

Tidal response of seismic wave velocity at shallow crust in Japan

Tomoya Takano¹ and Kiwamu Nishida²

¹Graduate School of Science and Technology, Hirosaki University

²Earthquake Research Institute, the University of Tokyo

Key Points:

- The spatial distribution of seismic velocity changes caused by tides was determined using dense network of seismic stations in Japan.
- The tidal response to velocity changes was extracted from ambient noise using an extended Kalman filter with a Maximum Likelihood method.
- Strain–velocity sensitivities tend to increase at a low S-wave velocity in the shallow crust.

Corresponding author: Tomoya Takano, takanot@hirosaki-u.ac.jp

Abstract

Microcracks in geomaterials cause variations in the elastic moduli under applied strain, thereby creating seismic wave velocity variations. These are crucial for understanding the dynamic processes of the crust, such as fault-zone damage, healing, and volcanic activities. Solid earth tides have been used to detect seismic velocity changes responding to crustal-scale deformations. However, no prior research has explored the characteristics of the seismic velocity variations caused by large-scale tidal deformation. To systematically evaluate the tidal response to velocity variations, we developed a new method that utilized the flexibility of a state-space model. The tidal response was derived from hourly stacked noise autocorrelations using a seismic interferometry method throughout Japan. In particular, large tide-induced seismic velocities were observed in the low S-wave velocity region of the shallow crust. Overall, the tidal responses to velocity variations can provide new insights into the response mechanisms of the shallow crust to applied strain.

Plain Language Summary

Rock deformations can open or close microcracks in rocks along with varying their elastic moduli under an applied strain. The temporal variations in the elastic moduli of rocks alter the seismic wave velocity, which can be monitored to provide information on the strain applied to the crust. This is crucial for understanding the geological processes in the fault zones and volcanic regions. To utilize the seismic velocity variations for monitoring how much the Earth's structure deforms, the response of the seismic velocity to the deformations must be assessed. The deformation of the Earth's surface caused by the gravity of the Moon and Sun, which is called solid Earth tides, has been used to study seismic velocity variations in response to crustal deformation. However, only a limited number of regions have been studied for the tidal response of the seismic velocity, and the characteristics of its variations caused by tidal deformation were not yet apparent. This study measured the tidal responses to seismic velocity variations throughout Japan with reliable estimations. Notably, the tide-induced seismic velocity variations tend to increase in the low S-wave velocity region. Overall, these results provide new insights into the response mechanisms of the shallow crust to deformations.

1 Introduction

The temporal evolution of the stress or strain applied to the crust provides essential information for understanding the dynamic processes in fault zones and active volcanoes. This is because the occurrence of earthquakes depends on the stress or strain state of the Earth and the fluid distribution around the fault. Volcanic eruptions occur because of the pressure accumulation under volcanic fluid pressurization and magma supply from deeper regions. Upon observing the response of the crust to the applied stress or strain, the in-situ stress or strain variations in the crust can be estimated, which generally involve limitations in the case of direct measurement. The geomaterials in the crust are nonlinearly elastic (Walsh, 1965), and their elastic moduli vary with the applied strain. As the seismic wave velocity depends on the elastic moduli, the applied strain induces variations in the seismic wave velocity. Therefore, tracking the seismic wave velocity variations can adequately serve as a proxy for examining the temporal variations in the elastic constants caused by the applied strain in the crust. Previous studies have reported that the temporal variations in the seismic wave velocity are associated with the static strain variations induced by, for instance, large earthquakes (e.g. Brenguier, Campillo, et al., 2008) and volcanic activities (e.g. Brenguier, Shapiro, et al., 2008; Takano et al., 2017). To monitor the applied strain in the crust and its responses, the variations in the seismic wave velocity response to a given strain perturbation must be examined.

As we can precisely compute the static strain caused by a solid earth tide, the seismic velocity variations associated with the tidal strain provide information on the strain-velocity relationships on Earth. Earlier, in controlled active seismic experiments, the seismic velocity observably varied with the tidal strain (e.g. De Fazio et al., 1973; Reasenberg & Aki, 1974; Yamamura et al., 2003). However, active seismic experiments do not yield temporal resolution and constrained locations in repeated experiments. Recently, a passive noise-based technique (e.g. Obermann & Hillers, 2019) observed seismic velocity variations related to tides (e.g. Takano et al., 2014, 2019; Sens-Schönfelder & Eulenfeld, 2019; Hillers et al., 2015; Mao et al., 2019). To estimate the velocity variations caused by tides, two strategies have been employed using ambient noise correlations. The first one involves the stacking of ambient noise correlations according to the tidal deformation amplitude and measuring the phase differences between the noise correlations during the dilatation and contraction of the crust (Takano et al., 2014; Hillers et al., 2015; Takano et al., 2019). After stacking the noise correlations for a long time period, the ve-

