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1 Migration of tremor locations before the 2008 eruption
2 of Meakandake Volcano, Hokkaido, Japan

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7 **Abstract**

8 We estimate the locations of three tremor sequences, denoted A, B, and
9 C, that occurred before the 2008 eruption at Meakandake volcano, eastern
10 Hokkaido, Japan, using the spatial distribution of seismic amplitudes of vol-
11 canic tremor. Although we used only five seismic stations, we could estimate
12 the location of three tremor sequences. We find two source areas. The loca-
13 tion area of tremor A is about 2 km southwest of the erupted crater (area
14 a) and the other about 1 km southeast to south of the crater as that of the
15 tremor B (area b). For the tremor B, the location of its early phase is esti-
16 mated in area a while the location of its later phase appears to connect the
17 area a and b. This location migration of tremor B occurred simultaneously
18 with other important geophysical phenomena before the eruption event. Our
19 findings of two location areas of tremor sequences and location migration

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20 of tremor B are important for monitoring activities at Meakandake volcano,
21 particularly the location of a possible phreatic eruption.

22 *Keywords:* tremor location, migration, phreatic eruption, Meakandake

23 **1. Introduction**

24 Volcanic tremor is one of the most important phenomena to interpret
25 on-going volcano activity. Tremor generally occurs when volcano activity
26 is enhanced. Despite its significance, tremor location is rarely estimated,
27 because phase arrivals of tremor are not clear in most cases, so that con-
28 ventional hypocenter determination approaches cannot be applied directly.
29 Recently, with the improvement of portable seismometers, broadband seismic
30 observation around active volcanoes has been widely conducted. Analysis of
31 long-period (LP) or very long-period (VLP) signals including volcanic tremor
32 reveal not only location but also source mechanism of such events. On the
33 other hand, seismic array is another powerful tool to estimate tremor loca-
34 tions. Instead of reading phase arrivals, correlation among waveform records
35 is used in array analysis (e.g. Kawakatsu and Yamamoto, 2007). These two
36 techniques require special equipments, that is, broadband seismometers and
37 seismic array composed of many adjacent stations, respectively. In most
38 cases of active volcanoes, however, seismic observation networks are consti-

39 tuted by only a very limited number of short-period seismometers that can
40 be used only for locating earthquakes with clear phase arrivals.

41 New approaches for tremor location determination have been recently
42 proposed even with a small number of stations of short-period seismometers.
43 Battaglia and Aki (2003) estimated locations of volcanic earthquakes and
44 tremor at Piton de la Fournaise, an active volcano in Reunion Island, using
45 the spatial distribution of seismic wave amplitudes. Their basic idea is the
46 same as an old study of Yamasato (1997) or Jolly et al. (2002) who estimated
47 the source location of pyroclastic flows. Kumagai et al. (2010) revealed that
48 the method of Yamasato (1997), Jolly et al. (2002) and Battaglia and Aki
49 (2003) are based on the characteristics of intensive scattering of seismic waves
50 in volcanic areas, and the location method of seismic events is useful to mon-
51 itor volcanic activities even if there are a limited number of seismic stations.
52 This location method does not need special equipments and appears to be
53 applied to a conventional seismic network with short-period seismometers in
54 most of active volcanoes.

55 Meakandake volcano, located in an eastern part of Hokkaido, Japan, is
56 one of the volcanoes that erupt repeatedly at present in Japan. On November
57 18 and 28, 2008, small phreatic eruptions occurred at Meakandake. Before

58 and associated with the eruption activity, significant seismicity and tremor
59 activity were observed. In this study, we estimate the locations of tremor
60 sequences that occurred during the eruption activities of Meakandake in 2008,
61 using the location method of Battaglia and Aki (2003) or Kumagai et al.
62 (2010). We then discuss the relationship of the estimated tremor locations
63 and the observed eruption activity on the surface.

64 **2. Eruption sequences of Meakandake in 2008**

65 In this section, we shall introduce the overall features of Meakandake
66 volcano and its 2008 eruption sequence.

67 There are three main craters on Meakandake volcanic body, called Pon-
68 machineshiri, Naka-machineshiri and Akanfuji, as shown in Fig. 1. The latest
69 intensive magmatic eruption at Meakandake volcano occurred about 1000
70 years ago. All of the eruptive activities of Meakandake volcano recorded
71 since 1954 were phreatic eruptions at either Pon-machineshiri and Naka-
72 machineshiri (Japan Meteorological Agency, 2005). Especially, recent erup-
73 tions starting in 1988 occurred at a southeastern side of Pon-machineshiri,
74 except for the 2006 eruption which occurred at its northwestern side.

