Validating applicability of strain data calibrated in geodetic range on the seismic frequency band by examining Rayleigh wave amplitude ratio

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Abstract

Although borehole strain observatories provide useful information about crustal deformation, they need to be calibrated because their highly sensitive observations are affected by structural heterogeneity around the sensors used for settlements. The Geological Survey of Japan developed a strain observation network in southwest Japan to monitor slow-slip events in the Nankai Trough. In this study, the vertical axial strain data of this network as well as the horizontal strain of two newly constructed stations were calibrated using the method used in a previous study. Although this calibration was performed in the tidal frequency band, this study confirmed that the calibration is also valid in the frequency band of seismic surface waves at \( \sim 0.01 \) Hz. This fact enables us to use strain data not only for geodetic analysis of crustal deformation but also for the analysis of surface waves from moderate regional earthquakes. This study shows that synthetic strain waveforms are consistent with observed ones, suggesting the usefulness of strain data in centroid moment tensor analysis to improve the resolution of seismic moment tensors.

1. Introduction

Seismic phenomena generate ground deformation not only through translational motion but also strain and rotational motions (Aki & Richards, 2002). Seismometers usually measure the translational motion of the ground and have been used for various analyses. However, strain and rotational motions are expected to provide additional information on seismic sources or subsurface structures. For example, Donner et al. (2016) showed in numerical simulations that centroid moment tensor (CMT) analysis incorporating rotational motion in addition to translational motion improves the resolution of the analysis compared with CMT analysis with the same number of only translational seismic traces. The same is true for strain motion (e.g., Vera Rodriguez and Wuestefeld, 2020). Therefore, much effort has been made to observe strain and rotational deformation.

Although strain observation has recently been the focus of many researches with the advancement of distributed acoustic sensing techniques (e.g., Lindsey and Martin, 2020), strain has been observed for many years using strainmeters settled in boreholes for geodetic measurements. For example, Gladwin (1984) developed Gladwin Tensor Strainmeters (GTSMs) and Ishii et al. (2002) developed Ishii-type strainmeters. These borehole strainmeters are sensitive enough to detect signals of less than 1 nanostrain. Since strain measurements are influenced by local material heterogeneities, such as the cement used for fixing the equipment to bedrock, the calibration of strain data is required to be used for crustal deformation monitoring. Calibration is usually conducted by comparing the tidal components of the observed strain data with the synthetic tidal deformation (King et al., 1979; Gladwin & Hart, 1985; Hart et al., 1996, Roeloffs, 2010; Hodgkinson et al., 2013).

The Geological Survey of Japan (GSJ), National Institute of Advanced Industrial Science and Technology (AIST), has developed a strain observatory network in the southwest of Japan since 2009, which is used to monitor slow slip (Obara et al., 2004) in this region (Itaba et al., 2010). As of 2022, this network
included 4 GTSMs and 14 Ishii-type strainmeters (Table 1). These strainmeters measure horizontal strains along four axes and Ishii-type strainmeters measure additional vertical axial strain. Matsumoto & Kamigaichi (2021) conducted the calibration of 15 strainmeters except for three newly constructed stations by comparing tidal components of observed data with synthetic tidal deformation, where oceanic tidal loadings were accurately incorporated. As a result, they obtained a calibration matrix for horizontal strain tensors for all Ishii-type strainmeters and one GTSM. Calibrated strain data from Ishii-type strainmeters are currently used for geodetic signal analysis of slow slips (e.g., Yabe et al., 2021).
Table 1
Station information for four GTSMs and fourteen Ishii-type strainmeters of GSJ strain observatories (As of 2022)

