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Aseismic Slip during the 1996 Earthquake Swarm in and around the Onikobe Geothermal Area, NE Japan

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Abstract

In August 1996, an earthquake swarm including 4 earthquakes of magnitudes greater than 5 occurred in and around the Onikobe geothermal area, northeast Japan. While earlier studies detected ground displacements by L-band JERS radar interferograms, there remained certain puzzling discrepancies between the observations and the predictions from a seismologically inferred model, and no satisfactory models have yet been proposed. Here, by identifying that a steep gradient in the radar line-of-sight changes exists in seismicity gaps to the east and west of the Torage area (just to the north of the Onikobe caldera), we propose two aseismic reverse faults, which turn out to contribute to a local topographic growth as well as explaining the pre-existing discrepancies. Another significant signal is detected in a region in which no large earthquakes occurred; we account for this signal by another aseismic fault. This fault adds to another evidence for the recent hypothesis that aseismic fault motion drives earthquake swarm.

Key words: Aseismic Slip, InSAR, Earthquake Swarm, Topographic Growth, Onikobe Caldera

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1 Introduction

The Onikobe area in the backbone range running through NE Japan is one of the most active geothermal areas in Japan (Fig. 1) and has been experiencing M5-class earthquakes roughly every 10 years (Umino et al., 1998; Nakajima and Hasegawa, 2003). Four earthquakes of $M \geq 5.0$ occurred in and around the Onikobe caldera on 11 and 13 August 1996; these earthquakes accompanied over 1700 earthquakes of $M > 1$ by the end of September (Japan Meteorological Agency (JMA), Fig. 1), and thus we may regard the entire earthquake sequences as a swarm event. Fig.2 shows the focal mechanisms of major earthquakes determined by Umino et al. (1998), including previous earthquakes in 1970s and 1980s. This area is characterized by very complex rock types (mainly volcanic), and densely distributed faults as evidenced by those earthquake mechanisms (Fig.2); it should also be noted that none of those faults were exposed on the surface. While earthquake fault models were estimated for those large earthquakes from seismic data (Umino et al., 1998; Okada et al., 2001) and previous studies had already reported a detection of associated ground displacements by Interferometric Synthetic Aperture Radar (InSAR) (Aoki et al., 2002; Yagai et al., 2002), the ground displacement pre-

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dicted from the seismologically derived fault models did not fit the observed InSAR data as we discuss in Section 2.

Here we propose a new fault model that consists of not only seismic but also aseismic faults and discuss its two distinct implications. First, with the aid of a digital elevation model (DEM) and pre-existing geological and aftershock distribution data, we demonstrate that the InSAR data cannot be explained unless we introduce aseismic faults, and that the aseismic slip contributes to local topographic growth. Second, we test an emerging hypothesis that aseismic fault motion drives earthquake swarms, by employing detailed hypocenter data (Vidale and Shearer, 2006; Lohman and McGuire, 2007). Aseismic slip has been documented only in well-developed faults such as oceanic subduction zones (e.g., Heki et al., 1997; Thatcher, 2001), and large scale continental transform faults (e.g., Rosen et al., 1998; Bürgman et al., 2000b; Fielding et al., 2004; Furuya and Satyabala, 2008). Although there are no well-developed faults, the Onikobe area is characterized by very high heat flow and hydrothermal activities, which is preferable for the aseismic slip as demonstrated by some recent studies (McGuire et al., 2005; Vidale and Shearer, 2006; Lohman and McGuire, 2007). Our study verifies that aseismic slips can also occur in undeveloped fault under particular conditions.

