures 9b and 11a). The BSFZ is constituted of the NNE–SSW-trending west-dipping faults separeated by clear small gaps or steps (e.g., the Chinai, Aibano, Kamidera, Katsuno, Hira, Katata, Hiei, and Zeze faults; Figure 11a), and is reported to have a reverse fault sense of east side subsidence (Takemura et al., 2013). The location of some members of the BSFZ clearly coincide with the boundary between the low-velocity zone (Lake Biwa) and the high-velocity zone in our topography-corrected S-wave velocity models (Hira-Hiei mountains; Figure 11). On the western side of Mt. Hira, the observed NNE-SSW trending linear low-velocity anomaly is consistent with the location of the Hanaore Fault (HOF; Figure 11a), which is a ~50-km-long, right-lateral strike-slip Fault (Noda & Shimamoto, 2009). However, the effects of BSFZ and HOF are not clearly visible in S-wave velocity models without topography correction (Figure S7).

Both the western and eastern sides of the NKTZ are characterized by conspicuous fault systems, with some major faults running through geological units, such as the Yagi-Yabu faults (YGF-YBF) and the Mitoke Fault (MTF) (Mogi et al., 1991). The intervening spaces between fault pairs such as the YGF–YBF and MTF faults are often located in the terrace and the alluvial plain (Katsura, 1990). In our results, the low-velocity anomaly observed between the YGF–YBF and the MTF (Figure 11) probably represent sedimentary units within and around the Fukuchiyama basin (FB), but may also be indicating a possibility of the existence of active faults interconnecting these fault pairs. Besides the gaps between pairs of known active faults, several narrow, elongated low-velocity anomalies are identified, which, to some extent, coincide well with the locations of known active faults. Nonetheless, there are cases where the observed elongated low-velocity anomalies and the locations of known active faults do not correspond (Figures 10-11). Such elongated low-velocity anomalies are largely trending to the NW–SE and NE–SW directions. We posit that some of these linear low-velocity zones are likely to be attributable to the weathering effects and sediments associated with the activity of undocumented concealed faults or fault zones. On the southern side of the Kinki region, alternating linear low- and high-velocity zones are pronounced (demarcated by a blue curly bracket in Figure 10b), attributable to the presence of sediments within meridional deep valleys and moutain ranges, respectively (Matsushita, 1963).

Distinct low-velocity anomalies occur at the Sanda basin (SB), FB, Osaka basin (OB), Nara basin (NB), Kyoto basin (KB), Oomi basin (OoB), and the Ise basin (IB) (Figure 9). The OB manifest as a near-elliptical low-velocity zone, with the northern and southern edges of this zone appearing to be oriented ENE–WSW and NE–SW, respectively. The low-velocity values in this area are likely to be representing the Plio-Pleistocene Osaka Group sediments, which are discordantly overlain by terrace and alluvial deposits (Itihara et al., 1997). The ENE–WSW trending northern boundary of the OB coincides with the location of the Arima-Takatsuki Tectonic Line (ATTL; blue line in Figure 9a). The ATTL is reported to be striking in an ENE–WSW direction, nearly parallel to the MTL (Mitchell et al., 2011), and is characterized by a linear fault zone and steep fault surfaces. Based on this notion, the ATTL marks the boundary between high-velocity zones (mountainous regions; e.g., the Hokusetsu Mountains) and low-velocity zones (basins; e.g., the SB and OB in Figure 9a).