<table>
<thead>
<tr>
<th>Station name</th>
<th>Latitude (ºN)</th>
<th>Longitude (ºE)</th>
<th>Type</th>
<th>Bed rock</th>
<th>Deployment</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>TYS</td>
<td>35.0405</td>
<td>137.3578</td>
<td>Ishii</td>
<td>Tonalite</td>
<td>2008</td>
<td></td>
</tr>
<tr>
<td>TYE</td>
<td>34.7659</td>
<td>137.4695</td>
<td>Ishii</td>
<td>Mudstone</td>
<td>2004</td>
<td></td>
</tr>
<tr>
<td>NSZ</td>
<td>34.8442</td>
<td>137.1057</td>
<td>Ishii</td>
<td>Pelitic gneiss</td>
<td>2013</td>
<td></td>
</tr>
<tr>
<td>ANO</td>
<td>34.7870</td>
<td>136.4019</td>
<td>Ishii</td>
<td>Granodiorite</td>
<td>2010</td>
<td></td>
</tr>
<tr>
<td>ITA</td>
<td>34.4534</td>
<td>136.3129</td>
<td>GTSM</td>
<td>Tonalite</td>
<td>2008</td>
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</tr>
<tr>
<td>MYM</td>
<td>34.1123</td>
<td>136.1815</td>
<td>Ishii</td>
<td>Granodiorite-porphyry</td>
<td>2008</td>
<td>Broken since 2019 (Not used)</td>
</tr>
<tr>
<td>ICU</td>
<td>33.9001</td>
<td>136.1379</td>
<td>Ishii</td>
<td>Welded tuff</td>
<td>2007</td>
<td></td>
</tr>
<tr>
<td>KST</td>
<td>33.5201</td>
<td>135.8363</td>
<td>Ishii</td>
<td>Sandstone/Mudstone</td>
<td>2008</td>
<td></td>
</tr>
<tr>
<td>HGM</td>
<td>33.8675</td>
<td>135.7318</td>
<td>Ishii</td>
<td>Shale/Sandstone</td>
<td>2007</td>
<td></td>
</tr>
<tr>
<td>HDW</td>
<td>33.8862</td>
<td>135.1988</td>
<td>Ishii</td>
<td>Sandstone/Shale</td>
<td>2022</td>
<td>Wait for data stabilization (Not used)</td>
</tr>
<tr>
<td>ANK</td>
<td>33.8661</td>
<td>134.6045</td>
<td>GTSM</td>
<td>Sandstone</td>
<td>2008</td>
<td>Not used</td>
</tr>
<tr>
<td>MUR</td>
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<td>134.1563</td>
<td>Ishii</td>
<td>Mudstone</td>
<td>2008</td>
<td></td>
</tr>
<tr>
<td>SSK</td>
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<td>133.3229</td>
<td>Ishii</td>
<td>Mudstone</td>
<td>2010</td>
<td></td>
</tr>
<tr>
<td>KOC</td>
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<td>GTSM</td>
<td>Sandstone</td>
<td>2008</td>
<td>Not used</td>
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<tr>
<td>NHK</td>
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<td>Ishii</td>
<td>Granodiorite</td>
<td>2013</td>
<td></td>
</tr>
<tr>
<td>MAT</td>
<td>33.8422</td>
<td>132.7393</td>
<td>GTSM</td>
<td>Granodiorite</td>
<td>2008</td>
<td>Not used</td>
</tr>
<tr>
<td>TSS</td>
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<td>132.9757</td>
<td>Ishii</td>
<td>Granite</td>
<td>2008</td>
<td></td>
</tr>
<tr>
<td>UWA</td>
<td>33.3859</td>
<td>132.4823</td>
<td>Ishii</td>
<td>Mudstone</td>
<td>2008</td>
<td></td>
</tr>
</tbody>
</table>

