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Revisiting the source rupture process of the mainshock of the 2016

Kumamoto earthquake and implications for the generation of near-

fault ground motions and forward directivity pulses

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ABSTRACT

Strong near-fault ground motions associated with the M_{JMA} 7.3 mainshock of the 2016 Kumamoto earthquake sequence, central Kyushu, Japan, have received attention by seismological and engineering communities. In this study, the kinematic source rupture process was reanalyzed based on an improved approach for the representation of source faults. The slips at densely distributed point sources were defined via the bilinear interpolation of those at surrounding control points. The result shows that the rupture started on the Hinagu fault with a small initial rupture and propagated beyond the junction to the Futagawa fault. The rupture on the Futagawa fault mainly propagated up- and northeastward. A large slip area with a peak slip of 4.9 m and peak slip-velocity of 3.1 m/s was detected at depths ranging from 3 to 15 km in the central part of the Futagawa fault. This asperity spatially coincides with a body with moderate seismic velocity ($V_{\mathrm{P}} \sim 6$ km/s) and low seismic attenuation. The slips on the Futagawa fault have significant normal-slip components, whereas the slip vectors of the Hinagu fault represent almost pure right-lateral strike-slip. The shallower part of the fault segments in the western Aso caldera is characterized by relatively large normal slips. The estimated slip velocity functions at shallower depths (<3 km) are almost temporally symmetric and relatively long. The shallower portion of the source fault significantly contributes to the velocity and displacement waveforms at near-fault sites. On the contrary, the slip velocity functions at deeper depths (>3 km) are temporally asymmetric and have a sharp peak. The simulation of

35   the ground motion evolution suggests that the lateral flow in the Aso Valley was primarily

36   triggered by the strong forward-directivity pulse generated from the asperity on the Futagawa

37   fault.

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INTRODUCTION

The M_{JMA} 7.3 mainshock of the 2016 Kumamoto earthquake sequence occurred in the Kumamoto Prefecture, central Kyushu, Japan (Figure 1) at 01:25 Japan Standard Time (JST; UTC+09:00) on April 16, 2016, ~28 h after the first M_{JMA} 6.5 earthquake on April 14, 2016. It was a crustal earthquake that occurred along major active faults of the Futagawa and Hinagu fault systems (e.g., Kumahara *et al.*, 2016; Shirahama *et al.*, 2016), which are developed in areas with high inelastic strain rates in the center of Kyushu Island (Matsumoto *et al.*, 2016). This right-lateral strike-slip earthquake had a normal-slip component, which is thought to be due to lateral heterogeneous and depth-dependent stress fields (Matsumoto *et al.*, 2018). Surface ruptures associated with this earthquake were widely observed in the area from the Kumamoto Plain to the Aso caldera. Shirahama *et al.* (2016) surveyed surface ruptures in the entire area and reported that they occurred along ~34 km of the Futagawa fault and the northernmost portion of the Hinagu fault (northern part of the Takano–Shirahata segment). They also reported a large slip, with a maximum dextral slip of 2.2 m, in the central section of the rupture zone along the Futagawa fault and that the surface rupture along the Futagawa fault extended to the western rim of the Aso caldera.

The ground motions during the 2016 Kumamoto earthquake were recorded by nationwide strong motion networks (e.g., Suzuki *et al.*, 2017) as well as local observation networks of seismic intensity (Japan Meteorological Agency, 2018). Suzuki *et al.* (2017) summarized the

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58   features observed in strong motion records from strong motion networks K-NET and KiK-net

59   of the National Research Institute of Earth Science and Disaster Resilience (NIED) and

60   reported that large peak ground accelerations and velocities extended northeastward from the

61   epicentral area. Near-fault ground motions observed at several strong motion stations along the

62   Futagawa and Hinagu faults have received attention by seismological and engineering

63   communities (e.g., Furumura, 2016; Iwata and Asano, 2016; Kobayashi *et al.*, 2017; Ikutama

64   *et al.*, 2018; Kidoh and Nagano, 2020). Many studies have been conducted to investigate the

65   physical causes of extremely strong ground motions and building damage distribution

66   considering the effects of shallow soft sedimentary layers using aftershock and microtremor

67   observations in the damaged area (e.g., Yamanaka *et al.*, 2016; Kawase *et al.*, 2017; Yamada *et*

68   *al.*, 2017; Sun *et al.*, 2020).

69   Regarding the source effects, Asano and Iwata (2016) estimated a source rupture model for

70   the mainshock by the kinematic inversion of waveform data at 15 strong motion stations. They

71   revealed that the rupture started in the deep portion of a northwest-dipping fault in the northern

72   part of the Hinagu fault and propagated beyond the junction to the Futagawa fault. Most of the

73   significant slips (up to 5.1 m) occurred on the Futagawa fault. The source rupture process has

74   also been estimated in several other studies using kinematic waveform inversion of strong

75   motion data (Kubo *et al.*, 2016; Uchide *et al.*, 2016; Nozu and Nagasaka, 2017; Yoshida *et al.*,

76   2017; Hallo and Gallovič, 2020). The number of assumed fault segments varied from one to

77   four, except for Kubo *et al.* (2016) who used a curved fault model. Somei *et al.* (2020) modeled

78   three strong motion generation areas to simulate broadband strong ground motions in 0.2–10

79   Hz. Yagi *et al.* (2016) analyzed teleseismic P-wave data to estimate the rupture process of a

80   single fault plane along the Futagawa fault. Hao *et al.* (2017) jointly analyzed teleseismic and

81   strong motion displacement waveform data to estimate the kinematic rupture process of two

82   fault planes corresponding to the Futagawa and Hinagu faults. Fukahata and Hashimoto (2016)

83   and Himematsu and Furuya (2016) obtained slip models based on the geodetic inversion of

84   interferometric synthetic aperture radar (InSAR) data. Fukahata and Hashimoto (2016)

85   assumed two fault planes along the Futagawa and Hinagu faults and detected the largest slip of

86   >5 m with a significant normal-slip component close to the center of the Futagawa fault

87   (130.9°E). Himematsu and Furuya (2016) established a fault model with three segments and

88   concluded that slip partitioning occurred in an oblique extensional stress regime. Ozawa *et al.*

89   (2016) estimated four rectangular faults to explain InSAR and Global Navigation Satellite

90   System (GNSS) data using a dislocation model. Two were subparallel faults along the

91   Futagawa fault, one was a right-lateral strike slip fault with a high dip angle, and the other one

92   included a large normal-slip component with a low dip angle. Zhang *et al*. (2018) analyzed

93   InSAR, Global Positioning System (GPS), and strong motion data by assuming three fault

94   segments and Yue *et al.* (2017) analyzed GPS, strong motion, InSAR, and surface offset data

95   to obtain the kinematic rupture process for three curved fault models. Kobayashi *et al.* (2017)

96   carried out the joint inversion of teleseismic, strong motion, and GNSS data assuming five fault

97   segments including two subparallel faults in the central part of the Futagawa fault. Published

98   source models commonly include nearly unilateral rupture propagation from the hypocenter

99   near the junction or bend of the Hinagu and Futagawa faults into the northeastern direction

100   along the Futagawa fault. With respect to the generation mechanism of near-fault strong ground

101   motions, Kobayashi *et al.* (2017) demonstrated that extreme pulse-like ground motion

102   waveforms at near-field stations are attributed to the upward rupture directivity, slip rate, and

103   nearly simultaneous slip of two subparallel faults. However, the highest frequency of their

104   waveform analysis was limited to 0.4 Hz. The target frequency range in such studies should be

105   extended to higher frequencies to discuss the correlation between the source description and

106   the resultant ground motions.

107   Some of the above-mentioned studies focused on the correlation between the source rupture

108   process and geothermal structure of the Aso volcano. It has been concluded that the rupture of

109   the mainshock was terminated by a material barrier with high temperature and low seismic

110   velocity, which might be related to the presence of partial melting (Ozawa *et al.*, 2016; Yagi *et*

111   *al.*, 2016; Yue *et al.*, 2017; Zhang *et al.*, 2018). Seismic velocity tomography (Shito *et al.*, 2017;

112   Zhao *et al.*, 2018; Aoyagi *et al.*, 2020), seismic attenuation tomography (Komatsu *et al.*, 2017;

113   Wang *et al.*, 2017), and electrical resistivity structure surveys (Aizawa *et al.*, 2021) have been

114   conducted to quantitatively determine the correlation between the heterogeneous earthquake

115   source process and crustal structure. Information about this correlation is important for future

116   seismic hazard assessments because it could reduce the diversity in the source rupture scenarios,

117   which largely affects the ground motion prediction, and thus lead to more seismologically

118   reasonable ground motion predictions and seismic hazard assessments.

119   In this study, we used kinematic waveform inversion to reveal the fault rupture process of the

120   mainshock of the 2016 Kumamoto earthquake sequence. The multiple time-window linear

121   waveform inversion method (e.g., Hartzell and Heaton, 1983) was used to estimate the

122   kinematic rupture process, which was also employed in Asano and Iwata (2016). Based on this

123   method, the fault is divided into several subfaults and the point source is assumed to be in the

124   center of each subfault. The subfault dimensions are defined after considering the spatial

125   resolution of the data (e.g., Ammon *et al.*, 2020). However, the point sources should be densely

126   distributed across the fault plane to accurately reproduce the finite source effects in the near-

127   fault area because near-fault ground motion is quite sensitive to the discretization of the source

128   fault (e.g., Hisada and Bielak, 2003; Kidoh *et al.*, 2020; Wu *et al.*, 2021). The above-mentioned

129   papers suggest that the subfault size or point-source interval should be as small as 0.5 times the

130   fault distance to simulate accurately near-fault ground motions. On the other hand, the subfault

131   size used in Asano and Iwata (2016) was 2.0 km and the top depth of the source fault model

132   was about 2 km, and such setting was not necessarily efficient to reproduce ground motion

133   including permanent offset in the area very close to the surface ruptures. Therefore, we tried to

134   improve the representation of the source fault model in this study.

135   The improvement on that point is also required for quantitatively discussing the shape

136   and duration of slip velocity time function because the moment-rate function with a coarse

137   subfault size (2.0 km both in length and width) obtained in Asano and Iwata (2016) is a

138   convolution of local slip-velocity function and rupture propagation effect within the finite-sized

139   subfault. The shallowest discretized point-source in Asano and Iwata (2016) was located at

140   about 2.9 km depth. Therefore, it was not easy to discuss about depth-dependency of slip

141   velocity function directly from the obtained subfault moment-rate functions across the source

142   fault because long duration of the moment-rate function could also be explained to some extent

143   by long rupture duration. The characteristics of the slip velocity function can be analyzed in-

144   depth by setting dense distribution of element point-sources in the kinematic inversion scheme,

145   but it is necessary to avoid handling of an excessively large number of unknown parameters.

146   Thus, we propose an improved representation of the source fault model, which is suitable for

147   the reproduction of near-fault ground motions, and we also used the latest velocity structure

148   models in the target region, which was not available in 2016. The source fault planes were

149   extended to near the ground surface considering the geometry based on the latest study of the

150   aftershock distribution and the surface rupture distribution. The target frequency range was also

151   extended to higher frequencies than used in our previous work because the absolute amount of

152   slip-velocity function would also be affected by the width of the analyzed frequency range. Of

153   course, we expected that the general feature of the source process such as the location of large

154   slips will not change significantly because the original result was appropriately constrained by

155   many stations distributed in wide area.

