



Title	Persistent Shallow Background Microseismicity on Hekla Volcano, Iceland: A Potential Monitoring Tool
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Publication date	2014-12-01
Publication information	Eibl, Eva P. S., Christopher J. Bean, Kristín Vogfjörð, and Aoife Braiden. "Persistent Shallow Background Microseismicity on Hekla Volcano, Iceland: A Potential Monitoring Tool." Elsevier, December 1, 2014. https://doi.org/10.1016/j.jvolgeores.2014.11.004 .
Publisher	Elsevier
Item record/more information	http://hdl.handle.net/10197/6281
Publisher's statement	This is the author's version of a work that was accepted for publication in Journal of Volcanology and Geothermal Research. Changes resulting from the publishing process, such as peer review, editing, corrections, structural formatting, and other quality control mechanisms may not be reflected in this document. Changes may have been made to this work since it was submitted for publication. A definitive version was subsequently published in Journal of Volcanology and Geothermal Research (VOL 289, ISSUE#, (2014)) DOI: 10.1016/j.jvolgeores.2014.11.004
Publisher's version (DOI)	10.1016/j.jvolgeores.2014.11.004

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Persistent Shallow Background Microseismicity on Hekla Volcano, Iceland: A Potential Monitoring Tool

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1 Abstract

Hekla is one of Iceland's most active volcanoes. Since 1970 it has erupted four times with a period of quiescence of 14 years since the last eruption. We detected persistent levels of background microseismicity with a temporary seismic network in autumn 2012. An amplitude based as well as an arrival-time based location method was applied to two populations of events and located them at shallow depths on the northern flank, close to the summit. This seismicity has not been identified previously by the permanent seismic network in Iceland as it is below its detection threshold. The detected events were either short, higher frequency events with distinct arrivals located beneath the summit on the northern flank of Hekla or longer, emergent, lower frequency events about 4 km northeast of the summit at 200–400 m depth below the surface. Estimated moment magnitudes were $MW = -1.1$ to -0.1 and $MW = -0.9$ to -0.0 and local magnitudes $ML = -0.5$ to $+0.3$ and $ML = -0.3$ to $+0.3$, respectively. This seismicity does not show any correlation with gas output but is located at the steepest slopes of the edifice. Hence we suggest that the current shallow microseismicity at Hekla is structurally controlled. This offers a possible opportunity of using near summit microseismicity as a tool for monitoring emerging unrest at Hekla. Microseismicity rates will be very sensitive to small stress perturbations due to magma migration at depth. Currently in the absence of microseismicity monitoring, Hekla switches from apparently quiescent to fully eruptive on the order of only 1 h.

2 Introduction

Hekla is one of Iceland's most active volcanoes, located on the Mid-Atlantic plate margin. Its activity is related to its position at the connection between the South Iceland Seismic Zone striking east-west and the Eastern Volcanic Zone striking north-south (Einarsson, 1991). Hekla is elongated in WSW-ENE direction with similarly trending fractures through its summit. Previous eruptions had a repose time of about 60 years before 1970 and about 10 years after 1970 (Soosalu and Einarsson, 1997) with eruptions in 1970 (Einarsson and Björnsson, 1976), 1980/81 (Grönvold et al., 1983), 1991 (Gudmundsson et al., 1992) and its latest in 2000 (e.g. Höskuldsson et al. (2007); Soosalu et al. (2005)). After an eruption on Heimaey Island offshore south Iceland in 1973 (Thórarinnsson et al., 1973) a permanent, analogue seismic network was set up in South Iceland including a station 31 km southwest of Hekla in 1974 (Einarsson and Björnsson, 1987). This station was the closest to Hekla until 1982 when an additional permanent station 22 km west of Hekla was installed (Einarsson and Björnsson, 1987). Those stations recorded tremor

43 during and shortly before the subsequent eruptions which was analysed in further detail
44 (Grönvold et al., 1983). Detailed studies of the earthquakes around Hekla were based
45 on the network operated by the Icelandic Meteorological Office (IMO) that has been
46 recording since the beginning of 1990. Currently the closest permanent seismometer is
47 a digital, 3-component Lennartz 5s instrument in Haukadalur 15 km west of Hekla (see
48 e.g. Jakobsdóttir (2008)). From March 1982 (Einarsson and Björnsson, 1987) until 2010
49 (pers. comm. Páll Einarsson, May 2014) an analogue, vertical-component seismometer
50 was additionally operated 2 km north of the summit. Two other temporary digital, 3-
51 component, broadband stations are operated by IMO since late 2011/ early 2012 about 4
52 km north and 6.5 km south of Hekla (pers. comm. Martin Hensch, May 2014).

53 It has been observed that seismic activity at Hekla is strongly linked to its eruptions (e.g.
54 Einarsson (1991)). The visual beginning of an eruption is accompanied by low-frequency
55 (0.5 - 1.5 Hz with dominant peaks at 0.7 - 0.9) volcanic tremor that decreases the detection
56 threshold for earthquakes classified as high-frequency and low-frequency in Soosalu et al.
57 (2005). Those high and low-frequency events comprise the sparse background activity
58 (Soosalu and Einarsson, 2002; Soosalu et al., 2005). Earthquake signals containing only
59 low-frequencies have been observed at Hekla during inter-eruption periods. They have
60 clear P and S wave onsets, occurred in the 8 - 14 km depth range and were interpreted as
61 tectonic events with low stress drop (Soosalu and Einarsson, 1997). The high-frequency
62 earthquakes were observed during eruptions and a few months before or after an eruption.
63 They occur in the 8 - 12 km depth range and are also interpreted as tectonic earthquakes,
64 but requiring higher strain relative to the low-frequency events (Soosalu et al., 2005).

65 Seismic precursors to the four eruptions since 1970 were detected 25, 23, 28 and about
66 80 minutes (Einarsson and Björnsson, 1976; Grönvold et al., 1983; Gudmundsson et al.,
67 1992; Soosalu et al., 2005) before the visible onset of the eruption. They were thought to
68 be related to movement of magma. In 2000, for instance, the seismicity started gradually,
69 growing in event frequency and intensity (M_L -0.5 to +2.1) over a timescale of several
70 tens of minutes. 80 to 45 minutes before the eruption some of the events could be located
71 in the depth range of 0 - 4 km. In the next ten minutes the seismicity was located at
72 up to 6 km depth and from 35 minutes before the eruption they occurred in up to 14
73 km depth mainly in 4 - 9 km depth. Most of the events also clustered north of the main
74 fissure on the summit of Hekla. With the beginning of the eruption the earthquakes in the
75 swarm became more infrequent and decreased in intensity (Soosalu et al., 2005). When
76 the seismicity reached 6 km depth the first contraction signal was observed at the nearest
77 strainmeter at 15 km distance (Sturkell et al., 2013). The contraction rate increased until
78 the start of the eruption. During previous eruptions a similar behaviour was observed.
79 The first seismicity was detected less than half an hour before the eruption at the same
80 time as strainmeters recorded a contraction signal (or expansion depending on the loca-
81 tion of the strainmeter) (Linde et al., 1993; Gudmundsson et al., 1992).

82 Before and after the most recent eruption in 2000 an inflation of an area with a radius
83 of 20 km around Hekla was observed. Shortly before the eruption in 2000 the ground
84 surface south of the eruptive fissure deformed upwards, north of the fissure it deformed
85 downwards. This was probably linked to the intrusion of a dike (Ofeigsson et al., 2011).
86 A GPS study (Geirsson et al., 2012) also detected an inflation signal at Hekla. This study
87 suggests that the observed inflation continued at least until 2010 or 2011.

88 In summary (i) Hekla is likely still in an inflating phase. (ii) Previous studies have shown
89 that significant levels of seismicity have only been detected on the order of one hour (or
90 less) prior to eruptions. (iii) For these earthquakes three to four times more events were
91 detected at a temporary station ca. 2 km north of the summit in comparison to per-

92 manent instrumentation 15 km to the west (IMO station) (Soosalu and Einarsson, 1997,
93 2002; Soosalu et al., 2005).
94 In this paper we find significant levels of microseismicity using a temporary deployment of
95 five broadband stations in the summit region (August to October 2012). We suggest that
96 such microseismicity might be used to track low level strain fluctuations. Due to the short
97 warning periods ahead of previous eruptions, the characterisation of level, location, size
98 and process of this seismic activity is important from a hazard perspective. We undertake
99 an analysis of these events using a standard amplitude and an arrival-time based location
100 method. The paper describes event characteristics, location estimates and magnitudes.
101 Locations are interpreted in the context of synthetic simulations, deformation and gas
102 observations.

103 3 Seismic network on Hekla

104 We augmented the IMO network from August until October 2012 with five Gralp 6TD
 105 (30 s - 100 Hz) sensors on the summit and the eastern flank of Hekla volcano. The first
 106 instrument became operational on August, 9th. Data sampled at 100 Hz were stored
 107 locally until the instruments were decommissioned on October, 10th. The instrument
 108 locations are given in figure 1, their coordinates in table 1. The network was configured
 109 with a focus on event detection and accessibility of the site. Locations could be improved
 110 with a different geometry and effects are discussed further in section 5.2. The distances
 111 between the stations vary between 1 and 4 km. They were buried up to 20 cm deep in
 112 unconsolidated volcanic material which was frozen soon after the station deployment.

