RAUNVÍSINDASTOFNUN HÁSKÓLANS

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Science Institute - University of Iceland
Dunhaga 3 - Reykjavík

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September 1977 deflation of the Krafla
central volcano in NE-Iceland**

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Abstract

The September 1977 deflation event in the Krafla caldera was one of a series of such events that has been in progress since December 1975. The operation of portable seismographs in the active region and favorable location of the main seismic activity with respect to the permanent seismograph network in NE-Iceland allow a more detailed study of this deflation event than most of the other events. Continuous volcanic tremor appeared on the local seismographs shortly before 16 h on September 8, 1977. Deflation of the volcano began at the same time. A small basaltic eruption broke out on a 0.9 km long fissure near the northern rim of the caldera at about 18 h. Earthquake activity increased soon after the beginning of the tremor and the first earthquakes were located in the caldera region. The earthquake activity then migrated southwards along the Krafla fault swarm with a speed of about 0.5 m sec^{-1} , and culminated shortly before midnight with 8 earthquakes larger than magnitude 3 that were located near the Námafjall geothermal area 8 km south of the center of the caldera. Shortly after the earthquake activity migrated into the Námafjall area small amounts of basaltic pumice were erupted through a 1138 m deep drill hole there. Depths of earthquakes were 0-6 km in the northern part of the hypocentral zone and 0-4 in the southern part. The first motion pattern of P-waves suggests dip-slip faulting on steeply dipping fault planes consistent with the extensive normal faulting observed on the surface

throughout the epicentral zone. The magnitude-frequency relationship was nonlinear and changed during the earthquake sequence. The seismological data strongly support the interpretation that deflation of the Krafla volcano is associated with horizontal migration of magma from the caldera region and formation of dykes in the Krafla fault swarm.

Introduction

The volcanic rift zone in NE-Iceland has been the scene of considerable tectonic and volcanic activity since 1975. This activity may provide the first well observed demonstration of a process that appears to play an important part in Icelandic tectonism. The structure of the rift zone in this part of Iceland is characterized by NNE trending fault swarms that are arranged en echelon within the N-trending zone (Fig. 1). Open fissures, normal faults and eruptive fissures are the most common types of faults found in these swarms. Each fault swarm runs through a central volcano (Saemundsson, 1974). Some of these volcanoes, such as the Krafla volcano, have developed calderas.

The present activity has so far been mostly confined to the Krafla volcano and the associated Krafla fault swarm. The volcano has been in a state of repeated inflation since at least early 1976. The inflation can be interpreted as the result of magma accumulating at the depth of about 3 km beneath the center of the volcano (Björnsson et al., 1978, Tryggvason, 1978 a). The inflation is interrupted by relatively short deflation events when the central region of the caldera may subside as much as 2.5 m. The onset of deflation is followed by earthquakes and rifting in the fault swarm. The earthquake activity starts within the caldera and then propagates along the fault swarm. This has been interpreted as being caused by horizontal magma migration away from the area of magma accumulation under the volcano.

The deflation events have many seismological similarities even though they are not all of the same size, and different sections of the fault swarm are affected each time. The

Fig. 1. Map of the volcanic rift zone in NE-Iceland with its fault swarms and central volcanic complexes. The locations of permanent seismograph stations used in this study are marked with dots. The inset shows the location of the map within the volcanic rift zone of Iceland and the Tjörnes Fracture Zone (TFZ). Adapted from a map by Kristján Saemundsson in Björnsson et al., (1978).

deflation event that started on September 8, 1977 was in many respects typical of these events. The epicentral area of the associated earthquakes was favourably located with respect to the permanent seismograph network, and 4 portable seismographs could be placed in the area within one and a half days. The seismic activity of this event could therefore be studied in more detail than that of most other events.

The data presented in this paper strongly support the interpretation of Björnsson et al. (1977) that deflation of the Krafla volcano and the current rifting of the plate boundary in NE-Iceland are associated with horizontal migration of magma from the volcano along the Krafla fault swarm.