76 velocity variations caused by the nontidal effects can be canceled. The second one exam-
77 ines the velocity variations corresponding to the tidal harmonics from the spectrum of
78 seismic velocity variations with high temporal sampling (Sens-Schönfelder & Eulenfeld,
79 2019). These previous studies estimated the strain–velocity sensitivities based on the ve-
80 locity variations induced by tidal deformation at depths shallower than a few kilometers.
81 As such, the estimated magnitudes of strain–velocity sensitivity may depend on the strength
82 of nonlinear elasticity at their location. Previous studies have employed several meth-
83 ods to detect the tidal responses of velocity variations and the spatial sensitivities of wave-
84 fields. The tidal responses to velocity variations have been detected in certain regions.
85 However, no existing research has reported the spatial features of the seismic velocity
86 variations observed in response to crustal-scale deformation. In order to estimate the spa-
87 tial distribution of velocity changes in response to tides, it is necessary to measure the
88 response using a uniform method and establish criteria for determining whether the seis-
89 mic wave velocity responds to tides. This study aims to calculate the tidal response of
90 the velocity variations in the shallow crust throughout Japan for proposing the criteria
91 for detecting these tidal responses.

92 To estimate the variations in only seismic velocity related to tidal strain, the tidal
93 response to the velocity variations must be accurately separated from the other causes
94 of seismic velocity variations. Recently, Nishida et al. (2020) developed a novel method
95 for estimating the seismic velocity variations using an extended Kalman filter based on
96 a state-space model (e.g. Durbin & Koopman, 2012). They utilized an extended Kalman
97 filter algorithm to estimate the seismic velocity variations as a state variable and used
98 the Maximum Likelihood method to estimate the hyperparameters describing the veloc-
99 ity variations related to precipitation and large earthquakes. The flexibility of the state-
100 space model for the time-series data can easily incorporate the seismic wave velocity vari-
101 ations induced by external perturbations into the model. As the period, phase, and am-
102 plitude of the tides were accurately determined in advance, the superposition of the pe-
103 riodic functions can model the tide-induced velocity variations. Thus, the tidal responses
104 to the velocity variations were incorporated into the state-space model as hyperparam-
105 eters to systematically estimate the tidal strain response using the extended Kalman fil-
106 ter and Maximum Likelihood method.

107 In this study, we investigated the seismic velocity variations in response to tidal
108 strains throughout Japan. First, the nine components of the ambient-noise autocorre-

109 lation functions were calculated using Japan’s dense seismic network. Thereafter, we ex-
110 tracted the seismic velocity variations related to the tide from the hourly stacked-noise
111 correlation functions using an extended Kalman filter with a Maximum Likelihood method.
112 The observed strain–velocity sensitivities were compared with the S-wave velocity struc-
113 ture at each station. The spatial distribution of the tide-induced velocity variations was
114 studied to characterize the mechanical properties of the shallow crust in response to de-
115 formation. As the tides and seismic ambient noise can be observed at any location and
116 instant, the flexibility of the state-space model will enable us to attain a higher spatial
117 resolution of the tidal strain–velocity sensitivities with a dense seismic network, such as
118 a Large N-array or Distributed Acoustic Sensing (DAS) observation.

119 **2 Data**

120 To detect the tidal response of seismic wave velocity variations, we computed the
121 autocorrelation functions of ambient noise at a single station using 796 Hi-net seismic
122 stations operated by the National Research Institute for Earth Science and Disaster Pre-
123 vention (NIED). As most of the Hi-net network stations include the same borehole-type
124 sensors, the characteristics among instruments will vary less. The location map of the
125 seismic stations with the maximum tidal volumetric strain at the ground surface com-
126 puted by GOTIC2 (Matsumoto et al., 2001) is illustrated in Figure 1. GOTIC2 computes
127 the tidal strain, including the solid earth tide and ocean load. In the GOTIC2 program,
128 ocean loading was computed with a five-minute resolution around Japan. The NIED de-
129 ployed three-component velocity meters with a natural frequency of 1 Hz at the bottom
130 of each borehole located at a depth of about 100 m or more at most stations. After sub-
131 tracting the common data logger noise (Takagi et al., 2015), the instrumental responses
132 of the seismometers were deconvolved using the inverse filtering technique (Maeda et al.,
133 2011). We resampled the data to 2 Hz to efficiently compute the correlation functions.
134 For each station every hour, we computed three components (north–north, east–east, upward–
135 upward) of an auto-correlation function and six components (east–north, east–upward,
136 north–east, north–upward, upward–east, and upward–north) of a single-station cross-
137 correlation (Hobiger et al., 2014). The correlation functions were filtered at frequency
138 bands of 0.2–0.5 Hz. In summary, we analyzed the ambient noise recorded from January
139 1, 2010, to December 31, 2011.