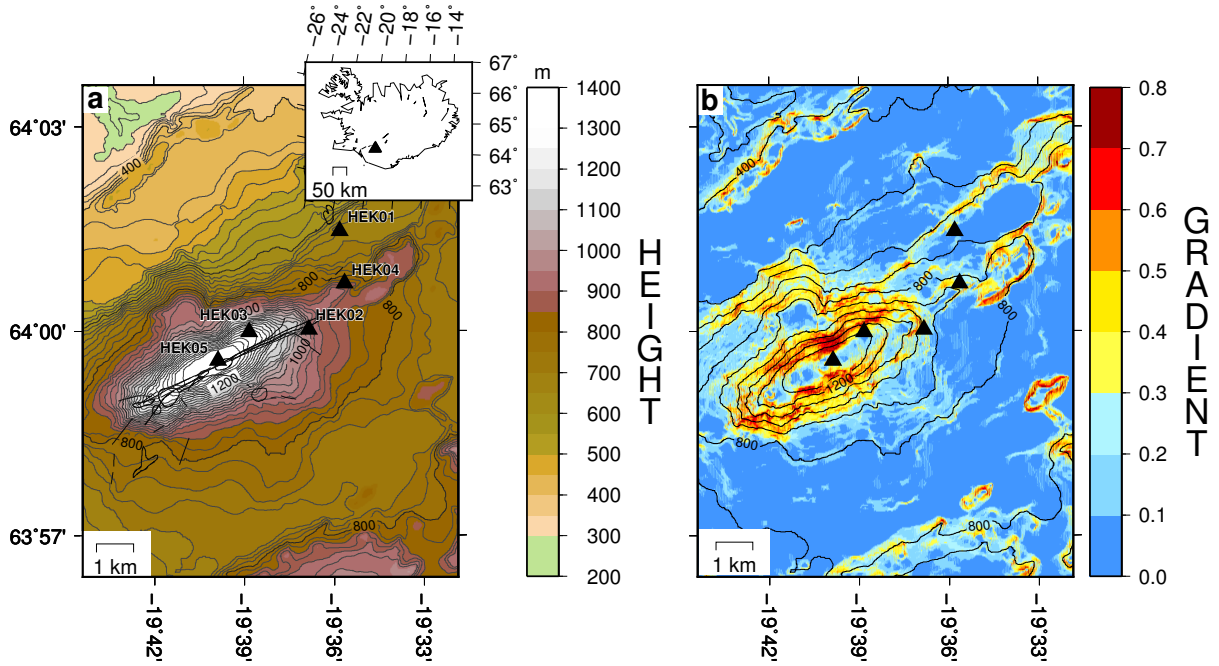


Figure 1: (a) The topography of Hekla volcano and the locations of the seismometers (black triangles). Thin black lines indicate the eruptive fissures of the eruptions in 1970, 1980/81, 1991 and 2000 Hskuldsson et al. (2007). (b) Maximum gradient of the topography indicating the steepest slopes and highlighting the elongation of Hekla along the main eruptive fissures at the summit.

Station	HEK05	HEK03	HEK02	HEK04	HEK01
Station coordinates	63.99281 N 19.66449 W	63.99973 N 19.64707 W	64.00353 N 19.61382 W	64.011646 N 19.593963 W	64.02444 N 19.59652 W
Site correction factor in the 4 - 7 Hz band	1.378350	0.663770	0.851536	0.776647	1.0
Site correction factor in the 7 - 10 Hz band	0.750484	0.360419	0.232864	0.341289	1.0

Table 1: Coordinates and site correction factors at the different stations in the 4 - 7 Hz and 7 - 10 Hz band derived from 53 regional events.

113 4 Observations

114 The seismic background activity detected in August to October 2012 consists of two
 115 apparently different types of events with different frequency content, signal lengths and
 116 onsets. We refer to them as type 1 and type 2 events and show their occurrence in a seven
 117 week period in figure 2. The events were picked automatically with a STA/LTA filter that
 118 triggered only when an event was visible on at least three seismic stations. Type 1 and type
 119 2 events were visually identified based on their signal length and frequency content. The
 120 vertical lines in figure 2 indicate times when seismometers started or stopped recording.
 121 All instruments were deployed within three days. Data gaps occurred in late September
 122 due to loss of power from snow and ice covering the solar panels. The number next to them
 123 corresponds to the number of recording seismometers. If there is a correlation between
 124 the number of type 1 and type 2 events it is weak and we do not regard it as significant.
 125 Station availability affected the detection threshold and local power failures generally led
 126 to an underestimation of the numbers of micro-earthquakes e.g. in late September.

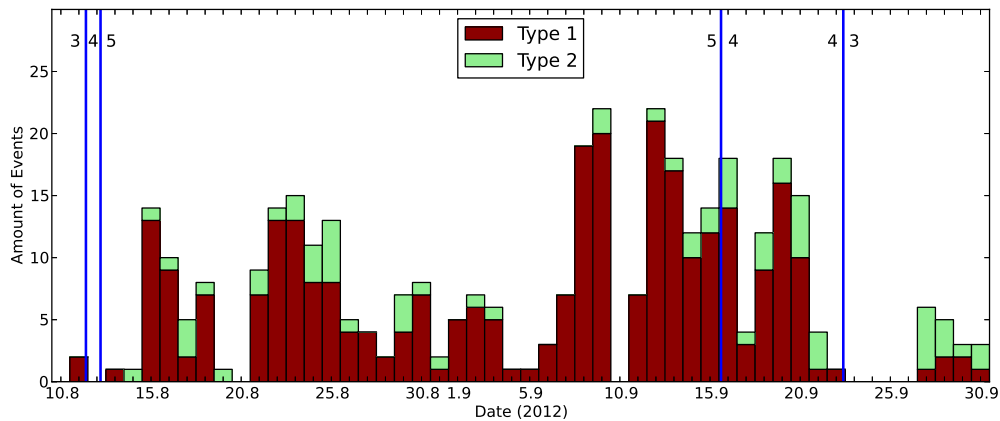


Figure 2: Occurrence of the type 1 and type 2 events during August and September 2012. The vertical lines indicate a change in the number of operational seismometers given by the number.

127 4.1 Type 1 Events

128 The shortest events we detected are 3 - 4 s long in duration, have distinct onsets and
 129 energy between 3 and 20 Hz, mostly around 10 Hz at the station with the strongest and
 130 shortest signal (figure 3a). 85% of the events are earliest (by up to 1 s) at HEK03 (see
 131 figure 1) and about 15% are earliest at HEK05. Some events are barely visible or not
 132 visible on HEK01 where noise levels are slightly higher.

133 Different seismic phases cannot be identified, possibly due to close proximity to the source.
 134 Soosalu et al. (2005) identified clear P and S phase onsets in events which were recorded at
 135 15 km distance from Hekla. They were classified as high- or low-frequency events, came
 136 from the same region as events identified here but had significantly higher amplitudes
 137 and associated signal to noise ratio (S/N). Figure 3 shows the three components of a
 138 typical event on two different stations and their spectra and spectrogram of the vertical
 139 component. Note the significantly lower frequency content on HEK02 although it is only
 140 1.6 km away from HEK03.

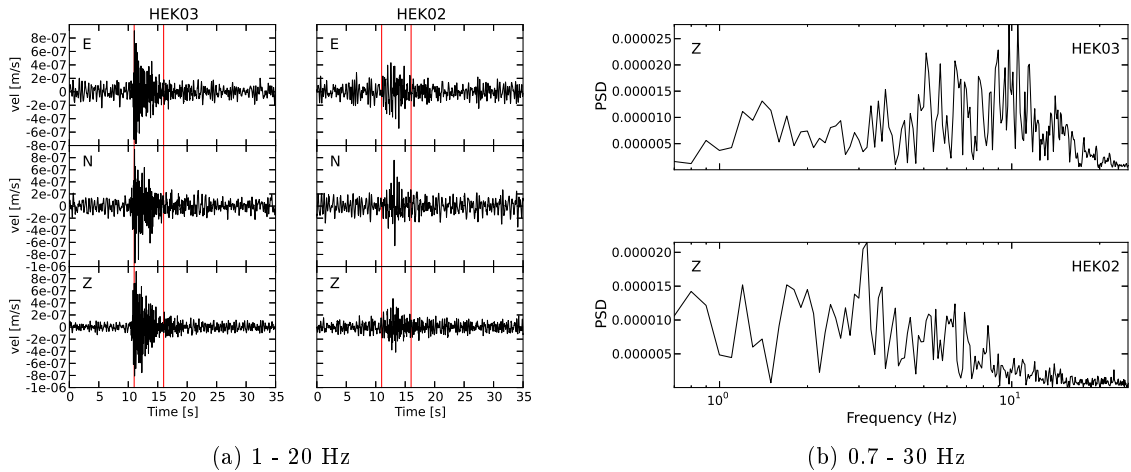


Figure 3: A typical 1 - 20 Hz filtered type 1 event which occurred on August 26th, 2012 at 5:52:19. (a) Instrument corrected seismograms on HEK02 and HEK03 on all three components show the earlier arrival on HEK03. The portion of the signal between the vertical red lines was used in the intensity location method. (b) Spectrum of the vertical component at HEK02 and HEK03 showing the high-frequency attenuation on HEK02. (c) Instrument corrected vertical component seismogram and spectrogram of the event at HEK03.

141 4.2 Type 2 Events

142 The type 2 events (figure 4) are mostly 10 to 30 s long in duration and emergent. The
 143 station closest to the source records frequencies between 1 and 14 Hz with most of the
 144 energy being between 2 and 5 Hz. Due to their emergent nature it is not possible to visually
 145 observe on which station the event arrives first. Figure 4 shows a typical event recorded
 146 on HEK01 and HEK03 on all three components and their spectra and spectrogram. In
 147 contrast to type 1 events, type 2 events have a slightly higher S/N and a higher absolute
 148 amplitude. Their frequency content is also lower while their duration in time is longer.
 149 The diffuse nature and longer duration might be a propagation effect i.e. caused by
 150 scattering if the events occurred outside our station network, at greater distances than
 151 type 1 events. This might also imply that these events resemble the type 1 events but
 152 are slightly bigger in amplitude and occurred farther away from the station. The similar
 153 distance travelled from source to the station might be the reason for the similar spectral

154 content at different stations in contrast to the spectral differences observed for type 1
 155 events. In order to improve the locations we considered stacking but as our events do not
 156 form families this was not possible.