The course of events

After the deflation of April 27-29, 1977, the Krafla volcano inflated at the usual rate of 5-7 mm per day near the center of the caldera (Björnsson et al., 1978). A new deflation event began on September 8 with volcanic tremor and a dense earthquake swarm. The tremor became visible on the most sensitive seismograph (GD) at 1547. At about the same time a recording tiltmeter within the caldera showed that deflation had started (Tryggvason, 1978 a). The tremor amplitude increased rapidly the first 15 minutes and then more slowly for an additional hour. Earthquakes accompanied the tremor, but they were small to begin with and difficult to locate because of the high tremor amplitude. They were clearly located in the central part of the caldera, however.

The deflation and the tremor were observed by the local people, and within half an hour a warning of impending volcano danger was issued in the Mývatn district.

An eruption of highly fluid basaltic lava started on a fissure near the northern rim of the caldera at 18 h. (Sigurdsson, 1978). The fissure was 0.9 km long and had a northerly trend. The outbreak of lava did not have a significant effect on the amplitude of the volcanic tremor,

Fig. 2. The fault system of the Krafla area, lavas and seismograph stations. HF, HD, HK, BF and LU were temporary stations. The Námafjall geothermal area is located 1-2 km SE of station BF. Roads are shown for coordination. Adapted from a map by Kristján Saemundsson in Björnsson et. al., (1978).

Fig. 3. The time sequence of the earthquakes. The magnitude determined from the seismograms of AKU, is plotted as a function of time. The data set is incomplete below magnitude 2.3, especially in the first part of the sequence.

which remained fairly constant in the time interval 17-19 h. The eruption was continuous until 20 h, but then its vigor began decreasing and by 2230 the lava production had ceased (Larsen et al., 1978). The area covered by the lava is estimated as 0.8 km^2 and the volume $2 \times 10^6 \text{ m}^3$ (Björnsson et al., 1978).

The tremor amplitude and the earthquake activity increased shortly after 19 h. Towards the evening it became clear that the earthquakes were propagating to the south, i.e. towards the inhabited area near lake Mývatn. The southern part of the fault swarm was put under observation in case an eruption would break out there. Faults in the Nāmafjall area (Fig. 2) began moving at 2240 and 2345 red flashes were seen emanating from the geothermal area there (Larsen et al., 1978). Inspection revealed that glowing pumice was being thrown out of a producing steam well. The total amount of basaltic pumice that came up through the 1138 m deep drill hole has been estimated 2500 kg (Björnsson and Sigurdsson, 1978) to 3000 kg (Larsen et al., 1978). The seismic activity culminated shortly before midnight with 8 earthquakes of magnitude 3 and larger (Fig. 3). After that there was a steady decrease in the earthquake activity, the tremor amplitude and in the deflation rate. Inflation of the caldera region resumed on September 9, and thus a new inflation-deflation cycle started.

Instruments and data

The seismograph station AKU, a member of the World Wide Standardized Seismograph Network, is located at the distance of 60 km from the Krafla area. The magnitudes shown in Fig. 3 were determined from the records at AKU and were kindly supplied by Þórunn Skaftadóttir of the Icelandic Meteorological Office.

A relatively dense network of short period seismographs was established in NE-Iceland in 1974-75, i.e. shortly before the present tectonic activity began. Three seismographs are located within 10 km of Krafla and additional three instruments are within 50 km distance. These stations are a part of a network that covers the active zones of Iceland and is operated by the Science Institute, the National Energy Authority and the Icelandic Meteorologic Office. In addition

to the permanent stations four portable seismographs were put into operation on September 9 and 10. For a part of the earthquake swarm, therefore, 10 seismographs were recording within 50 km of the epicentral area. More distant stations were also used when possible.

