

1 Systematic differences in energy radiation processes between regular and inland low-
2 frequency earthquakes in and around the focal area of the 2008 M_w 6.9 Iwate-Miyagi,
3 Japan, earthquake

4 Masaki Orimo^{1*}, Keisuke Yoshida¹, Toru Matsuzawa¹, Taka'aki Taira², Kentaro Emoto³, Akira
5 Hasegawa¹

6 ¹ Research Center for Prediction of Earthquakes and Volcanic Eruptions, Graduate School of
7 Science, Tohoku University, Sendai, Japan

8 ² Berkeley Seismological Laboratory, University of California, Berkeley, California, U.S.A.

9 ³ Institute of Seismology and Volcanology, Kyushu University, Fukuoka, Japan

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11 *Corresponding author: Masaki Orimo (masaki.orimo.s2@dc.tohoku.ac.jp)

12 **Key Points:**

- 13 • We conducted spectral analysis to quantify scaled energy for regular and low-
14 frequency earthquakes (LFEs)
- 15 • The scaled energy of LFEs are one to three orders of magnitude lower than that of
16 regular earthquakes.
- 17 • The local magnitudes for the LFEs show good agreement with the energy
18 magnitudes, but large discrepancy with the seismic moment magnitudes.

19

20

21

22 Abstract

23 Many unknowns exist regarding the energy radiation processes of the inland low-
24 frequency earthquakes (LFEs) often observed beneath volcanoes. To evaluate their energy
25 radiation characteristics, we estimated the scaled energy for LFEs and regular earthquakes
26 in and around the focal area of the 2008 M_w 6.9 Iwate-Miyagi earthquake. We computed
27 the source spectra for regular earthquakes, deep LFEs, and shallow LFEs by correcting for
28 the site and path effects from direct S-waves. We computed the radiated energy and
29 seismic moment, and obtained the scaled energy (e_R) for 1464 regular earthquakes, 169
30 deep LFEs, and 52 shallow LFEs. The e_R for regular earthquakes is in the order of 10^{-5} to
31 10^{-4} , typical for crustal earthquakes, and tends to become smaller near volcanoes and
32 shallow LFEs. In contrast, e_R is in the order of 10^{-7} and 10^{-6} for deep and shallow LFEs,
33 respectively, one to three orders of magnitude smaller than that for regular earthquakes.
34 This result suggests that LFEs are associated with a much lower stress drop and/or slower
35 rupture and deformation rates than regular earthquakes. Although the energy magnitudes
36 derived from radiated energy generally show good agreement with the local magnitudes for
37 the three types of earthquakes, the moment and local magnitudes show a large discrepancy
38 for the LFEs. This suggests that the local magnitude based only on the maximum amplitude
39 of the observed seismic records may not provide good information on the static sizes of LFEs
40 whose e_R values are substantially different from those of regular earthquakes.

41

42 Plain Language Summary

43 Low-frequency earthquakes (LFEs) beneath volcanoes have unique characteristics
44 such as a much lower dominant frequency than regular earthquakes with similar
45 magnitudes and long-lasting trailing parts. Although previous studies have suggested that
46 magma or fluid may be related to the generation of LFEs, the details of seismic wave
47 radiation processes are not as well understood for LFEs. One key parameter for evaluating
48 the energy radiation characteristics is the scaled energy (e_R). Here, we estimated e_R for
49 regular earthquakes and LFEs in and around the focal area of the 2008 M_w 6.9 Iwate-Miyagi
50 earthquake. We found that the e_R values for regular earthquakes were in the order of 10^{-5} ,
51 which is consistent with previous studies. However, the e_R values for the LFEs were one to
52 three orders of magnitude lower than those of regular earthquakes. These results suggest
53 that LFEs are associated with smaller stress drops and/or slower rupture and deformation
54 rates than are regular earthquakes. Furthermore, our results show that owing to this
55 difference in e_R , the local magnitude, which is based on the maximum amplitude of the
56 seismic waves, cannot adequately represent the static sizes of the LFEs.

57

58 1 Introduction

59 A special type of earthquake called a deep low-frequency earthquake (deep LFE) has
60 markedly different characteristics from regular earthquakes. They have a very low dominant
61 frequency compared with regular earthquakes of similar magnitude (e.g., Ukawa and
62 Ohtake, 1987; Hasegawa and Yamamoto, 1994; Obara, 2002; Katsumata and Kamaya, 2003;
63 Rogers and Dragert, 2003). Some occur in the transition zone between the unstable and
64 stable slip zones along the plate boundary (e.g., Obara 2002; Katsumata and Kamaya, 2003;
65 Rogers and Dragert, 2003), while others occur around the Moho discontinuity in the
66 continental plate, often beneath volcanoes (e.g., Ukawa and Ohtake, 1987; Hasegawa and
67 Yamamoto, 1994). The first type of deep LFEs represent shear faulting along the plate
68 boundary (e.g., Shelly et al., 2006, 2007; Ide et al. 2007b), whereas many unknowns exist
69 regarding the seismic wave radiation processes of latter deep LFEs called inland deep LFEs.
70 The Japan Meteorological Agency (JMA) routinely locates regular earthquakes and deep
71 LFEs in and around the Japanese Islands and determines their local magnitudes (M_{jma} ; M_{jma}
72 indicates the JMA magnitude scale) using a nationwide seismic network (e.g., Kamaya and
73 Katsumata 2004). This study focuses on inland deep LFEs.

74 Inland deep LFEs have a very low dominant frequency compared to regular
75 earthquakes with similar magnitudes and characteristic long-lasting high-amplitude trailing
76 parts (e.g., Hasegawa and Yamamoto, 1994). They often occur in or at the margins of low
77 seismic wave velocity regions, which may reflect magma or fluid dehydration from magma
78 (e.g., Hasegawa et al., 1991; Hasegawa and Yamamoto, 1994; Nakajima et al., 2001).
79 Previous studies have suggested that deep LFEs may be related to magmatic or fluid
80 activities such as magma movement (e.g., Aki and Koyanagi, 1981; Ukawa and Ohtake, 1987;
81 Hasegawa and Yamamoto, 1994; Aso et al., 2013), fluid movement (e.g., Hasegawa et al.,
82 2005; Wech et al., 2020), fluid-induced oscillations (e.g., Aki et al., 1977; Julian, 1994; Aso
83 and Ide, 2014), cooling magma (e.g., Aso and Tsai, 2014) and failure along a tensile-shear
84 crack (e.g., Nakamichi et al., 2003).

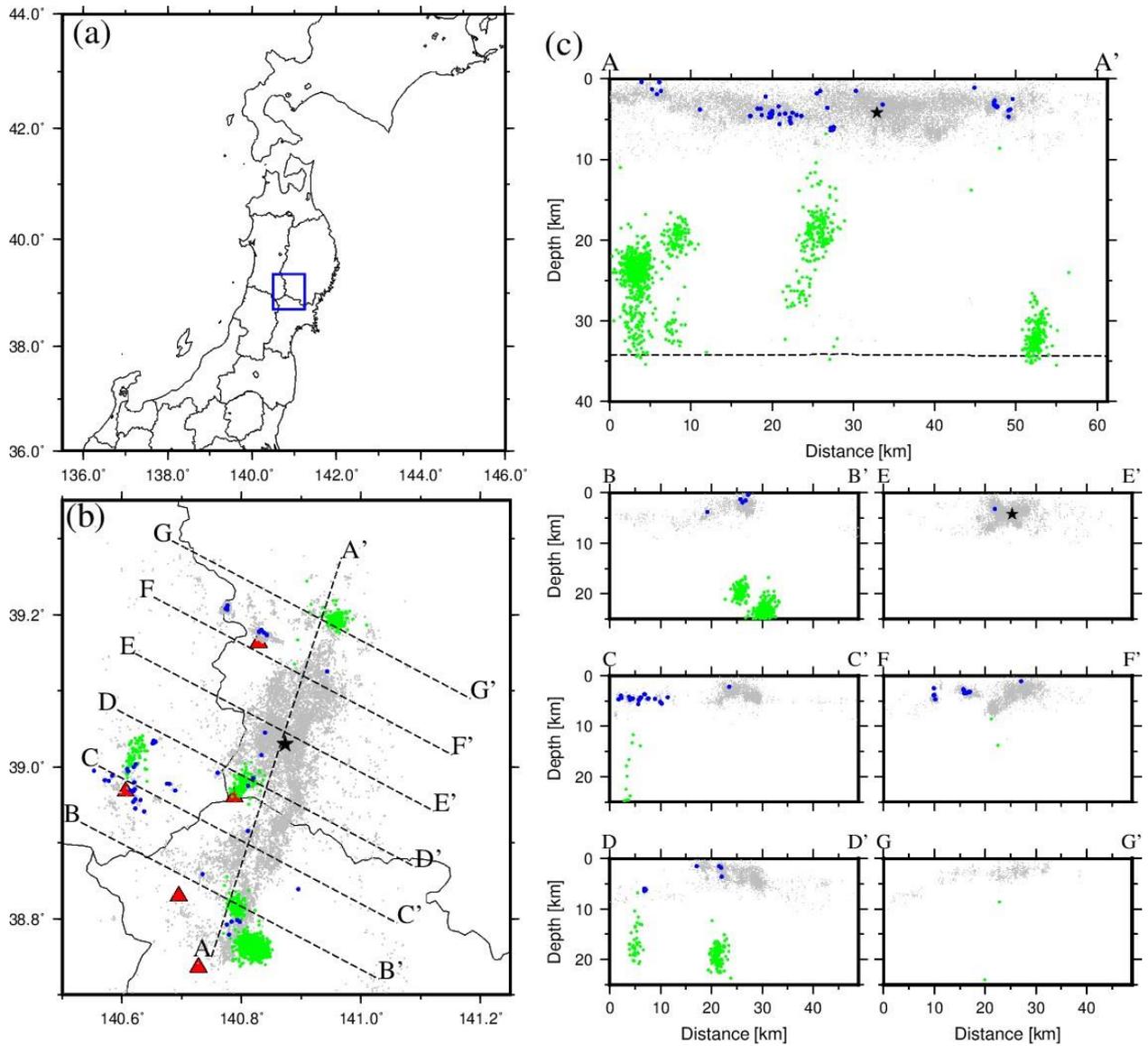
85 Most deep inland LFEs have been observed beneath volcanoes around the Moho
86 discontinuity (e.g., Ukawa and Ohtake, 1987; Hasegawa and Yamamoto, 1994). However,
87 some recent studies have reported that earthquakes with dominant frequencies much
88 lower than those of regular earthquakes of similar magnitude also occur in the upper crust,
89 very close to the locations of regular earthquakes (e.g., Baltay et al., 2011; Kosuga and
90 Haruyama, 2018; Yoshida et al., 2020; Nakajima and Hasegawa, 2021; Tsuchiyama et al.,
91 2022). Yoshida et al. (2020) showed that deep LFEs with long-lasting high-amplitude trailing
92 parts in Hakodate, Hokkaido, Japan, are distributed continuously from the lower crust to the
93 upper crust, and that the LFEs in the upper crust occur within a few kilometers of regular
94 earthquakes. Nakajima and Hasegawa (2021) systematically analyzed small earthquakes
95 with $0 \leq M_{jma} \leq 2.5$ in the upper crust in Japan. They found that several earthquakes,
96 whose dominant frequencies were much lower than those of regular earthquakes of similar
97 magnitudes, occurred very close to regular earthquakes. We refer to these earthquakes as
98 "shallow low-frequency earthquakes (shallow LFEs)," according to Nakajima and Hasegawa
99 (2021) and Hasegawa and Nakajima (2022). Previous studies have proposed that fluid

100 movement stemming from the subducting slab (Yoshida et al., 2020) or crustal fluid
101 redistribution due to recent large earthquakes (Nakajima and Hasegawa, 2021; Tsuchiyama
102 et al., 2022) may be related to the generation of these shallow LFEs in the upper crust.