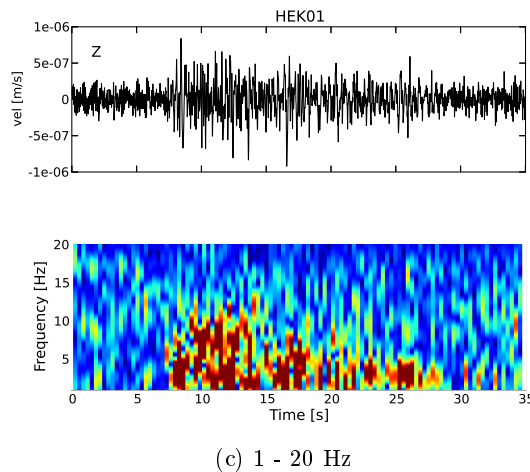
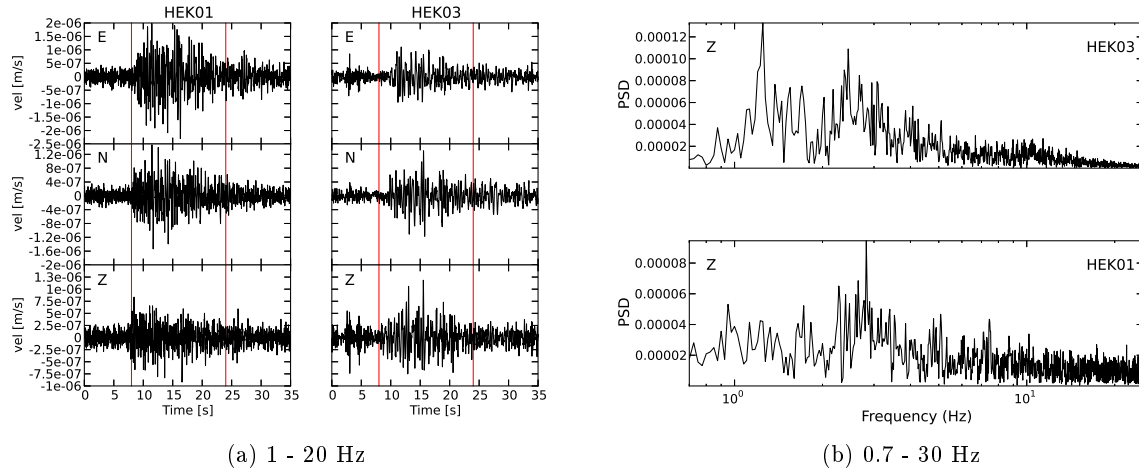


Figure 4: A typical 1 - 20 Hz filtered type 2 event which occurred on August 29th, 2012 at 19:01:13. Subfigures as in figure 3. Note the lower frequency content, the emergent onset and longer duration in comparison to the type 1 events.

157 We have checked the nearest permanent IMO station 15 km west of Hekla for the type 1
 158 and 2 events detected by our network. None of the type 1 events were recorded and about
 159 10% of type 2 events are weakly visible. On the temporary stations at 4 km and 6.5 km
 160 distance only 26 type 2 events and no type 1 events could be detected due to data gaps
 161 and low signal to noise ratios respectively.

162 5 Event Locations: Methodology

163 Arrival-time methods cannot be used for determining locations of our entire database as
 164 most of our events are either emergent or have onsets that are often hidden in noise. We
 165 based our locations on the intensity location method described in (Taisne et al., 2011)
 166 which is based on a study of Battaglia and Aki (2003). In this method the intensity
 167 ratios for all station pairs are calculated using the median of the absolute value of the
 168 Hilbert Transform of the instrument corrected seismogram. These intensity ratio pairs
 169 are then compared with the expected intensity ratios assuming a source at a grid point
 170 in a predefined 3D grid:

$$\frac{I_i(r_i)}{I_j(r_j)} = e^{\frac{\pi \cdot f}{Q \cdot \beta} \cdot (r_j - r_i)} \cdot \left(\frac{r_j}{r_i}\right)^n \quad (5.1)$$

171 where I_i and I_j indicate the amplitudes of the signal at station i and j , r_i and r_j the
 172 distance between the source and seismometer i and j , f the dominant frequency of the
 173 signal, Q the quality factor for attenuation and β the wave velocity in m/s. n is set to
 174 0.5 for surface waves and 1 for body waves. For $n=0.5$ the distance r_i is only calculated
 175 with respect to x and y as we assume a source on the surface. For $n=1$ we inverted for x ,
 176 y and depth (see also Battaglia and Aki (2003)).

177 To calculate the error percentage (RES) of each grid point the square root of the sum of
 178 the squared absolute errors between the observed and calculated intensity ratios divided
 179 by the sum of the squared observed intensity ratios is calculated (Battaglia et al., 2005):

$$RES = 100 \cdot \sqrt{\frac{\sum_i \sum_{j>i} \left(\frac{I_{isyn}(r_i)}{I_{jsyn}(r_j)} - \frac{I_i(r_i)}{I_j(r_j)} \right)^2}{\sum_i \sum_{j>i} \left(\frac{I_i(r_i)}{I_j(r_j)} \right)^2}} \quad (5.2)$$

180 Assumptions are that there is only one source at a given time, that for one event each
 181 station record is dominated by the same seismic phase and that each station has the
 182 same quality. Noise or a station with a bad fit will lead to a low error percentage and
 183 worsen the result. Equation 5.2 is a far field approximation and might create errors for
 184 the events closest to the stations especially in the lower frequency band. S waves will
 185 have wavelengths of around 360 m in the lower frequency band and 235 m in the higher
 186 frequency band. The grid point with the minimum percentage error is assumed as source.
 187 This method was previously used to estimate size, length and velocity of pyroclastic flows
 188 (Jolly et al., 2002), locate volcanic tremor (Battaglia et al., 2005; Battaglia and Aki, 2003),
 189 VT, long-period events (Battaglia and Aki, 2003) and non-volcanic tremor in subduction
 190 zones (Husker et al., 2012) and track lahars (Kumagai et al., 2009). Locations were
 191 initially either visually confirmed by rocks, flow deposits or eruptive vents or by locations
 192 from another location method. Problems with the locations were attributed to more than
 193 one active source, anisotropically radiated seismic energy or trapped seismic energy (Jolly
 194 et al., 2002), saturation problems at the stations, low signal to noise ratios (Battaglia
 195 et al., 2005) and heterogeneities or a magma chamber in the wave path (Battaglia and
 196 Aki, 2003).

197 5.1 Data processing

198 Altogether, 210 type 1 and 40 type 2 events recorded by 5 stations were located. As a first
 199 processing step the seismogram of each event was corrected for the instrument response

200 and filtered to the 4 - 7 Hz or 7 - 10 Hz frequency band. We chose those frequency bands
201 based on the spectral signal strength and isotropic radiation effects at frequencies above
202 5 Hz (Kumagai et al., 2010).

203 In order to perform site corrections we calculated coda amplitudes for 53 regional events.
204 Coda waves are routinely used for site amplification estimations (Kumagai et al., 2009;
205 Battaglia and Aki, 2003; Aki and Ferrazzini, 2000b). The regional events occurred in 15
206 to 150 km distance from Hekla, were less than 8 km deep with moment magnitudes mostly
207 between 1 and 3 and an azimuthal range between 90 and 270°. We used a time window that
208 started at a time that was twice the arrival of the S phase (Aki and Ferrazzini, 2000a;
209 Aki and Chouet, 1975). The coda of the seismogram was instrument corrected before
210 root mean square (RMS) values in 5 s long, non-overlapping windows were calculated
211 (Aki, 1969). We used HEK01 as reference and averaged all RMS values. HEK01 was
212 chosen as reference station as it had the longest seismic dataset and the least local, high-
213 frequency noise given its low elevation and a more sheltered location. RMS values were
214 discarded if a regional event was time-coincident with a local Hekla microseismic event in
215 the corresponding frequency band. The site correction factors are given in table 1.

216 After applying site corrections a grid search assuming body and surface wave propagation
217 was performed. The grid was a rectangular cuboid extending 7 km east-west, 9 km north-
218 south and from sea level up to 1500 m a.s.l. For the grid search we assumed a surface
219 wave velocity of 1.8 km/s and a shear wave velocity of 2.0 km/s. Based on the results
220 from our sensitivity tests (see paragraph 2 in 5.2) we assumed a quality factor of 100. As a
221 comparison, on Piton de la Fournaise volcano shear wave velocities of 2.3 km/s and quality
222 factors of 50 (Battaglia and Aki, 2003) or shear wave velocities of 1.0 km/s and quality
223 factors of 170 (Taisne et al., 2011) were used for locations. The resulting $\beta \cdot Q$ products
224 are consistent with ours and give the least event location scatter based on our sensitivity
225 test (see 5.2). Each grid point was then compared to a topographic map, all points above
226 the topography were excluded and the remaining grid point with the lowest percentage
227 error was picked as the source. Because other studies (e.g. Pálmason (1971)) found lower
228 seismic velocities that might seem more appropriate in shallow volcanic environments, we
229 show locations for a lower quality factor and seismic velocity in the Appendix.

230 5.2 Synthetic Tests

231 The accuracy of the method has been tested previously on visible rockfall, tremor from an
232 eruptive fissure and on a hybrid event and compared with a travel-time location method.
233 For details see Battaglia and Aki (2003), Taisne et al. (2011) and Battaglia et al. (2005),
234 respectively. Locations assuming body and surface waves were similar with better loca-
235 tions for body waves located in the 5 - 10 Hz band (Battaglia and Aki, 2003).

236 We tested the sensitivity of the intensity method with respect to seismic velocity and
237 quality factor using five type 1 events. The seismic velocities varied between 1 km/s and
238 4 km/s and the quality factors from 40 to 200. The influence on the locations was only
239 slight and seems to be best for a $Q \cdot \beta$ product of 180 km/s. For higher values the im-
240 provement is negligible, for lower values the scattering of the locations increases.