75 Starting in the end of September 2008, seismic activities were intensi-

76 fied at Meakandake (Fig. 2). On November 18, substantial ash fall around
77 the 96-1 crater, located in the southeastern side of Pon-machineshiri, was
78 observed by Volcanic Observations and Information Center, Sapporo Dis-
79 trict Meteorological Observatory (hereafter Sapporo VOIC), recognizing it
80 as the eruption of Meakandake in an official manner. Subsequently, gray
81 plume from the 96-1 crater was observed from November 28 to 29, as inten-
82 sive volcanic activity continued. All the eruptions in November 2008 were
83 phreatic with the amount of ash flows estimated about 12,000 tons in total
84 (Ishimaru et al., 2009). During these eruption activities, no significant visual
85 changes were observed around the 2006 craters located in a northwestern side
86 of Pon-machineshiri (Sapporo VOIC, 2008). Figure 3 shows the hypocentral
87 distribution at Meakandake volcano from July to December 2008. A focal
88 area of seismic swarms starting in September was around the 96-1 crater, in
89 the southeastern side of Pon-machineshiri. No significant horizontal migra-
90 tions of hypocenters were observed before eruptions in November. After these
91 eruptions, hypocenters became slightly shallower than those before them.

92 Some sequences of tremor occurred during the 2008 activities of Meakan-
93 dake (Fig. 2). During the first seismic swarm in the end of September, the
94 first clear tremor with the duration time of 4 minutes occurred on September

95 29. The maximum amplitude of this tremor at station V.MEAB is $2.4\mu\text{m}$,
 96 which is the largest tremor amplitude during the entire 2008 activity. Seis-
 97 micity became high after this tremor, as shown in Fig. 2. Daily frequency of
 98 earthquakes was 788 on September 29, which is also the largest during the
 99 2008 activity. Hypocenters of earthquakes did not change clearly before and
 100 after this tremor (Fig. 3). We call it tremor A. Two days before the eruption
 101 on November 18, tremor with the duration time of 30 minutes occurred on
 102 November 16. Seismic activity did not enhance before and after the occur-
 103 rence of this tremor. After this tremor, however, some continuous tremor
 104 sequences with low amplitude and long duration time occurred frequently.
 105 Especially, the tremor began at 3:37 on November 18 had the longest dura-
 106 tion time of 36 hours. A small eruption is inferred to have occurred during this
 107 tremor sequences (Ishimaru et al., 2009). We call the tremor that occurred
 108 on November 16 as tremor B, and the latter continuous tremor as tremor C.

109 In this study, we estimate the locations of the above three characteris-
 110 tic tremors A, B, and C, using the spatial distribution of tremor amplitudes
 111 (Battaglia and Aki, 2003; Kumagai et al., 2010). These three tremors are im-
 112 portant, because they appeared to be closely related to the intensive seismic
 113 activity and the eruptions.

114 3. Method of estimation of tremor location

115 In this section, we summarize the method of locating tremor with the
116 spatial distribution of its seismic amplitudes (Battaglia and Aki, 2003; Ku-
117 magai et al., 2010). Assuming body waves in a waveform record, seismic
118 amplitude of the band limited seismogram around frequency f at the i -th
119 station, $A_i(f)$, is expressed in the frequency domain as follows;

$$A_i(f) = A_0 \cdot \frac{\exp(-Br_i)}{r_i} \cdot S_i(f). \quad (1)$$

120 where A_0 is the amplitude term at the source, r_i is the distance between the
121 source and the i -th station, $S_i(f)$ is the site amplification factor at the i -th
122 station, and B represents the total attenuation coefficient expressed by the
123 quality factor Q due to scattering and intrinsic attenuations and the velocity
124 of a medium β , as follows;

$$B = \frac{\pi f}{Q\beta}. \quad (2)$$

125 Strictly speaking, the radiation pattern of seismic source, volcanic tremor
126 in this case, is far from isotropic. Kumagai et al. (2011) showed that the
127 observed radiation pattern in 5 Hz or higher frequencies is generally not
128 noticeable, using the simulation of scattering waves with heterogeneous media

129 and topography, so that we may assume isotropic radiation in the present
 130 case. In contrast, the site amplification factor, $S_i(f)$, seems very variable in
 131 this frequency range, so that we need to correct it, as introduced in eq. (1).