According to Hallo et al. (2019), the OB is bounded by two near-parallel reverse faults on its eastern margin, the Uemachi Fault Zone (UFZ) and the Ikoma Fault Zone (IFZ). However, the effect of these fault zones is not clear in our results. Even so, our results show a low-velocity anomaly stretching to deeper parts of the displayed vertical sections (Figure 9c) occurring between known locations of the UFZ and IFZ. This low-velocity anomaly corresponds to a sub-basin of the OB between the elevated areas of Ikoma and Uemachi Upland (Figure 9c), designated the Kawachi plain (Hatayama et al., 1995). The high-velocity basement material exhibits undulating topographic pattern, with some synclinal parts representing depressional areas in which deep sedimentary basins occur and anticlinal parts corresponding to the basement upheavals or mountain ranges (Figures 9c-f and 10c-d). Since surface wave inversion is strongly sensitive to the presence of sediments, the low-velocity anomalies observed at depressional areas are postulated to be representing the prevailing thick sediments (Miyamura et al., 1981; Nakayama, 1996; Takemura, 1985). At the Ise basin (IB, Figure 9e), the high-velocity material appears to have subsided significantly. This subsidence may be reflecting the effects of the Kuwana and the Yokkaichi reverse faults, which form part of the nearly N–S trending Yoro fault system (Research Group for Active Faults of Japan, 1991). Similar discontinuities within the high-velocity material are evident beneath the OB and KB low-velocity material (Figure 9c-f), likely to be representing the effects of the NKTZ and/or BSFZ.