156   The main focus of this study was placed on the depth dependency of the slip velocity

157   function of the source fault and its correlation with the crustal structure. The spatial relationship

158   between the whole rupture process and the known crustal structure or tectonic settings were

159   also addressed. Such studies may help to define the rupture area and the rupture scenario for

160   the seismic hazard modeling in advance. In addition, the forward-directivity pulse originating

161   from the heterogeneous source process of this earthquake and its contribution to the lateral flow

162   widely observed in the Aso Valley was discussed based on the estimated kinematic source

163   model and simulated ground motions.

164

165   **KINEMATIC WAVEFORM INVERSION AND FAULT PLANE MODEL**

166   Wald and Heaton (1994) proposed a method based on which the Green's function of the

167   subfault was obtained by summing up the responses of many point sources that were uniformly

168   distributed across the subfault to consider rupture front propagation within the subfault. Their

169   method requires the summation of Green's functions of individual point sources within a

170   subfault with appropriate rupture time delays based on the assumption of the velocity and

171   direction of the rupture front propagation in advance of starting the source inversion analysis.

172   Sekiguchi *et al.* (2002) alternatively proposed a convolution method in which a function

173   representing a bi-directional moving dislocation on each subfault was convolved with the

174   point-source synthetic wave from each subfault center.

175   In this study, we employed a slightly different approach. A schematic illustration of the

176   source fault representation is shown in Figure 2. Point sources were densely distributed across

177   the assumed fault planes, at intervals of 0.22 km. The relationship between the point-source

178   slip vector $\mathbf{m}_P$ and data vector $\mathbf{d}$ can be expressed as follows:

179                                  $$\mathbf{d} = \mathbf{G}\mathbf{m}_P, \tag{1}$$

180   where the matrix $\mathbf{G}$ contains the Green's functions from all point sources to the stations.

181   Unknown model parameters (slips in two orthogonal directions in each time window) were

182   assigned only to control points (larger circles in Figure 2) on the fault planes distributed at

183   intervals of 2.0 km in the strike and dip directions to avoid the introduction of excessive

184   complexity to the inverted source model. The slip of each individual point source in each time

185   window was then obtained by spatial bilinear interpolation of the slip amounts at the control

186   points. For example, the slip amount $m_{jk}(x, y)$ in the k-th direction in the j-th time window

187   of a point source located at (x, y) on a fault is given by the slip amounts of the four

188   surrounding control points $(x_1, y_1), (x_1, y_2), (x_2, y_1),$ and (x_2, y_2):

$$m_{jk}(x, y) = (1 - a)(1 - b)m_{jk}(x_1, y_1) + (1 - a)bm_{jk}(x_1, y_2)$$
$$+a(1 - b)m_{jk}(x_2, y_1) + abm_{jk}(x_2, y_2),$$
$$a = \frac{x - x_1}{x_2 - x_1},$$
$$b = \frac{y - y_1}{y_2 - y_1}, \tag{2}$$

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190 where x and y are coordinates along the strike and dip directions of the fault, respectively. Thus,

191 $\mathbf{m}_P$ can be linearly expressed with a model parameter vector $\mathbf{m}_C$ and matrix $\mathbf{A}$ containing

192 the coefficients required for spatial interpolation shown in Eq. (2).

193 $$\mathbf{m}_P = \mathbf{A}\mathbf{m}_C. \tag{3}$$

194 The observation Eq. (1) becomes:

195 $$\mathbf{d} = \mathbf{G}\mathbf{A}\mathbf{m}_C. \tag{4}$$

196 Therefore, the total number of unknown model parameters, which is the size of $\mathbf{m}_C$, is the

197 same as that of conventional kinematic source inversions and overparameterization can be

198 avoided.

199 We referred to the fault geometry models proposed by Mitsuoka *et al.* (2020) to set up the

200 fault plane model for the waveform inversion. They located hypocenters from 1993 to 2017

201 using permanent and temporary seismic stations in central Kyushu and derived the fault

202 geometry from the spatial distribution of the aftershocks. The fault plane model in our inversion

203 analysis was slightly modified to be consistent with surface rupture traces. It was composed of

204 five planar rectangular fault planes (Figure 1, Table 1). The surface projection of the fault

205 models is plotted with the aftershocks within 6 hours located by Mitsuoka *et al.* (2020) and the

206   surface rupture distribution compiled by Kumahara *et al.* (2016) in Figure 3a. Most aftershocks

207   in this period distributed in the depth range from 3 km to 18 km. Faults H1 and H2 have the

208   same strike angle but different dip angles. Fault H1 corresponds to the deeper portion of the

209   Hinagu fault and fault H2 is a vertical fault segment corresponding to the shallower portion of

210   the Hinagu fault as illustrated in Figure 3b. The aftershocks along the Hinagu Fault (H1 and

211   H2) did not distribute to the east of the Hinagu fault, therefore, we decided to assume that the

212   dip angle is nearly vertical in the shallow part of the Hinagu fault (H2). Faults F1, F2, and F3

213   are located along the Futagawa fault, and they are dipping northwestward in accordance with

214   the aftershock distribution in Figure 3c.

215   The rupture starting point was fixed at the hypocenter (32.758179°N, 130.766740°E,

216   12.808 km depth) reported by Mitsuoka *et al.* (2020) and was assumed to be on fault H1. The

217   total length of the assumed fault model was 44 km. All fault parameters are summarized in

218   Table 1. The total numbers of control points and point sources were 220 and 17,028,

219   respectively. The number of time windows was nine. The basis function of each time window

220   was a smoothed ramp function with a rise time of 1.0 s and each successive time window was

221   shifted by 0.5 s in the time domain. The slip direction at each control point was limited to −142°

222   ± 45° using the non-negative least squares method (Lawson and Hanson, 1974). The rupture-

223   front propagation velocity triggering the first time-window was searched between 2.0 km/s and

224   3.0 km/s, and found a velocity giving the minimum residual among them. Spatiotemporal

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225 smoothing constraint equations were introduced following Sekiguchi *et al.* (2000). The

226 appropriate weight of the smoothing constraints with respect to the observation equation was

227 determined by minimizing the Akaike Bayesian Information Criteria (ABIC, Akaike 1980).

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WAVEFORM DATA AND GREEN'S FUNCTIONS

230   Strong ground motion waveform data around the source region were collected from the K-

231 NET, KiK-net, and F-net of NIED (Aoi *et al.*, 2020; NIED, 2019a, b), JMA earthquake

232 observation network, and seismic intensity observation network of the Kumamoto prefectural

233 government (Japan Meteorological Agency, 2018). Three-components of velocity waveforms

234 in a 30-s window starting 1 s before the S-wave onset at twenty strong motion stations were

235 used in this study (Figure 1). We used the records from the downhole sensors of the NIED KiK-

236 net stations and the records on the ground surface of other stations. Original acceleration

237 waveform data were integrated into velocity in the time domain, except for stations belonging

238 to NIED F-net, which was equipped with velocity-type strong motion sensors (TOKYO KEIKI

239 TSM-1). All velocity waveforms were bandpass-filtered between 0.05 and 1 Hz using the

240 Chebyshev filter and then resampled at 5 Hz.

241   The theoretical Green's function from each point source to a strong motion station was

242 calculated using the discrete wavenumber method (Bouchon, 1981) and reflection and

243 transmission matrix method (Kennett and Kerry, 1979). A station-dependent layer-cake model

244   was assumed for each station to consider local site amplification effects of the sedimentary

245   layers (e.g., Asano and Iwata, 2009). These one-dimensional velocity structure models were

246   extracted from the latest three-dimensional velocity structure models of the target region. The

247   velocity model for the sedimentary layers was based on three-dimensional models recently

248   developed by Senna *et al.* (2018) and Asano *et al.* (2019). Senna *et al.* (2018) constructed a

249   three-dimensional velocity model for the Kumamoto Plain and surroundings using phase

250   velocity dispersion curves and H/V spectra from microtremor observations, boring logs, and

251   gravity data. Asano *et al.* (2019) constructed a three-dimensional velocity model covering the

252   Yatsushiro Plain, Tamana Plain, Amakusa Islands, and other areas in the Kumamoto Prefecture

253   using P-wave reflection surveys, phase velocity dispersion curves and H/V spectra from

254   microtremor observations, boring logs, and gravity data. Because these velocity models were

255   developed above the seismic bedrock (S-wave velocity 3.1 km/s), we referred to the Japan

256   Integrated Velocity Structure Model Version 1 (JIVSM, Koketsu *et al.*, 2012) for the crust and

257   upper mantle structures. Thus, the velocity model used to calculate the Green's function

258   differed from that used in Asano and Iwata (2016). The lowest S-wave velocity of the velocity

259   model in the present study is 0.1 km/s, which corresponds to alluvial deposits in Quaternary

260   sedimentary plains. The S-wave velocity of the topmost layer at each station varies from 0.10

261   km/s to 0.35 km/s. The velocity models assigned to each station are summarized in the

262   Electronic Supplement (Tables S1–S20, available in the supplemental material to this article).

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263 The theoretical Green's functions were bandpass-filtered and resampled with the same

264 procedure as that used for the observed waveforms.

265

266 **RESULTS**

267 The final slip distribution including slip vectors is shown in Figure 4a. The propagation

268 velocity of the first time-window front of this model is 2.4 km/s, yielding the minimum residual.

269 The average and maximum slip amounts over the entire fault are 1.9 and 4.9 m, respectively.

270 The slip velocity functions estimated for each control point are shown in Figure 4b and the

271 maximum peak slip velocity is 3.1 m/s. The characteristics of the slip velocity functions of the

272 faults are discussed in the following section. The total seismic moment is 4.89×10^{19} Nm (M_W

273 7.06). The fitting results for the waveforms, which are satisfactory for most stations, are shown

274 in Figure 5.

275 The overall slip distribution on the fault is similar to that of our previous model (Asano and

276 Iwata, 2016). The slip amount in the vicinity of the hypocenter of the mainshock is relatively

277 small, which indicates that the rupture started with a small initial rupture and intensified after

278 propagating along the Futagawa fault. The temporal rupture evolution presented in Figure 6

279 shows that the rupture of the Futagawa fault started after ~3 s in the deep portion of the fault

280 and propagated up- and northeastward. The slip in the shallow portion of the fault was delayed

281 compared with the deeper portion but not isolated. A large slip of 4–5 m was observed in the

282   depth range from ~3 to 15 km in the central part of the Futagawa fault. This large-slip area or

283   asperity has significant normal-slip components and a high slip velocity of 2–3 m/s. The

284   shallower part of the fault segments inside the western part of the Aso caldera also has a

285   relatively large normal slip (Figure 7) and it is consistent with surface displacements identified

286   during field surveys (e.g., Shirahama *et al.*, 2016; Toda *et al.*, 2019) and by satellite-based

287   geodetic measurements (e.g., Fujiwara *et al.*, 2016; Fukahata and Hashimoto, 2016;

288   Himematsu and Furuya, 2016; Ozawa *et al.*, 2016).

289   The slip of the Hinagu fault segments is nearly pure right-lateral strike-slip. The slip of the

290   Hinagu fault segments is concentrated in the northeastern part of the vertical fault segment,

291   which is consistent with the field observation of the surface rupture distribution along the

292   Hinagu fault (e.g., Shirahama *et al.*, 2016). The Hinagu fault also ruptured during the first

293   foreshock on April 14, 2016. Asano and Iwata (2016) established a slip model for this foreshock,

294   but the slips in the shallow portion of the foreshock and mainshock look complementary to

295   each other.