132 After correcting the site amplification factor $S_i(f)$, we calculate the co-
 133 efficient A_0 from the observed amplitude at the i -th station ($i = 1 \cdots N$);

$$A_0 = \frac{1}{N} \sum_{i=1}^N A_i r_i \exp(Br_i). \quad (3)$$

134 Normalized residual errors in location estimation is defined as follows;

$$Error \equiv \frac{\sum_{i=1}^N \{A_i - A_0 \exp(-Br_i)/r_i\}^2}{\sum_{i=1}^N A_i^2}. \quad (4)$$

135 To estimate each assigned tremor location, we would minimize the above
 136 error in eq. (4). As seen in eq. (1), we have to assume the medium velocity β
 137 and quality factor Q . Since the majority of tremor at frequency higher than 5
 138 Hz is composed by direct S waves and scattered by structural heterogeneities
 139 along the ray path of direct S waves (Kumagai et al., 2010), we use $\beta = 1.44$
 140 km/s in this study, same as the S wave velocity used by Sapporo VOIC for
 141 earthquake location at Meakandake. No investigations has been conducted
 142 about the estimation of Q around Meakandake. We assume Q to be constant

143 of 50, which Koyanagi et al. (1995) estimated around volcanic areas in Hawaii.

144 4. Data

145 Sapporo VOIC operates six seismic stations with seismometers with the
146 natural period of 1 s to monitor seismic activities of Meakandake volcano
147 (Fig. 1). While two stations, V.NSYM and V.PMNS, have only an UD
148 component, the other stations have three components. For the estimation of
149 tremor locations, we selected tremor waveforms from continuously recorded
150 waveforms at five stations, V.MEAA, V.MEAB, V.PMNS, V.NSYM and
151 V.MNDK (Fig. 1). Since two stations, V.NSYM and V.PMNS, have only an
152 UD component, we used this component at each record in this study. We
153 filtered tremor waveforms with a bandpass filter of 5-10 Hz ($f = 7.5$ Hz in eq.
154 1), because we may assume an isotropic radiation due to scattering effect of a
155 medium in this frequency range (Kumagai et al., 2011). Next, we calculated
156 each RMS amplitude using three time windows (20 s, 60 s, and 100 s) in the
157 duration time of the tremor. Since the radiation pattern of tremor source is
158 collapsed by structural heterogeneities along ray path, the energy of direct S
159 waves from tremor source is mainly shared in its early coda waves. Although
160 such coda waves does not radiation pattern, the energy of coda waves are

almost the same as those of direct S waves. Calculating RMS amplitude emphasize the energy of those coda waves, so it is reasonable to apply eq. (1) which is correct for direct S waves in estimation of tremor location. The spatial resolution scale that we focus is about 2 or 3 km, corresponding to the travel time of S wave to be about a few seconds in record. Although Kumagai et al. (2010) proposed that it is effective to shift the time window for RMS calculation according to the travel time from assumed source to the station, we does not shift the time windows for RMS calculation, because travel time of S wave is smaller than the time windows.

After calculating RMS amplitudes and correcting site amplification effect at five stations, we conduct a grid search in the three-dimensional space to minimize eq. (4) by changing source locations. Intervals of the grid search were 0.001 degree in horizontal directions and 0.1 km in a depth direction.

All the date and time used in this paper are based on its local time system called Japan Standard Time.

5. Estimation of site amplification factors

As explained in section 3, the correction of site amplifications is important in the present location analysis. The coda normalization method (Phillips

179 and Aki, 1986) has been widely applied to estimate site amplification factors.
180 Coda waves are generally defined in the time window after twice as travel
181 time of the S waves and are composed of backscattered waves from structural
182 heterogeneities spread over a large area. In this case, the amplitude of coda
183 waves depends only on its source factor and site amplification factor, because
184 all the records show a common-value of coda Q in high frequency, including
185 our seismic stations around Meakandake. Thus, we can estimate the relative
186 site amplification factor of each station stably with respect to a given master
187 station by the coda normalization method.