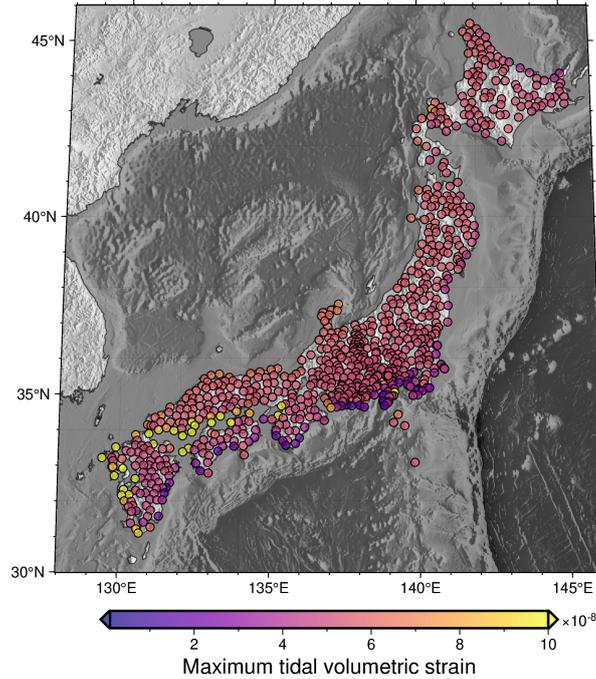


Figure 1. Location map of seismic stations. The color scale displays the maximum volumetric tidal strain during the observation period.

3 Method

The tidal response of seismic velocity variations was determined from the hourly stacked noise correlations using the extended Kalman filter with the Maximum Likelihood method based on the state-space model (Nishida et al., 2020). In Kalman filter processing, we minimized the squared differences between the model correlation function predicted from one previous step and the observed correlations. Assuming that the temporal variations of the seismic wave velocity in a given medium occur homogeneously, a model function of the observed correlations can be expressed by altering the amplitude and stretching factor of the reference correlation function using a stretching method in the time domain (Weaver & Lobkis, 2000). The stretching method has been linearized for application to a Kalman filter (Nishida et al., 2020). The tidal response of the velocity variations was determined as the explanatory variables in a state-space model in two steps. First, the temporal variations of amplitude and the stretching factor of the correlations were estimated as state variables in a state-space model with Kalman Filter processing. Second, the tidal response to the velocity variations was determined as an explanatory variable, referred to as a hyperparameter, using the Maximum Likelihood

156 method. Thus, we constructed a state-space model as follows:

$$157 \quad \mathbf{y}_t^p = \mathbf{m}^p (\boldsymbol{\alpha}_t + \mathbf{R}_t) + \boldsymbol{\epsilon}_t, \quad \boldsymbol{\epsilon}_t \sim \mathcal{N}(0, \mathbf{H}_t) \quad (1)$$

158

$$159 \quad \boldsymbol{\alpha}_{t+1} = \boldsymbol{\alpha}_t + \boldsymbol{\eta}_t, \quad \boldsymbol{\eta}_t \sim \mathcal{N}(0, \mathbf{Q}_t), \quad (2)$$

160 where \mathbf{y}_t^p denotes the data vector of the observed correlations for the p th component,
 161 $\boldsymbol{\alpha}_t$ represents the state vector, \mathbf{R}_t symbolizes the explanatory variable related to the tides,
 162 and $\boldsymbol{\epsilon}_t$ and $\boldsymbol{\eta}_t$ indicate the mutually independent random variables subject to a normal
 163 distribution (\mathcal{N}) with zero means and covariance matrix \mathbf{H}_t and \mathbf{Q}_t , respectively. The
 164 equations 1 and 2 have been elaborately expressed in the supplemental information (re-
 165 fer to Text S1).

166 To compute the reference correlation at each station, we first estimated the state
 167 variables of the stretching factor and amplitude common across all nine components with-
 168 out any explanatory variables. Thereafter, the reference correlation was estimated by
 169 averaging the observed correlations stretched with the estimated amplitude and stretch-
 170 ing factor for the observation duration. Prior data covariance, h_0 , was estimated based
 171 on the time average of the squared differences between the observed and reference cor-
 172 relations. The validation of prior data covariance is described in the supplemental in-
 173 formation (Text S2). The state variables were estimated through the recursive linear Kalman
 174 filter and smoother (Durbin & Koopman, 2012) by adjusting the explanatory variables,
 175 initial stretching factor, and prior model covariance for the initial value.