296

DISCUSSIONS

Spatial variation of the source rupture characteristics and crustal structure

299   The slip velocity on the fault is one of the key physical quantities controlling the generation

300   of a strong ground motion pulse (e.g., Miyake *et al.*, 2003; Gombert *et al.*, 2019; Wang *et al.*,

301   2019). As briefly mentioned in the introduction, strong ground motion pulses and permanent

302   displacements at near-fault sites during the mainshock of the 2016 Kumamoto earthquake

303   sequence have been modeled in many studies (e.g., Kobayashi *et al.*, 2017; Ikutama *et al.*,

304   2018; Tanaka *et al.*, 2018; Irikura *et al.*, 2020; Kidoh and Nagano, 2020). Kobayashi *et al.*

305   (2017) explained near-fault ground motions by combining the effect of the up-dip rupture

306   direction with the high slip velocity. Ikutama *et al.* (2018) proposed an approach based on

307   which the entire ruptured fault from the ground surface to the bottom of the fault was modeled

308   to reproduce strong ground motions and permanent displacements at sites close to the surface

309   fault trace. In their approach, the asperity also extends to the near-surface layer and the source

310   parameters of the entire fault follow the strong motion prediction recipe by Irikura and Miyake

311   (2011). Tanaka *et al.* (2018) proposed the use of regularized Yoffe-type slip velocity functions

312   in the shallower region (<3 km) above the seismogenic layer to evaluate long-period ground

313   motions with permanent displacements in the near-fault region. However, the asperity extends

314   into the shallow and seismogenic layers with a constant slip amount, similar to that reported

315   by Ikutama *et al.* (2018). Irikura *et al.* (2020) proposed an extreme source model in which the

316   rectangular area, that is, the long-period motion generation area (LMGA), is distributed in the

317   shallow depth range above the seismogenic layer. However, the physical meaning of the LMGA

318   remains unclear and its location and size depend on the availability of near-fault strong motion

319   stations. Our final slip model (Figure 4a) does not indicate any isolated slip patch in the

320   shallower depth range, but it seems reasonable to consider the slips at shallow depths as spatial

321   extensions of the large fault slip in the seismogenic depth.

322   The slip velocity functions estimated for the control points of the Futagawa fault segments

323   are plotted in Figure 8. All control points belonging to one depth bin were plotted together in

324   one panel. Because the peak time of the slip velocity varies among subfaults due to the variation

325   in the rupture time, the slip velocity functions of each bin are aligned by the peak time. The

326   shapes of the slip velocity functions in the first two depth bins (depth <3 km) are almost

327   temporally symmetric and resemble the temporal derivative of a ramp function. The duration

328   is also relatively long compared with that of deeper depth bins. The peak time relative to the

329   onset time is distributed around 2 s for the shallow two depth bins. The average peak time

330   weighted by the peak slip velocity is 2.0 s for the depth bin of 0.3–0.5 km and 1.3 s for the

331   depth bin of 2.2–2.4 km. There are some subfaults where the peak time of slip velocity is

332   relatively faster than others, which might reflect spatial variation in material and frictional

333   parameters in this depth range. Larger slip-weakening distance (D_C) and small strength excess

334   in the shallow layer zone are possible factor making the slip-velocity function smooth and long

335   (e.g., Dalguer *et al.*, 2020). The free surface effect will also take additional role in generating

336   long rise time (e.g., Wang and Day, 2020). On the other hand, the slip velocity functions of the

337   rest of the eight depth bins, which are thought to be within the typical depth of the seismogenic

338   layer of Japanese crustal earthquakes, are temporally asymmetric and have a sharp peak. They

339   are similar to the Kostrov-type slip velocity function, which is typically expected based on

340   dynamic fault rupture modeling. The weighted average peak time is 0.7–1.0 s, and the majority

341   of the subfaults with large peak slip velocity have peak time of approximately 0.5 s. It could

342   also be noted that some subfaults in the westernmost part of the Futagawa fault, where slip

343   amount is relatively less, tend to have symmetric slip velocity function with long duration

344   (Figures 3 and 8), suggesting existence of crustal heterogeneity in the seismogenic zone and its

345   relationship to the rupture process. We think the depth-variable shape/duration of slip velocity

346   function should be considered for source model setting in future strong motion prediction or

347   earthquake hazard analysis of inland active faults because assumption of slip-velocity functions

348   in shallower portion of the fault would strongly affect simulated ground motions in earthquake

349   hazard modeling.

350   Figure 9 shows the comparison between the observed and synthetic velocity and

351   displacement waveforms at three near-fault strong motion stations. Stations 93011 and 93048

352   were used for the kinematic source inversion, but station 93051 was excluded from the source

353   inversion analysis because a borehole station of KiK-net KMMH16 was located 660 m

354   northeast of this station, which is preferable to use for avoiding any possibility of nonlinear

355   response. Our present simulation does not include any effect of nonlinear site response, which

356   is also important for analyzing large ground motions. The locations of these stations are shown

357   in Figure 7. One-dimensional velocity models for those stations are given in Tables S13, S14,

358   and S21 in the supplemental material to this article. The permanent displacement of the

359   synthetic waveforms reproduce well the observed displacements. To compare the contributions

360   of shallower and deeper parts of the fault rupture, the synthetic ground motions were separately

361   calculated for the shallower and deeper parts of the fault and plotted in this figure. The

362   shallower part corresponds to 13 point-sources (i.e., ~3 km depth) along the dip direction from

363   the top of the fault plane and the deeper part corresponds to the rest of the fault plane. The

364   velocity pulse and permanent displacement at stations 93048 and 93051 are mostly generated

365   in the shallower part of the fault, which is consistent with previous work by Kobayashi *et al.*

366   (2017). Contribution of the deeper portion is relatively large in the fault-normal component.

367   The forward-directivity pulse generated in the deeper portion of the fault significantly

368   contributes to the velocity waveforms at another station (93011) located in the direction of the

369   forward rupture propagation, and its contribution is comparable to that of the shallower portion

370   as seen in the velocity waveforms, which has comparable peak amplitude between fault-parallel

371   and fault-normal components.

372   It is quite important to investigate relationships between rupture process and crustal

373   structure in-depth. The rupture progression and slip-velocity function could be controlled by

374   the nature of seismogenic layer such as material heterogeneity, temperature, strength, frictional

375   parameters etc. For example, Shito *et al.* (2017) demonstrated seismic velocity tomography in

376   the source region using travel time data of many small events recorded at permanent and

377 temporary seismic stations. They found that low V_P (~5.0 km/s), low V_S (~2.5 km/s), and high

378 V_P/V_S (~1.9) were widely distributed at shallow depths (0.0–2.5 km). The upper depth limit of

379 the aftershock hypocenters was ~3 km. Thus, the boundary between the surface low-velocity

380 layer and the seismogenic layer ($V_P \sim$ 6.0 km/s, $V_S \sim$ 3.5 km/s) will be at ~3 km in this area.

381 Such structural difference might make difference in dynamic source parameters between the

382 surface low-velocity layer and the seismogenic layer.

383   Shito *et al.* (2017) also suggested that the high seismic velocity at a depth ranging from 5.0

384 to 12.5 km along the source fault of the 2016 Kumamoto earthquake compared with the

385 surrounding area on Kyushu Island is associated with high crustal strength and thus is

386 indicative of regions that can sustain high tectonic stress. The depth range correlates with the

387 location of the large-slip area of the Futagawa fault. This depth range in the central part of the

388 Futagawa fault is also characterized by a high-Q_P and high-Q_S (low attenuation) zone, as

389 revealed by seismic attenuation tomography in Komatsu *et al.* (2017).

390   Aizawa *et al.* (2021) imaged the electrical resistivity structure around the source faults of

391 the 2016 Kumamoto earthquake sequence using broadband magnetotelluric data from 200 sites

392 and found that the rupture of the mainshock arrested along the western edge of a low-resistivity

393 zone with high-temperature magmatic fluids beneath the Aso volcano. Komatsu *et al.* (2017)

394 detected a low-Q_P and low-Q_S (high attenuation) zone beneath the Aso caldera, suggesting high

395 temperature in this region. Miyakawa *et al.* (2016) also reported that the magma reservoir and

396 rupture termination are correlated based on three-dimensional gravity modeling of the Aso

397 caldera area. The lack of large slips in the deep part of our source model within the caldera rim

398 agrees with their findings. Aizawa *et al.* (2021) also concluded that the southward rupture of

399 the mainshock along the Hinagu fault terminated along the northern edge of another low-

400 resistivity zone along the Hinagu fault. Their finding is also supported by Aoyagi *et al.* (2020)

401 who concluded that the southern termination of the mainshock rupture along the Hinagu fault

402 was caused by a barrier with high V_P/V_S, which corresponds to the existence of the Rokkoku

403 Tectonic Line (RTL) by the seismic velocity structure around the Hinagu fault. RTL is a

404 geologic boundary intruded by serpentine and it makes vertical offset in the seismogenic layer

405 (Yanagida, 1958; Aoyagi *et al.*, 2020). We also think that such crustal structure is a primary

406 factor for the fact that the rupture of the 2016 Kumamoto earthquake did not propagate

407 southwestward along the Hinagu fault. The electric conductivity can be used to image

408 heterogeneous source structures prior to future destructive earthquakes because it does not

409 require many aftershocks in the source region. The spatial relationship between the crustal

410 structure and slip heterogeneity must be understood to provide geophysical constraints for

411 scenario source models used for the strong motion prediction of future inland crustal

412 earthquakes.

413

414 **Lateral flow in the Aso caldera triggered by a forward-directivity pulse**

415   Another phenomenon observed for the 2016 Kumamoto earthquake is the emergence of

416   many open fissures and associated horizontal sliding of geologic blocks in the Aso Valley, the

417   northern part of the Aso Caldera. Fujiwara *et al.* (2017) reported large-scale surface

418   deformation in three areas with diameters of 500 m to 2 km in the Aso Valley, which is underlain

419   by thick lake-bottom deposits consisting of saturated silt. Each area was horizontally displaced

420   by more than 2 m in the north–northwest direction. Based on differential interferometric

421   synthetic aperture radar data, field observations, descriptions of the temporal and spatial

422   variations of the hot spring supplies, and data from a borehole camera, Tsuji *et al.* (2017)

423   suggested that the shallow geologic block of the Aso hot spring slid more than 1 m in the

424   northwest direction along a specific geologic layer at a depth of ~50 m.

425   Two strong motion stations (KMMH04 and 93002 in Figure 7) are located within the above-

426   mentioned displaced geological blocks. The distance between the two stations is 3.1 km. Figure

427   10 shows the velocity and displacement waveforms recorded at these two stations 30 s after

428   the rupture. The horizontal ground motion at station 93002 was rotated by 36° with respect to

429   the original records to correct for the misorientation of the sensor located inside a kitchenette

430   in the Uchinomaki branch of the Aso city office. The ground motions measured at these two

431   sites are similar. The ground displacement is characterized by a significant horizontal

432   permanent displacement toward the north to north–northwest direction and subsidence was not

433   observed. Such a displacement field cannot be explained by the fault movement of strike-slip

434   faults or normal faults striking in the northeast–southwest direction. Thus, it is reasonable to

435   interpret these displacements as lateral flow or horizontal sliding beneath those sites.