188 We estimate site factors by the coda normalization method with regional
189 events shown in Fig. 4 because the source factor of each earthquake can be
190 assumed common. Using the unified hypocenter catalogue of Japan Mete-
191 orological Agency and the JMA2001 travel time table (Ueno et al., 2002),
192 we calculated the arrival time of the direct S phase in each record. Next,
193 we selected seismic waveforms in the duration of 10.24 s after twice as the
194 calculated S arrival as our coda window, then calculated power spectrum
195 ratios to that of the reference station V.MEAB in which the absolute site
196 effect seem to be minimum for all the stations. A parzen window of 0.4 Hz
197 was applied to obtain smooth power spectrum. In an active volcano area, the

198 coda normalization method would not give proper site amplification factors
199 in some cases, probably because of the existence of localized heterogeneities
200 such as a magma chamber (Kumagai et al., 2009). In order to test this, we
201 also calculate the power spectrum ratios of direct S waves, with 20.48 s in
202 duration starting at 0.5 s before the predicted S arrivals, in order to check
203 the stability of site amplification factors estimated by the coda normalization
204 method.

205 Figure 5 shows each measured coda spectrum ratios in gray, and the
206 averaged ratio in black for each station. The averaged spectrum ratio of the
207 direct S wave is also shown in Fig. 5. In spite of all the stations located
208 within about 3 km from the summit of Meakandake, spectral ratios vary
209 significantly among stations. For example, V.PMNS and V.NSYM are in the
210 similar environment (in the steep slope of a volcanic body), but the factors
211 differ significantly from each other. While a relatively steep spectral peak
212 between 4 and 5 Hz is seen in V.PMNS, no such peaks are found in V.NSYM.
213 On the other hand, V.MNDK, located in the saddle of volcano peaks have
214 no significant spectral peaks below 10 Hz, but almost twice with respect to
215 V.MEAB. These results indicate that we have to correct site amplification
216 effects carefully to apply the present method to estimate tremor locations.

217 The spectral ratios of the direct S wave and coda show almost the same
218 features, for example, a steep spectral peak around 4 Hz at V.NSYM. This
219 fact suggests that there would not be any scatterers distributed in a system-
220 atic manner but they are rather randomly distributed beneath Meakandake
221 volcano, and the coda normalization method would give us proper site am-
222 plification factors.

223 We averaged the spectral ratio between 5 to 10 Hz of each earthquakes,
224 then averaged among all earthquakes to obtain the site amplification factor
225 for this study. The values of the site amplification factor and its standard
226 deviation are shown in table 1. The amplitude of tremor sequences we analyze
227 is large enough so we can ignore the influence of certainty of site amplification
228 factors in eq. (3).

229 **6. Estimation of Tremor locations**

230 In this section, we explain the results of our location estimation for the
231 three tremor sequences mentioned in section 2: Tremors A, B and C, using
232 the method described in section 3.

233 *6.1. Estimation of volcano-tectonic earthquake*

234 First, we estimate the volcano-tectonic (VT) earthquake using the method
235 in section 3 and compare the location with the routine location estimated by
236 phase arrivals to check the accuracy of our location estimation.

237 We selected the VT earthquake occurred at 18:02 on March 23, 2009.
238 Bandpass filtered (5-10 Hz) waveform recorded at station V.MEAB is shown
239 in Fig. 6. Origin time and theoretical arrival times of P and S phases calcu-
240 lated from routine location are also shown in the figure. Small earthquake
241 occurred just before the earthquake, but it did not affect to phase reading
242 and our location estimation. The time window of calculating RMS ampli-
243 tude is 3 s from the appearance time of maximum amplitude. As seen in
244 Fig. 6, the time window corresponds to the direct S wave and its early coda
245 wave. Figure 7 shows the estimated location of this earthquake and routine
246 location. Estimated location by the amplitude is about 1 km south and 1
247 km deeper than the routine location. The result means that our assumption
248 of isotropic radiation pattern of direct S wave and its early coda wave is gen-
249 erally appropriate, but we have to take the difference into account to discuss
250 the absolute location. We think that the horizontal heterogeneity of Q causes
251 such the difference in location estimation. The low Q medium spread under

252 the Pon-machineshiri crater. The wave to the northern station (V.NSYM or
253 V.MEAA) attenuate strongly than that to the southern station (V.MNDK).
254 So the location from our method is southerly compared with that from phase
255 arrivals.

256 Next, we check the influence of the value of Q with the same earthquake.
257 Assuming three values of Q (50, 75 and 100), we conducted the location
258 estimation. Figure 8 shows the result of location with three different Q
259 values. The location of three Q values are almost the same, so that the
260 values of Q does not have significant effect in this location method. So we
261 use 50 for the value of Q , mentioned at section 3.