The short period seismographs are most sensitive in the frequency band 3-30 Hz, and record with pen and ink on paper, except the portable instruments that record on smoked paper. Radio time is put directly on the record. Time corrections are thus eliminated and timing accuracy is better than 0.1 sec. To keep the time signals visible during periods of high volcanic tremor it is sometimes necessary to turn down the magnification of the seismograph by a factor of 32.

The computer program HYPOELLIPSE (Ward and Lahr, 1977) is used for the location of the hypocenters. The input data to the program are the P- and S-wave arrival times, the crustal structure and station corrections. The location procedure is described in more details by Einarsson (1978 a) and Brandsdóttir (1978).

The epicentral area of the September 1977 swarm was in a close proximity to the most dense part of the permanent seismograph network. It is therefore possible to locate this swarm with higher accuracy than most of the other swarms.

The hypocentral zone

The epicentral area as delineated by the most accurately located epicenters is shown in Fig. 4. Open circles denote epicenters located with standard error of 1 km or less in all directions, epicenters marked with closed circles have standard error of 0.5 km or less. The main epicentral zone is about 9 km long and less than 2 km wide. It extends from the center of the caldera southwards along the central part of the fault swarm to the Námafjall geothermal area. All the earthquakes of magnitude 3 and larger were

Fig. 4. The epicentral zone as delineated by the most accurately located epicenters. Circles mark the epicenters located with standard error of 0.5-1.0 km in all directions. Dots are epicenters located with standard error less than 0.5 km. Roads and the caldera fault are shown for coordination.

Fig. 5. A vertical section along the hypocentral zone. All hypocenters are projected on a vertical plane through the points 65°44'N, 16°48'W and 65°37'N, 16°52'W. Symbols as in Fig. 4. The vertical bar shows the drillhole in the Námafjall geothermal area where the small pumice eruption took place.

Fig. 6. A vertical cross section of the northern half of the hypocentral zone. All hypocenters in the northern half are projected on a vertical plane through the points $65^{\circ}41.91'N$, $16^{\circ}53.88'W$ and $65^{\circ}41.0'N$, $16^{\circ}45.0'W$. Symbols as in Fig. 4.

Fig. 7. A vertical cross section of the southern half of the hypocentral zone. All hypocenters in the southern half are projected on a vertical plane through the points $65^{\circ}38.92'N$, $16^{\circ}55.58'W$ and $65^{\circ}38.04'N$, $16^{\circ}46.74'W$. Symbols as in Fig. 4. The Námajall drill-hole is shown.

Fig. 8. The propagation of the earthquake activity. The latitude of the epicenter is plotted as a function of time of occurrence. The latitude and time of the volcanic eruptions is also shown. Epicenters with location errors smaller than 2 km are included. Most of them have location errors smaller than 1 km. The triangles mark the earthquakes of magnitude 3 and larger. The activity begins within the caldera near 65°43'N and propagates southwards along the Krafla fault swarm to 65°38' with a speed of about 0.5 m sec⁻¹.

located in the southernmost 2 km of the epicentral zone.

The hypocenters are plotted on a vertical plane striking along the earthquake zone in Fig. 5. The depth of the earthquakes is between 0 and 6 km with a concentration of hypocenters between 3 and 4 km depth. In the northern part of the zone the hypocenters reach a depth of about 6 km but near the southern end they only go down to 4 km. This is further demonstrated in the cross sections of the northern part (Fig. 6) and the southern part (Fig. 7). The hypocentral zone may be described as a vertical wall, less than 2 km wide, that extends from the surface down to about 6 km near its northern end and down to 4 km near its southern end.

The earthquake activity started at the northern end of the epicentral zone and then migrated southwards along the fault swarm (Fig. 8) until most of the epicentral zone was active. The migration speed was highest during the hours before 22 h but then it slowed down considerably. The average speed during the first few hours was about 0.5 m sec^{-1} .