436   We simulated ground motions at those two stations using our source model and velocity

437   structure models in Tables S22 and S23 in the supplemental material to this article. The

438   simulation method is the discrete wavenumber method, which was also used in the waveform

439   inversion study. The simulated velocity waveforms derived from the estimated source model

440   show a strong forward-directivity pulse in the northwestward fault-normal direction as

441   demonstrated in Figure 10, which originated from the asperity in the central part of the

442   Futagawa fault. The synthetic and observed velocity waveforms match well before the peak

443   ground velocity is reached. After the forward-directivity pulse passed, the discrepancy between

444   the observed and synthetic waveforms became nonnegligible. The observed velocity

445   waveforms show a remarkably slow deceleration after the peak, which is particularly noticeable

446   in the north–south component. The waveform comparison suggests that this peculiar ground

447   motion might be due to a force applied to a superficial geological block during lateral flow.

448   There are thick low-velocity sediments in the Aso valley because caldera lakes existed for long

449   years in the Holocene and Pleistocene (e.g., Miyabuchi and Sugiyama, 2012). The current

450   velocity models for these two stations (Tables S22 and S23) include thick low-velocity

451   sediments based on the subsurface structure model by Senna *et al.* (2018), thus, the synthetic

452   waveforms were affected by these lakebed sediments as basin-induced surface waves. However,

453 such phenomena were not recorded in the observed waveforms because the horizontal sliding

454 by liquefaction at a shallow depth (e.g., Tsuji *et al.*, 2017) might not allow seismic waves

455 propagate from deep valley to the surface, and the seismometer moved with the sliding block.

456 Unfortunately, the downhole record of KMMH04 at a depth of 127 m, which is much deeper

457 than the expected bottom depth of the displaced block, could not be used because the downhole

458 sensor was out of order at the time of the earthquake. Therefore, we could not directly observe

459 the relative movement between the surface and downhole records. Our source model and the

460 ground motion simulation enabled us to provide evidence that the lateral flow in the Aso Valley

461 area was primarily dynamically triggered by the northwest motion of the forward-directivity

462 pulse, which is preponderance in the fault-normal component over the fault-parallel component.

463 The same forward-directivity pulse might also have contributed to the remote triggering of a

464 *M* 5.9 event in the Yufuin geothermal area (e.g., Miyazawa, 2016).

465

466 **CONCLUSIONS**

467 In this study, an improved approach is proposed for the parameterization of the source fault

468 during kinematic waveform inversion analysis to enhance the reproducibility of the near-fault

469 ground motions during the mainshock of the 2016 Kumamoto earthquake sequence, central

470 Kyushu, Japan (M_{JMA} 7.3). Point sources, which were densely distributed at intervals of 0.22

471 km, were utilized in this work. The slip amount of each point source was obtained via the linear

472   correlation with the slip amounts at surrounding control points, which were uniformly

473   distributed across the source fault, at intervals of 2.0 km. A source fault plane model consisting

474   of five fault segments was set up using fault geometries recently proposed by Mitsuoka *et al.*

475   (2020) based on the distribution of aftershocks. The estimated source rupture process started

476   with a small initial rupture in the vicinity of the hypocenter of the Hinagu fault within the first

477   three seconds. The rupture then propagated up- and northeastward along the Futagawa fault. A

478   large slip with a peak of 4.9 m and peak slip-velocity of 3.1 m/s was detected at depths ranging

479   from 3 to 15 km in the central part of the Futagawa fault, which could be characterized as a

480   body with high seismic velocity and low seismic attenuation from published seismic velocity

481   and attenuation tomography studies. The slips in the shallower portion of the fault were not

482   isolated and can be regarded as extensions of the asperity into the seismogenic depth, but the

483   rupture was delayed compared with the slips in the deeper portion. The slip in the Aso caldera

484   is concentrated at shallow depth, which is consistent with surface ruptures, geodetically

485   measured displacement fields, and the existence of high-temperature magma fluid beneath the

486   Aso volcano. The slip velocity functions show depth-dependent features. The slip velocity

487   functions at the top 3 km are almost temporally symmetric and have relatively long durations

488   (4–5 s), although the slip velocity functions of the rest of the fault plane (>3 km depth) inside

489   the seismogenic layer have typical Kostrov-type characteristics with short durations. The slips

490   and slip velocities in the shallow portion of the source fault significantly contributed to the

491 pulse wave velocity and permanent displacement, except for sites located in the forward-

492 rupture direction.

493   Geophysical information on the crustal structure would be beneficial for the characterization

494 of the behaviors of source fault ruptures and is necessary for advancing source models used for

495 scenario-based strong motion prediction and seismic hazard assessment. Based on our results,

496 lateral flow of superficial layers observed in wide areas of the Aso Valley, northern Aso caldera,

497 was primarily triggered by the northwestward ground motion of a strong forward-directivity

498 pulse generated by the asperity on the Futagawa fault.

499

500 **DATA AND RESOURCES**

501 Strong motion data of K-NET and KiK-net were downloaded from the strong-motion

502 seismograph network (https://www.kyoshin.bosai.go.jp/, last accessed September 2016)

503 operated by the National Research Institute for Earth Science and Disaster Resilience (NIED),

504 Japan. The strong motion waveform data of F-net were downloaded from the NIED Hi-net

505 database (https://www.hinet.bosai.go.jp/, last accessed April 2016). The strong motion data

506 from the seismic intensity observation networks of the JMA and Kumamoto prefectural

507 government were obtained from JMA's website

508 (http://www.data.jma.go.jp/svd/eqev/data/kyoshin/jishin/1604160125_kumamoto/index.html,

509 last accessed July 2016). The 10 m-mesh digital elevation model used in the maps was provided

510 from the Geospatial Information Authority of Japan as the Fundamental Geospatial Data

511 (https://www.gsi.go.jp/kiban/). The supplemental material contains the 1D velocity structure

512 model of each strong motion station used for calculating Green's functions. All figures were

513 drawn using Generic Mapping Tools version 6 (Wessel *et al.*, 2019).

514

515 **DECLARATION OF COMPETING INTERESTS**

516 The authors acknowledge there are no conflicts of interest recorded.

517

518 **ACKNOWLEDGEMENTS**

523

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Tables

779   **Table 1.** Representation of fault models during kinematic waveform inversion analysis.

Fault segment	H1	H2	F1	F2	F3
Upper south corner	32.6550°N 130.7554°E	32.6550°N 130.7554°E	32.7722°N 130.8072°E	32.8668°N 130.9697°E	32.8890°N 130.9937°E
Strike/dip	201°/62°	201°/90°	235°/66°	228°/67°	239°/69°
Length	16.0 km	16.0 km	18.0 km	4.0 km	6.0 km
Width	13.1 km	6.0 km	19.1 km	19.1 km	19.1 km
Top depth	6.6 km	0.6 km	0.4 km	0.5 km	0.2 km
Bottom depth	18.2 km	6.6 km	17.9 km	17.9 km	18.1 km
Point-source interval	0.22 km × 0.22 km				
Number of point sources	4248	1944	6966	1542	2322
Control-point interval	2.0 km × 2.0 km				
Number of control points	56	24	90	20	30
Number of time windows	9				
Duration / time shift of the time window	1.0 s / 0.5 s				

780

Figure Captions

Figure 1. Index map of the study area. Sold triangles indicate the locations of strong motion stations used for the kinematic source inversion. The epicenter of the 2016 Kumamoto mainshock (Mitsuoka *et al.*, 2020) is represented by the solid star. Black broken rectangles correspond to the surface projection of source fault models used for the kinematic waveform inversion analysis and thick black solid lines indicate the top of the fault plane. Red lines represent active fault traces compiled by Nakata and Imaizumi (2002). RTL indicates the Rokkoku Tectonic Line (Yanagida, 1958). The inset map indicates the location of the study area in the Japanese archipelago.

Figure 2. Schematic illustration of the discretization of the source fault plane and locations of the control points (large solid circles) to which unknown model parameters were assigned during the kinematic waveform inversion. The dense distribution of point sources for the calculation of Green's functions is depicted by small solid circles.

Figure 3. (a) Epicenters of the aftershocks determined by Mitsuoka *et al.* (2020) within 6 hours from the mainshock (colored circles) and the assumed source fault model (solid rectangles). The purple lines represent the surface rupture traces compiled by Kumahara *et al.* (2016). (b) The vertical cross section of the aftershocks and the source fault plane around the Hinagu

800   Fault. (c) The vertical cross section of the aftershocks and the source fault plane around the

801   Futagawa Fault.

802

803   **Figure 4.** (a) Distribution of the final slip amounts on the source fault planes with slip vectors.

804   The contour interval of the slip is 1 m. The arrow shows the slip vector of the hanging wall

805   relative to the foot wall. The open star indicates the hypocenter or rupture starting point.

806   Dashed lines correspond to boundaries of fault segments with different strike and dip angles.

807   (b) Slip velocity functions obtained for every control point. The maximum slip velocity is

808   3.1 m/s.

809

810   **Figure 5.** Comparison between observed velocity waveforms (gray) and synthetic waveforms

811   (black) in the frequency range of 0.05–1 Hz. The maximum absolute amplitude of the

812   observed waveform is shown above each trace (unit: cm/s). EW: east–west, NS: north–

813   south, UD: up–down.

814

815   **Figure 6.** Snapshots of the slip velocity every 1 s. The open star indicates the rupture starting

816   point. Dashed lines correspond to boundaries of fault segments with different strike and dip

817   angles.

818

819   **Figure 7.** Map view of the estimated final slip distribution. The solid triangles indicate strong

820   motion stations in the near-source area, which were referred to in this paper. Red lines

821   represent active fault traces compiled by Nakata and Imaizumi (2002), and blue lines

822   represent surface ruptures associated the 2016 Kumamoto earthquake compiled by

823   Kumahara *et al.* (2016).

824

825   **Figure 8.** (a) Slip velocity functions at control points on the Futagawa fault segments for each

826   depth bin. Slip velocity functions of the same depth bin were aligned by the peak time,

827   which correspond to 0 s in the plots. (b) Relationship between peak time from the onset

828   time and peak slip velocity for each depth bin. The weight average of the peak time is

829   indicated in each plot.

830

831   **Figure 9.** Simulated and observed (a) velocity and (b) displacement waveforms at three near-

832   fault strong motion stations (93011, 93048, and 93051). The numerical value above each

833   trace indicates the maximum absolute amplitude of the observed waveform.

834

835   **Figure 10.** (a) Displacements observed during 30 s after the origin time at two strong motion

836   stations in the Aso Valley (KMMH04 and 93002). Crosses and associated numbers indicate

837   the lapsed time in seconds with respect to the origin. The thick arrow indicates the direction

838   normal to the strike angle of the Futagawa fault. (b) Simulated and observed velocity

839   waveforms for stations KMMH04 and 93002. The numerical value above each trace

840   indicates the maximum absolute amplitude of the observed waveform in cm/s.

841

Figure 1. Index map of the study area. Sold triangles indicate the locations of strong motion stations used for the kinematic source inversion. The epicenter of the 2016 Kumamoto mainshock (Mitsuoka *et al.*, 2020) is represented by the solid star. Black broken rectangles correspond to the surface projection of source fault models used for the kinematic waveform inversion analysis and thick black solid lines indicate the top of the fault plane. Red lines represent active fault traces compiled by Nakata and Imaizumi (2002). RTL indicates the Rokkoku Tectonic Line (Yanagida, 1958). The inset map indicates the location of the study area in the Japanese archipelago.