262 6.2. Tremor A on September 29

263 Figure 9(a) shows the waveform of tremor A at station V.MEAB. Tremor
264 began at 14:11 (81 s in the figure) and ended at 14:15 (321 s in the figure) on
265 September 29. We analyzed an early part between 85 and 180 s, including
266 a part of large amplitude, because a later part seems to be contaminated by
267 some earthquakes that occurred late. Figure 9(b) shows the power spectrum
268 of tremor A at station V.MEAB. No significant spectral peaks are noticed,
269 and the power of 5-10 Hz is sufficient for the present analysis. We calculated
270 RMS amplitudes with the time window of 20 s and each window shifted by

271 10 s. We used the five seismic stations shown in Fig. 1. In the time of large
272 amplitude (about 105 s to 155 s in Fig. 9a), the record of V.PMNS was off
273 scale, so we excluded the data of this station.

274 Figure 10 shows the spatial distribution of errors in our location estima-
275 tion defined by eq. (4) at the time window of 95 s. In a horizontal direction,
276 the area of small errors spreads about 1 km in the NE-SW direction but less
277 than 1 km in NW-SE. It is because the distribution of seismic stations is
278 limited in the NW-SE direction from the possible location, as seen in Fig.
279 1. We may conclude that the precision of tremor locations is about 1 km in
280 both horizontal and vertical directions.

281 The estimated tremor locations for analysis time window of tremor A are
282 shown in Fig. 11. They are concentrated in the western part of Akanfuji,
283 about 2 km southwest from the 96-1 crater. Because of the exclusion of
284 data at V.PMNS, some locations are estimated at almost surface, although
285 locations in the horizontal direction are estimated stably. Figure 12 shows
286 temporal variations of RMS amplitude ratios with respect to V.PMNS. The
287 record of V.PMNS was off scale in the shaded area. The amplitude ratio
288 between V.NSYM and V.MEAB is almost the same during the sequence of
289 tremor A. That is, no significant temporal changes of tremor location are

290 found.

291 Only the ray path from source area of this tremor to the station V.NSYM
292 passes under the Pon-machineshiri crater. So the influence of horizontal het-
293 erogeneity of Q would not affect so much to the absolute location compared
294 with the case of VT earthquake in the previous section.

295 6.3. Tremor B on November 16

296 Tremor B started at 00:56 (410 s in Fig. 13a) on November 16, then strong
297 motions continued for about 8 minutes, followed by weak motions for about
298 20 minutes, as shown in Fig. 13(a) at station V.MEAB. The total duration
299 time of tremor B was about 30 minutes. Visual observations of activities
300 around the craters were not available because the craters were covered by
301 cloud. Seismicity did not change noticeably before or after tremor B (Fig.
302 2).

303 We divided this tremor into three phases, phases 1, 2 and 3, based on the
304 temporal variation of tremor amplitude, as observed, for example, at station
305 V.MEAB (Fig. 13a). The end of phase 3 is set to be at 1440 s, because some
306 earthquakes are overlapped after this timing. Figure 13(b) shows the power
307 spectra of the three phases. The overall spectral shape of phase 2 is similar
308 to that of phase 1. No significant spectral peaks are found during the three

309 phases of this tremor. We calculated RMS amplitudes with the time window
310 of 60 s with each window shifted by 30 s. No seismograms are off scale during
311 this tremor, so that we could use all the five stations for location estimation.

312 The distribution of location errors for the time window at 1190 s (phase
313 3) is shown in Fig. 14. Similar to tremor A (Fig. 10), the area of small errors
314 tends to spread in the NE-SW direction. The precision of our tremor location
315 is estimated to be as 1 km in both horizontal and vertical directions. Figure
316 15 shows the estimated tremor locations for three phases. We plot red, green
317 and blue stars for phases 1, 2 and 3, respectively. Compared with the case of
318 tremor A (Fig. 11), the precision in the depth direction is improved by the
319 additional data at V.PMNS. While the locations of phase 3 are distributed
320 from the north of Akanfuji to the south of the 96-1 crater, those of phases 1
321 and 2 are distributed in the NW area of Akanfuji, almost same as those of
322 tremor A. Although errors in our location estimation spreads in the NE-SW
323 direction (Fig. 14), the differences in locations between phase 1 or 2 and
324 phase 3 does not be caused from the limited spatial resolution in present
325 analysis.