103 Previous studies have suggested that magma or crustal fluids may be responsible for
104 the generation of inland deep and shallow LFEs (e.g., Ukawa and Ohtake, 1987; Hasegawa
105 and Yamamoto, 1994; Hasegawa et al., 2005; Aso et al., 2013; Yoshida et al., 2020; Nakajima
106 and Hasegawa, 2021; Tsuchiyama et al., 2022). However, the details of the seismic wave
107 radiation processes of LFEs are still poorly understood compared with those of regular
108 earthquakes, which are essentially fault ruptures. Although dominant frequencies of LFEs
109 are much lower than those of regular earthquakes of similar magnitudes (e.g., Hasegawa
110 and Yamamoto, 1994), few quantitative evaluations have been conducted on the source
111 properties of LFEs. To understand the seismic wave radiation processes of LFEs, it is
112 necessary to quantify their seismic radiation energy. Traditionally, stress drops based on
113 earthquake corner frequencies have been used to characterize the source properties of
114 small regular earthquakes (e.g., Shearer et al., 2006; Trugman and Shearer, 2017; Yoshida et
115 al., 2017; Trugman, 2020; Shearer et al., 2022), assuming the ω^2 -model (Aki, 1967; Brune,
116 1970). However, some previous studies have suggested that the radiation processes of LFEs
117 may not be consistent with the ω^2 model (e.g., Ide et al., 2007a; Yoshida et al., 2020). Thus,
118 we here use seismic radiated energy to characterize the radiation properties of LFEs
119 because any assumptions on radiation pattern, such as a specific shape of source spectrum
120 like the Brune-type ω^2 model (Brune, 1970) or a specific value of rupture speed, are not
121 required to estimate radiated energy.

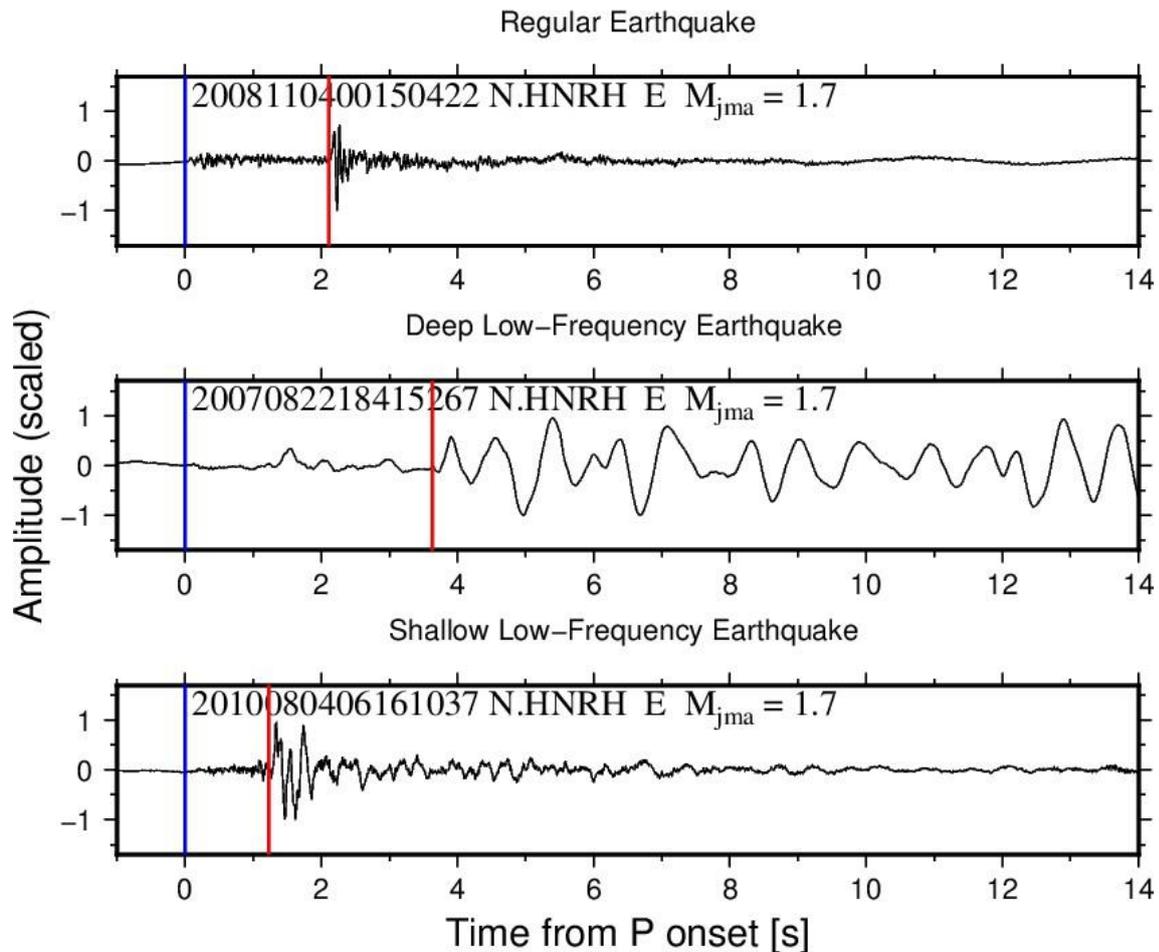
122 In and around the focal area of the 2008 M_w 6.9 Iwate-Miyagi earthquake, many
123 deep LFEs (20–40 km) have been observed beneath volcanoes around the Moho
124 discontinuity (e.g., Hasegawa and Yamamoto, 1994) (Figure 1 and Figure 2). This area is
125 surrounded by many borehole seismic stations, and recent studies have identified relocated
126 hypocenters for small earthquakes (e.g., Okada et al., 2012; Yoshida et al., 2014a) in this
127 region, making it suitable for studying the source properties of earthquakes (Figure 1 and
128 Figure 3). Nakajima and Hasegawa (2021) reported that shallow LFEs (~ 10 km) also occur in
129 and around this area (Figure 1 and Figure 2). The nature of shallow LFEs is also not yet clear,
130 but it is easier to study their source characteristics than to study deep LFEs because they
131 occur in the upper crust and the path effects are similar to those of regular earthquakes. In
132 this study, the scaled energy of regular earthquakes, deep LFEs, and shallow LFEs was
133 estimated to investigate their radiation properties.

134



135

136 Figure 1. Hypocenter distribution in and around the focal area of the 2008 M_w 6.9 Iwate-
 137 Miyagi earthquake. (a) Map of NE Japan. (b) Epicentral distribution in the region shown by a
 138 blue rectangle in (a). (c) Vertical cross-section along the line A–A' to G–G'. Gray and blue
 139 circles indicate regular earthquakes ($M_{jma} \geq 1.0$) and shallow LFEs, respectively (Nakajima
 140 and Hasegawa, 2021) for the period from 1 March, 2003 to 31 December, 2010. Green
 141 circles indicate deep LFEs identified by JMA for the period from 1 March, 2003 to 31
 142 December, 2020. A black star indicates the hypocenter of the 2008 M_w 6.9 Iwate-Miyagi
 143 earthquake. Red triangles indicate volcanoes. The black dashed line in line A–A' indicates
 144 the Moho discontinuity (Zhao et al., 1990). Note that we used the hypocenter data from the
 145 relocated catalog by Yoshida et al. (2014a) for the regular earthquakes and the shallow LFEs,
 146 and the JMA unified catalog for deep LFEs.



147

148 Figure 2. Examples of E-W-component waveforms of the regular earthquake, the deep LFE
 149 and the shallow LFE at N. HNRH. Waveforms are 0.2 Hz high-pass filtered and normalized by
 150 their respective maximum amplitudes. The blue and red lines indicate the onsets of the P-
 151 and S-waves respectively.

152 2 Data and Methods

153 2-1 Station and events

154 We analyzed S-wave spectra to obtain the scaled energy for 3560 regular
 155 earthquakes ($M_{jma} \geq 2.0$) and 52 shallow LFEs (Nakajima and Hasegawa, 2021) for the
 156 period from 1 March, 2003 to 31 December, 2010 and 1086 deep LFEs for the period from 1
 157 March, 2003 to 31 December, 2020. We used hypocenter data from the relocated catalog by
 158 Yoshida et al. (2014a) for regular earthquakes and shallow LFEs, and the JMA unified catalog
 159 for deep LFEs. Note that we did not analyze regular earthquakes and shallow LFEs whose
 160 hypocenters were not relocated by Yoshida et al. (2014a), even though the earthquakes
 161 were listed in the JMA unified catalog.

162 We obtained three-component (E-W, N-S, and U-D) waveform data from the
 163 velocity-type seismometers surrounding the focal area of the 2008 Iwate-Miyagi earthquake
 164 (Figure 3). Their natural frequencies were 1 Hz (except for TU.KWT, whose natural frequency
 165 was 0.05 Hz), and the sampling frequency was 100 Hz. We corrected the instrumental
 166 characteristics of the seismometers to obtain velocity waveforms.

182 circles indicate the regular earthquakes ($M_{jma} \geq 1.0$) and the shallow LFEs for the period
 183 from 1 March, 2003 to 31 December, 2010, respectively. Green circles indicate the deep
 184 LFEs identified by JMA for the period from 1 March, 2003 to 31 December, 2020. Black
 185 circles indicate the 14 regular earthquakes used to estimate the site response and
 186 attenuation factors. The beach balls indicate the focal mechanisms of the 14 events
 187 determined by the National Research Institute for Earth Science and Disaster Resilience
 188 (NIED) F-net moment tensor catalog. Note that we used the hypocenter data from the
 189 relocated catalog by Yoshida et al. (2014a) for the regular earthquakes and the shallow LFEs,
 190 and the JMA unified catalog for the deep LFEs

191 2-2 Source spectrum

192 We estimated the seismic-scale energy e_R for regular earthquakes, deep LFEs, and
 193 shallow LFEs using their source spectra to investigate the radiation properties of the LFEs. e_R
 194 is the ratio of the radiated energy E_R to the seismic moment M_0 ,

$$195 \quad e_R = \frac{E_R}{M_0} \quad (1)$$

196 and has been used to investigate the seismic radiation process between earthquakes of
 197 different magnitudes (e.g., Kanamori et al., 1993; Abercrombie, 1995; Prieto et al., 2004;
 198 Yoshida and Kanamori, 2023). In this study, we separately estimated E_R and M_0 using the
 199 source spectrum to obtain e_R .

200 We determined the source spectrum from the observed velocity spectra for each
 201 earthquake to calculate E_R and M_0 . The observed velocity spectrum $v_i(f)$ at the i th station
 202 at frequency f can be expressed using its source spectrum $S(f)$

$$203 \quad v_i(f) = \frac{2\pi f S(f) G_i(f) P_i(f) R_{\theta\phi i} F_{si}}{4\pi\rho\beta^3 r_i} \quad (2)$$

204 where $G_i(f)$ is the site response factor for the i th seismic station, $P_i(f)$ is the attenuation
 205 term, $R_{\theta\phi i}$ is the radiation pattern of the S-wave, F_{si} is the reflection effect from the free
 206 surface, r_i is hypocentral distance, β is the S-wave velocity around the source and ρ is the
 207 mean crustal density at 2.7 g/cm^3 , respectively. $P_i(f)$ can be expressed using the
 208 frequency-dependent attenuation term along a ray path from the source to the station
 209 ($t^*(f)$):

$$210 \quad P_i(f) = \exp(-\pi f t^*(f)) \quad (3)$$

211 where

$$212 \quad t^*(f) = \int_{source}^{receiver} \frac{Q^{-1}(f, s)}{\beta(s)} ds \quad (4)$$

213 where $Q^{-1}(f, s)$ and $\beta(s)$ are the attenuation factor at frequency f and S-wave velocity
 214 along the ray path s , respectively. We used the 1-D S-wave velocity structure proposed by
 215 Ueno et al. (2002). We assumed $R_{\theta\phi i}$ as the root mean square (RMS) value of a point shear
 216 dislocation $R_{\theta\phi i} = R_{\theta\phi i}^{RMS} = \sqrt{2/5}$ and F_{si} as 2 for all frequency points at all stations.

217 The source spectrum was calculated by correcting for site and path effects from the
 218 observed velocity spectra (Takahashi et al., 2005; Yoshida et al., 2017). We determined the
 219 site response factor $G_i(f)$ and the attenuation factor Q^{-1} using the coda normalization
 220 method (Aki and Chouet, 1975; Aki, 1980). We describe the data and methods used to
 221 estimate the site response factors in sections 2-3 and 2-5-2, the data and method to
 222 estimate the attenuation factor in section 2-4 and the method for calculating the source
 223 spectrum, radiated energy, seismic moment, and scaled energy for each earthquake in
 224 section 2-5-3.