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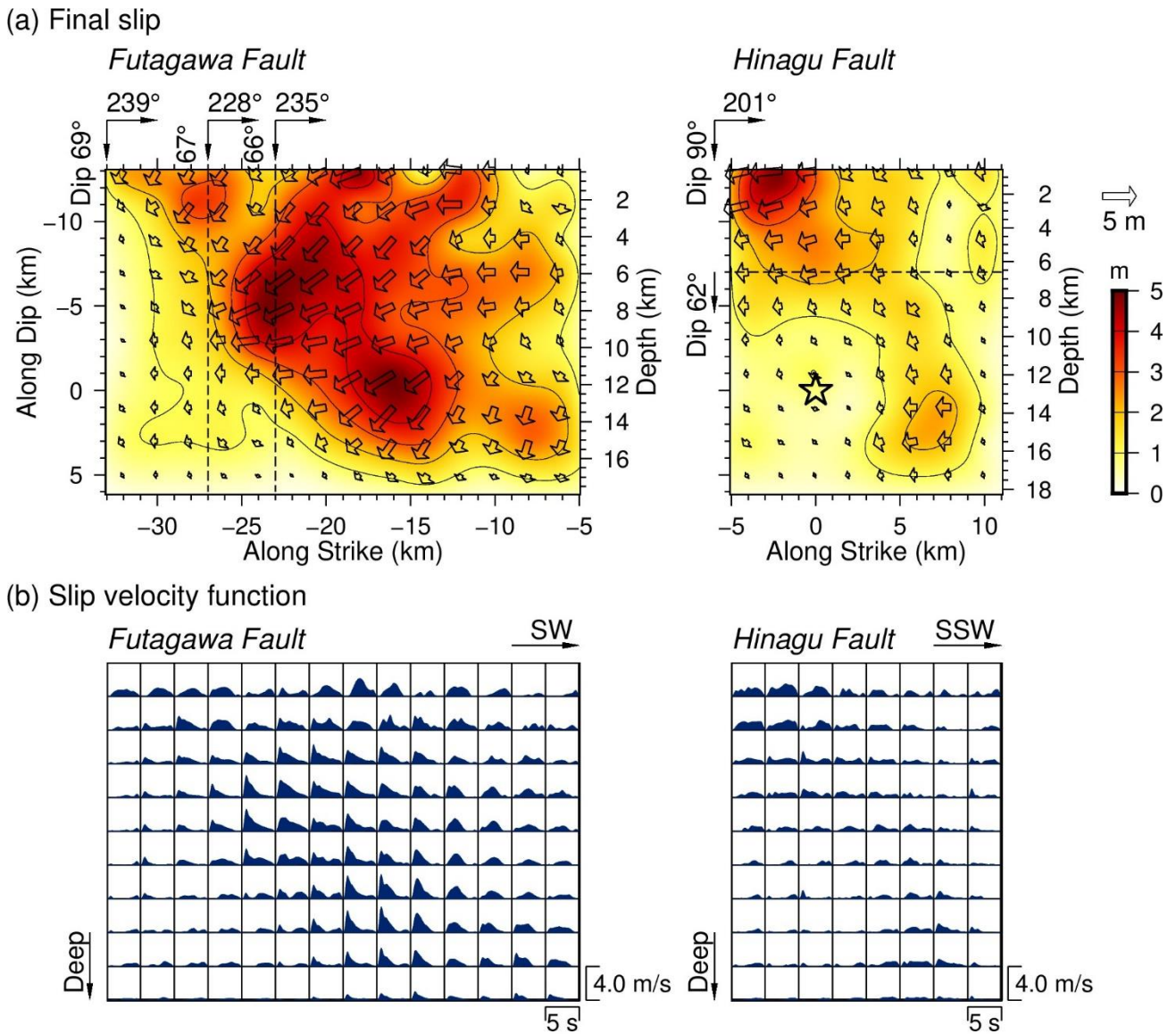
(a) Final slip