326 Figure 16 shows RMS amplitude ratios of tremor B with respect to
327 V.PMNS. For phases 1 and 2, the amplitude ratios of V.MEAB and V.NSYM

328 are almost the same. In contrast, those of V.NSYM become larger than those
 329 of V.MEAB for phase 3. Considering eq. (1), the amplitude ratio should re-
 330 flect the relative distance between source and station. Nearly similar values
 331 of V.MEAB and V.NSYM mean that the sources for phases 1 and 2 are lo-
 332 cated at the same distance from V.MEAB and V.NSYM. On the other hand,
 333 the large ratio of V.NSYM implies that the distance became short from the
 334 source of phase 3 to the station V.NSYM. In addition, we show the error
 335 distribution of phase 1 and 2, and the time series of errors in Fig. 17. As
 336 seen in Fig. 17(a), (b) and Fig. 14, the peak of small error is only one among
 337 all three phases. This fact means that there may be only one source at the
 338 time. Also, Fig. 17(c) shows that the value of minimum error does not change
 339 significantly among three phases. If there are two sources of tremor, the peak
 340 of small error is broadened and the value of minimum error should become
 341 large compared with the case of one source. It is possible that tremor occurs
 342 simultaneously at two source areas, but the source where waves radiated
 343 strongly should be different between phase 1 or 2 and phase 3. Although
 344 their absolute locations cannot be estimated precisely due to the limitation
 345 of the station distribution in the present analysis, we can clearly conclude
 346 that relative locations of phases 1 and 2 should be different from those of

347 phase 3. In addition, the locations of phase 2 are determined deeper than
348 those of phase 1 (Fig. 15). As seen in Fig. 16, the amplitude ratios of phase
349 2 at all the stations are larger than those of phase 1 in general. This fact
350 means that the tremor source appears to have migrated downwards during
351 the main part of tremor B, from phase 1 to phase 2.

352 The influence of heterogeneity of Q may not be significant at the absolute
353 location area of phase 1 and 2, although the absolute location area of phase
354 3 may be northerly and shallow because of the influence of Q .

355 6.4. Tremor C on November 17 to 19

356 Figure 18 shows the continuous waveform recorded at V.MEAB from
357 09:00 on November 17 to 16:00 on November 19. Beginning and ending of
358 this continuous tremor are showed in blue and red lines, respectively after
359 the determination of Sapporo VOIC. We select the following three parts of
360 the continuous tremor: (a) beginning at 10:15 on Nov. 17, (b) beginning
361 at 21:04 on Nov. 17, and (c) beginning at 3:37 on Nov. 18. The power
362 spectra of five parts of this tremor are shown in Fig. 19. All the spectra
363 were calculated with the time window of one hour and the beginning time of
364 window showed in the legend of Fig. 19. No spectral peaks are noticeable,
365 and the spectral structures are similar to each other. Different from tremor A

366 or B, the amplitude of continuous tremor C are slowly varied. It is therefore
367 difficult to define the beginning and ending of tremor C. We estimated tremor
368 locations for the following three time windows: (a) from 10:30 to 12:30 on
369 Nov. 17, (b) from 21:30 on Nov. 17 to 01:30 on Nov. 18 and (c) from 04:00
370 on Nov. 18 to 06:00 on Nov. 19. We selected these time windows, because
371 the tremor signals were clearly recorded at all the five seismic stations. The
372 time window for RMS amplitude calculations spans 100 s, with each window
373 shifted by every 50 s. Estimated tremor locations are shown in Fig. 20.
374 The locations spread from about 1 km south to southeast of the 96-1 crater.
375 Location depths are distributed from 0.5 km to -0.5 km with the average of
376 0 km (sea level), commonly among the three tremor parts.

377 Fig. 21 shows the amplitude ratios with the time windows of Fig. 18
378 with respect to V.PMNS. Blue lines indicate the time that we estimated the
379 location in our present analysis while not, but counted as tremor by Sapporo
380 VOIC in black lines. The average of amplitude ratios at V.MEAA is about
381 0.3 and that at V.MEAB is about 0.4, which does not vary so much during
382 the analyzed time. These features are similar to those of phase 3 of tremor B,
383 implying that the location of tremor C is clearly different from those of tremor
384 A and phase 1 or 2 of tremor B. The spreading of the estimated locations

385 in the NE-SW direction is probably artificial owned to the geometry of the
386 present station distribution (Fig. 1).

387 The absolute location of tremor C may be northerly and shallow because
388 of the influence of Q, so tremor C would occur just under the 96-1 crater.