225

226 **2-3 Site response factor**

227 The site response factors for the three components were estimated based on the
 228 coda normalization method proposed by Aki and Chouet (1975). We assumed that the coda
 229 wave at a lapse time longer than twice the direct S-wave travel time did not depend on the
 230 radiation pattern or hypocentral distance because the seismic energy was spatially
 231 homogeneously distributed at that time (Aki and Chouet, 1975). The coda-wave spectral
 232 amplitude $A_i(f, t_c)$ at the i th station at frequency f at lapse time t_c ($t_c > 2t_{si}$, where t_{si} is
 233 the S-wave travel time at the i th station) can be written in terms of its source spectrum
 234 $S(f)$ and site response factor $G_i(f)$ as

$$235 \quad A_i(f, t_c) = S(f)C(f, t_c)G_i(f) \quad (5)$$

236 where $C(f, t_c)$ is the coda excitation factor characterizing medium heterogeneity, which is
 237 only dependent on frequency and lapse time and is independent of the source station
 238 locations. Taking the ratio of the coda spectral amplitudes between different stations for the
 239 same event at a common lapse time (t_c) from the original time, the ratio is equivalent to the
 240 ratio of the site response factors between the two stations.

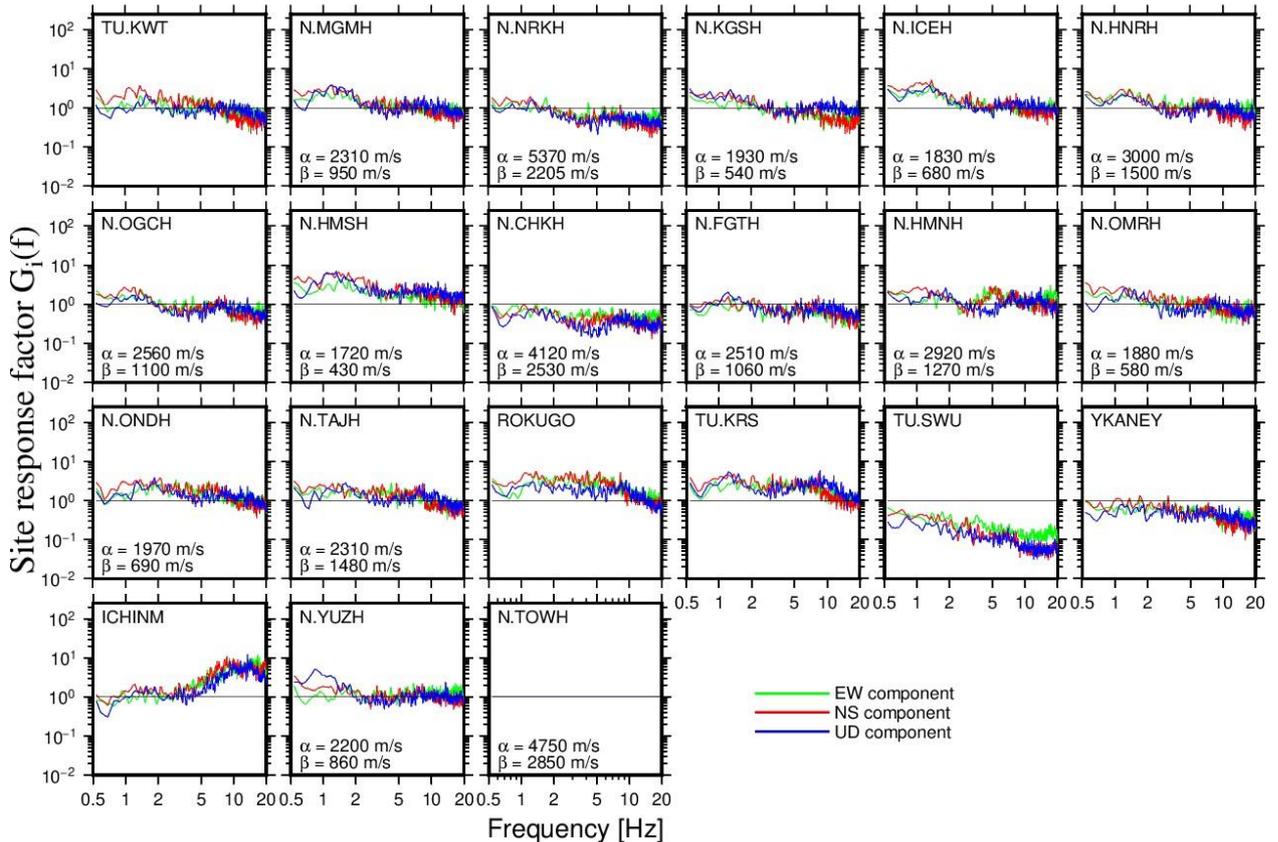
$$241 \quad \frac{A_i(f, t_c)}{A_j(f, t_c)} = \frac{S(f)C(f, t_c)G_i(f)}{S(f)C(f, t_c)G_j(f)} = \frac{G_i(f)}{G_j(f)} \quad (6)$$

242 To estimate the site response factors for three components, we selected 14 regular
 243 earthquakes with $M_{jma} \geq 3.5$ whose rupture processes are simple, based on apparent
 244 source time functions determined by Yoshida and Kanamori (2023), and used their coda
 245 spectral amplitudes. We fixed t_c at 45 s for all stations, components, frequency points, and
 246 events, and set the time window length to be 5.12 s. We removed the waveform records
 247 contaminated by other earthquakes beforehand and used only the waveforms whose SNRs
 248 at all frequency points from 0.5 Hz to 20 Hz were larger than 3.0.

249 Because we could not obtain the absolute value of the site response factors based
 250 on this method, we set a reference station located at a hard-rock site. The site response
 251 factor for the reference station was assumed to be 1 for each frequency point and
 252 component. We selected the station installed in the rock with the highest S-wave velocity
 253 (Yoshida et al., 2017), according to the core log data of the NIED Hi-net as the reference
 254 station (N. TOWH). For each component, we estimated the relative site response factors for
 255 the other stations by taking the logarithm of equation (6) for all station pairs and applying

256 the least-squares method. Verification of the site response factor at the reference station is
 257 described in section 2-5-2.

258 Figure 4 shows the site response factors obtained for each station and their
 259 components. Most stations in this study were borehole stations. Most stations on the hard
 260 rock site ($V_s > 1000$ m/s) showed a site response factor similar to that of the reference
 261 station. In contrast, stations located on soft sites ($V_s < 1000$ m/s) tended to have a larger
 262 site response factor at low frequencies ($f < 2$ Hz) which decreased with frequency at
 263 higher frequencies ($f > 2$ Hz).



264
 265 Figure 4. Estimated relative site response factors in comparison to that for the reference
 266 station. The green, red, and blue lines indicate the estimated relative site response factors
 267 for the east-west, north-south and vertical components respectively. α and β represent the
 268 P- and S-wave velocities at the stations listed on the Hi-net columnar maps, respectively.

269
 270 **2-4 Attenuation factor**

271 We assumed a spatially homogeneous attenuation factor Q^{-1} in the entire region
 272 and estimated the attenuation factor Q^{-1} for each frequency point based on the coda
 273 normalization method proposed by Aki (1980). This method is based on the distance decay
 274 of the S-wave amplitude, other than on geometrical spreading. By normalizing the direct S-
 275 wave spectral amplitude by the coda wave spectral amplitude, we can cancel the site and
 276 source effects on the observed waveforms and isolate the attenuation factor Q^{-1} (Aki,
 277 1980) as

278

$$\ln\left(\frac{v_i(f)r_i}{A_i(f,t_c)}\right) = -\pi f Q^{-1}(f) t_{si} + const \quad (7)$$

279

280 We determined $-\pi f Q^{-1}$ from the linear slope of the left side of equation (7) against
 281 the S-wave travel-time t_{si} (Figure 5-a) using least-square method and estimated Q^{-1} . We
 282 used three components of the S-wave and coda waveforms of the 14 earthquakes described
 283 in section 2-3 and calculated the left side of equation (7) for each station and component.
 284 We adopted the same method for selecting the spectra as described in sections 2-1 and 2-3.
 We estimated the Q^{-1} -value from 0.5 Hz to 20 Hz at 0.1 Hz intervals.

285

286 Figure 5 shows the estimated frequency-dependent attenuation factor Q^{-1} based on
 287 the coda normalization method. The estimated Q^{-1} decays with the frequency (Figure 5-b).
 288 We approximated the frequency-dependent attenuation factor using the Q-value at 1 Hz
 (Q_0^{-1}) and the frequency-dependence factor (a) by fitting

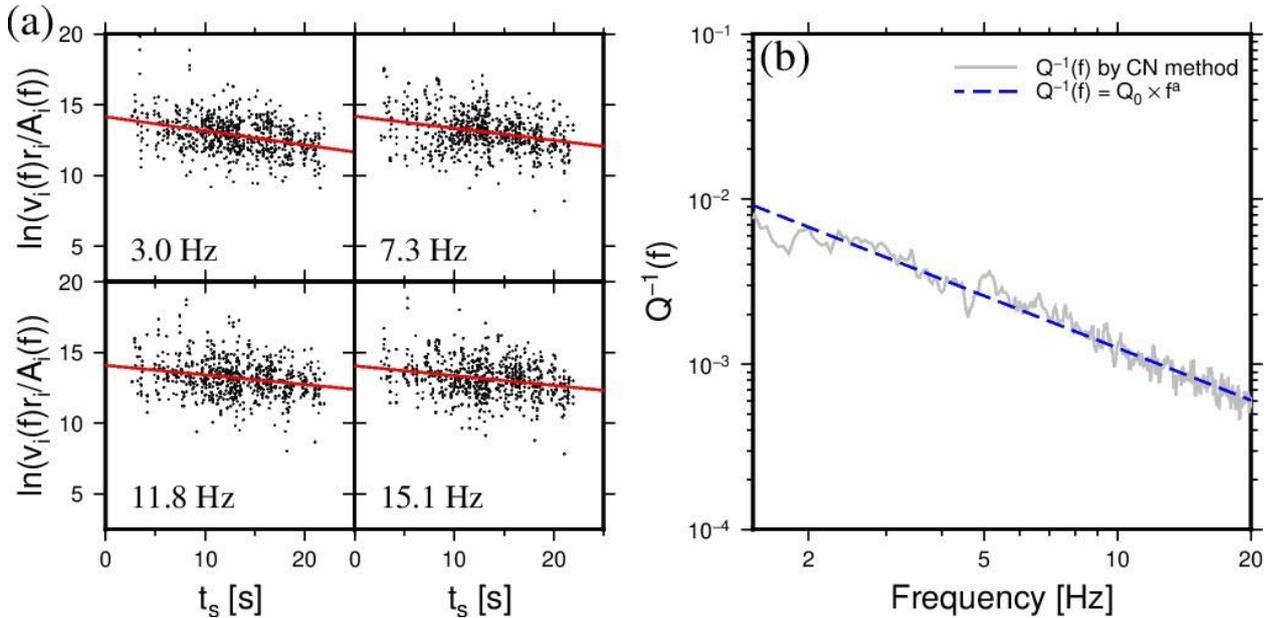
289

$$Q^{-1}(f) = Q_0^{-1} f^{-a} \quad (8)$$

290

291 However, after evaluating the estimated error of Q^{-1} at each frequency point using
 292 bootstrapping, we found that the estimates for the low-frequency component (~ 1.5 Hz) of
 293 Q^{-1} were not stable. This may indicate that the focal mechanism or sediment layer affects
 294 the low-frequency amplitude of the seismic waveforms. Therefore, we only used the high
 295 frequency component (1.5 Hz \sim 20 Hz) of Q^{-1} to estimate Q_0^{-1} and a . Q_0^{-1} and a were
 296 estimated at 0.015 ± 0.007 , 1.0 ± 0.02 respectively, with the least-square method (Figure
 297 5-b).

297



298

299 Figure 5. Estimated frequency-dependent attenuation factor Q^{-1} by coda normalization
 300 method. (a) The left-hand side of equation (7) versus direct S-wave travel-time t_{si} for four
 301 frequency points. The black dots indicate the individual data for each station and
 302 component for 14 earthquakes. The red line represents the best fit curve of equation (7). (b)

303 Estimated frequency-dependent Q^{-1} . The gray and blue dashed lines indicate the estimated
 304 Q^{-1} by coda normalization method and the best fit curve of equation (8), respectively.

305

306 **2-5 Determination of source spectrum and quantification of scaled energy**

307 **2-5-1 Tentative estimation of the source spectrum**

308 We tentatively determined the source spectrum for each earthquake using the site
 309 response and attenuation factors obtained in sections 2-3 and 2-4, according to equation
 310 (2). For each earthquake, the tentative source spectrum $S_0(f)$ was computed by correcting
 311 the site and path effects from the observed velocity spectra $v_i(f)$ for each station and
 312 component and taking their geometrical mean over the stations and components as follows:

$$313 \quad S_0(f) = \frac{4\pi\rho\beta^3}{R_{\theta\phi}^{RMS}} \prod_{i=1}^N \left[\frac{v_i(f)r_i}{2\pi f G_i(f) F_{si} \exp(-\pi f t^*(f))} \right]^{\frac{1}{N}} \quad (9)$$

314 where N denotes the number of velocity spectra used in the analysis. However, the site
 315 response factors estimated by the coda normalization method in section 2-3 are relative to
 316 those for the reference station. Therefore, we evaluated the site response factor for the
 317 reference station in the following section.