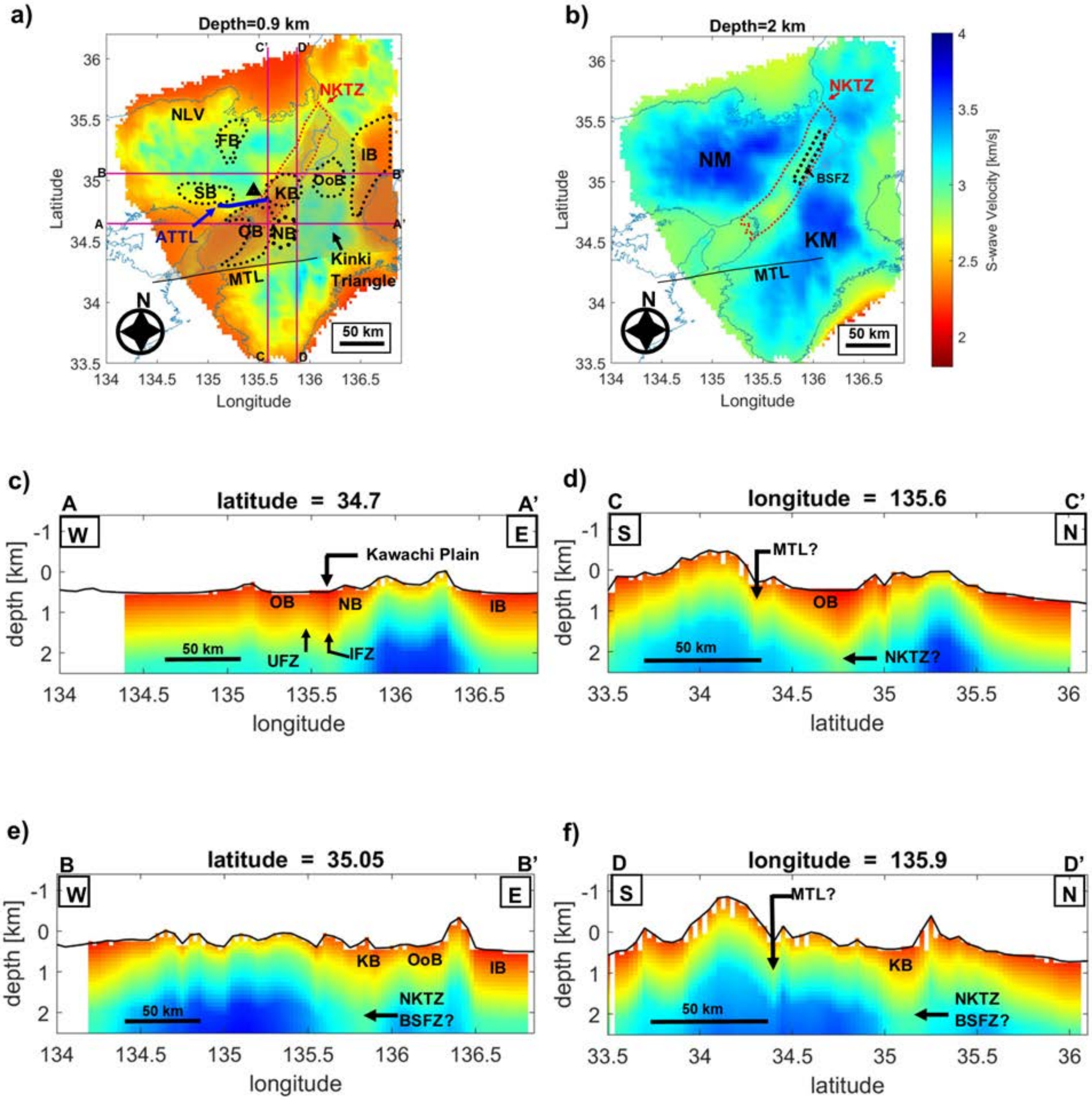

Figure 9. S-wave velocity structure at 0.9 km (a) and 2 km (b) depths below sea level, and vertical slices (c–f) through the S-wave velocity structure along the profiles marked in **a**, showing the variation of S-wave velocity with depth (bottom panels) and their respective elevation models in km (top panels). The solid black line represents the Median Tectonic Line (MTL), the dashed red closed-curve represents the Niigata-Kobe Tectonic Zone (NKTZ), and the thick solid blue line indicates the location of the Arima-Takatsuki Tectonic Line (ATTL). The gray-shaded area represents the Kinki triangle. Black triangle represents the location of Hokusetsu mountains. Thin, dashed black closed-curves show the locations of major sedimentary basins (SB, Sanda basin; FB, Fukuchiyama basin; OB, Osaka basin; NB, Nara basin; OoB, Oomi basin; KB, Kyoto basin; and IB, Ise basin). Also shown on the depth slices are the probable locations of the Kawachi plain, Nara basin (NB), Uemachi Fault Zone (UFZ), Ikoma Fault Zone (IFZ), Biwako-seigan Fault Zone (BSFZ), NKTZ, Osaka basin (OB), and Ise basin (IB).

To assess the seismic activity correlating to the distribution of anomalous zones identified in this study, we superimposed earthquake hypocenters for the period 2001–2012 (Yano et al., 2017) on the S-wave velocity model (Figures 10a, 11b). Numerous earthquake hypocenters are distributed across the high-velocity zone on the western side of Hira mountain (Mt. Hira in Figure 11b). By contrast, hypocenter clusters are evident on the low-velocity zone occurring between the eastern margin of the TA and MA tectonic blocks, western part of the HOF and the northern side of ATTL (Figures 10 and 11). Besides these notable clusters, the northwestern part of the Kinki region has a wide distribution of hypocenters, some of which are aligned in the same trend as elongated low-velocity zones or along the contact between low- and high-velocity zones (Figures 10a and 11b). Some of the linear low-velocity zones which are not consistent with the location of known active faults but exhibiting chains of earthquake hypocenters (Figure 11b) may be representing the weathering effects and sediments associated with the activity of undocumented concealed faults or fault zones.

A dense distribution of earthquake hypocenters is observed in the low-velocity zone along the western part of the Kii Mountainland (blue arrow in Figure 10a). These conspicuous seismic events appear to terminate to the north at the near ENE-WSW trending low-velocity zone, consistent with the location of the MTL. According to Kanamori and Tsumura (1971), increased seismicity to the south of the MTL is related to the regional structural heterogeneities associated with the past activity of the MTL, rather than to the local geological structures.