176 The tidal response of the seismic wave velocity was modeled by adding the cosine
 177 functions related to the tidal constituents as follows:

$$178 \quad r_t = \sum_{m=1}^M A_m \cos(\omega_m t + \varphi_m + \theta_m) \quad (3)$$

179 where m denotes the index of the tidal constituents, A_m indicates the sensitivity of the
 180 seismic wave velocity to tidal strain, φ_m represents the phase angle of the tide, ω_m de-
 181 notes the angular frequency of the tide, and θ_m represents the difference between the tidal
 182 strain and the observed variations in seismic velocity. A phase delay may occur in the
 183 response of the velocity variations to tidal strain caused by the nonlinear elasticity of
 184 the rock (Sens-Schönfelder & Eulenfeld, 2019). φ_m was estimated from the theoretical
 185 tidal strain computed using GOTIC2 at each station, whereas ω_m was obtained from the
 186 table of tidal constituents (Cartwright & Edden, 1973). The velocity variations related
 187 to only M_2 tide were incorporated into the modeled tidal response of the seismic wave

188 velocity, which most significantly contributed to the seismic wave velocity variations in
 189 the tidal constituents (Sens-Schönfelder & Eulenfeld, 2019). As the M_2 tide originated
 190 from the moon, the thermoelastic effects did not contribute to the seismic velocity vari-
 191 ations.

192 The logarithmic likelihood $\ln L$ was maximized with respect to the hyperparam-
 193 eters. $\ln L$ was computed following the Kalman filtering processes (e.g. Durbin & Koop-
 194 man, 2012); $\ln L$ is a function of hyperparameter β as

$$195 \quad \beta = (p_0, p_1, \gamma_1, A_{M_2}, \varphi_{M_2}), \quad (4)$$

196 where p_0 and p_1 represent the covariance of the initial value of the amplitude and stretch-
 197 ing factor, respectively, γ_1 denotes the initial value of the stretching factor, and A_{M_2} and
 198 φ_{M_2} denote the tidal strain-velocity sensitivity and phase shift between the tidal strain
 199 and velocity variations for the M_2 tide, respectively. The covariance of the initial value
 200 was assumed to be equal to that of the prior model. Using the quasi-Newton method (Zhu
 201 et al., 1997), the logarithmic likelihood $\ln L$ was maximized with respect to the hyper-
 202 parameters, searching for tidal strain responses from 0.001 to 1 % and phase shift from
 203 -180° to 180° , respectively. Considering a large covariance of the stretching parameter
 204 creates large short-term variations in seismic wave velocity, which could mask the neg-
 205 ligible velocity variations induced by the tides. We set the search range of p_1 to account
 206 for long-term seasonal variations and short-term tidal responses. In particular, p_1 rang-
 207 ing from 2×10^{-13} to 5×10^{-10} varied the stretching parameters from 0.0001 % to 0.05 %
 208 linearly for three hours, whereas they were estimated up to 1 % for one month. Here we
 209 give an example of a station where the tidal response of velocity change is significant.

210 The time series of the observed velocity variations without the explanatory variables at
 211 the N.SIKH station is presented in Figure 2 (a) and (b). The long-term variations in seis-
 212 mic wave velocity ranged from a few days to tens of days (Figure 2 (a)). Focusing on the
 213 velocity variations caused over a few days, the velocity variations can be observed with
 214 a half-day cycle (Figure 2 (b)). The power spectrum of the seismic wave velocity varied
 215 over two years and was estimated without the explanatory variables. In particular, they
 216 displayed a spectral peak corresponding to the semi-diurnal tidal variation (Figure 2 (c)).
 217 With the Maximum Likelihood method applied to determine the velocity variation re-
 218 lated to the M_2 tide as an explanatory variable, the tidal response was estimated with
 219 statistical reliability. In Figure 2 (d), the spectral peak of the velocity variations during

220 the period of M_2 tide disappeared because the velocity variation caused by M_2 tide was
 221 extracted as the explanatory variable.