2 Observation Results and Fault Modeling

InSAR images can detect mm to cm order changes in radar line-of-sight (LOS) through differential measurement of the phase component of temporally separated SAR signals (e.g., Massonnet and Feigl, 1998; Bürgmann et al., 2000a; Hanssen, 2001). In this study, we used L-band (wavelength: 23.6 cm) JERS

data to overcome interferometric coherence loss because the studied area is densely vegetated. Fig. 3 shows the three interferograms with good coherence created from six independent acquisitions. To suppress atmospheric noise and to evaluate measurement errors, we averaged those three independent interferograms (Fig. 4a) and defined the standard deviation σ for each pixel of the averaged interferogram (Fig. 4d) (e.g., Goumelen and Amelung, 2005; Furuya et al., 2007) by the following equation:

$$\sigma(j) = \sqrt{\frac{\sum_{i=1}^3 (X_{avr}(j) - X_i(j))^2}{3 - 1}} \quad (1)$$

where j is the number of pixel, $X_{avr}(j)$ and $X_i(j)$ stand for the phase of the averaged and the i -th interferograms, respectively. The denominator is 2 considering the unbiasedness. Our aim is to construct a fault model which can explain the averaged interferogram (Fig.4a) within the standard deviation at each pixel (Fig.4d).

Fig. 4a is characterized by the following three distinct signals. First, the largest signal is observed in an inverted triangle-shaped area to the north of the Onikobe caldera, where the radar LOS distance decreases by 14–18 cm around the Torage (TRG) area (Fig. 1). Particularly remarkable is the very steep gradient in Δ LOS at the eastern and western margin of this area, which reaches 10 cm/km (equivalent to a strain of 10^{-4}) to locally 20 cm/km on both sides (Fig. 5 top). Second, a broad positive Δ LOS up to 5 cm is observed in the northeastern part of the Onikobe caldera. Finally, a negative Δ LOS area is found to the south of the TRG area, where no earthquakes larger than M4.6 took place; this signal is robust in view of the two independent interferograms (Figs. 3b and 3c). It was concluded that all ground deformations occurred

67 during the earthquake swarm event because no deformation signal could be
 68 detected in the interferograms before and after this event (Fig. 6).

69 Based on the focal mechanisms of large earthquakes and aftershocks, Umino
 70 et al.(1998) proposed a fault model that consists of three faults; we denote
 71 Umino et al’s three faults as U1, U2, and U3 (Table 1). To represent the four
 72 largest earthquakes (Fig. 1), Umino et al. (1998) put U1 for two consecutive
 73 M5.9 and M5.4 events on Aug 11, U2 for the M5.7 event on Aug 11, and U3
 74 for the M5.0 event on Aug 13. Fig. 4b shows the calculated ΔLOS based on
 75 Umino et al’s model; we used the elastic dislocation model (Okada, 1992) to
 76 compute the ΔLOS . Fig. 4e shows the root mean square (RMS) of the misfit
 77 between Figs. 4a and 4b, which can be defined as follows:

$$78 \quad RMS(j) = \sqrt{(X_{avr}(j) - X_s(j))^2} \quad (2)$$

79 where j is the number of pixel, $X_{avr}(j)$ and $X_s(j)$ stand for the phase of the
 80 averaged and the synthetic interferograms, respectively. Fig. 4e shows large
 81 discrepancies to the east and the west of the TRG area, because this fault
 82 model cannot explain the steep gradient in ΔLOS noted above. Even if we
 83 considered a heterogeneous slip distribution on U1, we could not explain the
 84 steep gradient in ΔLOS (Supplemental Fig. S1) not only because U1 is too
 85 deep (Table 1) to cause such localized signals but also because we assume a
 86 plane surface for the slip distribution. U2 and U3 are too far to affect the
 87 observed steep gradient in ΔLOS . Also it is unlikely that heterogeneities in
 88 elastic properties, if they exist, make such very steep gradient in ΔLOS . Thus
 89 we are led to prescribe two reverse faults at the western and eastern margins of
 90 the TRG area, which we hereafter denote F4 and F5. We cannot use seismicity
 91 to estimate the geometry of F4 and F5 because few earthquakes occurred

92 around these faults (Fig. 1). Instead, when comparing Fig.4a and 4c, the 2-
93 dimensional pattern of ΔLOS tightly constrains the geometry and location of
94 these faults.