241 In order to try to further quantify the effects of station geometry we performed synthetic
242 tests assuming a source in a homogeneous, isotropic medium. The amplitudes of the sig-
243 nal at the seismometers were calculated for body waves using the formula from (Battaglia
244 and Aki, 2003) describing the amplitude decrease with respect to distance:

$$A(r_i) = \frac{A_0}{r_i} \cdot e^{-\frac{\pi \cdot f}{Q \cdot \beta} \cdot r_i} \quad (5.3)$$

245 For the forward calculations of the amplitudes a frequency band of 7 to 10 Hz, a quality
246 factor of 100, a seismic velocity of 2.0 km/s consistent with the above mentioned inversion
247 settings and an arbitrary amplitude at the source A_0 of 2000 for body waves were assumed.
248 A 6 s long Gaussian wavelet sampled at a rate of 100 Hz was used as the source. The
249 synthetic seismograms at the five Hekla stations were then inverted in the 4 - 7 Hz and 7
250 - 10 Hz band assuming $\beta=2.0$ km/s when considering body waves and $\beta=1.8$ km/s when
251 considering surface waves and a quality factor of 100 (figure 5).

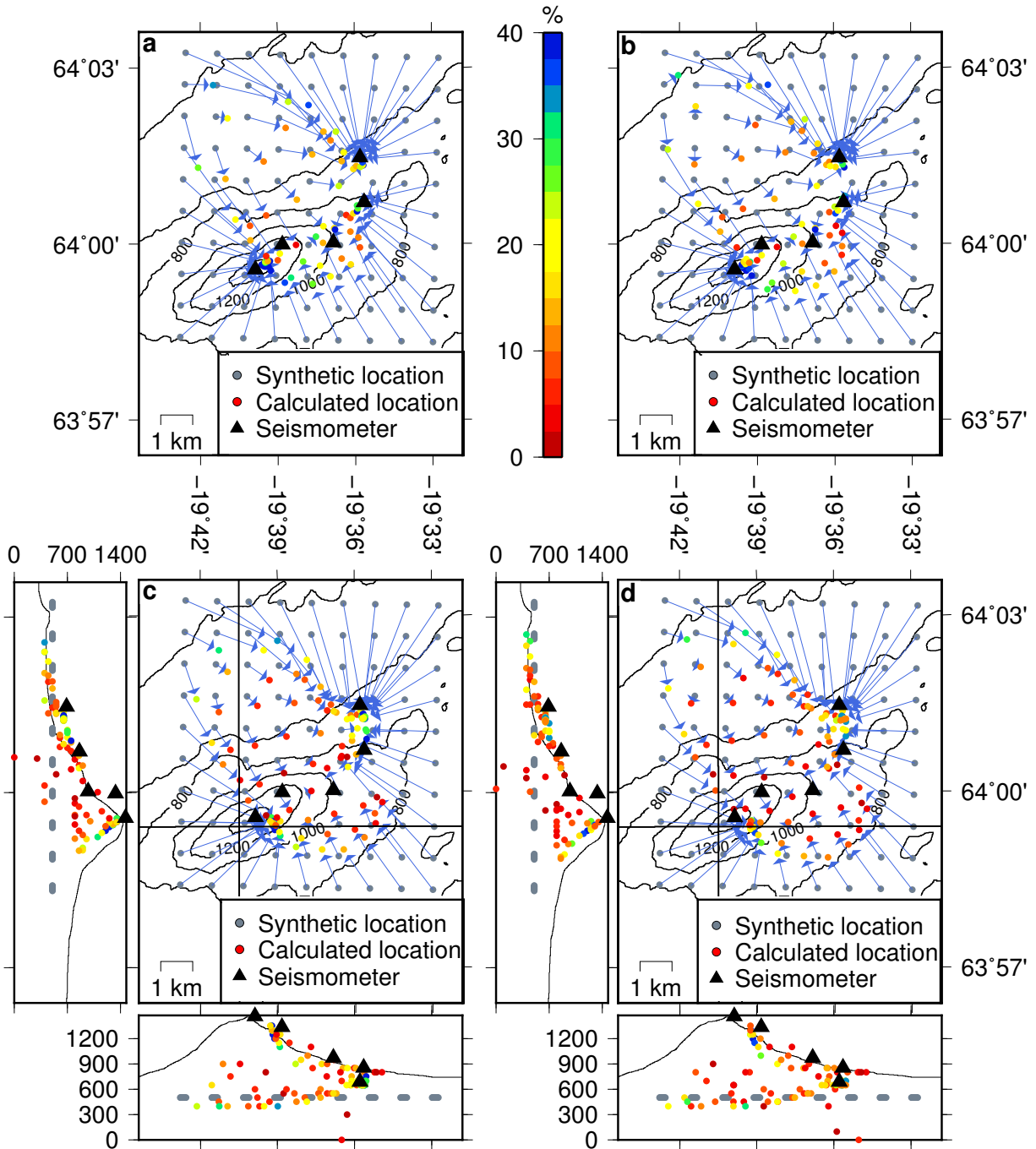


Figure 5: Synthetic tests of the intensity location method on a dataset created with a Gaussian pulse as source travelling as body waves. The plots show the locations of the original sources in grey and the locations of the best fitting source linked with a blue arrow. The best fitting sources are marked with a colored point according to their error percentage. The straight black lines indicate the location of the cross sections. The intensity location method was performed (a) in the 4 - 7 Hz band assuming surface waves, (b) in the 7 - 10 Hz band assuming surface waves, (c) in the 4 - 7 Hz band assuming body waves and (d) in the 7 - 10 Hz band assuming body waves.

252 The initially rectangular grid of sources at 500 m elevation was extending 2 to 3 km farther
 253 farther in each direction than the stations at the lowest/ highest latitude/ longitude. Our
 254 tests reveal that due to our station configuration (which were designed with a focus on
 255 event detection but not location) the general locations migrate systematically towards the

256 stations in the inversion. Grid points northeast and southwest of our stations move the
 257 most and cluster near HEK01 and HEK05, respectively. Grid points within or closer to
 258 our network migrate significantly less. Most locations stayed at approximately the same
 259 depth or migrated to a shallower location.
 260 It is important to note that although points move they remain on the initial side of the
 261 stations. That is, an event south of the stations is located south of it, an event north
 262 of the stations remains in the north during the location procedure. Exceptions include
 263 a location very close to HEK03 where a grid point slightly north of the station migrated
 264 slightly south and near HEK02 where an individual grid point in the south migrated to the
 265 north. High error percentages of points near HEK01 moreover seem to indicate locations
 266 of the real sources northeast of the network, which is where, in fact, they are located.
 267 We tested the quality factor and seismic velocity for recoverability. Quality factors be-
 268 tween 10 and 190 (stepsize 10) and seismic velocities between 0.5 km/s and 7 km/s
 269 (stepsize 500 m/s) were assumed for a location midway between HEK02 and HEK03 and
 270 one location outside the network 3 km north of HEK03. For the summit event between
 271 HEK01 and HEK03 the results were closest to the real location for $Q=90$ and $\beta=2.0$
 272 km/s for a varying quality factor and $Q=100$ and $\beta=2.0$ km/s for a varying velocity. For
 273 the location outside the network the best locations were $Q=100$ and $\beta=1.0$ km/s for a
 274 varying velocity and $Q=40$ and $\beta=2.0$ km/s for a varying quality factor. Those values are
 275 slightly lower than the parameters $Q=100$ and $\beta=2.0$ km/s that were used in the forward
 276 calculations.

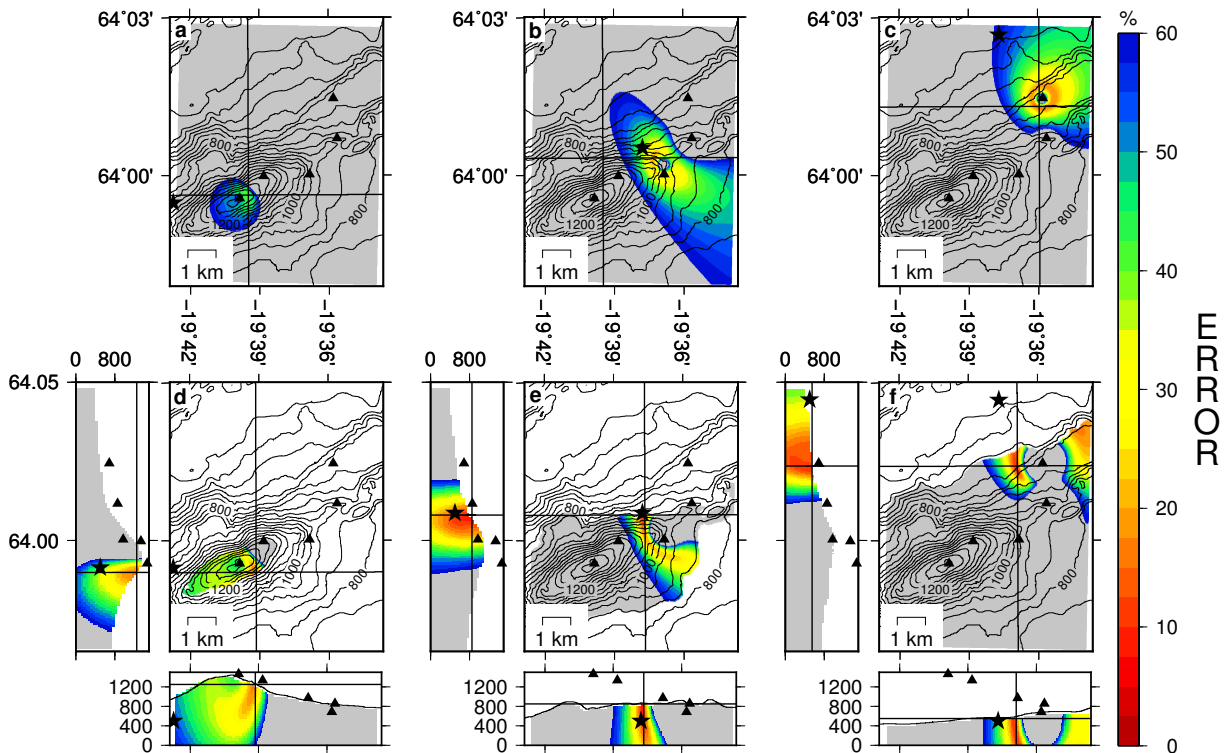


Figure 6: Typical error percentage distributions for three synthetic events. Grey points have errors higher than 60%. All figures show the results in the 7 - 10 Hz band as the figures in the 4 - 7 Hz band were nearly identical. The straight black lines indicate the location of the cross sections and also the location of the point with the lowest error percentage. The black star indicates the original location. The elongation of the error ellipse perpendicular to the line of stations is visible as well as its elongation in depth. (a-c) located assuming surface waves, (d-f) located assuming body waves.