Since strain is effective data in the geodetic range, it should also be useful in the seismic frequency band. Takeda et al. (2011) reported surface waves of the 2011 Tohoku earthquake (Mw 9) recorded in Ishii-type strainmeters of the GSJ. They showed that the vertical strain of surface waves is proportional to vertical translational velocity seismograms of nearby broadband seismometers (F-net; National Research
Institute for Earth Science and Disaster Resilience, 2019). Okubo et al. (2004a, 2004b) also reported the proportionality between strain seismogram and broadband translational velocity seismogram. This proportionality is valid because both types of seismograms are determined by the second derivative of source time function (Okubo et al., 2004a). When using strain data for seismic analysis, one may wonder whether the calibration matrix derived from tidal frequency band is still effective at higher frequency bands of seismic waves. Matsumoto & Kamigaichi (2021) validated this question by comparing surface waves of the 2010 Chile (Maule) earthquake (Mw 8.8) observed in calibrated strain data with synthetic strain surface waves calculated with the Preliminary Reference Earth Model (PREM) structure (Dziewonski and Anderson, 1981). In this study, we conducted an additional validation for this question by comparing calibrated strain data of GSJ with the nearby translational velocity data of F-net. The amplitude ratio of these components was calculated by assuming that Rayleigh waves from distant earthquakes can be approximated as a plane wave. The observed amplitude ratio was compared with synthetic amplitude ratio to validate the strain calibration in seismic frequency band.

In the following manuscript, we have shown that the calibrated strain data of the GSJ is useful in the seismic frequency band as well. The next section introduces the data used in this study. In Section 3, we have calibrated the vertical strain of Ishii-type strainmeters in the GSJ strain network as well as the horizontal strain of the two newly constructed strainmeters using the method used by Matsumoto & Kamigaichi (2021). In Section 4, we have derived analytical expressions of the amplitude ratio between the strain and translational velocity of Rayleigh waves from far-field earthquakes in a homogeneous half-space elastic body, and compared them with the observed amplitude ratio. In Section 5, examples of strain surface waves from near-field earthquakes are presented.

2. Data

In this study, we have used strain data from the Ishii-type strainmeters in the GSJ network (Fig. 1a). The strainmeters were settled at the bottom of the boreholes at a depth of approximately 600 m. The strainmeters measured four horizontal axial strains, which were separated by 45° from each other, and one vertical axial strain. Among the 14 Ishii-type strain observatories, the Station TYE is the oldest, which was constructed in 2004. Ten of these were constructed in 2007–2010, and two (Stations NSZ and NHK) were constructed in 2013 (Table 1). We did not use Station MYM in the analysis because it broke down in 2019. Station HDW, which was constructed in 2022 and has just started observations, was also excluded from the analysis because we had to wait until the cement used to settle the sensor in the borehole solidified. Therefore, we used 12 out of 14 Ishii-type strain observatories (Fig. 1a and Table 1). We also used translational velocity data from broadband F-net stations for comparison with strain data (Fig. 1a). Strain and translational velocity data were acquired using 20 and 100 Hz sampling, respectively. Only the strain Station TYE was acquired with 50 Hz sampling. These data were low-passed below 0.1 Hz and down-sampled to 1 Hz. Then, the instrumental response of the nearby F-net station was convolved with the strain data when comparing the strain and translational velocity data.
Velocity logging was conducted while constructing the 600 m depth borehole. We used the harmonic average of the logging data to define the average P- and S-wave velocities ($c_p$ and $c_s$, respectively) at the station, except for Station TYE (Fig. 1b). At Station TYE, only P-wave velocity logging was conducted, although the digital data were not preserved. Hence, we assumed a P-wave velocity of 5.0 km/s and an S-wave velocity of 2.5 km/s at this station, referring to values at neighboring stations (Stations NSZ and TYS).

### 3. Calibration For Vertical And Horizontal Strain

Matsumoto and Kamigaichi (2021) obtained calibration matrices of horizontal strain for strain meters in GSJ networks, except for the Stations NSZ and NHK. However, they did not obtain calibration coefficients for the vertical axial strain of Ishii-type strainmeters. We conducted calibrations for the vertical axial strain of all stations and the horizontal strain of the Stations NSZ and NHK using the same method as that used by Matsumoto and Kamigaichi (2021). Strain data are composed of signals from tidal deformation, atmospheric pressure response, and other deformations (such as earthquakes). We extracted the $M_2$ and $O_1$ tidal components using observed data (November 2021–March 2022 for vertical strain, May 2015–April 2016 for horizontal strain of Station NHK, and May 2014 – August 2014 for horizontal strain of Station NSZ) using the BAYTAP-G program (Tamura et al., 1991). The observed tidal deformations were compared with those predicted using the modified GOTIC2 program (Kamigaichi et al., 2020, 2021). For a detailed methodology for the calibration of vertical and horizontal strain, refer to Appendices A and B, respectively, as well as the study by Matsumoto and Kamigaichi (2021). The calibration coefficients obtained for the vertical strain and the calibration matrices for the horizontal strain are summarized in Tables A1 and B3, respectively.