Focal mechanisms and faulting

An attempt was made to determine focal mechanism solutions from the first motions of P-waves. This can only be achieved when a sufficient number of local seismograph stations records the same event with a clear first motion. The maximum number of local seismographs available for this study was 7, which is not enough for single event focal mechanism solutions. Therefore only composite solutions were attempted. The focal mechanism study depends critically on the portable seismographs that were in operation after September 10, i.e. after the main earthquake activity was over and inflation had started again.

The most accurately located earthquakes were selected for this study and grouped according to the first motion pattern. Then it was attempted to combine the events of each group for a composite focal mechanism solution. This attempt was not very successful, the focal mechanisms of the earthquakes were apparently quite variable. Out of

Fig. 9. Composite focal mechanism solutions. The first motions of P-waves are plotted on a stereographic projection of the upper hemisphere. Dots are compressive and circles are dilatational first motions. First motions of nodal character are marked with an X. The preferred solutions are shown with continuous lines. Broken lines are alternative solutions.

55 events originally selected only 10 could be combined for the two kinds of solutions shown in Fig. 9 A and B. The preferred solutions are shown with continuous lines, alternative solutions are shown with broken lines. All the solutions have a steeply dipping nodal plane with a strike slightly east of north, which is also the strike of a large majority of the surface faults. This plane is therefore interpreted as the fault plane. In plot A the preferred solution has a fault plane with strike 5° and dip 84° to the west. Normal faulting with nearly vertical slip vector is indicated. The alternative solution has a fault plane with a strike 5° and dip 70° to the east. The other nodal plane cannot be determined uniquely, but reverse faulting with a variable component of right-lateral strike-slip movement is indicated. In plot B the preferred solution has a fault plane striking 10° and dipping 70° to the east. Nodal and inconsistent readings in the NW quadrant indicate that the other nodal plane has a strike close to 130° , and normal faulting with a component of left-lateral strike-slip is indicated. The alternative solution has a fault plane striking 10° and dipping 80° to the west. The other nodal plane has nearly the same strike and indicates primarily reverse faulting. In summary, the available focal mechanism solutions suggest dip-slip motion on steeply dipping faults. The faulting is normal or reverse depending on whether an easterly or westerly dip is assumed.

Extensive surface faulting occurred on September 8 and 9 in the fault swarm. The faulted area was about 3 km wide and 16 km long, and extended from the eruption fissure near the northern rim of the caldera to a few km south of the Námafjall area. The main types of faulting were normal faulting and fissuring. A component of left-lateral strike-slip motion could be seen in a few places. The movement was distributed on many parallel faults, sometimes several tens of centimeters and up to a meter on each fault. The structure of the faulting is a graben structure with the flanks uplifted (Björnsson et al., 1978, Tryggvason 1978, a).

A majority of the earthquakes had a P-wave first motion pattern consistent with normal faulting, even though a focal mechanism could not be determined. Other events were clearly inconsistent with normal faulting. The focal mechanisms were more variable than one would expect from the observations of the surface faulting.

The magnitude-frequency relationship.

The magnitude and frequency of events of an earthquake sequence commonly follow the law of Gutenberg and Richter:

$$\log N = A - bM$$

In this formula N is the number of earthquakes of magnitude M and greater, A and b are constants. Laboratory studies of microfracturing, theoretical considerations and earthquake data suggest that the value of the constant b is related to the effective stress in the source region (Scholz, 1968, Wyss, 1973). The world average of the b -value is close to 1 (Richter, 1958, Båth, 1973) but significantly higher values are sometimes found associated with volcanism (e.g. Minakami, 1960, Einarsson and Björnsson, 1976) and the spreading segments of the mid-Atlantic ridge (Francis, 1968). Temporal variations of the b -value have also been found (e.g. Utsu, 1971, Gibowicz, 1973 a and b) and interpreted as the result of changing effective stress.