277 Figure 6 gives an idea of the width and shape of the error in three perpendicular planes
278 through the location of the lowest error percentage. The three representative events were
279 located west of HEK05, northwest of HEK02 and north of HEK01. The grid point with
280 the best fit is marked by the black cross. The plots show the elongation of the lower
281 error percentages perpendicular to the linear trend formed by the deployed stations. The
282 uncertainty in the location is therefore highest in NW-SE direction and in depth. As
283 expected, events at the summit or in the vicinity of the network are well recovered.

284 **6 Location Results**

285 **6.1 Locations of the Type 1 Events**

286 The intensity location method was applied to 210 type 1 events for which the start and
287 end times were picked as demonstrated in figures 3a. The total time window was 2 to
288 8 s long. Locations were estimated in the 4 - 7 Hz band (figure 7) and 7 - 10 Hz band
289 (figure 9).

290 Most locations cluster around HEK05 and HEK03 which is consistent with the obser-
291 vations that the signals are strongest, shortest and arrive first at these stations. They
292 are also the most constrained events, located in the uppermost 400 m below the summit
293 (800 - 1200 m elevation) slightly north of the main ridge. This clustering of locations is
294 exclusively on the northern flank of the edifice. However, based on our synthetic tests we
295 are confident that this is not an artifact.

296 The locations near HEK01 have a high error percentage even for the best locations. This
297 might indicate that they occurred outside our station network and moved towards HEK01
298 as seen in the synthetic tests. The grey points aligning N-S or E-W in the northwestern
299 corner have very high error percentages, occurred outside the grid we set up and follow
300 the edge of the grid.

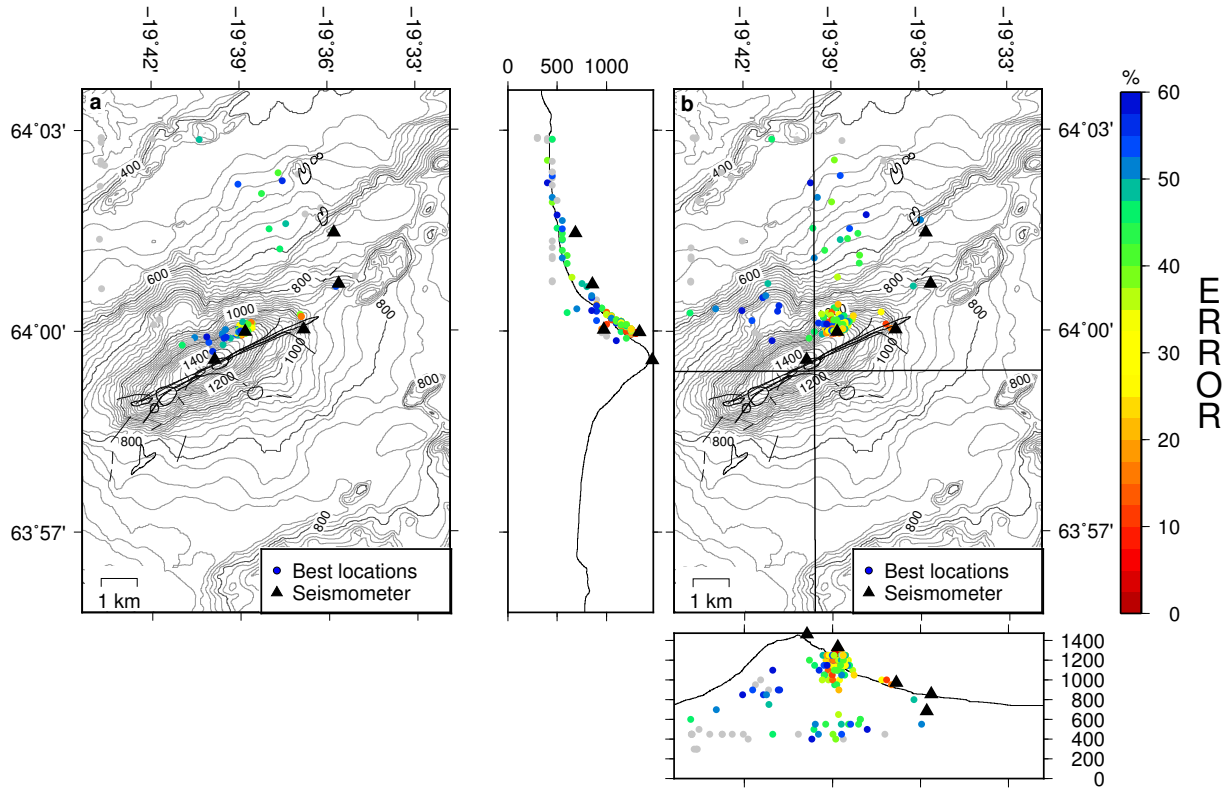


Figure 7: Best fitting locations of 210 type 1 events colored according to the error percentage at the best location. Grey points have errors higher than 60%. Locations are based on the intensity location method in the 4 - 7 Hz band. Each event is represented by one point. Thin black lines indicate the eruptive fissures of the eruptions in 1970, 1980/81, 1991 and 2000. The straight black lines indicate the location of the cross sections. The location method assumed (a) surface waves and (b) body waves.

301 The error percentage distribution is shown in three dimensions at the best fitting location
 302 in figure 8 for two representative events. One event was located at the summit of
 303 Hekla (figure 8a) the other one north of it (figure 8b). The uncertainty in depth is visible
 304 especially for the event north of the summit. The error ellipse is slightly elongated per-
 305 pendicular to the line of stations and quite broad northwest of the stations. These results
 306 are consistent with the synthetic tests in figure 6.

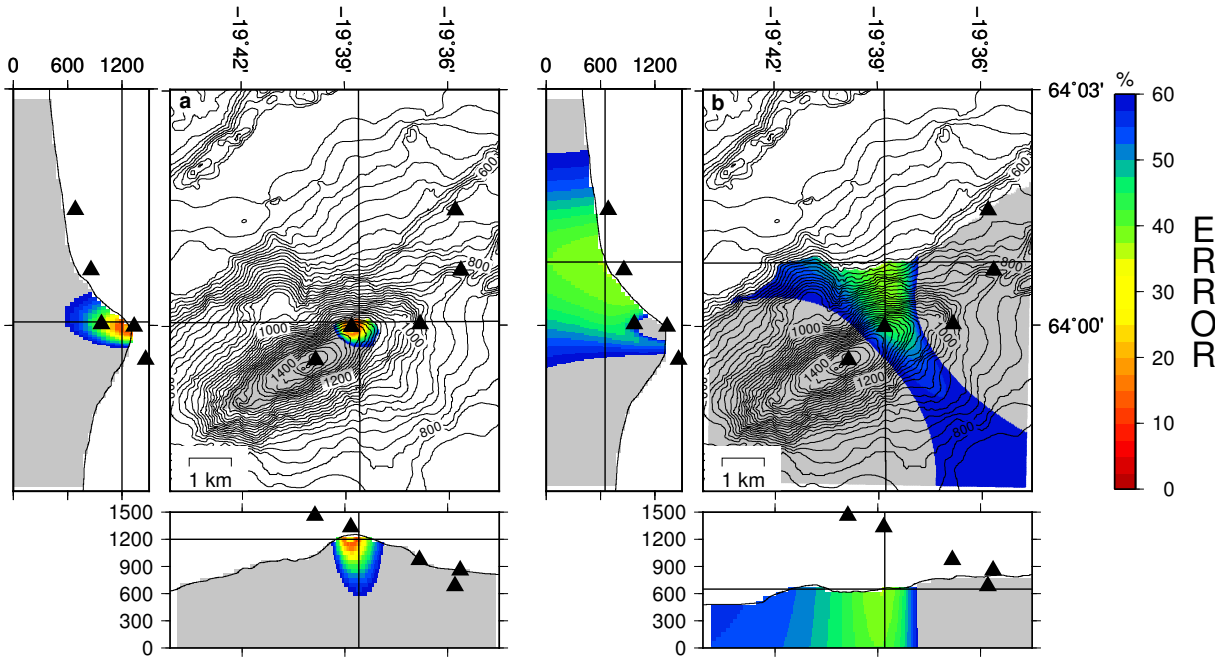


Figure 8: Error percentage distributions for two type 1 events located in the 4 - 7 Hz band assuming body waves. The straight black lines indicate the location of the cross sections as well as the location with the lowest error percentage. The events were located (a) near the summit and (b) northeast of the volcano.