Futagawa Fault

Hinagu Fault

(b) Slip velocity function

Futagawa Fault

Hinagu Fault

Figure 4. (a) Distribution of the final slip amounts on the source fault planes with slip vectors. The contour interval of the slip is 1 m. The arrow shows the slip vector of the hanging wall relative to the foot wall. The open star indicates the hypocenter or rupture starting point. Dashed lines correspond to boundaries of fault segments with different strike and dip angles. (b) Slip velocity functions obtained for every control point. The maximum slip velocity is 3.1 m/s.
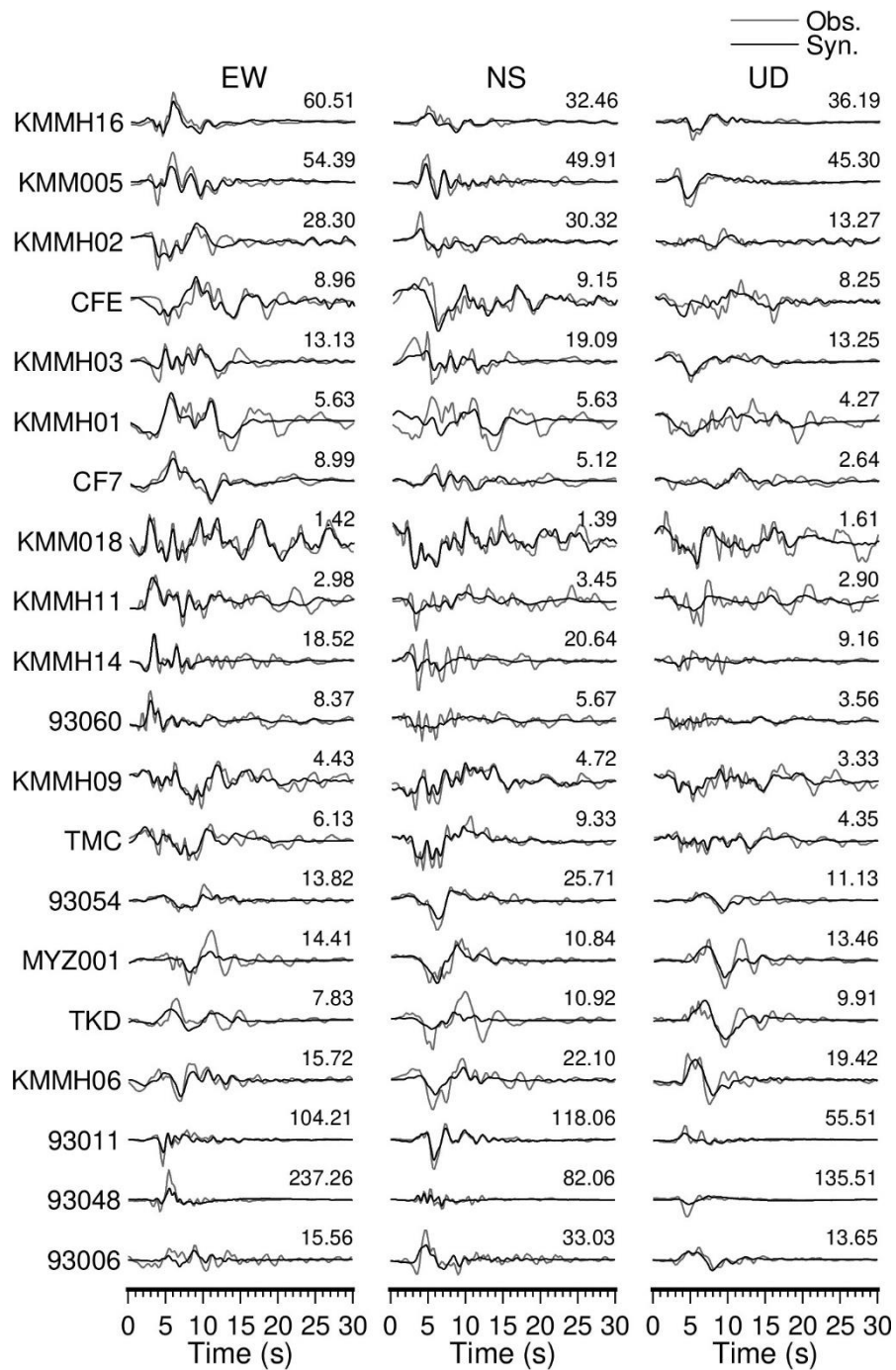
Figure 5. Comparison between observed velocity waveforms (gray) and synthetic waveforms (black) in the frequency range of 0.05–1 Hz. The maximum absolute amplitude of the observed waveform is shown above each trace (unit: cm/s). EW: east–west, NS: north–south, UD: up–down.
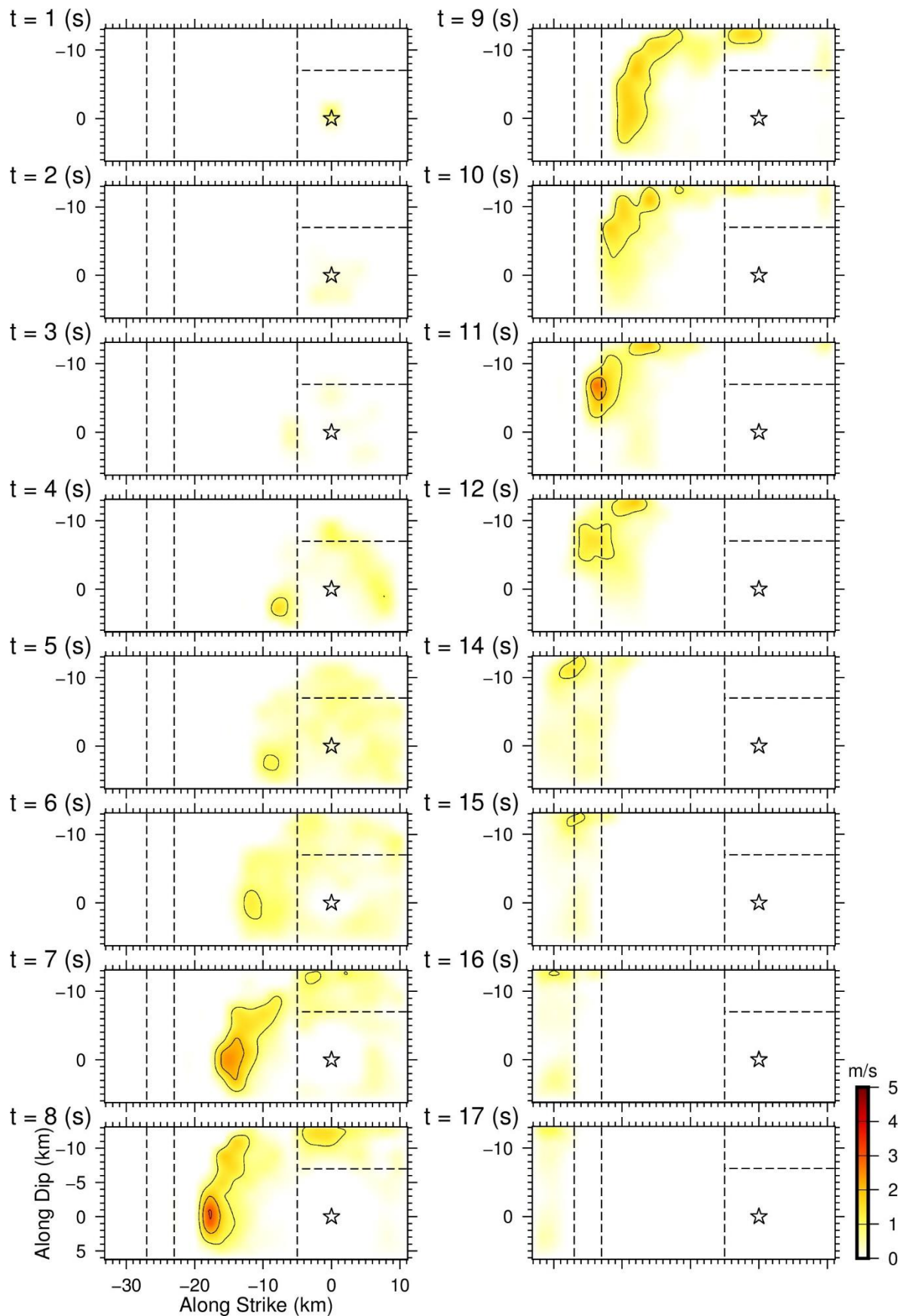
Figure 6. Snapshots of the slip velocity at every 1 s. The open star indicates the rupture starting point. Dashed lines correspond to boundaries of fault segments with different strike and dip angles.
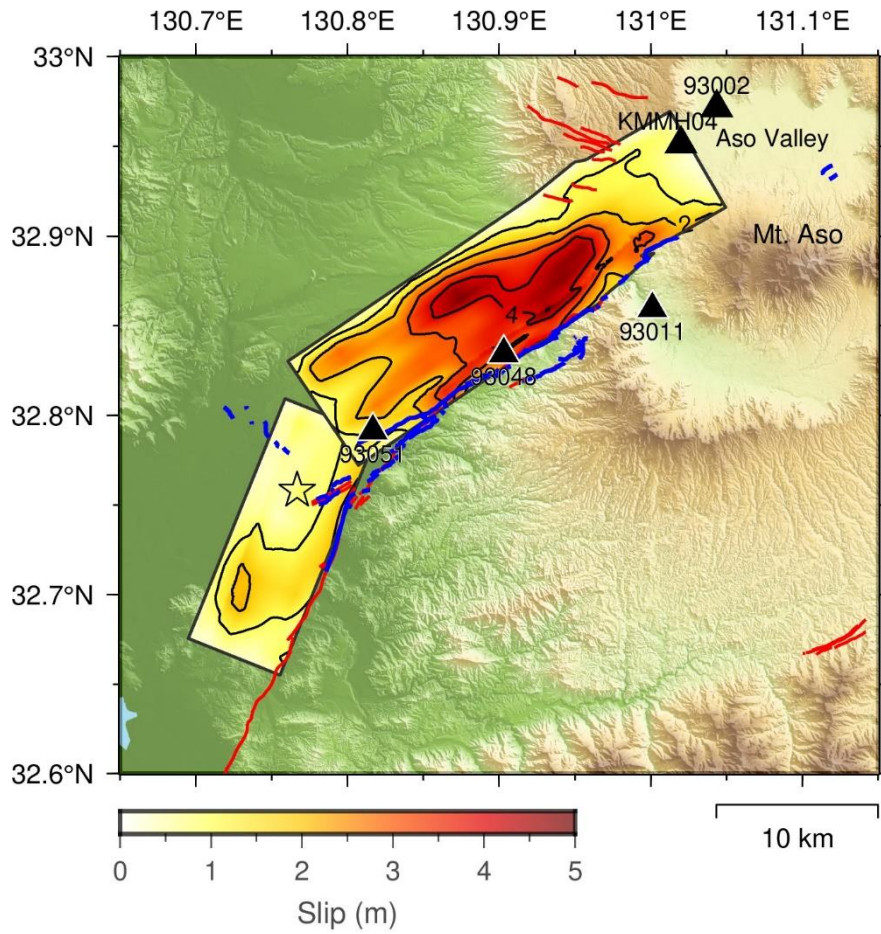
Figure 7. Map view of the estimated final slip distribution. The solid triangles indicate strong motion stations in the near-source area, which were referred to in this paper. Red lines represent active fault traces compiled by Nakata and Imaizumi (2002), and blue lines represent surface ruptures associated the 2016 Kumamoto earthquake compiled by Kumahara et al. (2016).
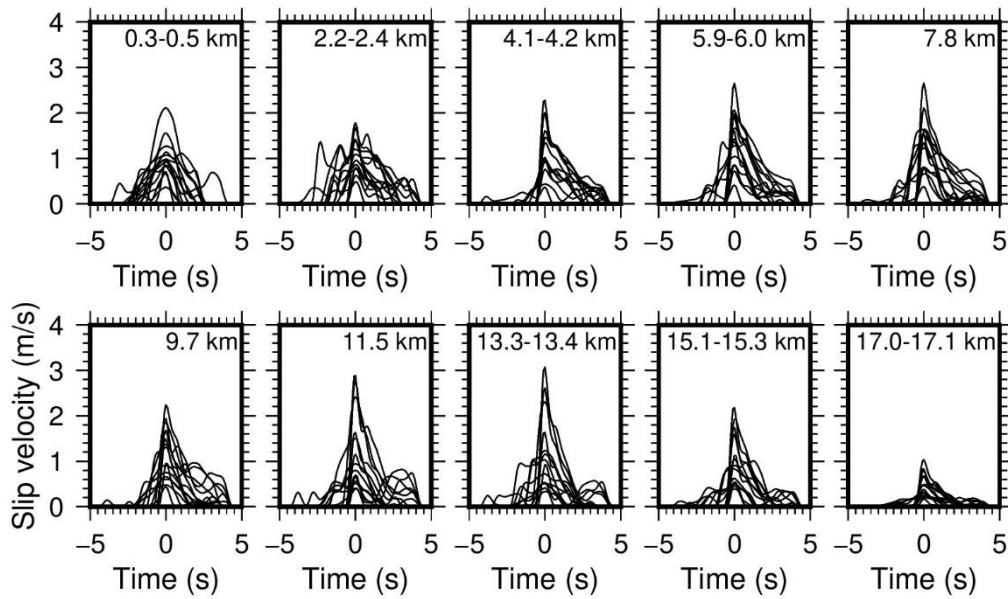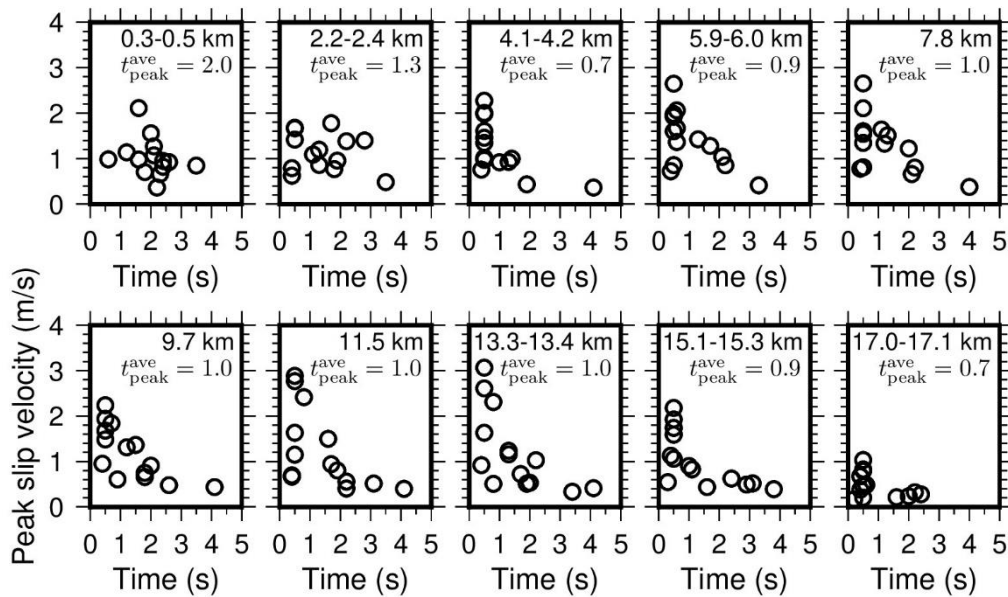
Figure 8. (a) Slip velocity functions at control points on the Futagawa fault segments for each depth bin. Slip velocity functions of the same depth bin were aligned by the peak time, which correspond to 0 s in the plots. (b) Relationship between peak time from the onset time and peak slip velocity for each depth bin. The weight average of the peak time is indicated in each plot.
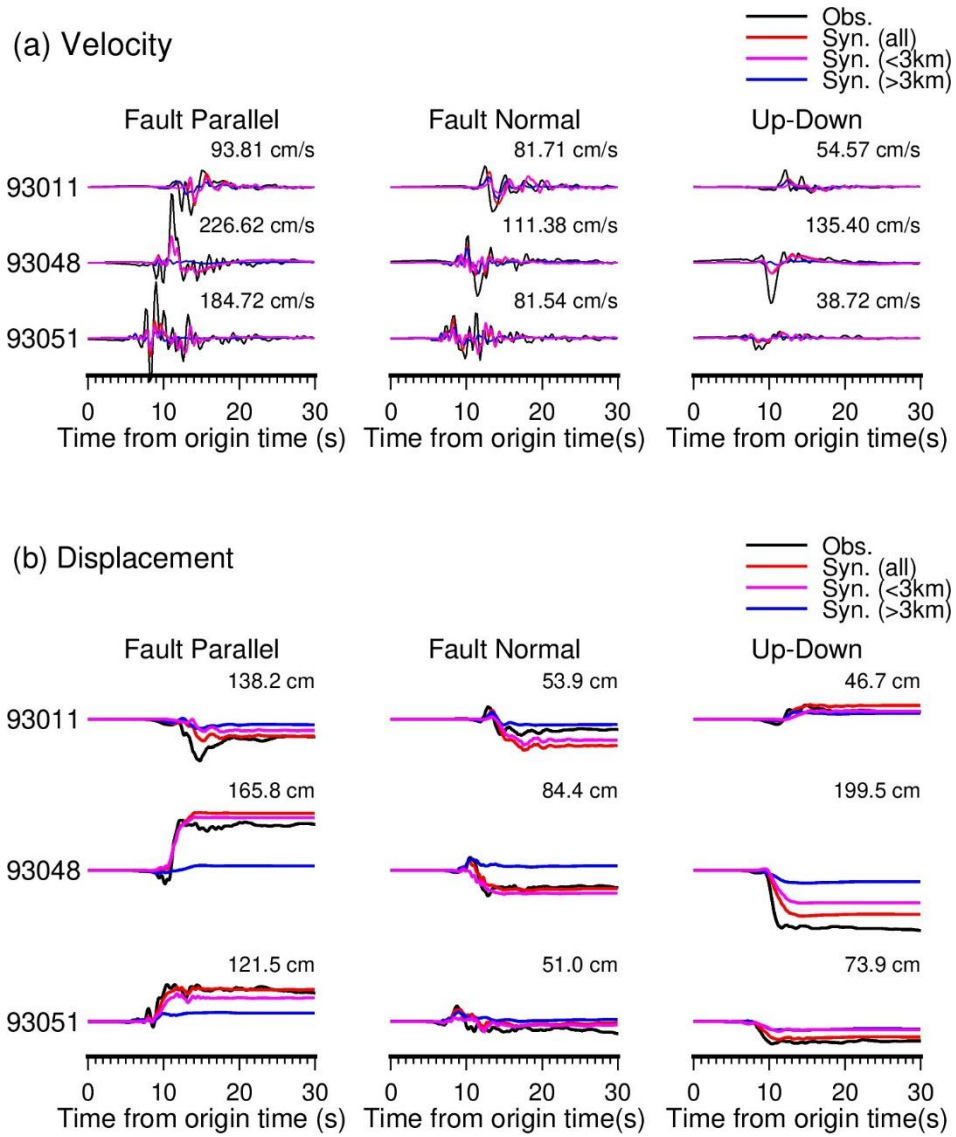
Figure 9. Simulated and observed (a) velocity and (b) displacement waveforms at three near-fault strong motion stations (93011, 93048, and 93051). The numerical value above each trace indicates the maximum absolute amplitude of the observed waveform.
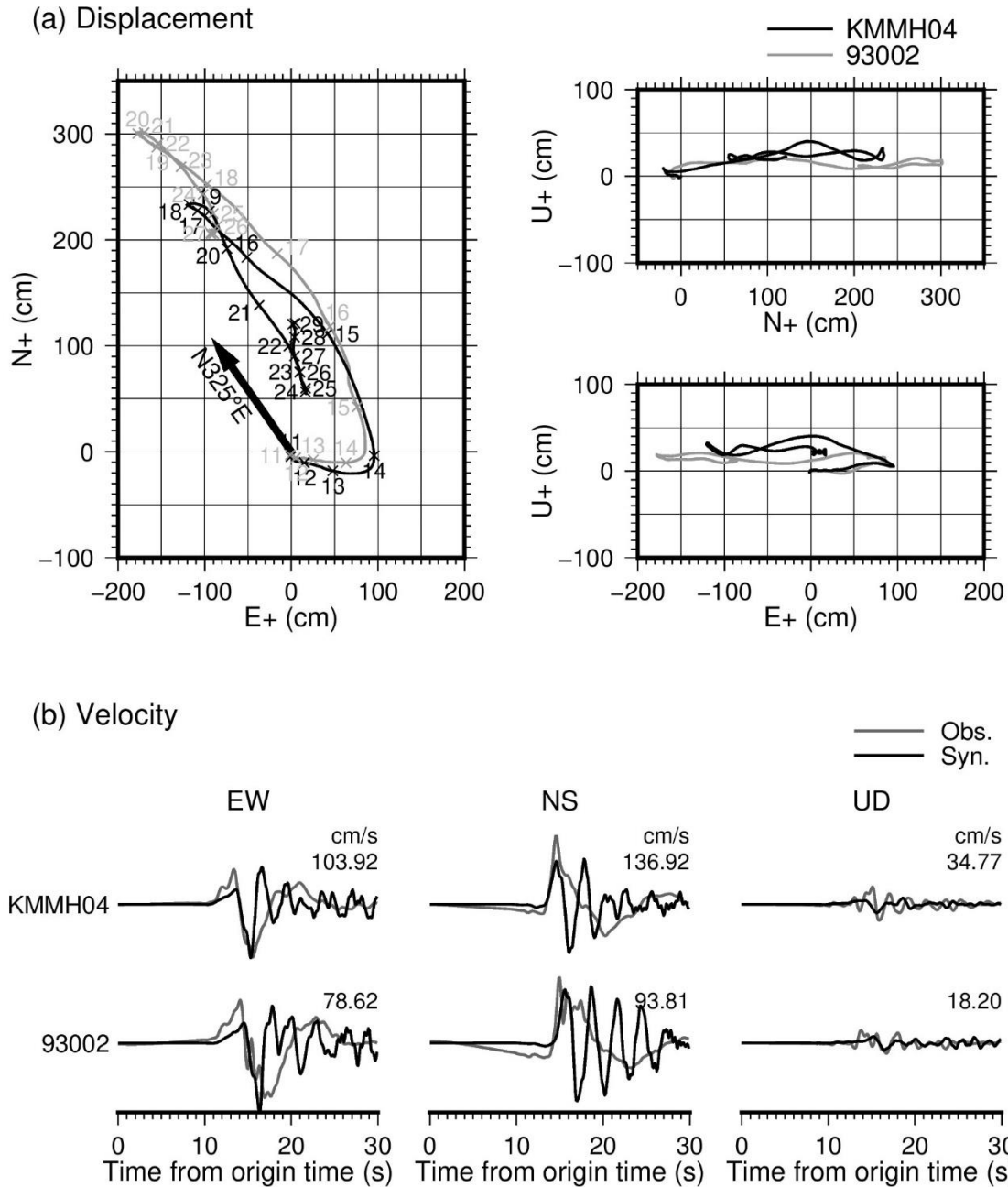
Figure 10. (a) Displacements observed during 30 s after the origin time at two strong motion stations in the Aso Valley (KMMH04 and 93002). Crosses and associated numbers indicate the lapsed time in seconds with respect to the origin time. The thick arrow indicates the direction normal to the strike angle of the Futagawa fault. (b) Simulated and observed velocity waveforms for stations KMMH04 and 93002. The numerical value above each trace indicates the maximum absolute amplitude of the observed waveform in cm/s.