95 It is unrealistic to constrain all the fault geometries only from seismicities
96 and faults on the surface, as the studied area is highly fractured due to past
97 caldera collapse and each fault is not exposed on the surface as noted above.
98 This is also evidenced by the mismatch between aftershock locations and the
99 geometry of the faults U1–U3. The essence of our modelling strategy is to
100 explain the InSAR data, using minimum number of fault planes. We named
101 each fault of our original model as F1 F2... to clearly distinguish our model
102 from Umino’s fault model U1–U3. Considering from different types of focal
103 mechanisms (Fig.2), at the very least, we have to introduce two faults F1 and
104 F2; F1 corresponds to M5.9 and M5.4, and F2 represent M5.7 earthquakes.
105 Meanwhile, comparing Figs. 4a and 4b, we cannot find any InSAR signals
106 associated with the M5.0 earthquake that led Umino et al to propose their U3,
107 which is presumably because the moment release is small (Table 1). Therefore,
108 we did not take this earthquake into account. Considering the good accuracy
109 of hypocenter determination, the negative ΔLOS to the south of the TRG
110 area is unlikely to be caused by the M5.0 event and therefore it needs to be
111 explained by another source F3, which is not a repositioned model of U3 but
112 totally new fault detected solely by InSAR as well as F4 and F5. It should
113 be noted that this signal is very robust in light of short wavelength and small
114 standard deviation σ (Fig.4d). We discuss the geophysical implication of F3
115 in Section 3.2. Thus, we introduce five faults as the minimum fault number
116 to account for the InSAR and seismological data. We stressed again that few
117 earthquakes occurred in and around the faults F4 and F5, and that there are

118 no specific earthquakes correspond to the faults F3–F5 (Figs. 1 and 2).

119 We first found plausible ranges of each fault parameter by a trial-and-error for-
120 ward modeling approach, and then adopted a grid-search approach to obtain
121 the fault parameters in our model (Table 2). Although non-linear inversion
122 approach is often taken to constrain nine parameters for a uniform slip model
123 (e.g., Aoki et al.1999), it is unrealistic to precisely determine the 45 indepen-
124 dent parameters by any non-linear inversion methods. Also, any non-linear
125 inversion methods need very good initial estimates predetermined by forward
126 modeling. We think that if the initial estimate explains the data (Fig.4a)
127 within the error level (Fig.4d) it should be very close to the true solution,
128 and that our approach is a realistic and reliable way at the moment. Con-
129 sidering the complex crustal structure in this area, we do not rely absolutely
130 on the fault parameters of U1 and U2 in making our model. Actually, we
131 could not explain the InSAR data when we incorporated U1 and U2 in our
132 model without any modifications. We used the fault parameters of U1 and U2
133 as the initial estimates of F1 and F2. Because of the strong dependence of
134 ΔLOS on each parameter, the acceptable ranges obtained in our model are
135 quite narrow, especially for the fault location. The synthetic ΔLOS calculated
136 from our preferred model and its RMS misfit are shown in Figs. 4c and 4f,
137 respectively. The RMS misfits are largely within the standard deviation of the
138 observed data (Figs. 4d and 4f). However, we can identify a spotty but signif-
139 icant misfit in Fig. 4f. We believe that the spotty misfit is caused by a local
140 small landslide that was probably triggered during the swarm event because
141 (1) the area closely matches the local landslide morphology (H. Yagai and T.
142 Ozawa, personal communication, 2008) and (2) the sign of ΔLOS is consistent
143 with the direction of the landslide.

144 It should be noted that the faults F4 and F5 are located within gaps in the
 145 swarm seismicity (Fig. 1), which would be explained by extraordinary high
 146 temperature to the east and west of the TRG areas estimated from the bore
 147 hole data (Tamanyu et al., 1996), low S-wave velocity anomaly, and anoma-
 148 lously high S-wave attenuation (Onodera et al., 1998). Actually, we also ob-
 149 served few seismicities in the southern part of the Onikobe caldera where
 150 geothermal gradients far exceed $200^{\circ}\text{C}/\text{km}$ and a geothermal power plant ex-
 151 ists (PP in Fig.1). The depth of very few earthquakes in the vicinity of F4
 152 and F5 are far deeper than the inferred depth ranges of F4 (0.5 km to 4.5 km)
 153 and F5 (0.6 km to 6.6 km), and the magnitudes of those earthquakes are very
 154 small ($M < 3.0$). Only one medium earthquake ($M4.7$) occurred close to the
 155 southern tip of F5, but it accounted for only $\sim 20\%$ of the slip amount of F5.
 156 Hence, F4 and F5 clearly represent faults that slipped aseismically, although
 157 the exact timing of the slip remains uncertain because of poor time resolution
 158 of the InSAR data.