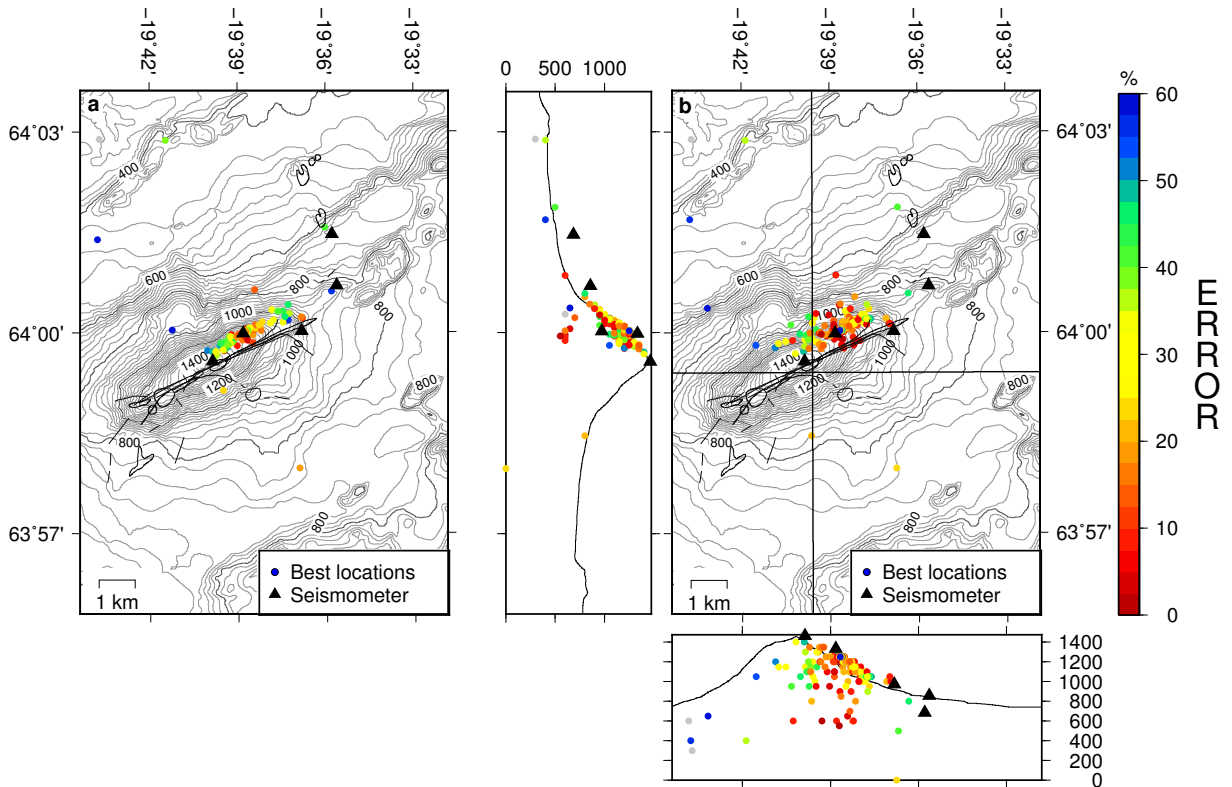


Figure 9: Same as figure 7 except in the 7 - 10 Hz frequency band.

307 In figure 9 we show the same analysis as in figure 7, except for 7 - 10 Hz frequency band.
 308 Here the locations have lower error percentages and are more clustered along the northern

309 flank, than the locations for the 4 - 7 Hz band. This is consistent with observations of
 310 Kumagai et al. (2010) that amplitude based location methods perform better for higher
 311 frequency data, where the wavefield is more isotropic through wave scattering.

312 6.2 Locations of the Type 2 Events

313 Forty type 2 events were located with overall window lengths of 12 to 25 s as indicated
 314 by the red lines in figure 4a. As the S/N ratio is better in the 4 - 7 Hz band (see figure 4)
 315 we only show the result in this band (figure 10).

316 The type 2 events are mostly located northwest of HEK01 beyond the flank of Hekla at
 317 400 to 600 m elevation. A few events were located near the summit and have a slightly
 318 higher error percentage. It is possible that these events are a mis-classification of type 1
 319 events. Error percentages are comparable to the type 1 events and are slightly lower in
 320 the 7 - 10 Hz band.

321 The type 2 events seem to occur outside our network and might therefore have high error
 322 percentages. A comparison to our synthetic tests shown in figure 5 suggests that they
 323 occur northeast of our stations and their apparent locations cluster near HEK01 due to
 324 our station geometry. They have nevertheless a different character than type 1 events and
 325 clearly have a different location, likely to the northeast of Hekla.

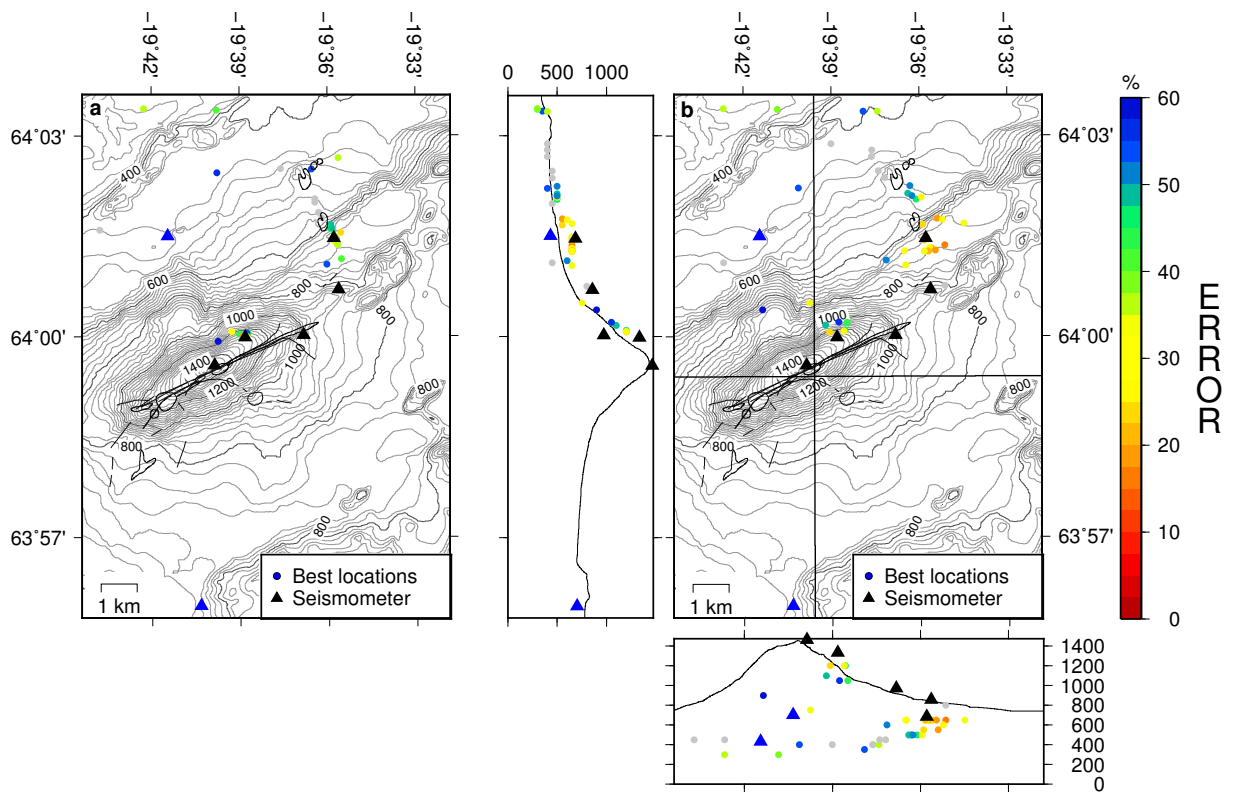


Figure 10: Same as figure 7 but for 40 type 2 events. Locations of the two closest temporary IMO stations marked by blue triangles.

326 A subset of 19 of these events was located using our five stations and the two additional
 327 temporary IMO stations (blue triangles in figure 10). Events in the east moved further
 328 eastwards away from the station network. Events north of the network moved towards the
 329 stations and westwards or stayed where they were. In fact the locations were consistent

330 with two arrival-time located type 2 events. See section 7 for the methodology. We
331 conclude that the IMO stations help to constrain locations and support our previous
332 suggestion that most of these events occurred outside the network. Although they also
333 suggest that some of them actually occurred west of the volcano in a greater distance.

334 **7 Arrival-Time Location Method**

335 For 23 type 1 events we picked P wave arrival times on all five stations and therefore
 336 were able to apply an arrival-time location method. The remaining 187 of the events
 337 were emergent or had high noise levels masking the onset. We expect P waves to arrive
 338 first and chose a P wave velocity of $2.0\text{km/s} \cdot \sqrt{3} \approx 3.4\text{ km/s}$ based on our shear wave
 339 velocity. Relative arrival-times T_{syn} from each grid point were compared with observed
 340 arrival-times T_{obs} for grid points below the topography. The error percentage RES was
 341 calculated and the minimum in the grid picked as best fitting location (Battaglia et al.,
 342 2005):

$$RES = 100 \cdot \sqrt{\frac{\sum_i (T_{syn} - T_{obs})^2}{\sum_i (T_{obs})^2}} \quad (7.1)$$

343 The locations from the arrival-time location method are shown in figure 11a. They are
 344 broadly consistent with the locations from the intensity location method for the same
 345 events (figure 11b). Using the arrival-time location method the locations are about 500
 346 m further east, slightly more scattered and were located a few hundred meters deeper.
 347 They also support the observation that the locations are north of the summit fissure on
 348 the northern flank of Hekla. The error percentage distribution in figure 12 clearly shows
 349 the error in the NW-SE direction and in depth.