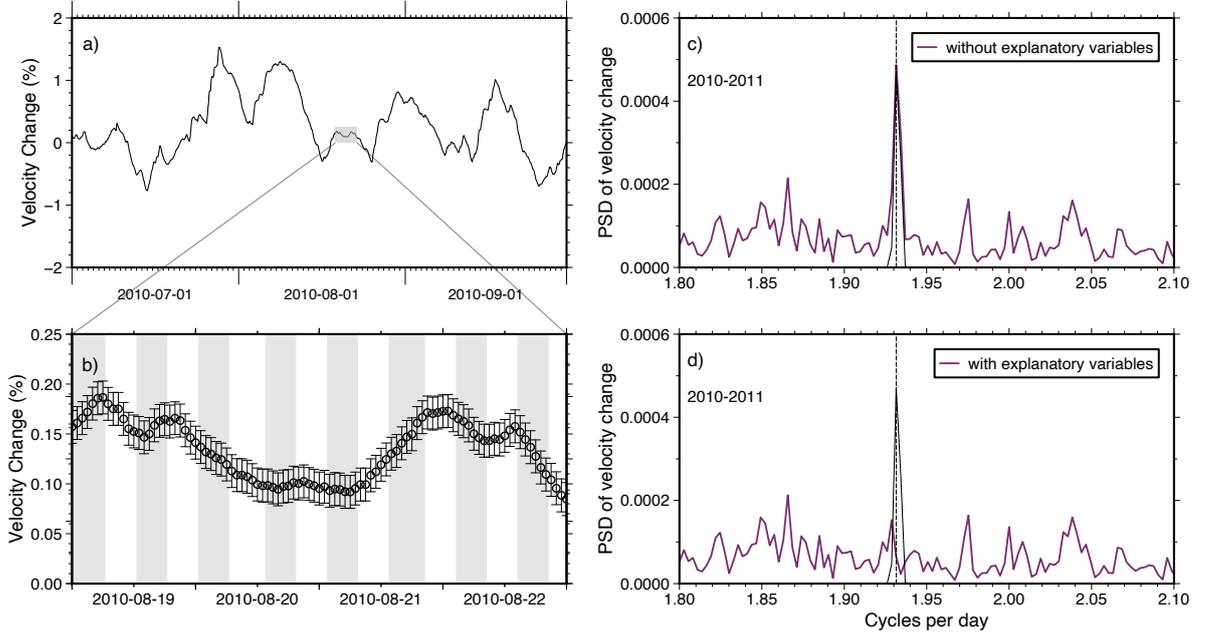


Figure 2. (a) Time series of velocity variations estimated without explanatory variables from 2010/7/1 to 2010/9/30 at N.SIKH station. (b) Enlarged view of the shaded region in Figure (a). The gray region depicts the period of contraction under M_2 tide. (c) The purple line denotes the power spectrum of velocity variations ($\%^2/\text{cycles per day}$). The black line displays the power spectrum of modeled velocity variations. Dashed-black line represents the cycles per day of the M_2 tide. The power spectrum was computed for the observed duration. (d) Power spectrum of velocity variations ($\%^2/\text{cycles per day}$) with the explanatory variables. The black line indicates the power spectrum of modeled velocity variations. Dashed-black line denotes the number of cycles per day of the M_2 tide.

221

222 To evaluate whether the observed velocity variations reliably respond to the tidal
 223 deformation, the appropriate number of hyperparameters was estimated using the *AIC*
 224 (Akaike, 1974) defined as

$$225 \quad AIC_K = -2 \ln \hat{L}_K + 2K \quad (5)$$

226 where K denotes the number of hyperparameters and $\ln \hat{L}_K$ represents the logarithmic
 227 likelihood for K hyperparameters. We compared the *AIC* between the hyperparam-
 228 eters, including the tidal response of the velocity variations. If the increment of *AIC* ($\Delta AIC \equiv$

229 $AIC_K - AIC_{K-2}$) was less than zero, the hyperparameters including M_2 tides were deemed
 230 as appropriate. Among all stations, 56.5% of the stations displayed a reliable tidal re-
 231 sponse to the velocity variations.

232 4 Results

233 The spatial distribution of the velocity variations in response to M_2 tide and the
 234 phase delay of the seismic velocity variations with respect to the tidal strain is illustrated
 235 in Figure 3 (a) and (c), respectively, which were estimated as the hyperparameters us-
 236 ing the Maximum Likelihood method. The stations with AIC increments less than 0 are
 237 displayed in the figure. The velocity variations in response to the M_2 tide were estimated
 238 up to 0.35%. At stations with ΔAIC greater than 0, the velocity variations were gen-
 239 erally estimated as less than 0.001%, indicating that the tidal response to the velocity
 240 variations was not statistically significant (Figure S2). The phase delay of velocity changes
 241 with respect to the M_2 tide is expressed in the supplemental information (refer to Text S4).

242 Based on the velocity variations, dv/v , related to the M_2 tide and maximum vol-
 243 umetric strain of the M_2 tide, ε , on the ground surface computed by GOTIC2, we can
 244 infer the strain-velocity sensitivity, $\frac{dv/v}{\varepsilon}$, at each station. The spatial distribution of strain-
 245 velocity sensitivity is illustrated in Figure 3 (b). The strain-velocity sensitivities varied
 246 from approximately 10^3 to 10^5 . The magnitude of the strain-velocity sensitivity was con-
 247 sistent with previous studies estimating the strain-velocity sensitivity in the shallow por-
 248 tion of the crust (Takano et al., 2014, 2019; Hillers et al., 2015; Sens-Schönfelder & Eu-
 249 lenfeld, 2019). Although the seismic velocity variations at each station were independently
 250 evaluated, the spatial distribution of the tidal response to the velocity variations displayed
 251 a characteristic spatial pattern. In addition, the spatial distributions of the tidal response
 252 to the velocity variations were compared with the geological setting of the Japanese is-
 253 lands. The locations of the active faults obtained from the digital map (Nakata & Imaizumi,
 254 2002) and active volcanoes are presented in Figure 3 (d). First, a large tidal response was
 255 observed in the Kyushu region, where active volcanoes along the Ryukyu arc volcanic
 256 front and the median tectonic line were located. Certain stations in the Shikoku region,
 257 intersecting with the median tectonic line, exhibited a large tidal response to velocity
 258 variations. The regions spanning from central Japan to the Kinki region, where several
 259 active seismic faults have been detected, were characterized by large tidal responses. Al-
 260 though the southern portion of the Chubu region facing the Pacific Ocean exhibits a small