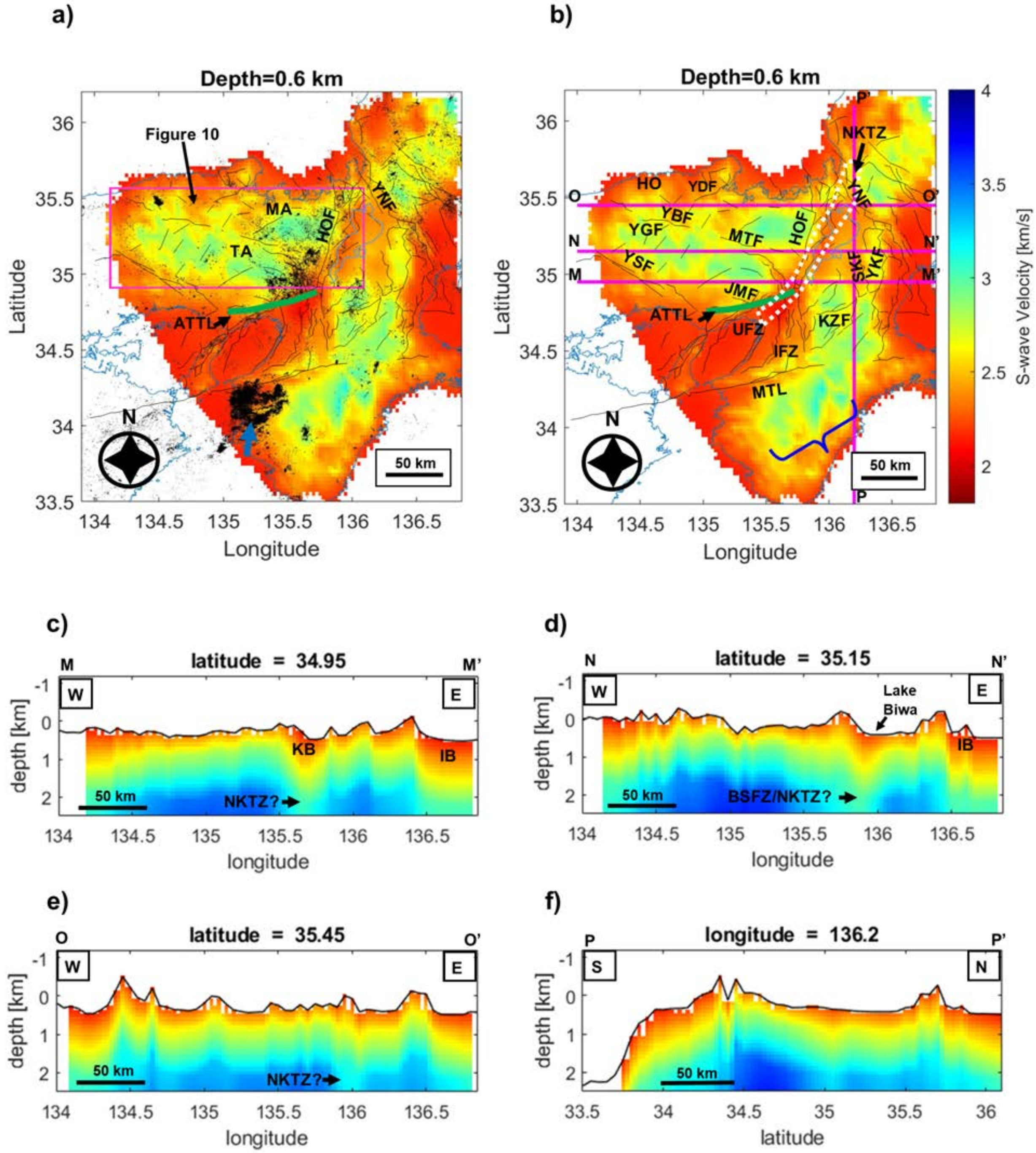

Figure 10. (a) Map of seismic events that occurred between January 2001 and December 2012 (Yano et al., 2017) superimposed on the S-wave velocity model horizontal slice at 0.6 km depth below sea level. Plotted hypocenters (black dots) are for earthquakes ranging from 0 to 6.5 in moment magnitude for depths shallower than 12 km. Blue arrow indicates the location of dense distribution of earthquake hypocenters along the western part of the Kii Mountainland. (b) Distribution of active faults superimposed on the S-wave velocity model horizontal slice at 0.6 km depth below sea level. Solid black lines represent active faults documented prior to this study (Research Group for Active Faults of Japan, 1991). Thick dashed white closed-curve and a solid green line indicate the locations of the Niigata–Kobe Tectonic Zone (NKTZ) and the Arima–Takatsuki Tectonic Line (ATTL), respectively. Also shown are the locations of the Median Tectonic Line (MTL), Yamada Fault (YDF), Yamasaki Fault (YSF),