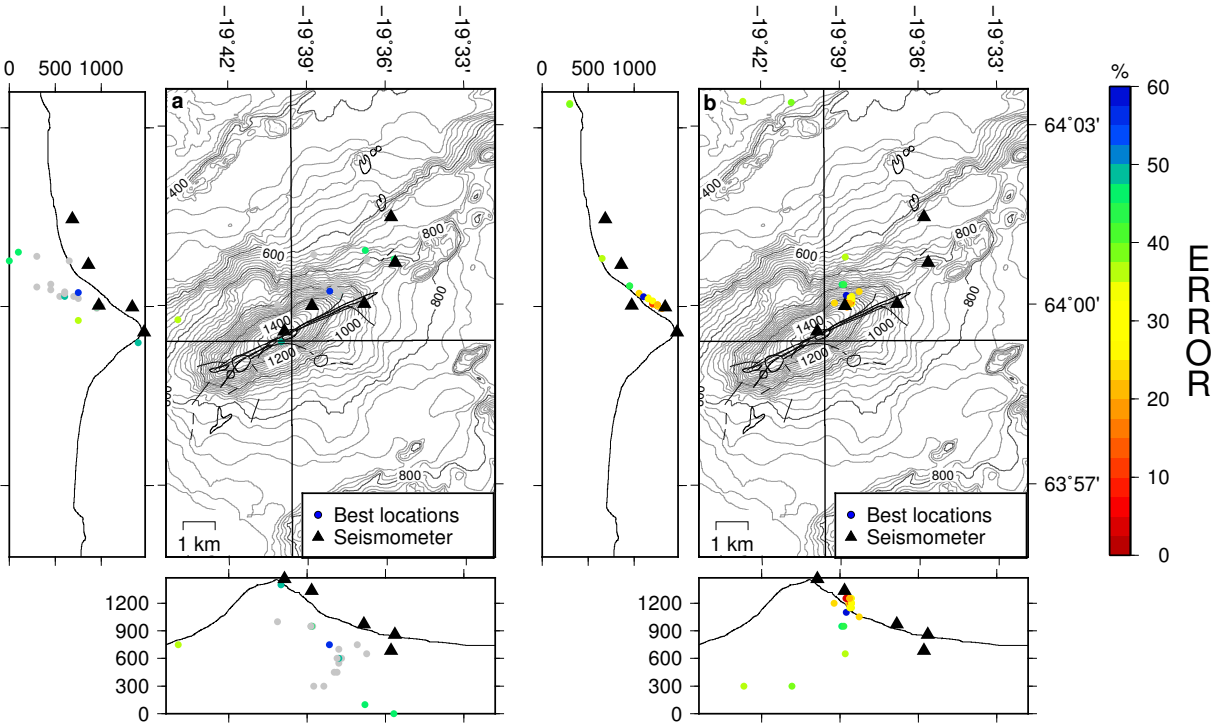


Figure 11: Arrival-time locations of 23 type 1 events (a) in comparison to the results from the intensity location method (4 - 7 Hz) (b). Note the higher error percentages and the locations north of the central fissure at Hekla for the arrival-time locations.

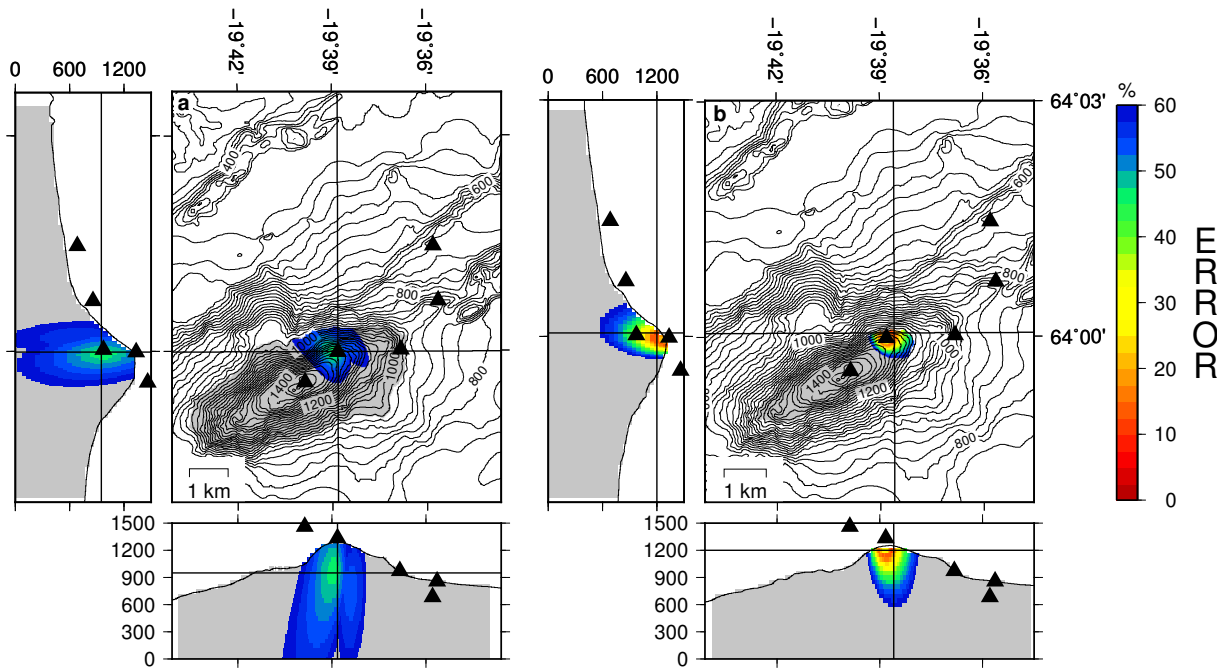


Figure 12: Typical error percentage distribution for a sample location calculated using arrival-times (a) and the corresponding error percentages for the same event located with the intensity location method (b). The straight black lines indicate the location of the cross sections as well as the location with the lowest error percentage.

350 8 Qualitative Estimates of Event Magnitudes

351 The regional events used to calculate site correction factors were also used to estimate the
 352 size of type 1 and type 2 events. 50 regional events were instrument and site corrected
 353 and filtered to the 4 - 7 and 7 - 10 Hz band. The maximum of the smoothed Hilbert
 354 Transform was used as maximum amplitude A_i at station i . The amplitude at the source
 355 A_0 was then calculated for all five stations based on (Battaglia and Aki, 2003):

$$A_{0i} = \frac{A_i \cdot r_i}{e^{-\frac{\pi \cdot f}{Q \cdot \beta} \cdot r_i}}, r_i = \sqrt{x^2 + y^2 + z^2} \quad (8.1)$$

356 We assumed $\beta = 2000$ m/s for body waves, $Q = 100$ and calculated the distance based
 357 on UTM coordinates of the Hekla stations and the IMO catalogue earthquake locations.
 358 A linear regression was then performed with the logarithm of the mean or median of the
 359 amplitudes at the source and the published magnitudes (dashed line, figure 13). The
 360 amplitudes at the source of type 1 and type 2 events were calculated similarly using the
 361 best fitting source location. These amplitudes were then converted to a magnitude using
 362 the regression line determined for the regional events (figure 13).

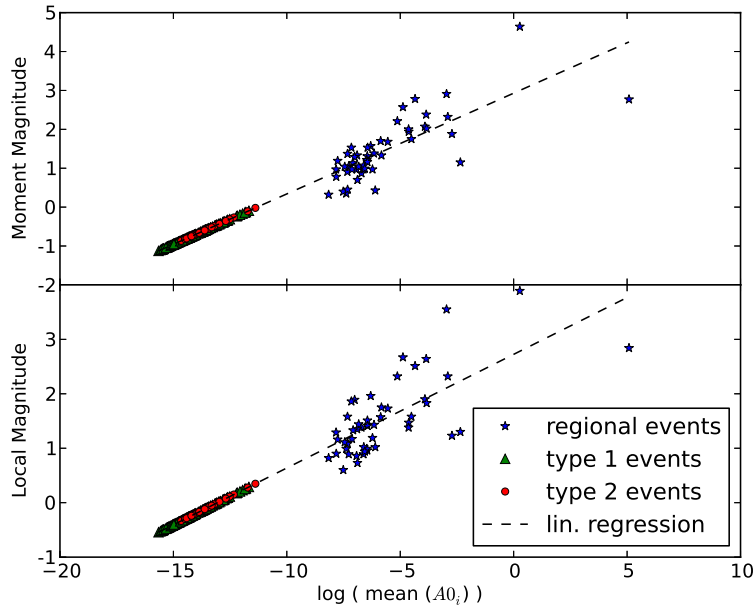


Figure 13: Moment and local magnitudes for type 1 and type 2 events were estimated based on a linear regression of the magnitudes of regional events and the mean amplitudes at the source derived from an amplitude location method.

363 The moment magnitudes of the regional events were 0.32 to 4.64, the local magnitudes
 364 0.6 to 3.89. As we assumed a straight wave propagation we underestimate A_0 for regional
 365 events which implies that the magnitudes of type 1 and type 2 events will be overestimated.
 366 A_0 for type 2 events might be underestimated as well if they actually lie outside our
 367 network. Despite the many assumption underlying this qualitative analysis it is clear
 368 that type 1 events are smaller than type 2 events. Estimated moment magnitudes are
 369 $M_W = -1.1$ to -0.1 for type 1 and $M_W = -0.9$ to -0.0 for type 2 and local magnitudes M_L

370 = -0.5 to +0.3 for type 1 and M_L = -0.3 to +0.3 for type 2 events. The influence of the
371 frequency band and the mean or median is insignificant.