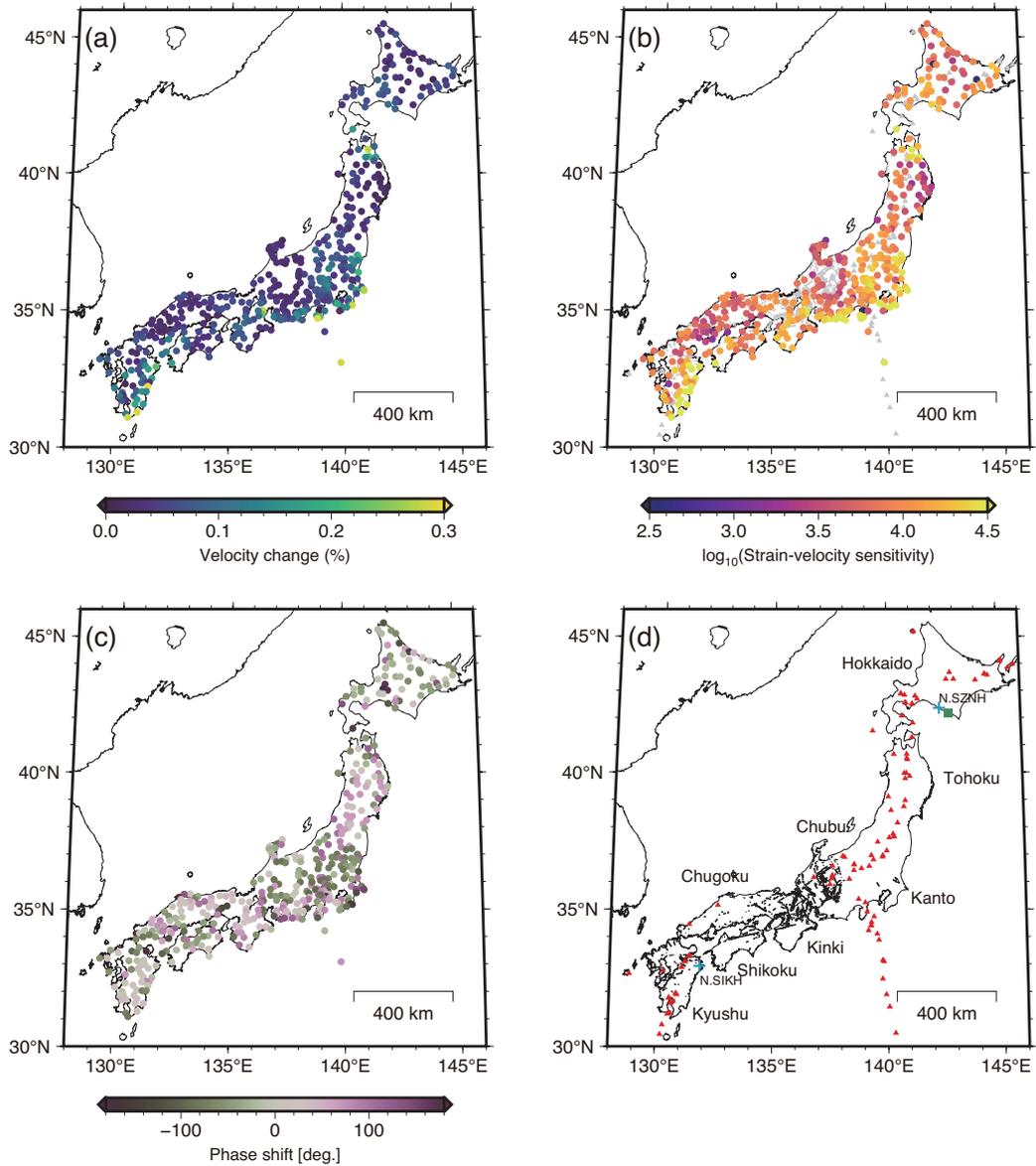


Figure 3. (a) Spatial distribution of velocity variations in response to the M_2 tide. (b) Spatial distribution of tidal strain response of velocity variations. Active volcanoes and faults are plotted in gray color. (c) Spatial distribution of phase delay of seismic velocity variations in response to tidal deformation. (d) Geological features in Japan. Red triangles display active volcanoes; solid-black lines represent active faults (Nakata & Imaizumi, 2002); the blue addition symbol indicates the location of N.SIKH station and N.SZNH station; the green square denotes the location of the tidal gauge. In (a), (b), and (c), the stations with AIC increments less than 0 are plotted.