The quality of the estimated calibration coefficients for vertical strain can be assessed by examining the phase lags between the observed and synthetic tides and the stability of the coefficients for the $M_2$ and $O_1$ tides. Phase lags were within a few degrees at most of the stations examined. However, the phase lags are moderately large at Stations MUR and UWA and markedly large at Station ICU. Therefore, the calibration coefficients for the vertical strain at these three stations should be carefully treated. Among other stations, calibration coefficients estimated from the $M_2$ and $O_1$ tides are coincident with each other, except for Station KST, where the calibration coefficient from the $O_1$ tide is approximately 10 times larger than that from the $M_2$ tide. Therefore, the calibration coefficient for vertical strain at this station should be also treated carefully. The estimated coefficients of horizontal strain were validated using the off-diagonal components of the perturbation matrices P (Appendix B; Matsumoto & Kamigaichi, 2021).

Matsumoto and Kamigaichi (2021) showed that the off-diagonal components of perturbation matrices at Station MUR have relatively large values compared with those at Stations TYS, TYE, ANO, MYM, ICU, HGM, KST, SSK, TSS, and UWA. The off-diagonal components of the perturbation matrices at Stations NSZ and NHK, which were newly obtained, were also relatively large compared to those at the 10 stations.

### 4. Amplitude Ratio Of Surface Waves
In this section, we first derive an analytical expression for the amplitude ratio of Rayleigh waves by assuming plane wave propagation on a homogenous half-space elastic body. Subsequently, the amplitude ratios and phase differences between the strain and translational velocity waveforms are calculated from observations and compared with synthetics.

4.1. Synthetic amplitude ratio

Here, we consider the plane wave of a Rayleigh wave propagating through a homogeneous half-space elastic body. $x_1$ and $x_2$ axes were taken in the horizontal and vertically down directions, respectively. Rayleigh waves are assumed to propagate toward the $-x_1$ direction. In these settings, the amplitudes of the Rayleigh waves are expressed as

$$u_1 = A \exp(-bx_2) \exp[i k (x_1 + ct)]$$

and

$$u_2 = B \exp(-bx_2) \exp[i k (x_1 + ct)]$$

where $A$ and $B$ are the amplitudes, $b$ is the skin depth, $k$ is the wavenumber, and $c$ is the Rayleigh wave velocity. Substituting equations (1) and (2) into an equation of motion for an elastic body, nonzero amplitudes $A$ and $B$ can exist in the following two modes: For mode 1,

$$b_1 = k \sqrt{1 - \frac{c^2}{c_s^2}}$$

and for mode 2,

$$b_2 = k \sqrt{1 - \frac{c^2}{c_p^2}}$$
In addition, the following relationship is required to satisfy the free-surface boundary condition at the ground surface:

\[ \frac{B_2}{A_2} = i \sqrt{1 - \frac{c^2}{c_p^2}} \]