372 9 Discussion and Conclusions

373 The type 1 events which were located near the summit of Hekla have significantly lower
374 error percentages at the best fitting location and cluster more than the type 2 events near
375 HEK01 off the flank. The amplitude based locations of 23 of those events chosen because
376 they have sharp onsets, are consistent with their arrival-time locations. According to
377 our synthetic tests some of the events clustering near HEK05 at the summit might have
378 occurred further south or further west. The locations near HEK01 have likely occurred
379 further north or further east. Thus, the cluster visible near HEK01 is most probably
380 an artificial cluster caused by the station geometry. Importantly our synthetic tests also
381 reveal that although a point might move towards the stations it still remains on the initial
382 side (either north or south of the station). This implies that all type 1 events likely occur
383 on the northern flank of the volcano, just north of the main eruptive WSW-ENE striking
384 fissure. This result is consistent with tectonic high-frequency events which were located
385 in the uppermost 0 to 4 km north of the 2000 eruptive fissure before the eruption in
386 2000 (see figure 5a in Soosalu et al. (2005)). A Synthetic Aperture Radar (SAR) study
387 revealed that shortly before this eruption the surface south of the eruptive fissure was
388 deformed upwards and north of it downwards. The SAR displacement was modelled with
389 a strike-slip fault reaching down to 5.8 km below sea level at a dip of 70-73° SE (Ofeigsson
390 et al., 2011). Another interesting feature is visible when comparing the gradient in fig-
391 ure 1b with the type 1 event locations near the summit in figure 7. The northern flank of
392 Hekla is slightly steeper than the southern flank and the locations cluster in the steepest
393 region north of HEK03 when body waves are assumed and in the steepest regions north
394 of HEK05 and HEK03 if surface waves are assumed. This suggests that edifice stability
395 might play a role in the generation of these events. Seismic signals will in this case be
396 created by minor failures on near surface edifice faults.

397 The type 2 events cluster mostly in a northwest-southeast striking line near HEK01. Based
398 on synthetic tests we expect that the clustering of events near HEK01 is not real and that
399 they are likely located farther to the NE. This is consistent with the high error percent-
400 ages, obtained for individual type 2 seismic events. However, the location of the cluster of
401 those events is consistent with the locations of a previous tectonic, high-frequency event
402 swarm. The swarm occurred in the uppermost 3 km in the first three months after the
403 eruption in 1991 that was thought to be linked to a dike intrusion in that region (see
404 figure 2 in Soosalu et al. (2005)). Between 1991 and 2000 some tectonic, low-frequency
405 events occurred in the same region. In Soosalu et al. (2005) figure 2 also shows faults
406 oriented in NNW-SSE and SW-NE direction near HEK01 that might be a possible source
407 of earthquakes.

408 Based on their frequency content and seismogram envelope shapes (diffuse-like), we in-
409 terpret type 2 events as tectonic (or volcano-tectonic) in nature, suffering strong path
410 effects. Although they are poorly located we are confident that they lie some distance
411 away, outside our network. As they propagate they will be affected by attenuation and
412 scattering effects which might hide a type 1 like event of slightly bigger magnitude. Type
413 1 events are clearly brittle-failure (volcano-tectonic like) in nature. Type 1 events locate
414 along a well defined structural trend. This trend mirrors the orientation of Hekla's 2000
415 eruptive fissure and lies about 200 m to 1 km NNW of its surface expression. This ob-
416 servation combined with their occurrence on the steepest portion of the northern flank of
417 Hekla would suggest that type 1 events are structurally controlled and related to ongoing
418 instability of the northern flank of the volcano.

419 Since we did not observe diurnal trends in the amount of seismicity we consider ice or

420 temperature changes an unlikely source. If the seismicity is related to magma such shal-
421 low location of magma would influence gas, GPS or INSAR measurements as well, none
422 of which was observed. Ongoing summit gas measurements during our seismic experi-
423 ment were undertaken every six hours for half an hour until September 6th (pers. comm.
424 Evgenia Ilyinskaya, April 2014) but changes were only slight and no correlation with the
425 number of seismic events per day was found.

426 Although we subdivided the events in two classes we would like to stress that they might
427 be the same type but of different size occurring in different locations. This seems to be
428 supported by our locations and magnitude estimations. In previous eruptions shallow
429 seismicity was detected whilst the magma was still at significant depth but rising towards
430 the surface (Sturkell et al., 2013; Soosalu et al., 2005). Monitoring near summit micro-
431 seismicity can help further constrain these observations. On Piton de la Fournaise shallow
432 microseismicity is shown to herald the location of future eruptive fissures (Barros et al.,
433 2013).

434 Our experiment detected a high level of shallow background seismicity primarily on the
435 northern flank near the summit. This seismicity is not detected by the permanent stations
436 of the IMO network as it is below their detection threshold. We demonstrate that perma-
437 nent stations closer to Hekla could improve the detection threshold of ongoing earthquakes
438 on the volcano. The high levels of background microseismicity at Hekla suggests that the
439 edifice is likely in a state of critical instability. If so, microseismicity levels will be very
440 sensitive to small future stress perturbations associated with magma migration at depth.
441 We suggest that continuous near-summit monitoring of microseismicity levels might offer
442 an earlier indication of imminent eruptions. Currently Hekla switches from apparently
443 quiescent to fully eruptive on the order of only one hour.

444 10 Acknowledgements

445 The data were collected and analysed within the framework of FutureVolc, which has re-
446 ceived funding from the European Union’s Seventh Programme for research, technological
447 development and demonstration under grant agreement No 308377. We thank Baldur
448 Bergsson and Thomas Eyre for support in the field. We are grateful to Ivan Lokmer,
449 Benedikt G. Ofeigsson, Evgenia Ilyinskaya, Páll Einarsson and Jean Battaglia for help-
450 ful discussions and comments. We also want to thank Heidi Soosalu and an anonymous
451 reviewer for their critical comments that helped to improve the manuscript.

452 **11 Appendix**

453 Figure 14 shows the locations using the intensity location method assuming a lower quality
454 factor $Q=50$ (Battaglia and Aki, 2003) and seismic velocity $\beta=1.2$ km/s. The latter value
455 is based on a seismic refraction study that found low P wave velocities in a refraction
456 profile north of Hekla (Pálmason, 1971). The influence on the locations is small although
457 the locations scatter more than for higher values (see figure 7b, 9b and 10b). Hence we
458 conclude that for our network geometry the broad locations of the events (on the northern
459 flank of the volcano) are insensitive to the details of velocity and Q structure.

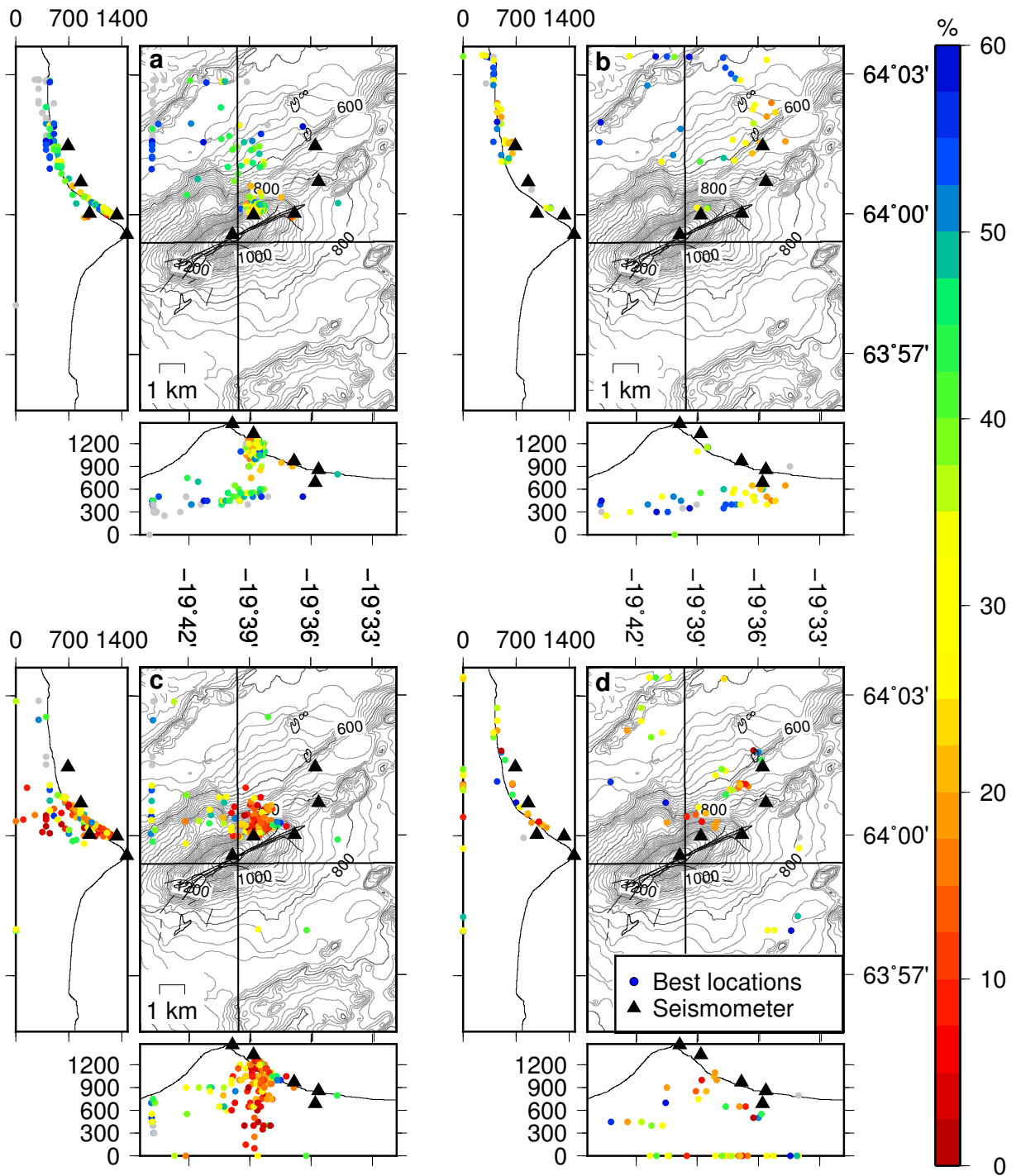


Figure 14: Intensity ratio locations assuming a $Q=50$ and $\beta=1.2$ km/s (a) in the 4 - 7 Hz band for type 1 events (4 - 7 Hz) (b) same as a but for type 2 events (c) same as a but in the 7 - 10 Hz frequency band (d) same as b but in the 7 - 10 Hz frequency band. Note the similarity to the locations with the higher quality factor and velocities.

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