261 tidal strain, it produced a larger tidal strain response compared to the surrounding area.
262 In Kanto and the eastern portion of the Chubu region, a large tidal response to veloc-
263 ity variations was observed in the area east of the Niigata-Kobe Tectonic Line, wherein
264 a high strain rate was observed to be dominated by a large contraction in the WNW—
265 ESE direction (Sagiya et al., 2000). In the Tohoku region, the back-arc area of the To-
266 hoku region generated a slightly larger tidal response than the island arc area of the re-
267 gion. In particular, the northern tip of the Tohoku region displayed a large tidal response.
268 Moreover, a small tidal response to the velocity variations was observed in the central
269 region of Hokkaido. However, the tidal response to the velocity variations was not strongly
270 correlated with the geological features.

271 To systematically investigate the tidal response characteristics based on the veloc-
272 ity variations, we compared the strain–velocity sensitivity with the S-wave velocity struc-
273 ture estimated from the cross-correlations of microseisms using Hi-net seismic stations
274 (Nishida et al., 2008). According to the depth sensitivity of the ambient noise correla-
275 tions observed in this study, we compared the strain–velocity sensitivity with the S-wave
276 velocity from the elevation at which the sensor was deployed to a depth of 1 km. The
277 strain–velocity sensitivity against the S-wave velocity at each seismic station is presented
278 in Figure 4, wherein the mean of 100 bootstrap-resamplings of strain-velocity sensitiv-
279 ity was plotted with a standard deviation of 0.06 km/s for bins with 50% overlapping
280 in S-wave velocity. For each bin, the average strain–velocity sensitivity increased linearly
281 as the S-wave velocity decreased. However, the tidal strain response to the velocity var-
282 ied considerably. Although the variations in the tidal response at the same S-wave ve-
283 locity may be caused by various geomaterials, the statistical trend of the tidal response
284 suggested certain common physical characteristics. Notably, the tomographic model of
285 S-wave velocity and the autocorrelation function in this study has different spatial res-
286 olution. At certain stations, the region in which the autocorrelation function propagated
287 does not necessarily correspond to the S-wave velocity structure estimated by cross-correlation
288 functions of ambient noise, which may alter the relationship between the strain and ve-
289 locity sensitivity and S-wave velocity.

290 **5 Discussion**

291 We extracted the tidal responses of the seismic velocity variations based on a state-
292 space model. The tidal responses to the seismic velocity variations exhibited a charac-

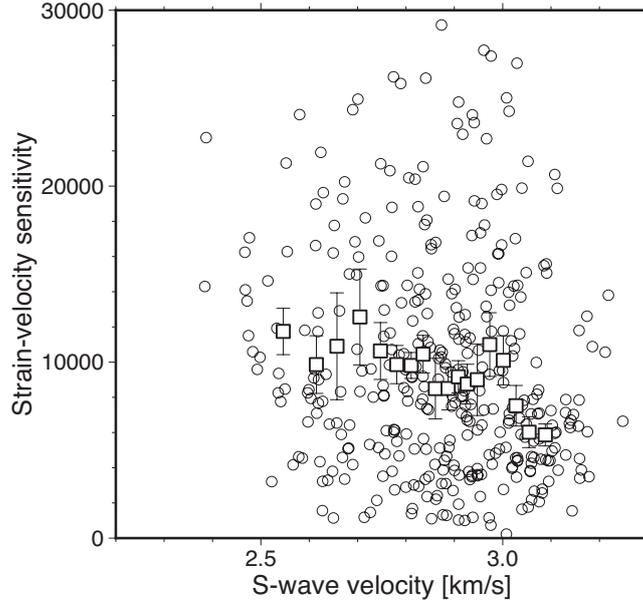


Figure 4. Relationship between S-wave velocity at depth of 1 km, and strain–velocity sensitivity at each station. White circles depict strain–velocity sensitivity at each station, and squares portray the mean of 100 bootstrap-resamplings of strain–velocity sensitivity for bins of 0.06 km/s in S-wave velocity.

293 teristic spatial pattern. In particular, the tidal strain response of velocity variations tended
 294 to increase in the low S-wave velocity regions in the shallow crust. The mechanism through
 295 which the seismic velocities respond to deformations is commonly interpreted as the open-
 296 ing and closing of microcracks in a medium (Walsh, 1965). If the rock strain indicates
 297 a nonhysteresis function of the confining pressure P_c and pore pressure P_o , the strain sen-
 298 sitivity of the velocity variations in the grain material can be formulated assuming a small
 299 aspect ratio, such as follows Shapiro (2003):

$$300 \quad \frac{1}{V_{S_0}} \frac{\partial V_S}{\partial \varepsilon} \sim \frac{1}{2\gamma^2} \phi_{c_0} \exp\left(-\frac{1}{\gamma} CP\right) \quad (6)$$