Jumantsuji Fault (JMF, a member of the ATTL), Yabu Fault (YBF), Yagi Fault (YGF), Mitoke Fault (MTF), Hanaori Fault (HOF), Kizugawa Fault (KZF), Suzuka Fault (SKF), Yokkaichi Fault (YKF), Yanagase Fault (YNF), Uemachi Fault Zone (UFZ), Ikoma Fault Zone (IFZ), Tanba Block (TA), Hokutan Block (HO), and the Maizuru Block (MA). Blue curly bracket marks the location of three alternating meridional deep valleys and mountain ranges. (c–f) Vertical sections of S-wave velocity along the profiles marked as solid magenta lines in Figure 9b. Inferred locations of the Kyoto basin (KB), Ise basin (IB), Lake Biwa and the Biwako-seigan Fault Zone (BSFZ) and/or Niigata-Kobe Tectonic Line (NKTZ) along the profile are also shown on the vertical sections.

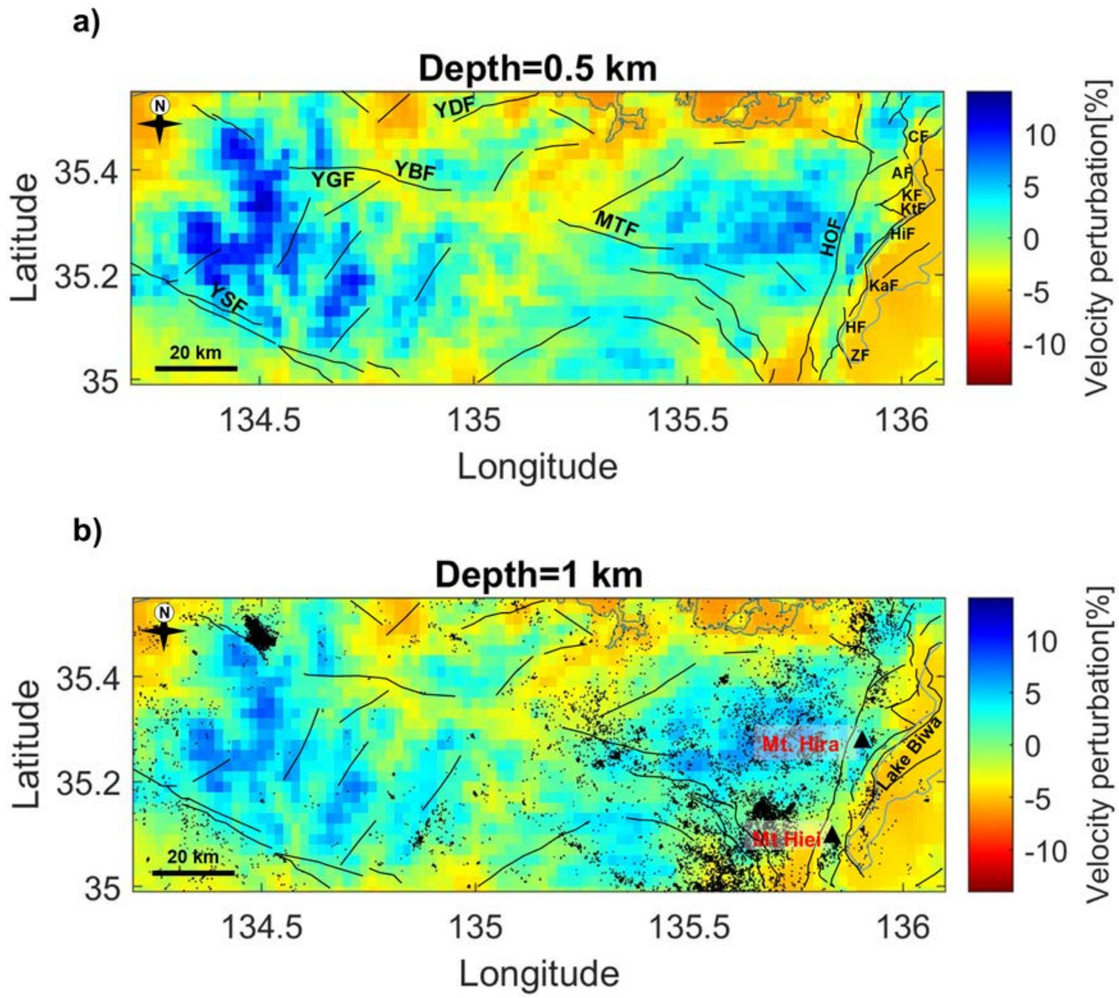

Figure 11. Enlarged view of the northern part of the Kinki region (shown in Figure 9). (a) S-wave velocity perturbation at a depth of 0.5 km below sea level. Also shown are the locations of the Yamada Fault (YDF), Yamasaki Fault (YSF), Yagi-Yabu Fault (YGF-YBF), Mitoke Fault (MTF), Hanaore Fault (HOF) and the Biwako-seigan Fault Zone members (Chinai Fault, CF; Aibano Fault, AF; Kamidera Fault, KF; Katsuno, KtF; Hira Fault, HiF; Katata, KaF; Hiei Fault, HF; Zeze Fault, ZF) (Kaneda et al., 2008). (b) S-wave velocity perturbation at a depth of 1 km below sea level, overlaid with earthquake hypocenters (black dots; Yano et al., 2017) and active faults. Black triangles represent the Hira and Hiei mountains. Solid black lines represent documented active faults (Research Group for Active Faults of Japan, 1991).

6 Conclusions

We constructed a high-resolution shallow 3D S-wave velocity structure of the Kinki region using data recorded by the 221 temporary and permanent seismic stations. We used the zero-crossing method to estimate S-wave phase velocity measurements in the frequency domain. We then applied a direct surface wave tomographic inversion method using high-frequency ambient noise data (0.083–0.67 Hz and 0.05–0.95 Hz). Our results revealed that S-wave velocities vary significantly in the vertical and horizontal directions, which is consistent with the heterogeneous geology of the Kinki region. Conspicuous high-velocity