Summarizing these conditions, the translational velocity and strain of Rayleigh waves can be expressed as:

\[ v_1 = i kc \frac{c^2}{2c_s^2} A_2 \exp [ik(x_1 + ct)] \]

\[ v_2 = kc \left( \frac{\frac{c^2}{2c_s^2} - 1}{\sqrt{1 - \frac{c^2}{c_s^2}}} + \sqrt{1 - \frac{c^2}{c_p^2}} \right) A_2 \exp [ik(x_1 + ct)] \]

\[ \epsilon_{11} = i k \frac{c^2}{2c_s^2} A_2 \exp [ik(x_1 + ct)] \]

\[ \epsilon_{22} = -ik \left( \frac{\frac{c^2}{2c_s^2} - \frac{c^2}{c_p^2}}{\sqrt{1 - \frac{c^2}{c_s^2}}} + \sqrt{1 - \frac{c^2}{c_p^2}} \right) A_2 \exp [ik(x_1 + ct)] \]
Here, we note that the sign of the vertical translational velocity $v_2$ is taken as positive for the vertically up direction to make comparison with observations easier. The Rayleigh wave velocity $c$ was determined using the following equation:

$$
\left(2 - \frac{c^2}{c_s^2}\right)^2 - 4\sqrt{1 - \frac{c^2}{c_p^2}} \sqrt{1 - \frac{c^2}{c_s^2}} = 0
$$

Therefore, amplitude ratios are written as follows,

$$\frac{\epsilon_{11}}{v_1} = \frac{1}{c}$$

$$\frac{\epsilon_{22}}{v_1} = -\frac{1 - \frac{2c_s^2}{c_p^2}}{c}$$

$$\frac{\epsilon_{22}}{v_2} = -\frac{2c_s^2}{c^2} \left(\sqrt{1 - \frac{c^2}{c_p^2}} - \frac{1}{2} \frac{2 - c^2/c_s^2}{\sqrt{1 - c^2/c_s^2}}\right)$$

$$\frac{\epsilon_{22}}{\epsilon_{11}} = -\left(1 - \frac{2c_s^2}{c_p^2}\right)$$

4.2. Observed amplitude ratio

Here, we calculated the amplitude ratios and phase differences of Rayleigh waves from far-field earthquakes in the strain and translational velocity data. We selected seven earthquakes larger than Mw 8.0 and shallower than 100 km depth that occurred in 2013–2021 (Fig. 2) listed in the Advanced National Seismic System (ANSS) Comprehensive Earthquake catalog (ComCat; U.S. Geological Survey, 2017). For
each far-field earthquake in Fig. 2, we calculated the back azimuths between stations and events and rotated the two horizontal components into radial and transverse components so that the horizontal axis matched the settings assumed in Section 4.1. We then used the radial ($\epsilon_{11}$ and $v_1$) and vertical ($\epsilon_{22}$ and $v_2$) components of the strain and nearby translational velocity data. We set time windows of 1200 s around Rayleigh waves from far-field earthquakes. The beginning of the time windows was set at 180 s before a travel time of 4 km/s. When comparing the time windows of the strain data and translational velocity data of the nearby F-net station, we calculated the cross-correlation functions of the radial waveforms to shift the time windows of the translational velocity data by adjusting the relative travel time differences between the strain and F-net stations. FFT was applied to these time windows, and amplitude ratios and phase differences at various frequencies were calculated from the complex amplitude ratio between components.