389 **7. Eruption Model for the 2008 event**

390 In the previous sections, we revealed two source areas among the tremor
391 sequences that we analyzed. Figure 22(a) summarizes these locations. The
392 locations of tremor A and phase 1 or 2 of tremor B are at the NW part
393 of Akanfuji (area a) while tremor C are near the 96-1 crater (area b). The
394 locations of phase 3 of tremor B appear to connect these two source areas.
395 The distance between two source areas is about 1 km. In addition, the depths
396 of phases 1 and 2 of tremor B are different from each other. Although the
397 precision of their absolute locations are rather unreliable, different relative
398 tremor locations can be confirmed by the systematic differences in RMS
399 amplitude ratio among stations (Figs. 12, 16 and 21).

400 Let us now discuss the relationship of the estimated tremor locations and
401 the results of other geophysical observations. Aoyama and Oshima (2009)
402 reported that a step in tilt was overlapped in tremor B by the analysis of

403 broad-band seismograms. They interpreted that this step was caused by dike
404 intrusions under Pon-machineshiri with its strike in the NW-SE direction
405 with the assumption of dip of 90 degree. They estimated the top of dike
406 should be 0.3 km above sea level to explain the tilt vector at the nearest
407 station to the dike. Hashimoto et al. (2009) reported that total geomagnetic
408 force significantly decreased by 2 nT at a station near the 96-1 crater after
409 tremor B occurred. They considered that heat demagnetization occurred in
410 association with tremor B.

411 Considering these results and our estimated tremor locations, we pro-
412 pose the following processes to the phreatic eruption on November 18 at
413 Meakandake, referring to Fig. 22(b). There was volcanic fluid such as hot
414 underground water and/or vapor under the NW part of Akanfuji. The acti-
415 vated fluid led to tremor A and high seismic activity on September 29. Such
416 fluid was activated again on November 16, exciting tremor B. Unlike tremor
417 A when it stayed there, the fluid moved to the the 96-1 crater, as shown by
418 the migration of locations of phase 3 of tremor B. The fluid intruded in the
419 form of dike with the NW-SE strike (Aoyama and Oshima, 2009), causing
420 heat demagnetization there (Hashimoto et al., 2009). Intruded fluid caused
421 the tremor C under the 96-1 crater on November 17, resulting in the surface

422 eruption on November 18. Alternatively, there are fluids already around the
423 96-1 crater. Tremor at source area a affects around the 96-1 crater. Tremor
424 A caused the seismicity while tremor B affects fluids to lead the demagneti-
425 zation and eruption. In this case, fluid migration was not necessary. In either
426 case, we detected the fluid activities at source area b, which was triggered
427 by the activities at source area a.

428 No significant changes found by visual observations in the 2006 craters
429 during 2008 activity suggest that fluid were not present at that time. It
430 was also confirmed by our analysis that no tremor location estimated around
431 the 2006 craters. The NW end of the dike may be located around the 96-1
432 crater, but not extending to the 2006 craters, which restricts the geometry
433 of intruded dike model of Aoyama and Oshima (2009). In addition, the
434 difference in depth between phases 1 and 2 of tremor B suggests some complex
435 mechanism of dike intrusion, and additional investigation about this intrusion
436 mechanism will be required in the future.

437 8. Conclusions

438 We estimated the locations of tremor sequences that occurred during the
439 2008 activity of Meakandake volcano with the spatial distribution of seismic

440 amplitudes. In spite of only five seismic stations, we found the evidence of the
441 migration of tremor locations before the eruption. The source area of tremor
442 A on September 29 and phases 1 and 2 of tremor B on November 16 is at the
443 NW part of Akanfuji. On the other hand, that of tremor C is around the 96-1
444 crater. The locations of phase 3 of tremor B appear to connect those two
445 source areas. Although the precision of absolute tremor locations is rather
446 low because of the limitation of station distribution, temporal variations of
447 RMS amplitude ratios clearly show that there were two source areas between
448 these three tremors and tremor location was clearly migrated during phase
449 3 of tremor B.

450 Considering other geophysical observations, this location migration may
451 be related to some dynamic behaviors of volcanic fluid which caused small
452 phreatic eruptions. In addition, a systematic difference in location depths
453 during phases 1 and 2 of tremor B appears to reflect some complex mecha-
454 nisms of dike intrusion.