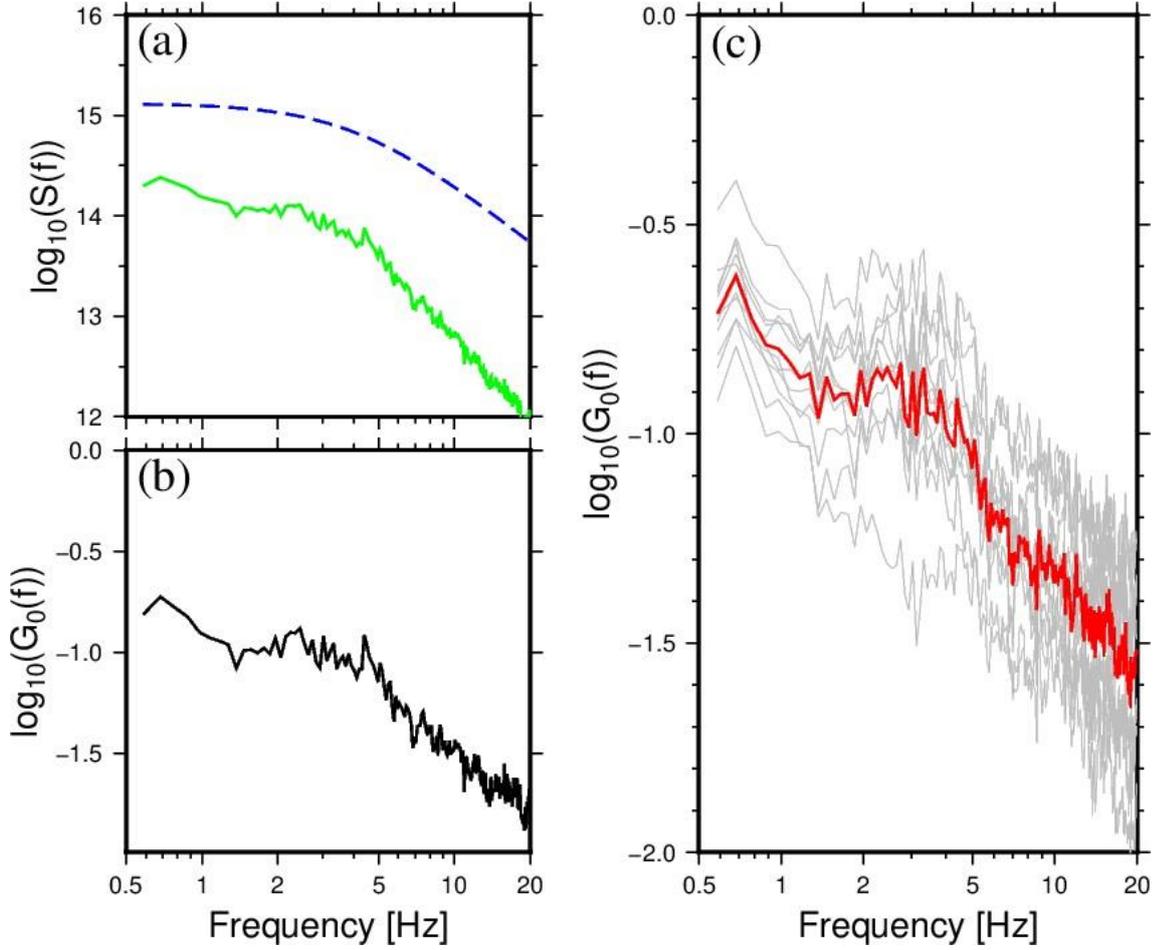
318 **2-5-2 Site characteristic factor of the reference station**

319 In section 2-3, we estimated the relative site response factors, assuming that the site
 320 response factor for the reference station is 1 for all frequency points for the three
 321 components. However, if the site response factor at reference station $G_0(f)$ deviated
 322 significantly from 1, the overall spectrum would deviate from the true one. In this study, we
 323 evaluated $G_0(f)$ based on the assumption that the logarithmic spectral residual of the
 324 tentative source spectrum $S_0(f)$ from the model source spectrum, assuming a Brune-type
 325 omega-square model (Brune, 1970) can represent the $G_0(f)$ (Figure 6-a and b) (Yoshida et
 326 al., 2017).

327 As in section 2-3 and 2-4, we used the 14 regular earthquakes because their rupture
 328 processes are simple, according to a previous study (Yoshida & Kanamori, 2023), thus, their
 329 source spectra are thought to be approximated by the omega-square model. We required
 330 the source corner frequencies and seismic moments to construct their model source
 331 spectra. For the seismic moments, we used the values listed in the F-net moment tensor
 332 catalog (NIED). We estimated the source corner frequencies of the 14 events based on the
 333 spectral ratio method using empirical Green's functions (EGFs). This method cancels site and
 334 path effects on the waveform by considering the spectral ratio of the earthquake to be
 335 suitable for a nearby smaller earthquake as an EGFs event (e.g., Mueller, 1985; Dreger,
 336 1994). The detailed procedure of the spectral ratio method is described in Appendix A.

337 The gray lines in Figure 6-c represent the estimated G_0 values for the respective
 338 events. Of the 14 events, we estimated the source corner frequencies and obtained model
 339 source spectra for 11 events. We omitted three events because we could not find

340 appropriate EGFs for them. The estimated $\log_{10} G_0(f)$ values decreased with increasing
 341 frequency. The similarity in the frequency characteristics of $G_0(f)$ values inferred from all
 342 11 earthquakes that occurred at different locations supports the idea that the logarithmic
 343 spectral residuals of the tentative and model source spectrum represent the site response
 344 factor for the reference station. We used the geometric mean of $G_0(f)$ estimated from the
 345 11 events as the representative value for subsequent analyses.
 346



347
 348 Figure 6. Estimated site response factor for the reference station $G_0(f)$. (a) Example of the
 349 model (blue dashed line) and estimated tentative source spectrum (green solid line). (b) The
 350 logarithmic spectra residual of the estimated tentative and model source spectrum (c)
 351 Estimated G_0 . The gray and red lines represent the individual and average G_0 , respectively.

352 **2-5-3 Final estimation of source spectrum and scaled energy**

353 Finally, we obtained the source spectrum $S(f)$ for each earthquake by correcting the
 354 site response factor for the reference station using the tentative source spectrum $S_0(f)$.

355

$$S(f) = \frac{S_0(f)}{G_0(f)} = \frac{4\pi\rho\beta^3}{G_0(f)R_{\theta\phi}^{RMS}} \times \prod_{i=1}^N \left[\frac{v_i(f)r_i}{2\pi f G_i(f) F_{si} \exp(-\pi f t^*(f))} \right]^{\frac{1}{N}} \quad (10)$$

356 From the source spectrum estimated using equation (10), we calculated the seismic
 357 moment and radiated energy separately to obtain the scaled energy in equation (1). We
 358 determined the seismic moment (M_0) by fitting the model source spectrum (S_n^{cal}) to the
 359 derived source spectrum ($S(f)$) to minimize the residual ($|\log_{10} S(f) - \log_{10} S_n^{cal}(f)|$)
 360 (Figure 7);

$$361 \quad S_n^{cal}(f) = \frac{M_0}{1 + (f/f_c)^n} \quad (11)$$

362 where f_c and n are the source corner frequency and spectral high-frequency falloff rate,
 363 respectively. We estimated f_c at 0.1 Hz intervals from 0.5 Hz to 20 Hz and n at 0.1 intervals
 364 from 0 to 4 with a grid-search method, respectively. Note that we did not use this result to
 365 calculate the radiated energy.

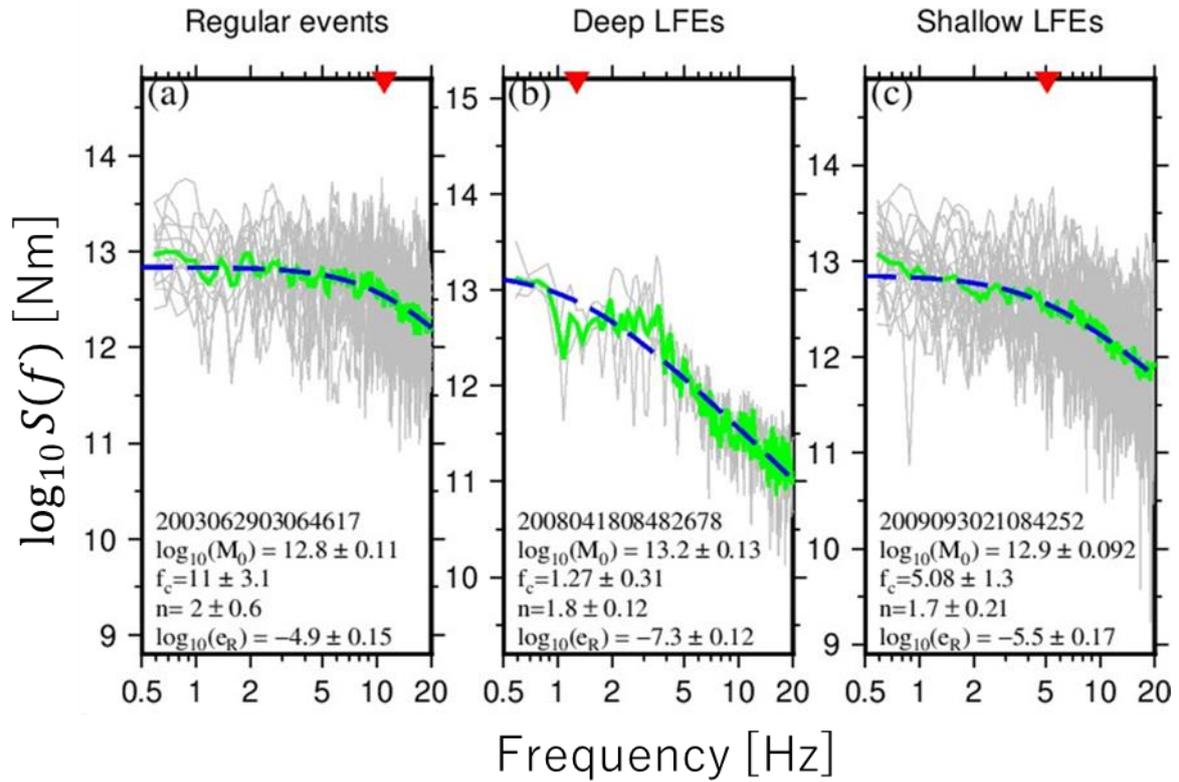
366 The radiated energy was calculated using the following equation (Vassiliou and
 367 Kanamori, 1982).

$$368 \quad E_R = \frac{8\pi}{10\rho\beta^5} \int_0^\infty (fS(f))^2 df \quad (12)$$

369 We ignored the P-wave radiated energy because most of the radiated energy was carried by
 370 S-wave (e.g., Kanamori et al., 2020). To calculate radiated energy based on equation (12),
 371 we used the derived source spectrum from 0.5 Hz to 20 Hz and extrapolated to the outside
 372 of the analyzed frequency band, assuming that the amplitude shows the value expected
 373 from the seismic moment for $f < 0.5$ Hz and decays in proportion to f^{-2} for $f > 20$ Hz,
 374 respectively (Snoke, 1987). It should be noted that we did not use the estimated n for
 375 extrapolation of the source spectrum and fixed the high-frequency falloff rate at 2 above 20
 376 Hz because the falloff rate n estimated by fitting above did not represent a high-frequency
 377 falloff rate above 20 Hz, only up to 20 Hz. Finally, we determined the scaled energy for each
 378 earthquake in equation (1) using the radiated energy and seismic moment estimated above.

379 To evaluate the estimation uncertainty, we resampled individual source spectra at
 380 different stations and components by bootstrapping and calculated their geometric mean,
 381 estimated source corner frequency, high-frequency falloff rate, seismic moment, radiated
 382 energy, and scaled energy 200 times. Standard deviations calculated for each factor were
 383 used as measures of uncertainty.