159 **3 Discussion**

160 *3.1 Insights Into Topographic Growth*

161 Cumulative crustal deformation due to repetitive fault motion and surface
 162 processes over a long time-scale are known to produce topographic expressions
 163 (e.g., Burbank and Anderson, 2001; Hallet and Molnar, 2001). By combining
 164 a DEM and the InSAR technique, we have found that the TRG area is an
 165 intriguing example of the coupling between tectonics and surface processes.

166 Fig. 5 (top) shows the cross-sectional profile of the surface topography and

167 the Δ LOS along a track across the TRG area; this figure shows that the
 168 location of abrupt changes in Δ LOS due to F4 and F5 coincides with that
 169 of the sharp ridges at the western and eastern margins of the TRG area.
 170 While the subsurface reverse fault motion on F4 and F5 does contribute to
 171 the growth of the sharp ridge, the surface processes, especially river incision,
 172 also affect topographic evolution because of the high annual rainfall in this
 173 region. Within the region bounded by the sharp ridge (indicated by a black
 174 line), there are two rivers running in deep V-shaped valleys (Fig. 5 top and
 175 middle), which should have denudated the inner flank of the sharp ridge. The
 176 combination of the steep and localized deformations due to the slips on F4
 177 and F5 and the river incision process has driven the topographic evolution of
 178 the sharp ridge. We also observe a domal topographic high above the central
 179 part of F1, where the ground displacement toward satellite culminates (Fig. 5
 180 top and Fig. 4c). River drainage patterns are widely used as good geomorphic
 181 markers of crustal deformation (e.g., Burbank and Anderson, 2001; Hallet
 182 and Molnar, 2001); Fig. 5 (middle) clearly shows that the two rivers (red
 183 lines) change their course around the domal topographic high, which suggests
 184 that the domal high grows rapidly at a rate comparable to or faster than the
 185 river incision rate. Our model explains this uplift pattern well in terms of the
 186 superposition of surface uplift due to F1, F4, and F5. The contributions of F4
 187 and F5 are large because the domal high is located on the hanging-wall side
 188 of both faults.

189 The very steep faults, F4 and F5, would have originated in the Mio-Pliocene
 190 caldera collapse, as revealed by detailed geological surveys (Ito et al., 1989;
 191 Yoshida, 2001). Although most surface topographies related to the caldera
 192 formation are now buried underneath subsequent volcanic sediments, many

193 subsurface structures have remained and currently serve as pre-existing weak-
194 nesses under the present regional tectonic regime. Umino et al. (1998) have
195 already pointed out the importance of old subsurface caldera structures as
196 pre-existing weaknesses for earthquake rupture; however, this study demon-
197 strates that the old subsurface caldera structures can also generate aseismic
198 slip, such as those found on F4 and F5, and contribute to the topographic
199 growth.

200 *3.2 Aseismic Fault Drives Earthquake Swarm?*

201 The stress triggering mechanism has been successfully exploited to account
202 for the spatiotemporal evolution of earthquake sequences (e.g., Stein, 1999;
203 Scholz, 2002). For earthquake swarm episodes around volcanic or geothermal
204 areas, however, Vidale and Shearer (2006) and Vidale et al. (2006) have sug-
205 gested that pore fluid pressure fluctuation and/or aseismic deformation would
206 be more likely drivers rather than the stress triggering mechanism. Lohman
207 and McGuire (2007) proposed an aseismic fault model to account for the In-
208 SAR data associated with the Obsidian Buttes swarm in 2005; their model
209 supports the hypothesis that aseismic creep drives some earthquake swarms.