Figure 3 shows an example of the amplitude ratios and phase differences calculated for the Station HGM and the 2015 Mw 8.3 Chile earthquake. We tested the amplitude ratios and phase lags shown in Equations (13)–(16): The phase differences obtained were consistent with the synthetic predictions in the surface wave frequency band (below 0.01 Hz). Amplitude ratios were also obtained stably below 0.01 Hz, although there were slight differences from the synthetic predictions. Figure S1 shows the results for the other stations for the same earthquake. Amplitude ratios and phase differences are stably estimated below 0.01 Hz at other stations as well. Figure 4 compares the observed and synthetic amplitude ratios and phase differences at Station HGM for all far-field earthquakes shown in Fig. 2. Here, the amplitude ratios and phase differences are averaged between 0.003 Hz and 0.01 Hz for each event (amplitude ratios are averaged on a logarithmic scale). The observed amplitude ratios were consistent for different events, which suggests the temporal stability of the strain data. Figure S2 shows the average amplitude ratios and phase differences at the other stations. The observed amplitude ratios and phase differences were stable, except for Station NHK. At this station, the amplitude ratios and phase differences of the first two earthquakes differed from those of the subsequent events. As Station NHK was constructed in 2013, we inferred that the data immediately after the settlement of the borehole strainmeter were affected by the solidification of the cement used for fixing the equipment to the bedrock. It is likely that the data for the two far-field earthquakes at this station are unstable owing to this effect. Although Station NSZ was also constructed in 2013, there was no evidence of instability in the observed amplitude ratios, as there was with Station NHK, was not observed. Figure 5 shows a comparison between the observed and synthetic amplitude ratios of $\epsilon_{22}/\epsilon_{11}$. At almost all the stations, the observed and synthetic amplitude ratios were almost consistent within a factor of two. Although they do not agree within one sigma, we can consider other sources of inconsistency. For example, we neglected spatial variations in the seismic structure and used a half-space homogeneous medium for synthetic calculations. Furthermore, there should be uncertainty in the P- and S-wave velocity values used in our synthetic calculation. We used these values based on logging data averaged over 600 m depth (Fig. 1b). However, the skin depth (the inverse of $b_1$ or $b_2$ in Section 4.1) is longer than a few kilometers, which is much larger than the depth of the boreholes. Therefore, the velocity values used in this study may not accurately represent the average values at which Rayleigh waves are affected. Only at the Station ICU was there a substantial difference.
between the observed and synthetic amplitude ratios. As documented in Section 3, the calibration quality of the vertical axial strain at the Station ICU is poor. A calibration coefficient that is too small for the vertical axial strain could be a reason for this large difference. Although the calibration coefficients at the Stations KST, MUR, and UWA are also questioned in Section 3, the observed amplitude ratios are consistent with the synthetic ones, suggesting that the calibration coefficients at these stations are suitable. Figure S3 shows a comparison between the observed and synthetic amplitude ratios of the other components. The ratios for $\epsilon_{11}/v_1$ were consistent with each other despite the different locations of the sensors, which could be another source of inconsistency. Although Stations MUR, NSZ, and NHK may have relatively worse quality calibration matrices, this result suggests that the horizontal strains for all stations were well calibrated. The ratios for $\epsilon_{22}/v_1$ and $\epsilon_{22}/v_2$ at all stations were similar to those for $\epsilon_{22}/\epsilon_{11}$ (Fig. 5), with slightly larger scattering.

5. Examples Of Strain Surface Waves From Near-field Earthquakes

Figure 6 depicts strain seismograms at Station HGM for the Mw 5.1 earthquake that occurred beneath Kii Peninsula on 2021/12/03, according to the F-net catalog (Fig. 1). We calculated synthetic strain seismograms from the earthquake using “Computer Programs in Seismology” (Hermann, 2013) and CMT information in the F-net catalog. One-dimensional seismic structure of F-net (Kubo et al., 2002) was used for this calculation. Although the observed and synthetic strain data are slightly different in terms of both amplitudes and phases, they are generally in good agreement. This coincidence suggests that strain data could be useful for CMT analysis in addition to translation data, which is usually used for such analysis. Good coincidence between observed and synthetic seismograms is confirmed at other stations as well (Figure S4). The development of joint CMT analysis using translation and strain data will be addressed in the future.

6. Conclusion

The strain observation network developed by the GSJ records broadband deformation from seismic frequency bands to the geodetic range. Due to their high sensitivity, calibration is required for strain data, and Matsumoto and Kamigaichi (2021) conducted calibration by comparing observed and synthetic horizontal strain data of tidal deformation. In this study, the vertical axial strain of all stations and the horizontal strain of newly constructed stations were determined using the same method. In addition, this study confirmed that the calibration in the tidal frequency band is also valid in the frequency bands of the surface waves (~ 0.01 Hz). Using Rayleigh waves from far-field earthquakes, we compared the observed and synthetic amplitude ratios of the translation and strain surface waves. We showed that they are consistent with each other, implying that the strain data from GSJ are available for seismic waveform analysis as well. Furthermore, we provided an example of the surface waves generated by moderate near-field earthquakes. This shows that the observed strain seismograms are consistent with synthetic strain seismograms based on F-net moment tensor information and 1D structure, suggesting that the strain
seismograms can be used to improve constraints on the seismic source information of regional earthquakes.