384

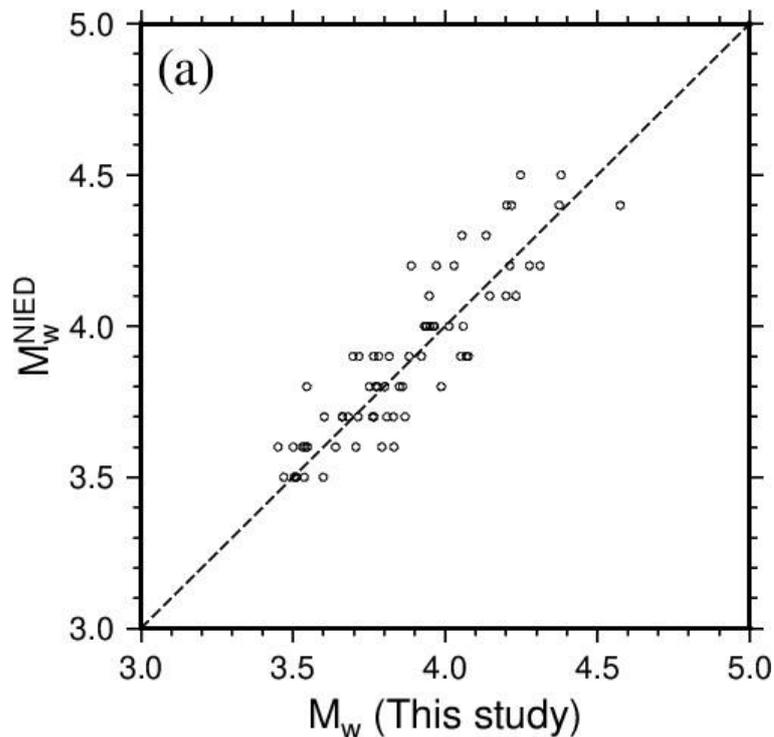


385

386 Figure 7. Examples of the estimated source spectra with the correction of the site response
 387 and attenuation factors. (a), (b), and (c) indicate source spectra for the regular earthquakes,
 388 the deep LFEs and the shallow LFEs, respectively. The gray lines indicate the individual
 389 source spectra from respective stations and components. The green solid and blue dashed
 390 lines indicate the geometric mean of the source spectra and the best-fit curve of equation
 391 (11), respectively. The red inverted triangles indicate the estimated source corner
 392 frequencies.

393 3 Results

394 We calculated the radiated energy (E_R) and seismic moment (M_0) using the derived
 395 source spectra and obtained the scaled energy (e_R) for 1464 regular earthquakes ($M_{jma} \geq$
 396 2.0), 169 deep LFEs, and 52 shallow LFEs detected by Nakajima and Hasegawa (2021). Figure
 397 8 shows the relationship between the moment magnitudes estimated in this study and
 398 those listed on the F-net moment tensor catalog (M_w^{NIED}) based on broadband data. We
 399 found a generally good agreement between them.

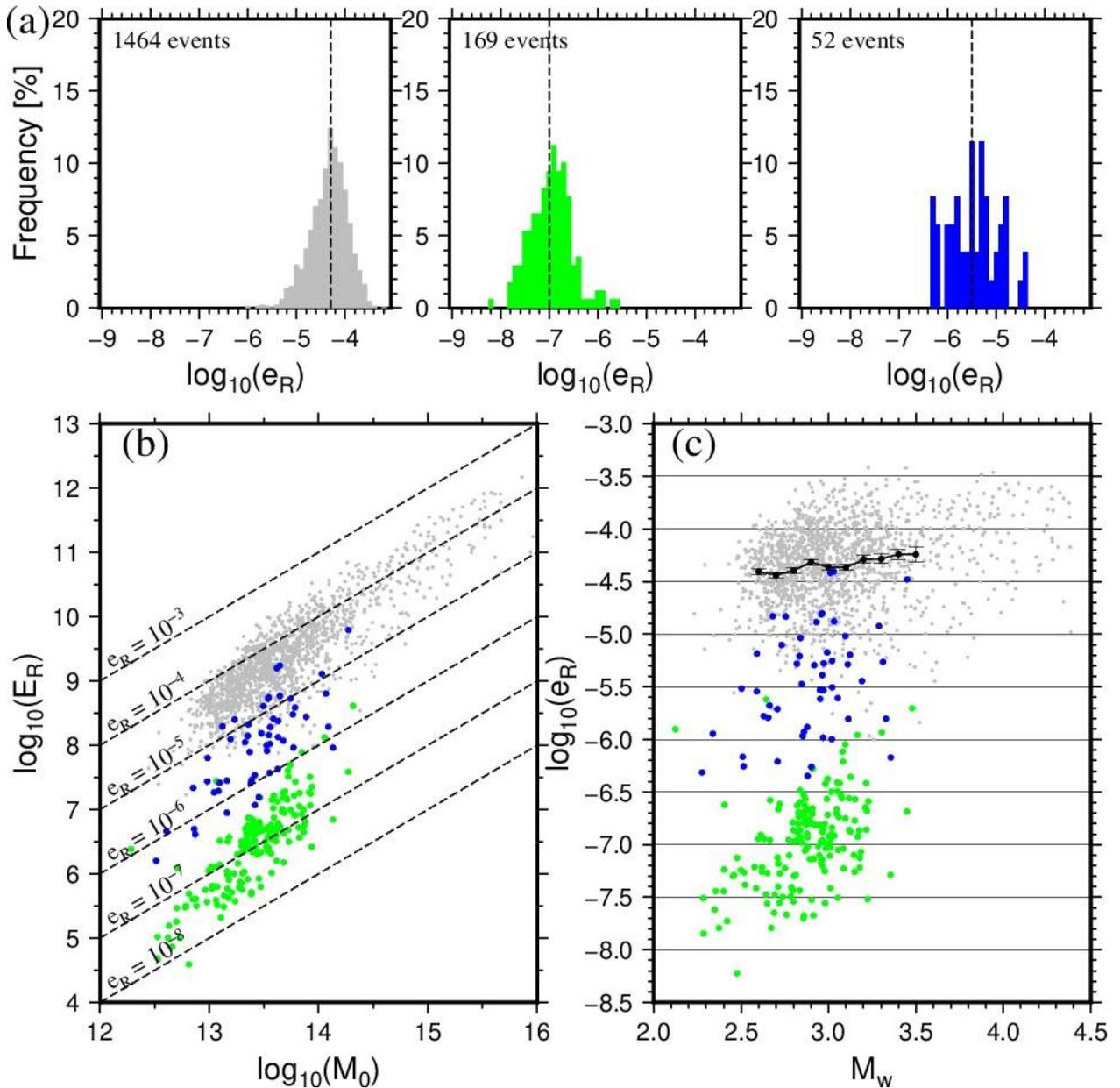


400

401 Figure 8. The relationship between moment magnitudes estimated in this study and those
 402 listed on F-net moment tensor catalog (M_w^{NIED}).

403 Figure 9-b shows the relationship between the estimated radiated energy and
 404 seismic moment. The radiated energy increased with seismic moment, and the radiated
 405 energy for LFEs was systematically lower than those for regular earthquakes with similar
 406 seismic moments (Figure 9-b).

407 Figure 9-a shows a histogram of the obtained e_R values and Figure 9-c shows e_R
 408 values as a function of the estimated moment magnitudes (M_w). The mean, standard error,
 409 and standard deviation of the estimated $\log_{10} e_R$ for regular earthquakes, deep, and shallow
 410 LFEs are shown in Table 1. For regular earthquakes, although e_R values are scattered, e_R
 411 values are on the order of 10^{-5} , which is consistent with previous studies on scaled energy
 412 for crustal earthquakes (e.g., Abercrombie, 1995; Ide and Beroza, 2001; Prieto et al., 2004;
 413 Kanamori and Brodsky, 2004; Yoshida & Kanamori, 2023). On the other hand, the e_R values
 414 for the deep and shallow LFEs are on the order of 10^{-7} and 10^{-6} , respectively. This result
 415 indicates that e_R values for LFEs are one to three orders of magnitude lower than those for
 416 regular earthquakes.



417

418 Figure 9. Estimated radiated energy (E_R), seismic moment (M_0) and scaled energy (e_R). (a)
 419 Histogram of e_R values. Gray, green, and blue histograms indicate the regular earthquakes
 420 ($M_{jma} \geq 2.0$), the deep and the shallow LFEs, respectively. (b) The relationship between the
 421 E_R values and M_0 values. (c) e_R values as a function of moment magnitudes (M_w) (Kanamori,
 422 1977; Hanks and Kanamori, 1979). Gray, green, and blue circles indicate the regular
 423 earthquakes, deep and shallow LFEs, respectively. Black dots and error bars indicate the
 424 average e_R values and standard errors at 0.1 bins in the obtained M_w , respectively.

425 Table 1. Mean, standard deviation, and standard error of the obtained $\log_{10} e_R$

	Mean	Standard error	Standard deviation
The regular earthquakes	-4.29	0.02	0.39
The deep LFEs	-6.97	0.06	0.43
The shallow LFEs	-5.46	0.14	0.50

426

427 Scaled energy (e_R) is proportional to the product of stress drop ($\Delta\sigma$) and radiation
 428 efficiency (η) (e.g., Kanamori and Brodsky, 2004; Rivera and Kanamori, 2005) as,

$$429 \quad e_R = \frac{\eta\Delta\sigma}{2\mu} \quad (13)$$

430 where μ is rigidity. Assuming that regular earthquakes and LFEs have the same radiation
 431 efficiency, the systematic difference in scaled energy indicates that the stress drops for LFEs
 432 are one to three orders of magnitude lower than those for regular earthquakes. For
 433 example, assuming that radiation efficiency is constant at 0.47 (e.g., Brune, 1970) and
 434 rigidity is constant at 30 GPa (e.g., Kanamori, 1977), average stress drops, based on our
 435 estimated mean scaled energy, are approximately 6.5 MPa, 0.013 MPa and 0.51 MPa for the
 436 regular earthquakes, deep, and shallow LFEs, respectively. However, for LFEs at plate
 437 boundaries, some studies have reported that not only the stress drop, but also the rupture
 438 velocity for LFEs is lower than that for regular earthquakes (e.g., Thomas et al., 2016; Supino
 439 et al., 2020). If LFEs are associated with slower rupture or deformation speeds than regular
 440 earthquakes, the radiation efficiency of LFEs should be lower than that of regular
 441 earthquakes (e.g., Venkataraman and Kanamori, 2004). In reality, LFEs are likely to exhibit a
 442 lower stress drop and radiation efficiency than regular earthquakes. Thus, our stress drop
 443 estimation based on the scaled energy yielded the minimum value of the stress drop for the
 444 LFEs.

445 Nakajima and Hasegawa (2021) detected shallow LFEs in this region based on the
 446 very low dominant frequencies of the observed direct S-waves compared to regular
 447 earthquakes with similar magnitudes. We showed that the e_R values of shallow LFEs are
 448 indeed systematically an order of magnitude smaller than those of regular earthquakes.
 449 However, it is uncertain whether shallow LFEs are essentially the same as deep LFEs or
 450 regular earthquakes with low dominant frequencies. Deep LFEs occur at distinctly greater
 451 depths than regular earthquakes and have different coda characteristics; however, the
 452 difference is not as obvious in the case of shallow LFEs versus regular earthquakes. To
 453 understand the source properties of shallow LFEs in more detail, it is necessary to
 454 investigate not only direct S-waves but also the characteristics of coda waves and their focal
 455 mechanisms (Yoshida et al., 2020).