By inspection of Fig. 3 two characteristics of the magnitude-frequency relationship of the September 1977 earthquake sequence stand out: The b -value must be high, and the largest earthquakes between 21 h and 23 h on September 8 appear to be exceptionally large. These observations are supported by Fig. 10 where the magnitude-frequency distribution of different parts of the sequence are plotted. Plot 10a shows the first 50 earthquakes of the sequence. Here the detection threshold is relatively high, which is partly caused by the high density of earthquakes on the seismogram. The difference between the maximum magnitude and the detection threshold

Fig. 10. The frequency-magnitude distribution of the earthquakes. M is the magnitude determined from the amplitude at the seismic station AKU, N is the number of earthquakes of magnitude M and larger.

a: The first 50 earthquakes. b: The first 100 earthquakes. c: Earthquakes no. 50-150. d: Earthquakes no. 100-200. e: Earthquakes no. 150-229. f: The whole earthquake sequence.

is too small for a statistically meaningful estimate of the b-value, but a very high b-value is indicated. Plot 10b shows the first 100 events of the sequence and includes the large earthquakes that occurred between 21 h and 23 h. The distribution has become highly non-linear and a b-value calculation is not meaningful. At lower magnitudes the distribution is similar to that shown in plot 10a, but at higher magnitudes the curve flattens out. Similar distribution appears in plot 10c, where events no. 50-150 are included. Plots 10d and 10e show the distribution of events no. 100-200 and 150-229 respectively. Here the large events are not included and the magnitude distribution is reasonably linear. Maximum likelihood estimates of the b-values and the 95% confidence limits are calculated according to the method of Page (1968). The b-values are 3.1 ± 0.9 and 2.7 ± 0.9 respectively and are not significantly different from each other. The magnitude distribution of the whole sequence is shown in Fig. 10f. As in Fig. 10b and 10c the distribution is non-linear.

The non-linearity of the magnitude distribution may be interpreted as being caused by the superposition of two earthquake sequences with different b-values. This interpretation is possible because the distribution curve is concave upwards. Convex curves cannot be interpreted in this way. The majority of the earthquakes belonged to an earthquake sequence with a high b-value, or about 3, indicating low effective stress. At about 21 h high stress had been built up near the end of the epicentral zone, which was released by an earthquake sequence with a low b-value, or about 0.6. Later events were primarily associated with adjustments within a low stress region.

Discussion

There is little doubt that the inflation and deflation of the Krafla area is associated with movements of magma. Gravity changes combined with the vertical movements require the transfer of material at least as dense as magma (Björnsson et al., 1978). The center of inflation

and deflation is located in the central part of the Krafla central volcanic complex and three of the deflation events so far have been accompanied by small basaltic lava eruptions. The close resemblance of the Krafla deformation with that observed at Kilauea Volcano in Hawaii (see e.g. Björnsson et al., 1978, Tryggvason, 1978 a and b, Kinoshita et al., 1974), the occurrence of S-wave shadows in the Krafla area (Einarsson, 1978 a) and the volcanic tremor that accompanies the deflation further strengthen the interpretation.

The deflation events are accompanied by earthquake activity that starts near the center of deflation and then migrates horizontally away from it. It is reasonable to assume that the earthquake hypocenters mark the pathway of the magma. If this assumption is accepted one can regard the hypocentral zone as shown in Figs. 4, 5, 6 and 7 as an approximate indicator of the shape and extent of the intrusion. The intrusion thus delineated is a horizontally fed, vertical dyke. The depth to the bottom of the dyke is about 6 km in the caldera, but only 4 km near its southern end. The depth to the top edge of the dyke is more difficult to ascertain, because some of the shallow seismicity is undoubtedly related to the extensive faulting above the dyke. The drillhole eruption demonstrates that the dyke reached a depth less than 1138 m near its southern end, and in the northern part of the caldera the dyke reached the surface in the fissure eruption. The high seismicity at the depth of 3-4 km suggests that the dyke has a maximum thickness there.