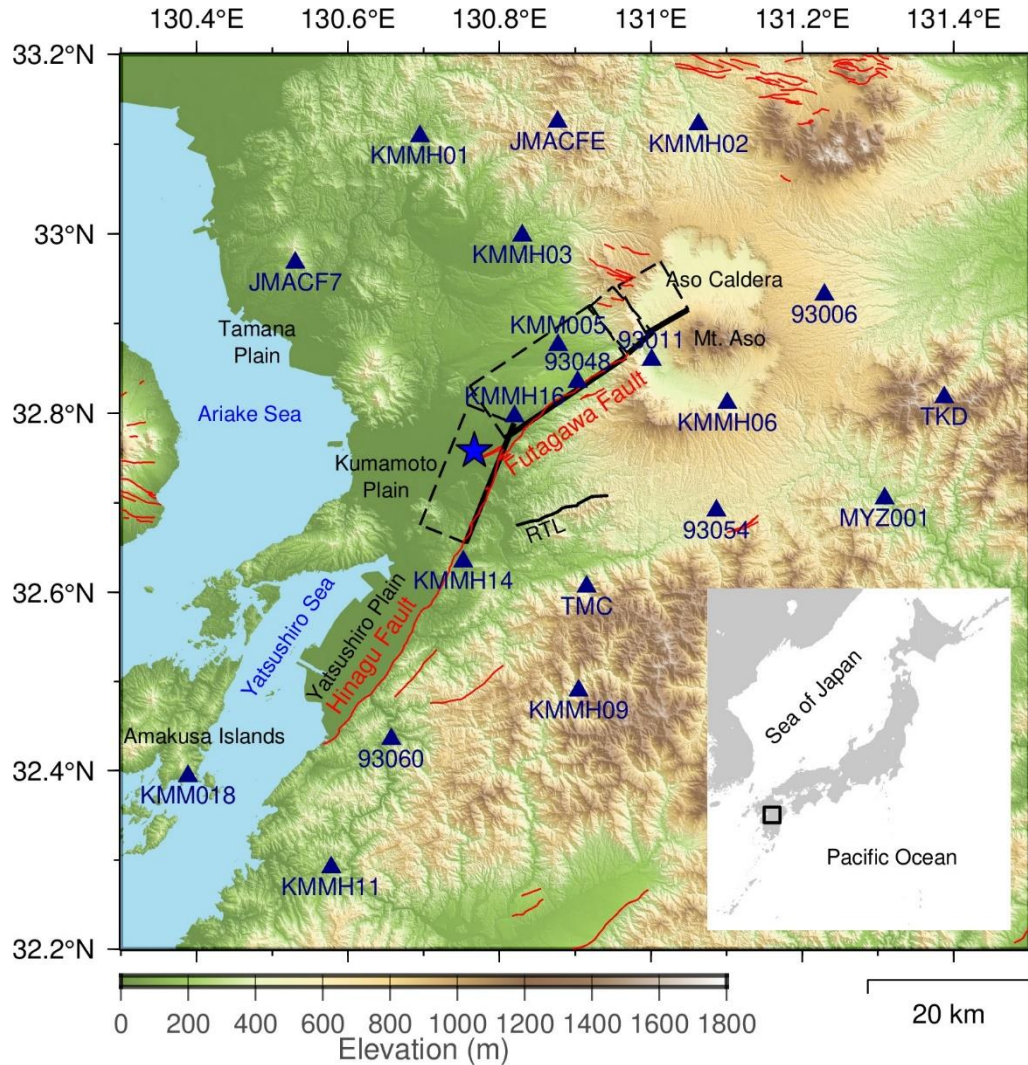# Revisiting the source rupture process of the mainshock of the 2016 Kumamoto earthquake and implications for the generation of near-fault ground motions and forward directivity pulses

Kimiyuki Asano and Tomotaka Iwata

This electronic supplement contains the velocity structure models of each strong motion station used for this study.

Table S1. Velocity structure model for KMM005.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
4.38	1429	125	1764	102	60
5.25	1456	150	1774	102	60
5.26	1484	175	1785	102	60
5.27	1512	200	1795	102	60
7.75	1540	225	1805	102	60
7.75	1568	250	1816	102	60
7.75	1595	275	1826	102	60
7.76	1600	300	1840	102	60
9.42	1600	350	1850	102	60
13.31	1600	400	1850	102	60
20.27	1700	450	1900	102	60
20.35	1800	500	1900	102	60
27.98	2000	600	1900	170	100
47.22	2500	1100	2150	255	150
425.91	4000	2100	2400	340	200
1020.79	5000	2700	2500	340	200
1220.80	5500	3100	2600	510	300
3981.30	5800	3400	2700	680	400
16471.80	6400	3800	2800	680	400
32641.00	7500	4500	3200	850	500

Table S2. Velocity structure model for KMM018.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1600	350	1850	102	60
2.70	1600	400	1850	102	60
5.83	1700	450	1900	102	60
6.98	1800	500	1900	102	60
24.49	2000	600	1900	170	100
45.86	2400	900	2050	170	100
192.07	3400	1600	2300	255	150
419.31	5000	2700	2500	340	200
657.64	5500	3100	2600	510	300
4203.10	5800	3400	2700	680	400
16930.40	6400	3800	2800	680	400
32079.30	7500	4500	3200	850	500

Table S3. Velocity structure model for KMMH01.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2000	600	1900	170	100
6.24	2500	1100	2150	255	150
8.42	3000	1400	2250	255	150
18.74	3500	1700	2300	255	150
23.50	4000	2100	2400	340	200
28.14	4000	2100	2400	340	200
56.01	5000	2700	2500	340	200
81.26	5500	3100	2600	510	300
3644.50	5800	3400	2700	680	400
13493.00	6400	3800	2800	680	400
32011.60	7500	4500	3200	850	500

Table S4. Velocity structure model for KMMH02.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2500	1100	2150	255	150
354.82	3000	1400	2250	255	150
1318.84	4000	2100	2400	340	200
1318.85	5000	2700	2500	340	200
1319.44	5500	3100	2600	510	300
4604.70	5800	3400	2700	680	400
14000.50	6400	3800	2800	680	400
33291.00	7500	4500	3200	850	500