456 **4. Discussion**

457 **4.1 The effect of assuming the homogenous attenuation factor and setting a time window** 458 **on e_R for deep LFEs**

459 We determined the attenuation factor Q^{-1} in section 2-3 assuming a spatially
 460 homogeneous Q^{-1} in and around the focal area of the 2008 Iwate-Miyagi earthquake.
 461 However, deep LFEs occur at greater depths (20–30 km) than regular earthquakes (10 km).
 462 Thus, the representative attenuation factor Q^{-1} for deep LFEs may differ from that for
 463 regular earthquakes, which may lead to a systematic bias in the estimation of the source
 464 spectra and scaled energy for deep LFEs. To evaluate the effect of the depth variation in the
 465 attenuation factor Q^{-1} on e_R for the deep LFEs, we tested a case with a two-layer structure
 466 of Q^{-1} bounded at a depth of 15 km. We compared e_R values assuming this model to e_R

467 values assuming the spatially homogeneous Q^{-1} structure. We considered the following
 468 two structures using an attenuation factor Q^{-1} at 1 Hz (Q_0^{-1}):

469 Model 1. Deep Q^{-1} is larger than shallow Q^{-1}

$$470 \quad Q_0^{-1} = \begin{cases} 0.015 & (\text{shallower than 15 km}) \\ 0.025 & (\text{deeper than 15 km}) \end{cases}$$

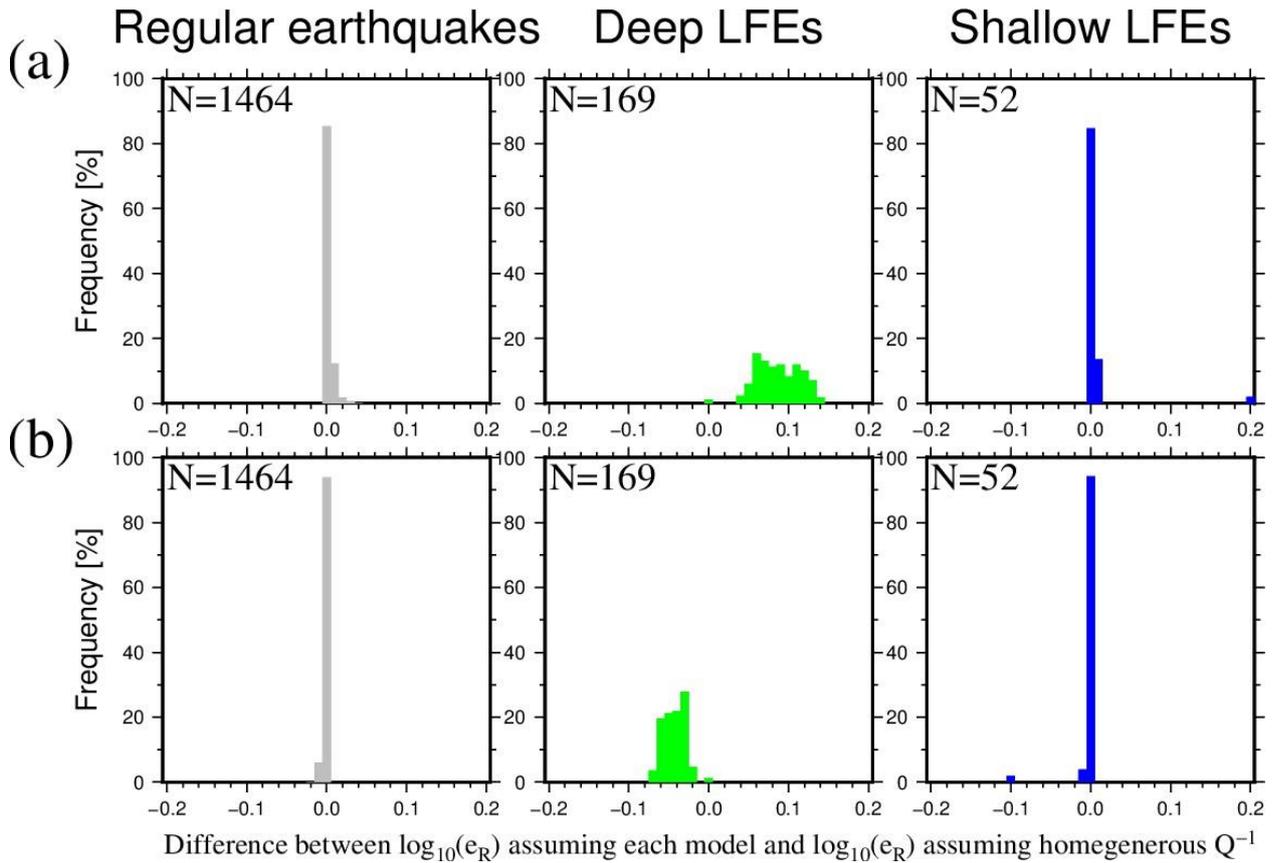
471 Model 2. Deep Q^{-1} is smaller than shallow Q^{-1} .

$$472 \quad Q_0^{-1} = \begin{cases} 0.015 & (\text{shallower than 15 km}) \\ 0.010 & (\text{deeper than 15 km}) \end{cases}$$

473 For the values of the frequency-dependent factor (a) and shallower Q_0^{-1} , we used the values
 474 estimated by the coda normalization method in section 2-4. The value of Q_0^{-1} for the deeper
 475 part of model 1 was determined based on the seismic attenuation structure beneath
 476 northeastern Japan, as estimated by Nakajima et al. (2013).

477 Figure 10 shows the difference between $\log_{10} e_R$ assuming each model and $\log_{10} e_R$
 478 assuming the spatially homogeneous Q^{-1} . For model 1, the differences in $\log_{10} e_R$ for the
 479 regular earthquakes and shallow LFEs that occurred in the upper crust were almost 0, and
 480 only approximately 0.1 to 0.2 for the deep LFEs. For model 2, the difference in
 481 $\log_{10} e_R$ between the regular earthquakes and shallow LFEs was approximately 0 and -0.1
 482 for the deep LFEs. This result suggests that the difference in $\log_{10} e_R$ between regular
 483 earthquakes and LFEs was not due to the depth variation in Q^{-1} .

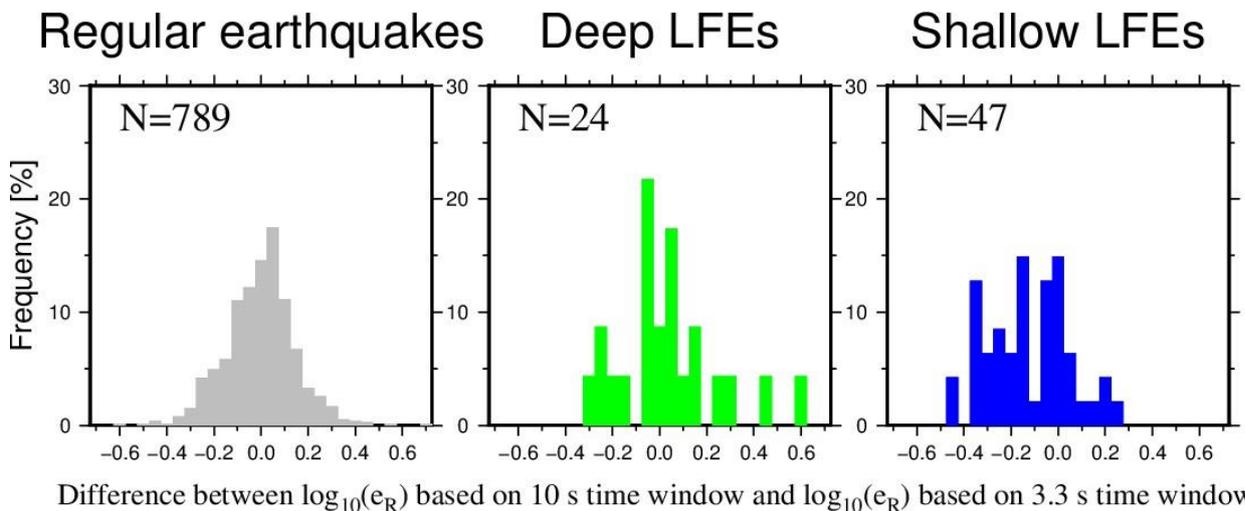
484 Another feature of deep LFEs is that they have a longer duration of seismic waves
 485 than regular earthquakes of a similar magnitude (e.g., Ukawa and Ohtake, 1987; Hasegawa
 486 and Yamamoto, 1994). The length of the time window may also affect the scaled energy
 487 estimates for deep LFEs. To investigate the effects of the time window, we estimated the
 488 scaled energy using different time window lengths (10.24 s) and compared them with those
 489 based on the original time window length (3.3 s). For comparison, we used only the
 490 observed spectra that met the SNR conditions described in section 2-2 in both time
 491 windows. Figure 11 shows the difference in $\log_{10} e_R$ using a 10.24 s and original time-
 492 window length. The difference in $\log_{10} e_R$ was only approximately 0.6, which suggests that
 493 the effect of time-window length is not large enough to explain the difference in the scaled
 494 energy between the regular earthquakes and the LFEs. These results suggest that the
 495 systematic difference in scaled energy between regular earthquakes and LFEs is not due to
 496 the depth variation in Q^{-1} or an inappropriate setting of the time window.



497

498 Figure 10. Difference in $\log_{10} e_R$ obtained from a two-layered Q^{-1} model (a: Model 1, b:
 499 Model 2) and from the spatially homogeneous Q^{-1} model. Gray, green, and blue histograms
 500 indicate the regular earthquakes, deep, and shallow LFEs, respectively.

501



502

503 Figure 11. Difference in $\log_{10} e_R$ using two different time windows (3.3 s and 10.24 s). (a),
 504 (b), and (c) indicate the result of the regular earthquakes, deep and shallow LFEs,
 505 respectively.

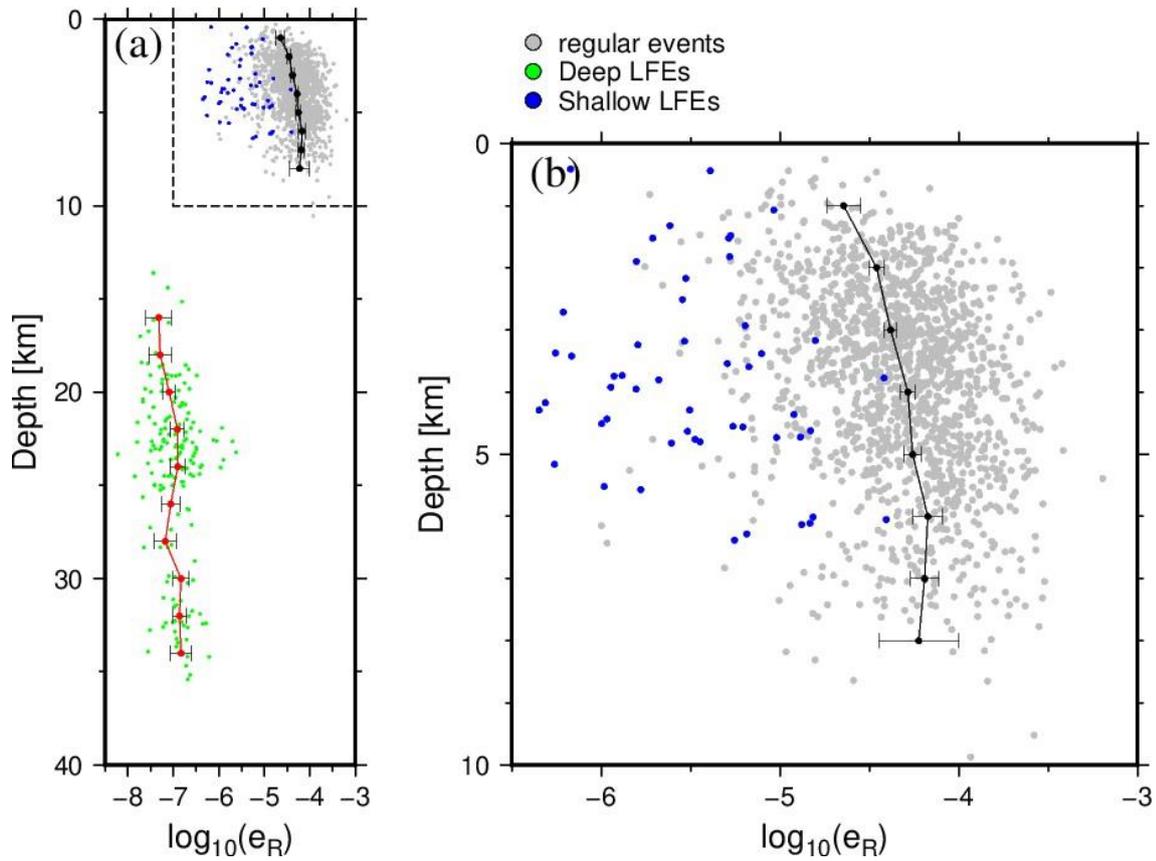
506

4.2 Depth and spatial distribution of scaled energy for the regular earthquakes

507 As shown in Figure 9, we found that the scaled energy (e_R) for regular earthquakes
508 was in the order of 10^{-5} to 10^{-4} . However, the estimated e_R values for the regular
509 earthquakes were scattered. To evaluate the diversity of e_R values among regular
510 earthquakes, we investigated the depth and spatial variation of e_R values. Figure 12 shows
511 e_R values plotted against focal depth. Although we considered the depth variation of the
512 seismic wave velocity in estimating e_R values, the values obtained for regular earthquakes
513 tended to increase with the focal depth. For regular earthquakes, previous studies have
514 suggested that stress drops tend to increase with focal depth (e.g., Hardebeck and Aron,
515 2009; Boyd et al., 2017; Huang et al., 2017; Trugman, 2020). Assuming that the radiation
516 efficiency is constant for all regular earthquakes, the scaled energy is proportional to the
517 stress drop (e.g., Kanamori and Brodsky, 2004; Rivera and Kanamori, 2005). The obtained
518 depth-dependent tendency may indicate that the stress drops increase with focal depth. In
519 contrast, some studies attributed the focal depth dependence of stress drop to the
520 assumption of a spatially homogeneous attenuation factor Q^{-1} (e.g., Abercrombie et al.,
521 2021). In section 4-1, we showed that the systematic difference in e_R values between
522 regular earthquakes and deep LFEs is not due to the depth variation of Q^{-1} using two-layer
523 structures of Q^{-1} . However, because the seismic attenuation Q^{-1} structure in the upper
524 crust in and around the focal area of the 2008 Iwate-Miyagi earthquake is not well-
525 constrained, we cannot rule out the possibility that the weak focal depth dependence of the
526 obtained e_R values is due to the assumption of a spatially homogeneous Q^{-1} structure in
527 the upper crust. Hereafter, for regular earthquakes, we compared e_R values only for
528 earthquakes that occurred at similar depths. Unlike regular earthquakes, e_R values for deep
529 LFEs exhibit no clear depth dependence.