455 Seismic observation in a volcanic area is far from an idealistic plan be-
456 cause of its severe environment. The distribution of seismic stations is limited
457 in general. In addition, there are rarely precursors related to small phreatic
458 eruptions in many cases. Our observation of the migration of tremor loca-

tions before the phreatic eruption is important in monitoring activities at Meakandake volcano, only with the limited seismic stations. Moreover, the depth migration of tremor B is important to study the fluid intrusion process. It is desired to apply the method of Battaglia and Aki (2003) or Kumagai et al. (2010) to estimate the locations of volcanic tremors in other volcanoes so that we will find new insight into both monitoring volcano activities and studying the mechanism of volcano eruptions.

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Station	Factor	S.D.
V.MEAB	1.0	(none)
V.MEAA	0.738	0.228
V.MNDK	2.761	0.249
V.NSYM	1.487	0.639
V.PMNS	2.213	0.507

Table 1: Site amplification factor and its standard deviation of each stations.

Figure 1: Locations of craters and seismic stations (inverted triangles) operated by Volcano Observation and Information Center, Sapporo District Meteorological Observatory around Meakandake volcano. Contours are plotted as each 50 m in elevation. V.AKBT in the southeast of the volcano is not used in this study.

Figure 2: (a) Daily frequency of earthquakes, and (b) temporal sequence of volcanic tremors in Meakandake from July to December, 2008. Diameter of circles represents the maximum amplitude at station V.MEAB. Triangles denote the eruptions on Nov. 18 and Nov. 28. (after the compilation of Sapporo VOIC.)

Figure 3: Hypocentral distribution from July to December 2008 at Meakandake volcano. Note that earthquakes of which hypocenters are determined are as the part of counted earthquakes (Fig. 2b).

Figure 4: Hypocentral distribution used for the estimation of site amplification factors.

Figure 5: Spectral ratios of coda and direct S waves at four stations. All the ratios are taken with respect to V.MEAB as the reference station. Black solid lines show the averaged spectrum ratios of coda waves while gray solid lines represent the ratios of each earthquake of Fig. 4. Black broken lines show the averaged spectral ratios of direct S waves.

Figure 6: Bandpass filtered waveform of VT earthquake recorded at station V.MEAB on March 23. Scales of time and amplitude are shown in the right of figure. O, P_{calc} and S_{calc} mean origin time, calculated arrival time of P and S waves from the routine location.

Figure 7: Estimated location (star) and routine location (circle) of the earthquake occurred at 18:02 at March 23, 2009. Distribution of error are also plotted.

Figure 8: The result of location estimation of three Q values (50, 75 and 100) for the earthquake of Fig. 6.

Figure 9: (a) Waveform of tremor A recorded at station V.MEAB on September 29. (b) Power spectrum of tremor A in the time window of (a) at station V.MEAB.

Figure 10: Spatial distribution of errors for tremor location estimation at the time window of 95 s.

Figure 11: Estimated locations of tremor A represented by the stars. The inverted triangles represent the used seismic stations.

Figure 12: Temporal variation of RMS amplitude ratios of tremor A with respect to V.PMNS. The shaded area represents the time when the seismogram was scaled out at V.PMNS.

Figure 13: (a) Same as Fig. 9(a) except for tremor B on November 16. (b) Power spectrum of tremor B in the each time window of (a) at station V.MEAB.

Figure 14: Same as Fig. 10 except for phase 3 at 1190 s.

Figure 15: Estimated locations of tremor B with three phases; red stars for phase 1, green stars for phase 2, and blue stars for phase 3, respectively.

Figure 16: Temporal variation of RMS amplitude ratios during tremor B with respect to V.PMNS

Figure 17: (a)(b) Same as Fig. 10 except for phase 1 at 470 s and phase 2 at 740 s, respectively. (c) Time series of minimum errors during the estimation of tremor B.

Figure 18: Continuous waveform at V.MEAB from 09:00, Nov. 17 to 16:00, Nov. 19. Start and end times of continuous tremors (after Sapporo VOIC) are denoted in blue and red lines, respectively. The labeled time windows, (a), (b) and (c), are the selected parts to estimate their locations.

Figure 19: Power spectra of continuous tremor C at station V.MEAB. Each time window for spectral calculations is one hour.

Figure 20: Estimated locations of tremor C at three times of Fig. 18.

Figure 21: Temporal variation of RMS amplitude ratios of tremor C from Nov. 17 to Nov. 19 with respect to V.PMNS. The three time windows of the present analysis are shown by blue lines, other tremor sequences by black lines, respectively.

Figure 22: (a) Comparison of locations among tremor sequences. (b) A model of fluid migration and intruded dike.