301 where γ represents the aspect ratio of the pore, V_{S_0} denotes the S-wave velocity in a static
 302 state, ϕ_{c_0} represents the porosity of intergranular pore defined as compliant porosity, C
 303 indicates the drained compressibility, and P represents the effective pressure. The effec-
 304 tive pressure is defined in terms of the pore pressure P_o and confining pressure P_c as fol-
 305 lows:

$$306 \quad P = P_c - P_o. \quad (7)$$

307 As related in Equation 6, the strain sensitivity decreases with the increasing effective
308 pressure. The aspect ratio of the pores or cracks also contributes to the sensitivity based
309 on the squared values. According to Equation 6, the aspect ratio of pores along with the
310 pore pressure and porosity of the intergranular pores contribute to the strain–velocity
311 sensitivity. Thus, a detailed comparison of the strain–velocity sensitivities with the ve-
312 locity structure of the crust is required to investigate the extent to which each factor con-
313 tributes to the observed strain–velocity sensitivities. V_P/V_S in the rocks is sensitive to
314 liquid compressibility, pore geometry, and liquid volume fraction. In contrast, the ratio
315 of the fractional variations in V_S and V_P is sensitive to liquid compressibility and pore
316 geometry (Takei, 2002). Upon comparing the present findings with V_P/V_S and the ra-
317 tio of the fractional variations in V_S and V_P , we can employ the constraints on these crustal
318 parameters.

319 Brenguier et al. (2014) estimated the seismic velocity susceptibility to the dynamic
320 stress induced by the 2011 Tohoku-Oki earthquake. They interpreted the spatial distri-
321 butions of the susceptibilities of seismic velocity based on the highly pressurized fluid
322 situated beneath the active volcanoes in eastern Japan. However, the tidal strain response
323 of the velocity variations surrounding the volcanic front in eastern Japan was not large.
324 According to the surface wave wavelength in the frequency band of 0.2–0.5 Hz or the sen-
325 sitivity kernel of diffused ballistic waves, the wavefield of autocorrelations is sensitive be-
326 tween the surface and a depth of a few kilometers, which is shallower than the depth sen-
327 sitivity reported by Brenguier et al. (2014). The S-wave velocity structure inferred from
328 the cross-correlations of microseisms (Nishida et al., 2008) situated 1 km beneath the vol-
329 canic front is not low in comparison with that of other regions because the spatial res-
330 olution of the correlations does not delineate the small-scale velocity perturbation of the
331 magma chamber (Nagaoka et al., 2012). As the Hi-net stations are sparsely located in
332 volcanic regions, autocorrelation analysis would create a small sample of information be-
333 neath the volcanoes. The lack of a sample of the tidal response of the velocity variations
334 beneath the volcano may create a difference from the stress susceptibility of the veloc-
335 ity variations. Although the stress susceptibility illustrates high-pressure fluid movement
336 activated by the Tohoku-Oki earthquake, the tidal response of the velocity variations ex-
337 hibits the response of the crust to static strain during its quiescent state. Brenguier et
338 al. (2014) assessed the transient response of the crust to the earthquake, whereas the cur-
339 rent results demonstrated the crustal response to the semi-diurnal deformation. The vari-

340 ations between the spatial features of strain–velocity sensitivity and stress susceptibil-
341 ity (Brenquier et al., 2014) suggest various response behaviors of the crust. In the fu-
342 ture, researchers need to consider both the transient and static responses of the crust
343 to more comprehensively understand the mechanical properties of the crust in response
344 to strain or stress.

345 **6 Conclusions**

346 In this study, we examined the seismic velocity variations in response to tides through-
347 out Japan. Utilizing the dense seismic network in Japan, we investigated the spatial ex-
348 tent of the tidal strain–velocity sensitivities. Accordingly, we extracted the tidal responses
349 to velocity variations from the hourly stacked noise autocorrelations by combining the
350 extended Kalman filter with the Maximum Likelihood method. The strain–velocity sen-
351 sitivities varied from approximately 10^3 to 10^5 . Upon comparing the strain–velocity sen-
352 sitivity with the S-wave velocity structure in Japan, the tidal response to seismic veloc-
353 ity variations was larger at low S-wave velocities in the shallow crust. Based on the strain–
354 velocity relationship in the grain material, the current results implied that the spatial
355 variations in the tidal response of seismic wave velocity can potentially characterize the
356 fluid pressure or shape of pores in the crust. The tidal responses to velocity variations
357 in various time periods were extracted to investigate the temporal variations in the me-
358 chanical properties of the shallow crust. Future studies can utilize dense seismic networks
359 such as a Large N-array or DAS observation to attain a higher spatial resolution of tidal
360 strain–velocity sensitivity.

361 **Open Research**

362 We used data from Hi-net (doi.org/10.17598/nied.0003) managed by the National
363 Research Institute for Earth Science and Disaster Prevention (NIED), Japan. The python
364 code of the extended Kalman filter is also available on the Zenodo web page

365 (https://zenodo.org/record/7476091#.Y8_BluzP20p).

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369 This work made use of ObsPy (Beyreuther et al., 2010), Numpy (Van Der Walt et al.,
370 2011) and SciPy (Virtanen et al., 2020), and GMT programs (Wessel & Smith, 1998).

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