210 Among the proposed fault sources, F3, F4, and F5 do not accompany large
211 earthquakes with equivalent moment magnitudes of these faults (Table 2)
212 and thus we conclude that these faults slip primarily aseismically. While F4
213 and F5 reside in seismicity gaps, we observed numerous small ($M < 4.6$)
214 earthquakes around F3 (Fig. 1). In contrast to the aftershocks around the
215 fault F1, we confirmed that those earthquakes around F3 did not indicate
216 mainshock-aftershock sequences (Fig. 7a). Despite the small magnitudes of the

217 earthquakes, significant deformation was detected by InSAR (Figs. 3 and 4a),
218 and the moment magnitude of F3 far exceeds that of the nearby earthquakes.
219 These observations thus support the emerging model that shallow aseismic
220 slip drives seismic swarms.

221 Although the studied area lies close to an active volcano, Mt. Kurikoma (Fig.
222 1), there is no evidence of magma intrusion during the 1996 Onikobe event, un-
223 like other reports, for example, on the eastern Izu-peninsula (Aoki et al., 1999)
224 and near Lake Tahoe (Smith et al., 2004). As another swarm mechanism, pore
225 pressure perturbation cannot be precluded because it can be caused by water
226 as well. The seismic tomography indeed indicates abundant water in the up-
227 permost crust of the Onikobe area (Nakajima and Hasegawa, 2003). It should
228 be noted, however, that such a water-rich environment is also favorable for
229 aseismic slip. In addition, as evidenced by low V_s anomaly over 10% (Onodera
230 et al., 1998) and a very high geothermal gradient (100–150 C°/km, or higher)
231 (Tanaka et al., 1999; Tamanyu et al., 1996), the Onikobe area is clearly under
232 high temperature, which suggests that the brittle-ductile transition zone is
233 shallow and that the aseismic slip on F3 is likely to occur across the transi-
234 tion zone. Considering that there is no direct observation of anomalous water
235 diffusion during the swarm, we speculate that abundant geothermal fluid does
236 not directly drive swarm events but assists aseismic slip. We need to check if
237 aseismic deformation always occurs for the other swarm events as well.

238 To quantify the amount of aseismic slip on F3, we plotted the depth ranges of
239 the fault slip and seismicity just below and above the fault F1 (Fig.7b) and
240 F3 (Fig.7c), respectively. The localized and robust signal detected by InSAR
241 near F3 (Fig. 4) requires the fault slip on F3 limited to shallow depth. Figure
242 7b indicates that the fault slip on F3 is much shallower than that of seismicity

243 unlike F1 (Fig.7c), which demonstrated that almost all the fault slip on F3
244 is aseismic. Considering observation errors of the moment magnitudes and
245 hypocentral depths, the aseismic slip on F3 is not obvious. However, such a
246 very shallow earthquake that accounts for the remarkable InSAR signal around
247 F3 is unlikely to occur because of low crustal strength. Thus we conclude that
248 the aseismic slip on F3 is highly likely. Further discussion is difficult because
249 the rock type distribution is very complex and F3 locates in the marginal area
250 of very high geothermal gradient in the southern part of the Onikobe caldera
251 as evidenced by the geothermal power plant, geysers, and few seismicities
252 (Fig.1).

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359 Figure Captions

360 **Fig. 1.** Tectonic map of the 1996 Onikobe earthquake swarm area. The circles
361 indicate the hypocenters given by JMA (11 Aug. 1996 – 30 Sept. 1996). The
362 magnitude and depth of the earthquakes are indicated by radius and color, re-
363 spectively. The major earthquakes are shown by different symbols: M5.9 (star,
364 11 Aug.), M5.4 (square, 11 Aug.), M5.7 (diamond, 11 Aug.), M5.0 (inverted
365 triangle, 13 Aug.). The red diamonds indicate major hot springs, geysers,

and fumaroles. The large red diamond denoted by PP indicates a geothermal power plant. TRG: Mt. Torage, KRK: Mt. Kurikoma (active volcano), ARD: Mt. Arao (active volcano, and the center of the Onikobe caldera). The white broken lines show the location of Fig. 5. The rectangles with solid lines indicate the surface projections of the aseismic faults (F4 and F5 in Table 2). The rectangle with white dotted line indicates the surface projection of another aseismic fault (F3 in Table 2).