In the light of the extensive surface faulting and rifting of more than one meter (Björnsson et al., 1978, Tryggvason, 1978b) the earthquake activity on September 8-9 was surprisingly small. The largest earthquakes were of magnitude 3.8, and were not locatable by seismograph networks outside of Iceland. Even more surprising is the lack of earthquake activity in the northern part of the caldera where the fissure eruption took place. Normal faulting of well over 1 m was observed locally near the eruption fissure, but no earthquake larger than

Fig. 11. Vertical section along the Krafla fault swarm, showing the eruptive fissure, the S-wave attenuation zone, the Námafjall drillhole and center of magma accumulation as determined from elevation and tilt changes. The propagation of the seismic activity along the fault swarm is shown with lines. At the time indicated all the seismic activity was north of the line. The final extent of the seismic zone is also shown.

magnitude 2.5 could be attributed to this area. Small earthquakes during the first two hours of deflation could have originated in this part of the fault zone, but they could not be located because of the high tremor amplitude. A dyke intrusion appears to be a fairly peaceful process in terms of seismic activity.

The spatial relationship between the events of September 8-9 is shown in a section in Fig. 11. The center of inflation and deflation as determined by elevation changes (Björnsson et al., 1978) and by tilt changes (Tryggvason, 1978 a) is shown with an arrow at about 3 km depth. S-waves are attenuated in the same general area (Einarsson, 1978 a). The top of the attenuating body is near 3 km depth and the bottom is shallower than 7 km, but is poorly constrained. The eruptive fissure is shown to the north of the inflation center. If one assumes that the magma that was erupted through the fissure started ascending from the depth of 3 km shortly before 16 h when the volcanic tremor started, the average speed of ascend is 0.5 m sec^{-1} .

The dyke then propagated to the south with the same speed of 0.5 m sec^{-1} as indicated by the earthquake activity. The position of the earthquake front at different times is shown in Fig. 11.

The high b-value of the main part of the earthquake sequence suggests that the earthquakes were caused by failure of the crust at low effective stress. The effective stress may be expected to be low near an intruding fluid body. Near the end of the dyke, however, a high effective stress region may develop where the magma does not reach. The large earthquakes between 21 h and 23 h occurred when the dyke had nearly reached its full length and they were probably associated with relatively high effective stress.

According to the tilt record (Tryggvason, 1978 a) only about two thirds of the final volume had flowed out of the magma storage under Krafla at the time of the large earthquakes. After that the hypocentral zone did not add significantly to its length, but was mainly extended to greater depth. It is therefore possible that the dyke

propagated downwards as well as horizontally.

The length of the dyke is probably controlled by a number of factors such as the pressure in the magma chamber, the tectonic stress, the viscosity of the magma and the width of the dyke. It is also possible that transverse tectonic structures effectively block the horizontal flow. The first deflation event that started on December 20, 1975, for example, was accompanied by seismic activity on the Krafla fault swarm and the Grímsey fault that joins it off shore (Björnsson et al., 1977). A focal mechanism solution of the largest earthquake of that sequence was consistent with transform faulting along the Grímsey fault (Einarsson, 1978 b). It therefore seems likely that transverse faults could similarly affect the length of other intrusions. Fresh ground cracks with NW orientation were seen in the road in the Námafjall area on September 9, and a prominent transverse lineament can be seen on aerial photographs in the Námafjall area near the end of the September 1978 intrusion, but this lineament cannot yet be identified as a fault.