**Abbreviations**

CMT  
Centroid Moment Tensor  
GTSMs  
Gladwin Tensor Strainmeters  
GSJ  
Geological Survey of Japan  
AIST  
National Institute of Advanced Industrial Science and Technology

**Declarations**

**Ethics approval and consent to participate**

Not applicable.

**Consent for publication**

Not applicable.

**Availability of data and materials**

Strain data from the AIST are available from the corresponding author upon request. The calibration matrices and coefficients are presented in the appendix (Tables A1 and B3).

**Competing interests**

The authors declare that they have no competing interests.

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**Authors’ contributions**

N. M. calibrated the strain. S. Y. analyzed the surface waves of the calibrated strain and F-net data. K. I. contributed to the study methodology. S. Y. drafted the manuscript, and all the authors read and approved the final manuscript.
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17. Matsumoto N, Kamigaichi O (2021) Tidal Calibration of Multicomponent Borehole Strainmeters and Validation of the Method Using Surface Waves, Research square. https://doi.org/10.21203/rs.3.rs-602970/v1


Figures

Figure 1

Distributions and seismic velocities of strain observatories in GSJ network. (a) Spatial distribution of strain observatories used in this study is shown by red triangles. Strain observatories not used are shown by gray triangle or squares. Broadband seismometers used in this study are shown by blue squares. Black lines show pairs of strain observatories and broadband seismometers used for amplitude ratio calculation in this manuscript. Distance of each pair are written in the map. A beach ball in this map shows CMT solution by F-net for an Mw 5.1 earthquake on 2021/12/03, mentioned in Section 5. (b) Average P- and S-wave velocities at strain observatories derived from logging data. Standard deviations are shown by gray lines.
Figure 2

Map for far-field earthquakes used in this study. A red triangle represents location of strain observatories used in this study. Red stars are far-field earthquakes used in this study. Origin time (UTC) and magnitude of these earthquakes are written in the map, which is based on ANSS Comprehensive Earthquake catalog (ComCat).
Figure 3

Amplitude ratios and phase differences at Station HGM for the 2015/09/16 event. (a) Amplitude ratio spectra of four combinations of strain and translational velocity data, which are shown with different colors. Solid curves and dashed lines are observations and synthetic values, respectively. (b) Phase difference spectra of four combinations of strain and translational velocity data, which are shown with the same colors used in (a). Arrows on the right represent synthetic phase differences.
Figure 4

Average amplitude ratios and phase differences at Station HGM for seven far-field earthquakes. (a) Average amplitude ratios for seven different earthquakes. Dates on the horizontal axis represent hypocenter time of earthquakes (see Figure 2). Four combinations of strain and translational velocity data are shown with different colors. Dots and dashed lines are observations and synthetic values, respectively. Vertical bars on dots represent standard deviation of averaged data. (b) Average phase differences for seven different earthquakes. Dots and dashed lines are observations and synthetic values, respectively.
Figure 5

Comparisons between observed and synthetic amplitude ratio of $\varepsilon_{22}/\varepsilon_{11}$. Gray horizontal bars represent variations in one sigma of observed amplitude ratio among different events. A black solid line represents a line where observed and synthetic values are consistent. A region surrounded by two dashed lines represents observed and synthetic values are consistent within a factor of two.
Figure 6

Strain seismograms from a near-field moderate earthquake (Figure 1) at Station HGM. Four panels show strain seismograms of $\varepsilon_{NN}$, $\varepsilon_{NE}$, $\varepsilon_{EE}$, and $\varepsilon_{ZZ}$ components. Black and red curves are observed and synthetic (based on F-net moment tensor) seismograms, respectively.

Supplementary Files

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- FigureS3.docx
- FigureS4.docx
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- TableB2.pdf
- TableB3.pdf