Fig. 2. Focal mechanisms of major earthquakes superposed on the geologic map (Geological Survey of Japan, 2004). The focal solutions (lower hemisphere projection) are given by Umino et al.(1998). We added two large earthquakes occurred on 1976 and 1985 for reference. Legend of the geologic map is simplified. The white triangles indicate main landmarks. TRG: Mt. Torage, KRK: Mt. Kurikoma, ARD: Mt. Arao. The solid lines indicate known faults and the broken lines denote concealed faults. The red circle indicates Onikobe Caldera (after Yoshida, 2001). Note that both focal mechanisms and geologic map reflect highly complex crustal structure.

Fig. 3. Interferograms acquired from three independent periods. a) 23 Jul. 1995 - 5 Oct. 1996, b) 12 Apr. 1996 - 19 Dec. 1997, c) 9 Jul. 1996 - 13 Jun. 1998. Data were taken along descending path 64 and rows 235 to 236. To remove the topographic fringes, we used a digital elevation map with a resolution of 50 m provided by the Geographical Survey Institute, Japan.

Fig. 4. Comparison between observed and synthetic Δ LOS. Positive (negative) Δ LOS stands for an increase (decrease) in the LOS distance. a) Observed Δ LOS. b) and c) are the synthetic Δ LOS calculated from the fault model of Umino et al. (1998) and this study, respectively. d) Standard deviation of a).

391 See text for details. e) Root mean square (RMS) misfit between a) and b).
392 f) RMS misfit between a) and c). The white frames and vectors indicate the
393 fault surface and slip vectors projected onto the surface, respectively.

394 **Fig. 5.** Spatial relationship between Δ LOS, topography, and inferred aseismic
395 faults. The location is indicated in Fig. 1. Top: Δ LOS (blue) and altitude (red)
396 along the white line in the middle figure. Middle: 3D view of the Torage area
397 from the north. Δ LOS value is draped onto the topography. The triangle
398 indicates the summit of Mt. Torage (1432.9 m). The marked negative Δ LOS
399 area is surrounded by a sharp ridge (black line). The domal topographic high
400 inside the sharp ridge shows good spatial correlation with the negative Δ LOS
401 area. The red lines indicate major rivers flowing in deep valleys. Bottom: 3D
402 view of the inferred high dip-angle aseismic faults (F4 and F5) running parallel
403 to the sharp ridge.

404 **Fig. 6.** Interferograms obtained before and after the earthquake swarm. (a)
405 12 Apr. 1996 - 9 Jul. 1996 (before), (b) 5 Oct. 1996 - 19 Dec. 1997 (after), (c)
406 19 Dec. 1997 - 13 Jun. 1998 (after).

407 **Fig. 7.** (a) Time sequence of the earthquakes around F1 and F3 during 11-
408 14 August 1996. The red and green crosses correspond to the earthquakes
409 with those epicenters inside the surface projection of F1 and F3, respectively
410 (Figs. 1 and 4c). The magnitudes are normalized by those of the largest earth-
411 quake for both sequences. The green symbols do not indicate the mainshock-
412 aftershock sequence unlike the red symbols. Depth distribution of earthquakes
413 with those epicenters inside the surface projection of (b) F1 and (c) F3 are
414 also plotted. Red bar in (b) and green bar in (c) indicate the depth range of
415 the fault F1 and F3, respectively.

416 **Fig. S1.** Fault modeling considering slip distribution. (a) Observed Δ LOS
417 (Fig.4a). (b) Synthetic Δ LOS based on distributed slip fault models, Um1 and
418 Um2. The location and geometry of Um1 and Um2 are slightly modified from
419 Umino et al.'s U1 and U2. Slip distribution on Um1 and Um2 are indicated in
420 (c) and (d), respectively. Color indicates the amount of slip in meter. Bottom
421 depth, length, width, dip, and strike of Um1 is 7.5 km, 10 km, 5 km, 45 degree,
422 and 350 degree, respectively. Those for Um2 is 8.0 km, 11 km, 5 km, 90 degree,
423 and 225 degree, respectively.