Table S5. Velocity structure model for KMMH03.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2000	600	1900	170	100
17.81	2400	900	2050	170	100
21.28	2500	1100	2150	255	150
381.37	4000	2100	2400	340	200
387.16	4000	2100	2400	340	200
389.32	5000	2700	2500	340	200
390.75	5500	3100	2600	510	300
4209.10	5800	3400	2700	680	400
16444.00	6400	3800	2800	680	400
32198.10	7500	4500	3200	850	500

Table S6. Velocity structure model for KMMH06.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1401	100	1753	102	60
3.81	1512	200	1795	102	60
5.76	1600	300	1840	102	60
7.05	1600	350	1850	102	60
8.17	1600	400	1850	102	60
11.25	1700	450	1900	102	60
16.58	1800	500	1900	102	60
61.61	2000	600	1900	170	100
76.54	2500	1100	2150	255	150
386.31	4000	2100	2400	340	200
518.46	5000	2700	2500	340	200
718.46	5500	3100	2600	510	300
3741.60	5800	3400	2700	680	400
15264.90	6400	3800	2800	680	400
34533.60	7500	4500	3200	850	500

Table S7. Velocity structure model for KMMH09.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2000	600	1900	170	100
148.66	4000	2100	2400	340	200
158.31	4000	2100	2400	340	200
445.12	5000	2700	2500	340	200
457.70	5500	3100	2600	510	300
1912.60	5800	3400	2700	680	400
17609.60	6400	3800	2800	680	400
35973.00	7500	4500	3200	850	500

Table S8. Velocity structure model for KMMH11.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2000	600	1900	170	100
8.50	2400	900	2050	170	100
12.87	2500	1100	2150	255	150
27.39	3000	1400	2250	255	150
34.17	3400	1600	2300	255	150
42.91	3500	1700	2300	255	150
50.11	4000	2100	2400	340	200
80.07	4000	2100	2400	340	200
104.04	5000	2700	2500	340	200
140.06	5500	3100	2600	510	300
3291.00	5800	3400	2700	680	400
17532.40	6400	3800	2800	680	400
33488.20	7500	4500	3200	850	500

Table S9. Velocity structure model for KMMH14.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
3.97	1429	125	1764	102	60
4.00	1456	150	1774	102	60
4.29	1484	175	1785	102	60
9.83	1512	200	1795	102	60
9.90	1568	250	1816	102	60
9.90	1600	300	1840	102	60
20.58	1700	450	1900	102	60
83.35	4000	2100	2400	340	200
359.90	4000	2100	2400	340	200
362.73	5500	3100	2600	510	300
1938.30	5800	3400	2700	680	400
16074.60	6400	3800	2800	680	400
33428.10	7500	4500	3200	850	500

Table S10. Velocity structure model for KMMH16.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
9.13	1484	175	1785	102	60
9.43	1512	200	1795	102	60
18.68	1595	275	1826	102	60
18.73	1600	300	1840	102	60
23.25	1600	350	1850	102	60
27.34	1600	400	1850	102	60
52.20	2000	600	1900	170	100
80.04	2400	900	2050	170	100
315.82	2500	1100	2150	255	150
577.81	4000	2100	2400	340	200
1441.82	5000	2700	2500	340	200
1641.82	5500	3100	2600	510	300
3293.50	5800	3400	2700	680	400
16250.70	6400	3800	2800	680	400
32284.70	7500	4500	3200	850	500

Table S11. Velocity structure model for MYZ001.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	3000	1400	2250	255	150
11.99	3500	1700	2300	255	150
21.98	4000	2100	2400	340	200
32.57	4000	2100	2400	340	200
77.83	5000	2700	2500	340	200
106.46	5500	3100	2600	510	300
2883.40	5800	3400	2700	680	400
16774.40	6400	3800	2800	680	400
33854.70	7500	4500	3200	850	500

Table S12. Velocity structure model for 93006.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
2.31	1456	150	1774	102	60
2.38	1484	175	1785	102	60
8.90	1512	200	1795	102	60
10.52	1600	300	1840	102	60
20.56	1600	350	1850	102	60
22.31	1600	400	1850	102	60
25.67	1700	450	1900	102	60
26.06	2000	600	1900	170	100
49.83	2500	1100	2150	255	150
191.22	3000	1400	2250	255	150
201.67	4000	2100	2400	340	200
524.48	4000	2100	2400	340	200
599.00	5500	3100	2600	510	300
3648.20	5800	3400	2700	680	400
14139.30	6400	3800	2800	680	400
34033.90	7500	4500	3200	850	500

Table S13. Velocity structure model for 93011.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
5.67	1512	200	1795	102	60
7.73	1540	225	1805	102	60
12.35	1568	250	1816	102	60
13.98	1595	275	1826	102	60
15.26	1600	300	1840	102	60
18.69	1600	350	1850	102	60
23.33	1600	400	1850	102	60
81.41	1700	450	1900	102	60
97.26	2000	600	1900	170	100
136.03	2500	1100	2150	255	150
617.97	4000	2100	2400	340	200
1403.37	5000	2700	2500	340	200
1603.37	5500	3100	2600	510	300
4173.90	5800	3400	2700	680	400
15757.20	6400	3800	2800	680	400
34002.20	7500	4500	3200	850	500

Table S14. Velocity structure model for 93048.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
10.01	1512	200	1795	102	60
18.75	1600	300	1840	102	60
23.92	1600	350	1850	102	60
29.88	1600	400	1850	102	60
46.15	1700	450	1900	102	60
46.15	2500	1100	2150	255	150
189.10	4000	2100	2400	340	200
1139.22	5000	2700	2500	340	200
1339.22	5500	3100	2600	510	300
3849.10	5800	3400	2700	680	400
16012.10	6400	3800	2800	680	400
33083.60	7500	4500	3200	850	500

Table S15. Velocity structure model for 93054.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m^3)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2000	600	1900	170	100
7.40	2500	1100	2150	255	150
84.98	4000	2100	2400	340	200
92.18	4000	2100	2400	340	200
364.85	5500	3100	2600	510	300
2933.30	5800	3400	2700	680	400
16525.90	6400	3800	2800	680	400
35317.10	7500	4500	3200	850	500

Table S16. Velocity structure model for 93060.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.56	1429	125	1764	102	60
3.17	1456	150	1774	102	60
4.46	1512	200	1795	102	60
6.27	1600	300	1840	102	60
7.76	2500	1100	2150	255	150
10.68	3000	1400	2250	255	150
21.09	3500	1700	2300	255	150
31.26	4000	2100	2400	340	200
79.78	5000	2700	2500	340	200
127.84	5500	3100	2600	510	300
2507.40	5800	3400	2700	680	400
16588.10	6400	3800	2800	680	400
33673.20	7500	4500	3200	850	500

Table S17. Velocity structure model for JMACF7.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2000	600	1900	170	100
13.07	2500	1100	2150	255	150
36.24	3000	1400	2250	255	150
36.41	3500	1700	2300	255	150
36.58	4000	2100	2400	340	200
371.04	4000	2100	2400	340	200
387.85	5000	2700	2500	340	200
391.41	5500	3100	2600	510	300
2861.80	5800	3400	2700	680	400
14888.30	6400	3800	2800	680	400
32164.30	7500	4500	3200	850	500

Table S18. Velocity structure model for JMACFE.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2500	1100	2150	255	150
1042.69	3000	1400	2250	255	150
1643.99	4000	2100	2400	340	200
1644.04	5500	3100	2600	510	300
4585.00	5800	3400	2700	680	400
13491.70	6400	3800	2800	680	400
32853.20	7500	4500	3200	850	500

Table S19. Velocity structure model for TMC.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2500	1100	2150	255	150
8.16	3000	1400	2250	255	150
19.27	3500	1700	2300	255	150
30.53	4000	2100	2400	340	200
35.19	4000	2100	2400	340	200
186.52	5000	2700	2500	340	200
230.31	5500	3100	2600	510	300
1862.60	5800	3400	2700	680	400
17081.50	6400	3800	2800	680	400
35413.20	7500	4500	3200	850	500

Table S20. Velocity structure model for TKD.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
0.10	1429	125	1764	102	60
0.25	1456	150	1774	102	60
0.50	1484	175	1785	102	60
0.75	1512	200	1795	102	60
1.00	1540	225	1805	102	60
1.25	1568	250	1816	102	60
1.50	1595	275	1826	102	60
2.00	1600	300	1840	102	60
3.00	1600	350	1850	102	60
3.50	1600	400	1850	102	60
4.00	1700	450	1900	102	60
5.00	2500	1100	2150	255	150
19.86	3000	1400	2250	255	150
29.30	3500	1700	2300	255	150
45.88	4000	2100	2400	340	200
64.02	4000	2100	2400	340	200
126.32	5000	2700	2500	340	200
201.00	5500	3100	2600	510	300
2051.10	5800	3400	2700	680	400
16223.50	6400	3800	2800	680	400
33680.70	7500	4500	3200	850	500

Table S21. Velocity structure model for 93051.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
12.83	1512	200	1795	102	60
21.75	1595	275	1826	102	60
21.75	1600	300	1840	102	60
25.64	1600	350	1850	102	60
30.26	1600	400	1850	102	60
40.97	1800	500	1900	102	60
42.71	2000	600	1900	170	100
82.34	2400	900	2050	170	100
337.33	2500	1100	2150	255	150
584.13	4000	2100	2400	340	200
1425.66	5000	2700	2500	340	200
1625.66	5500	3100	2600	510	300
3246.30	5800	3400	2700	680	400
16253.40	6400	3800	2800	680	400
32289.80	7500	4500	3200	850	500

Table S22. Velocity structure model for KMMH04.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
32.32	1456	150	1774	102	60
48.41	1484	175	1785	102	60
48.41	1512	200	1795	102	60
64.89	1595	275	1826	102	60
91.50	1600	300	1840	102	60
101.93	1600	350	1850	102	60
112.05	1600	400	1850	102	60
160.34	1700	450	1900	102	60
160.53	1800	500	1900	102	60
213.01	2000	600	1900	170	100
257.85	2500	1100	2150	255	150
412.61	3000	1400	2250	255	150
530.33	4000	2100	2400	340	200
1125.89	5000	2700	2500	340	200
1325.89	5500	3100	2600	510	300
4580.50	5800	3400	2700	680	400
16860.70	6400	3800	2800	680	400
33825.90	7500	4500	3200	850	500

Table S23. Velocity structure model for 93002.

Top depth (m)	V_P (m/s)	V_S (m/s)	Density (kg/m³)	Q_P	Q_S
0.00	1401	100	1753	102	60
32.70	1484	175	1785	102	60
87.08	1512	200	1795	102	60
95.14	1540	225	1805	102	60
100.91	1568	250	1816	102	60
104.89	1595	275	1826	102	60
110.74	1600	300	1840	102	60
116.21	1600	350	1850	102	60
128.43	1600	400	1850	102	60
140.86	1700	450	1900	102	60
156.65	2500	1100	2150	255	150
496.31	3000	1400	2250	255	150
871.68	3500	1700	2300	255	150
871.71	4000	2100	2400	340	200
1222.87	5000	2700	2500	340	200
1422.87	5500	3100	2600	510	300
4550.30	5800	3400	2700	680	400
16602.00	6400	3800	2800	680	400
33810.80	7500	4500	3200	850	500