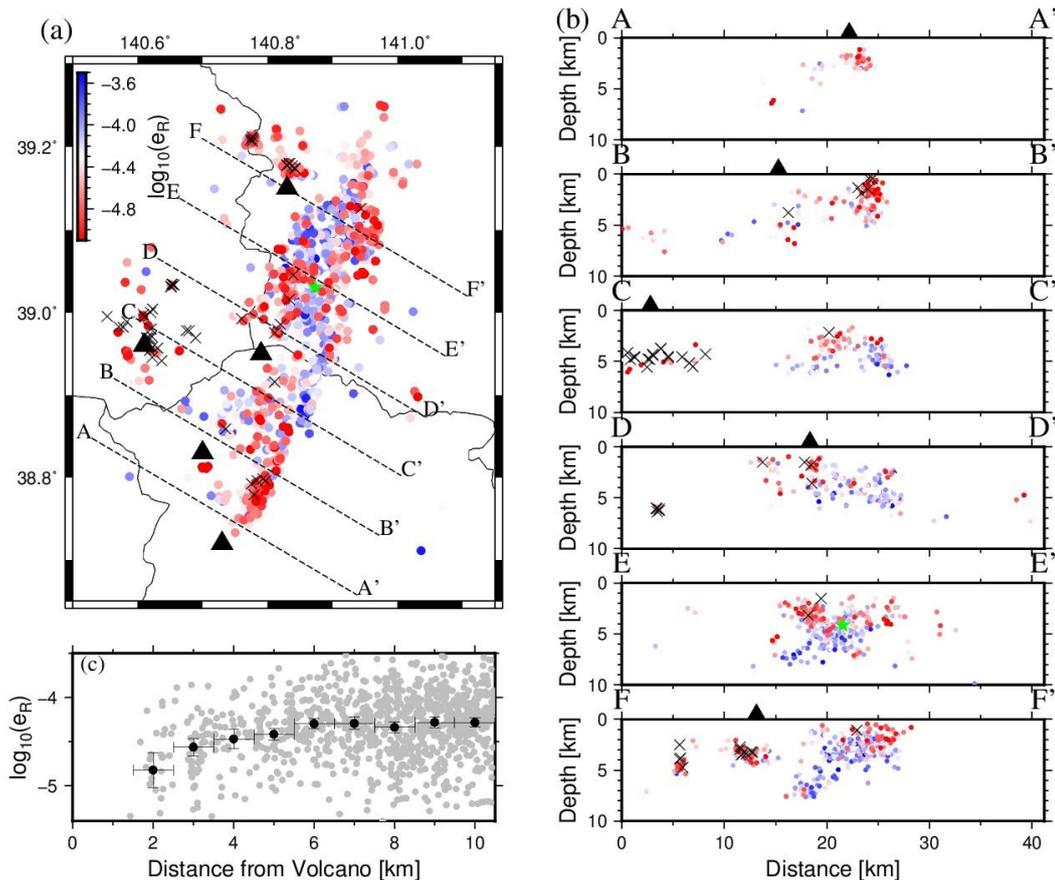
530 Figure 13 shows the spatial distribution of e_R values for the regular earthquakes and
531 as a function of distance from the nearest volcanoes. Although e_R values are scattered, they
532 tend to decrease with decreasing distance from the nearest volcanoes (Figure 13-c) and
533 tend to be small, close to the shallow LFEs (Figure 13-a, b). Low seismic wave velocities
534 (Okada et al., 2010) and relatively high pore pressure ratio regions (Yoshida et al., 2014b)
535 were estimated beneath the volcanoes in and around the aftershock areas of the 2008
536 Iwate-Miyagi earthquakes. Crustal fluids and high pore pressures may be related to the
537 diversity of e_R values between regular earthquakes. Some studies have indicated that the
538 stress drop decreases when the frictional strength decreases owing to the fluid and high
539 pore pressure (e.g., Goertz-Allmann et al., 2011b; Kwiatek et al., 2014; Yoshida et al., 2017).
540 In addition, fluid and high pore pressure may affect rupture and slip velocity (e.g., Liu and
541 Rice, 2005). These effects may be related to regular earthquakes with relatively small e_R
542 values. However, because we do not know the detailed spatial variation of the seismic
543 attenuation factor Q^{-1} in the upper crust, we cannot rule out the possibility that near-
544 source attenuation due to crustal fluid contributes to regular earthquakes with relatively
545 low e_R values. In both cases, this result suggests that we can obtain information regarding
546 the presence of fluid or near-source medium properties through the spectral analysis of
547 seismic waves.

548



549

550 Figure 12. The obtained e_R values as a function of focal depths. (a) Shallower than 40 km. (b)
 551 In the dashed rectangle in (a), gray, green, and blue circles indicate the regular earthquakes,
 552 deep and shallow LFEs, respectively. The black dots and error bars show the average e_R
 553 values and standard error of the regular earthquakes for each depth range of 1 km. The red
 554 dots and error bars indicate the average e_R values and standard errors of the deep LFEs for
 555 each 2 km depth range.



556

557 Figure 13. Spatial distribution of the regular earthquakes colored by e_R values. (a) Map view.
 558 (b) Cross-sectional views along lines A–A' to F–F' in (a). Circle colored by e_R values and cross
 559 marks indicate the regular earthquakes and the shallow LFEs, respectively. Black triangles
 560 indicate volcanoes. The green star indicates the 2008 Iwate-Miyagi earthquake. (c) e_R values
 561 plotted against the distance from the nearest volcano. Gray dots indicate individual data.
 562 Black dots and error bars indicate e_R values and standard errors in the range of 1 km.

563 4.3. Comparison of the seismic moment and radiated energy between the regular 564 earthquakes and the LFEs

565 In this study, we estimated the seismic moments for LFEs based on source spectra
 566 using a good observation network. Owing to the absence of such a good observational
 567 network, we could not estimate the seismic moment, and the only available measure of the
 568 LFE size was the local magnitude. In this study, we investigated the relationship between
 569 the radiated energy, seismic moment, and local magnitude (M_{jma}). M_{jma} is a magnitude
 570 scale based on the maximum amplitude of the seismic waveform determined by the JMA
 571 (Katsumata, 1999, 2004).

572 To compare M_{jma} and moment magnitude (M_w) for a wider magnitude range, in
 573 addition to the results of the regular earthquakes with $M_{jma} \geq 2.0$ obtained in section 3, we
 574 estimated M_w for 13619 regular earthquakes with $1.0 \leq M_{jma} \leq 1.9$ in and around the
 575 focal area of the 2008 Iwate-Miyagi earthquake. Because the SNRs at the low-frequency
 576 component (< 5 Hz) are not high for small earthquakes, we used the observed velocity
 577 spectra from 5 Hz to 20 Hz to estimate moment magnitudes for regular earthquakes with

578 $1.0 \leq M_{jma} \leq 1.9$. Figure 14-a shows the relationship between M_{jma} and M_w . M_w tended
 579 to be greater than M_{jma} for all three types of earthquakes: regular earthquakes ($M_{jma} <$
 580 3.5), deep LFEs, and shallow LFEs. However, the discrepancy between M_w and M_{jma} for the
 581 LFEs was larger than that for the regular earthquakes.

582 Some previous studies indicated that M_w for regular earthquakes with $M_L < 4$ (M_L
 583 indicates local magnitude) tends to be larger than M_L (e.g., Grünthal & Wahlström, 2003;
 584 Edwards et al., 2010; Goertz-Allmann et al., 2011a; Munafò et al., 2016; Ross et al., 2016;
 585 Uchide and Imanishi, 2018). Our results are consistent with these previous studies. We
 586 evaluated the approximate relationship between M_w and M_{jma} for small regular
 587 earthquakes with $1.0 \leq M_{jma} \leq 4.0$. Because it does not appear to have a linear
 588 relationship, we assumed the form of $M_w = aM_{jma}^2 + bM_{jma} + c$. We obtained $a = 0.49 \pm$
 589 0.03 , $b = 0.35 \pm 0.01$ and $c = 1.65 \pm 0.03$, respectively, which fits the observation well
 590 (Figure 14-a), by the least-square method.

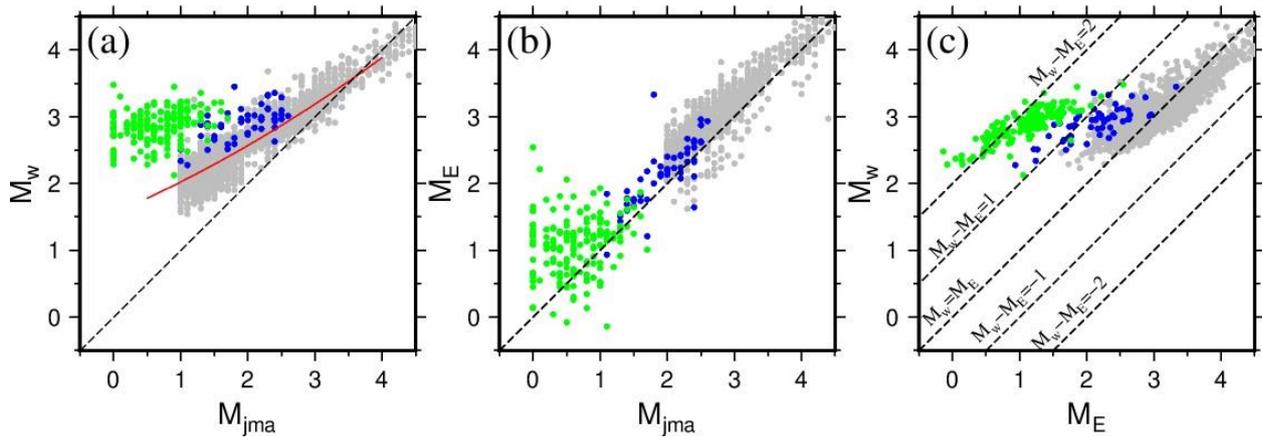
591 We estimated the energy magnitude (M_E), defined as $M_E = (\log_{10} E_R - 4.8)/1.5$
 592 (Choy and Boatwright, 1995) using the radiated energy (E_R) [unit: J]. Figure 14-b shows the
 593 relationship between M_{jma} and M_E . Although the obtained M_E values are scattered, we
 594 found generally good agreement between M_{jma} and M_E for the three types of earthquakes.
 595 The observed nonlinear relationship between M_w and M_{jma} , on the other hand, may
 596 indicate that M_{jma} based on the maximum amplitude of seismic waves, reflects the radiated
 597 energy rather than the seismic moment. The obtained M_w of LFEs is significantly larger than
 598 M_{jma} because LFEs radiate less energy than regular earthquakes with similar seismic
 599 moments, as indicated by e_R .