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References

- Båth, M., 1973. Introduction to Seismology.
Birkhäuser Verlag, Basel and Stuttgart. 395 pp.
- Björnsson, A., Saemundsson, K., Einarsson, P., Tryggvason, E., and Grönvold, K., 1977. Current rifting episode in north Iceland,
Nature, 266, 318-323.
- Björnsson, A., and Sigurdsson, O., 1978. Hraungos úr borholu í Bjarnarflagi.
Náttúrufræðingurinn, 48, 19-23 (in Icelandic with English abstract).
- Björnsson, A., Johnsen, G., Sigurdsson, S., Thorbergsson, G., and Tryggvason, E., 1978. Rifting of the plate boundary in north Iceland,
in the press.
- Brandsdóttir, B., 1978. Earthquakes associated with the deflation of the Krafla volcano on September 8, 1977 (in Icelandic).
University of Iceland, thesis.
- Einarsson, P., 1978. S-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust.
Bull. Volc., in the press.
- Einarsson, P., 1978. Seismicity and earthquake focal mechanisms along the mid-Atlantic plate boundary between Iceland and the Azores.
Tectonophysics, in the press.
- Einarsson, P., and Björnsson, S., 1976. Seismic activity associated with the 1970 eruption of volcano Hekla in Iceland,
Jökull, 26, 8-19.
- Francis, T.J.G., 1968. The detailed seismicity of mid-ocean ridges.
Earth Planet. Sci. Let. 4, 39-46.
- Gibowicz, S.J., 1973. Variation of frequency-magnitude relationship during Taupo earthquake swarm of 1964-65.
N.Z. J. of Geol. Geophys, 16, 18.

- Gibowicz, S.J., 1973. Variation of the frequency-magnitude relation during the 1931 Hawkes Bay, 1934 Pahiatna, and 1943 Wairarapa aftershock sequences.
N.Z. J. of Geol. Geophys., 16, 1009-1045.
- Kinoshita, W.T., Swanson, D.A., and Jackson, D.B., 1974. The measurement of crustal deformation related to volcanic activity at Kilauea volcano, Hawaii.
Phys. Volcanology, ed. L. Civetta, P. Gasparini, G. Lnongo and A. Rapolla
Elsevier Publ. Co, 333 pp.
- Lahr, J.C., and Ward, P.L., 1977. HYPOELLIPSE: A computer program for determining local hypocentral parameters, magnitude, and first motion pattern,
U.S. Geol. Surv. Open-File Report.
- Larsen, G., Grönvold, K., and Thorarinsson, S., 1978. Volcanic eruption through a geothermal borehole at Námafjall, Iceland.
Nordic Volcanological Institute, Report 78 12
Science Institute 78-10, 18 pp.
- Minakami, T., 1960. Fundamental research for predicting volcanic eruptions, Part 1.
Bull. Earthqu. Res. Inst. 38, 497.
- Page, R., 1968. Aftershocks and microaftershocks of the great Alaska earthquake of 1964.
Bull. Seismol. Soc. Am., 58, 1131-1168.
- Richter, C.F., 1958. Elementary Seismology.
Freeman & Co., San Francisco and London, 768 pp.
- Saemundsson, K., 1974. Evolution of the axial rifting zone in Northern Iceland and the Tjörnes fracture zone.
Geol. Soc. Amer. Bull., 85, 495.
- Scholz, C.H., 1968. The frequency-magnitude relation of microfracturing in rock and its relation to earthquakes.
Bull. Seismol. Soc. Am., 58, p. 399-415.

Sigurdsson, O., 1978. Náttúruhamfarir í Thingeyjarthingi 1976-1978 (in Icelandic, with English summary) Týli, in the press.

Tryggvason, E., 1978. Tilt observations in the Krafla-Mývatn area 1976-1977.
Nordic Volcanological Institute,
Report 78 02, 45 pp.

Tryggvason, E., 1978. Distance measurements in 1977 in the Krafla-Mývatn area and observed ground movements.
Nordic Volcanological Institute,
Report 78 10, 47 pp.

Utsu, T., 1971. Aftershocks and earthquake statistics III.
J. Faculty of Science, Hokkaido Univ., Series VII,
3, 379-441.

Wyss, M., 1973. Towards a physical understanding of the earthquake frequency distribution.
Geophys. J. R. astr. Soc., 31, 341.