Table 1. Fault parameters estimated from the seismological data.

Fault	Longitude ^a (deg)	Latitude ^a (deg)	Length (km)	Width (km)	Depth ^a (km)	Dip ^b (deg)	Strike ^b (deg)	Rake ^b (deg)	Slip (cm)	Mw ^c
U1	140.627	38.899	10.0	8.2	7.5	45.0	350.0	90.0	60	6.15
U2	140.649	38.855	10.0	6.0	7.0	90.0	47.0	180.0	40	5.94
U3	140.576	38.800	5.0	3.0	8.0	50.0	49.0	120.0	40	5.54

Each value was scanned from Figure 9 in Umino et al. (1998). We also referred to Okada et al. (2001) while obtaining some parameters from Umino et al. (1998).

^a The position is for the center of each fault plane.

^b We adopted the convention defined by Aki and Richards (1980).

^c Equivalent moment magnitude defined by Kanamori (1977).

Table 2. Fault parameters of our model and those acceptable ranges indicated by $\pm$.

Fault	Longitude ^a (deg)	Latitude ^a (deg)	Length (km)	Width (km)	Depth ^a (km)	Dip ^b (deg)	Strike ^b (deg)	Rake ^b (deg)	Slip (cm)	Mw ^c
F1	140.640 $\pm$ 0.003	38.905 $\pm$ 0.003	9.0 $\pm$ 0.5	5.3 $\pm$ 0.3	6.7 $\pm$ 0.3	40.0 $\pm$ 2.5	340.0 $\pm$ 5.0	110 $\pm$ 5	52 $\pm$ 3	5.95 $\pm$ 0.05
F2	140.650 $\pm$ 0.002	38.860 $\pm$ 0.002	7.8 $\pm$ 0.5	6.8 $\pm$ 0.2	5.5 $\pm$ 0.2	80.0 $\pm$ 2.5	47.0 $\pm$ 2.0	−169 $\pm$ 1	44 $\pm$ 3	5.93 $\pm$ 0.04
F3	140.648 $\pm$ 0.002	38.831 $\pm$ 0.003	5.8 $\pm$ 0.4	2.0 $\pm$ 0.2	2.1 $\pm$ 0.3	37.5 $\pm$ 2.5	12.5 $\pm$ 2.5	115 $\pm$ 3	30 $\pm$ 2	5.38 $\pm$ 0.07
F4	140.601 $\pm$ 0.001	38.909 $\pm$ 0.001	6.0 $\pm$ 0.5	4.0 $\pm$ 0.2	2.5 $\pm$ 0.1	86.0 $\pm$ 2.0	325.0 $\pm$ 2.5	80 $\pm$ 10	20 $\pm$ 2	5.48 $\pm$ 0.07
F5	140.663 $\pm$ 0.002	38.916 $\pm$ 0.002	5.0 $\pm$ 0.5	6.0 $\pm$ 0.2	3.6 $\pm$ 0.2	84.0 $\pm$ 2.0	21.0 $\pm$ 5.0	90 $\pm$ 10	20 $\pm$ 5	5.54 $\pm$ 0.11

^a The position is for the center of each fault plane.

^b We adopted the convention defined by Aki and Richards (1980).

^c Equivalent moment magnitude defined by Kanamori (1977).

Fig.1

Fig.2

a) 23 Jul.1995 - 5 Oct. 1996

b) 12 Apr.1996 - 19 Dec. 1997

c) 9 Jul.1996 - 13 Jun. 1998

Fig.3

Fig.4

Fig.5

a) 12 Apr. 1996 - 9 Jul. 1996

b) 5 Oct. 1996 - 19 Dec. 1997

c) 19 Dec. 1997 - 13 Jun. 1998

Fig.6

Fig. 7

(a) Observed LOS

(b) Synthetic LOS

(c) Slip distribution on U_{m1} (c) Slip distribution on U_{m2}

Fig. S1