600 Figure 14-c shows the relationship between M_w and M_E . For regular earthquakes,
 601 M_w and M_E showed good agreement. On the other hand, for the LFEs, M_w was 1–2, which
 602 was systematically larger than M_E . This trend can be understood by the difference in the
 603 scaled energy between regular earthquakes and LFEs. We can write the $M_w - M_E$
 604 relationship with e_R (Choy and Boatwright, 1995) as follows:

$$605 \quad M_w - M_E = -\frac{2}{3}(\log e_R + 4.3) \quad (14)$$

606 Equation (14) indicates that M_w is equal to M_E when e_R value is approximately 5×10^{-5}
 607 and the discrepancy between M_E and M_w increases as e_R values decrease. The generally
 608 good agreement between M_w and M_E for the regular earthquakes reflect that e_R values for
 609 regular earthquakes are in the order of 10^{-5} to 10^{-4} as shown in Figure 9. This is
 610 reasonable because Kanamori (1977) originally defined M_w to satisfy the Gutenberg-Richter
 611 magnitude-energy relation (Gutenberg, 1956) assuming e_R values to be approximately
 612 5×10^{-5} . However, the large discrepancy in the LFEs may reflect the fact that the e_R values
 613 of the LFEs is systematically one to three orders of magnitude smaller than those of regular
 614 earthquakes. This result suggests that for LFEs, the local magnitude may provide relatively
 615 good information for the radiated energy but not for the seismic moment.

616



617

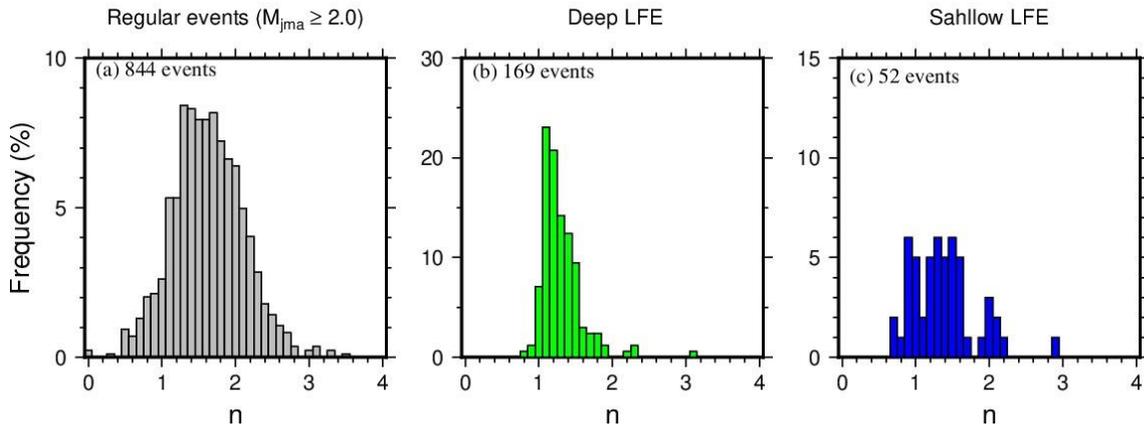
618 Figure 14. The estimated moment magnitude (M_w) and energy magnitude (M_E). (a) M_w as a
 619 function of M_{jma} . The red line indicates the best fit curve of $M_w = aM_{jma}^2 + bM_{jma} + c$. (b)
 620 M_E as a function of M_{jma} . (c) M_w as a function of M_E . Gray, green, and blue circles indicate
 621 the regular earthquakes, deep and shallow LFEs, respectively.

622 4.4. Spectral high-frequency falloff rate n as a clue to source complexity

623 Finally, the frequency characteristics of the derived source spectra were
 624 investigated. This section focuses on the high-frequency spectral falloff rate (n) estimated in
 625 section 2-5-3. Because of the narrow frequency band, we could not estimate n values stably
 626 for earthquakes whose estimated corner frequencies were larger than 10 Hz. Thus, we
 627 evaluated only n values for the earthquakes with estimated corner frequencies lower than
 628 10 Hz. We estimated n values for 844 regular earthquakes, 169 deep LFEs, and 52 shallow
 629 LFEs. Figure 15 shows the histograms of the estimated n values and Table 2 shows the
 630 mean, standard error, and standard deviation of the obtained n values. For regular
 631 earthquakes, the mean values of obtained n was 1.61, which was smaller than the value of
 632 n assumed in the omega-square model ($n = 2$). Previous studies reported the existence of
 633 regular earthquakes with $n < 2$ (e.g., Allmann and Shearer, 2009; Trugman and Shearer,
 634 2017; Shearer et al., 2022). The mean n value in this study was similar to that reported by
 635 Allmann and Shearer (2009). For earthquakes associated with complex fault ruptures such
 636 as multiple shock, the frequency-characteristic of source spectrum becomes more complex,
 637 deviating from the ω^2 model (e.g., Madariaga, 1979) and high-frequency energy radiation
 638 increases, which may be related to the smaller n values compared with the ω^2 model.
 639 Because of the presence of earthquakes associated with complex rupture processes, the
 640 average n values for regular earthquakes may be less than two.

641 In contrast, the mean values of the obtained n is 1.34 for 169 deep LFEs, and is 1.41
 642 for 52 shallow LFEs. Thus, LFEs tend to have smaller n values than regular earthquakes,
 643 which is consistent with previous studies on LFEs at plate boundaries (e.g., Ide et al., 2007a)
 644 and continental plates (e.g., Yoshida et al., 2020). However, it should be noted that n values
 645 must be greater than 1.5 because the radiated energy defined by equation (12) diverges to
 646 infinity if $n \leq 1.5$. Thus, the estimated n values smaller than 1.5 for the LFEs probably
 647 represent the characteristics of the source spectrum for a limited frequency band.

648 Compared to regular earthquakes, the average n values for the LFEs are smaller,
 649 which may suggest that the difference in frequency characteristics between regular
 650 earthquakes and LFEs results from not only the source corner frequency but also the
 651 spectral high-frequency falloff rate above the corner frequency. The smaller n values for the
 652 LFEs are seemingly inconsistent with their smaller radiated energy. However, the source
 653 corner frequencies of the LFEs were lower than those of regular earthquakes with similar
 654 magnitudes. Our results suggest that the high-frequency energy of LFEs is small owing to the
 655 small corner frequency, probably due to lower slip and/or deformation rates compared with
 656 regular earthquakes, rather than n value. Previous studies have suggested that deep LFEs
 657 may be related to multiple subevents such as a chain of tensile-shear cracks (Ikegaya and
 658 Yamamoto, 2021) and fluid resonance after shear faulting (Hensch et al., 2019). The
 659 systematically smaller n values for the LFEs suggest that LFEs are composed of complex
 660 rupture and/or deformation processes that radiate multiple low-frequency pulses. However,
 661 it should be noted that the upper limit of the currently available frequency band in this
 662 study is 20 Hz; therefore, the complexity of the spectrum on the high-frequency side
 663 becomes more distinct for events with a smaller source corner frequency (f_c) (Figure B-1).



664
 665 Figure 15. Estimated spectral high-frequency falloff rate n . (a), (b), and (c) indicate the
 666 histogram of the obtained n for the regular earthquakes, deep, and shallow LFEs,
 667 respectively.

668 Table 2. Mean, standard error and standard deviation of the obtained n

	Mean	Standard error	Standard deviation
The regular earthquakes	1.61	0.02	0.49
The deep LFEs	1.34	0.08	0.28
The shallow LFEs	1.41	0.10	0.43

669

670 5. Conclusion

671 In this study, we estimated the source spectra and obtained the scaled energy for
 672 1469 regular earthquakes with $M_{jma} \geq 2.0$, 169 deep LFEs, and 52 shallow LFEs detected by
 673 Nakajima and Hasegawa (2021) in and around the focal area of the 2008 M_w 6.9 Iwate-
 674 Miyagi earthquake. The scaled energy (e_R) for regular earthquakes are in the order of 10^{-5}
 675 to 10^{-4} , which is consistent with previous studies. Additionally, our results show that

676 regular earthquakes with relatively smaller e_R tend to occur near volcanoes and close to the
 677 location of shallow LFEs, suggesting that crustal fluid may be related to seismic radiation. On
 678 the other hand, e_R is of the order of 10^{-7} for deep LFEs, and 10^{-6} for shallow LFEs. The
 679 obtained systematic difference did not change even if accounting for the depth variation of
 680 the attenuation factor Q^{-1} . In addition, the spectral high-frequency falloff rates for LFEs
 681 tended to be lower than those for regular earthquakes. These results suggest that LFEs are
 682 associated with complex rupture and/or deformation processes with lower stress drops
 683 and/or slower rupture and/or deformation rates than those of regular earthquakes.

684 Although we found generally good agreement between the local magnitudes and
 685 energy magnitudes derived from the radiated energy for the three types of earthquakes, the
 686 discrepancy between the local magnitudes and moment magnitudes was significantly large
 687 for the LFEs. This result suggests that for LFEs, the local magnitude provides relatively good
 688 information about the radiated energy but does not provide information about the seismic
 689 moment.

690 **Acknowledgment**

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 692 Research Institute for Earth and Science and Disaster Resilience), Japan Meteorological
 693 Agency, and Tohoku University. M.O. thanks Mare Yamamoto for comments and discussions
 694 on the LFEs. All figures in this study were created using GMT (Wessel and Smith, 1998)

695 **References**

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938

939 Appendix-A

940 The procedure for the spectral ratio method

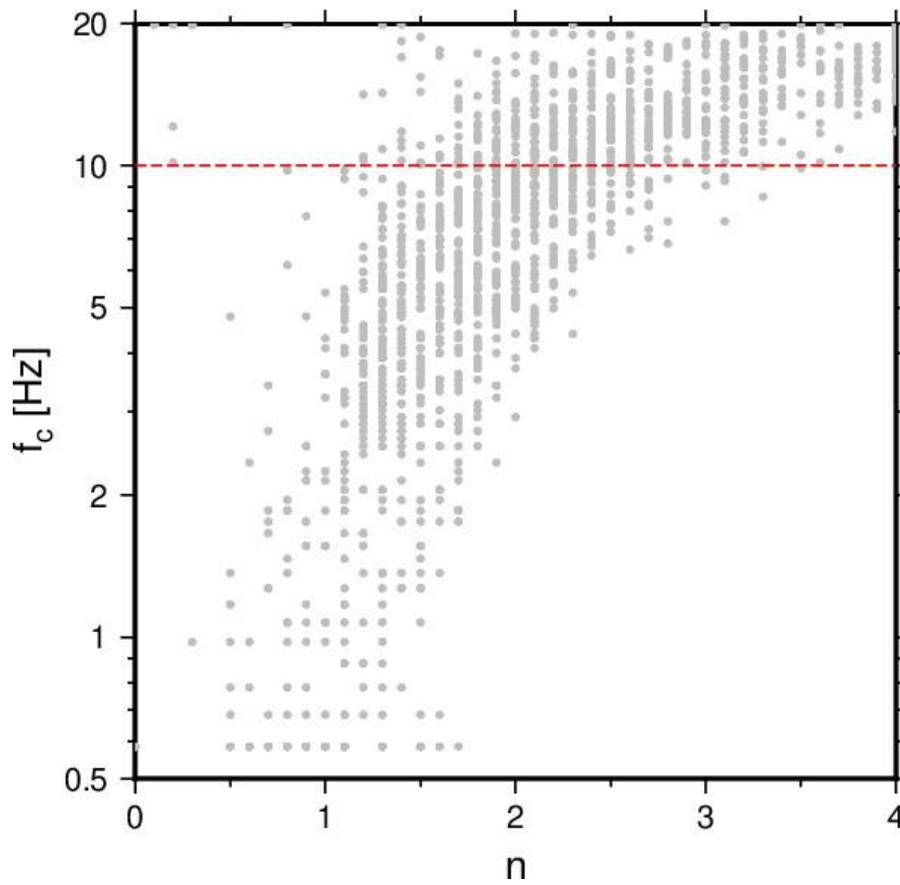
- 941 1. Select EGFs. We selected small events with magnitude differences greater than 1 which
942 occurred within 3 km from the reference events as EGFs. Prior to this, we removed the
943 reference events which have less than 10 EGFs and did not compute source corner
944 frequency.
- 945 2. Calculate spectral ratio. We calculated the spectral ratio of the reference-EGF pairs for
946 each station and component, and obtained representative spectral ratios using their
947 geometric mean.
- 948 3. Estimate source corner frequency. We estimated source corner frequency of the
949 reference events and EGFs by fitting model source spectra assuming the ω^2 model to
950 the derived spectral ratio. Source corner frequencies were estimated at 0.1 Hz intervals
951 from 0.5 Hz to 20 Hz using a grid search method.

952

953 Appendix-B

954 The relationship between estimated source corner frequency (f_c) and high-frequency falloff
955 rate n

956



957

958 Figure B-1 The relationship between the estimated source corner frequency (f_c) and

959 spectral high-frequency falloff rate (n) for regular earthquakes.