zones are identified in the northwestern and southeastern parts of the study area and are attributed to shallow basement material, mountainous regions, or sedimentary complexes. Sedimentary basins manifest as low-velocity zones. Using horizontal and depth slices of the S-wave velocity model, we estimated the locations of the recently reactivated Niigata-Kobe Tectonic Zone and the highly active Arima-Takatsuki Tectonic Line on the northern edge of the Osaka basin. In the western coast area of Lake Biwa, effects of the active Biwako-seigan Fault Zone are revealed clearly in our results (Figure 8e–f).

We also identified several fine-scale low-velocity tectonic structures, coexisting with known active faults, such as the N–S-, ENE–WSW-, and NE–SW-trending active faults on the eastern side of the Niigata–Kobe Tectonic Zone. In addition, our results revealed elongated low-velocity features that are not consistent with known active faults, likely to be indicating a possible existence of unidentified faults (largely NW–SE- and NE–SW-trending faults) across the Kinki region. These findings allude to the improved resolution of our S-wave velocity model compared with previous studies of the Kinki region. The observed probable concealed fault zones (linear low-velocity anomalies) characterized by aligned distribution of earthquake hypocenters will be useful for hazard assessment and disaster mitigation. The alternating pattern of subsided and uplifted zones observed in vertical slices of our S-wave velocity model is consistent with the tectonic history of the Kinki triangle, which has been dominated by the E–W compressional movement and is also an area characterized by a dense distribution of active faults. These results improve our understanding of shallow crustal structure in the Kinki region. Furthermore, a good correlation between heterogeneities in the S-wave velocity model and the spatial distribution of fault traces and other geologic features in the Kinki region suggests that the approach adopted in this study can be utilized as an effective method for unraveling the complex crustal structure of environments akin to the Kinki region.

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