The Grímsvötn Caldera, Vatnajökull:
Subglacial Topography and Structure of Caldera Infill

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ABSTRACT

A shallow seismic reflection survey was carried out on the ice shelf covering the subglacial lake in the Grímsvötn Caldera, Vatnajökull, in 1987. The survey showed that at the relatively low water level 9 months after the jökulhlaup in 1986, the area of the subglacial lake was 10 km^2 and the volume of the lake was 0.5 km^3. The ice shelf was 240-260 m thick in most parts and the water layer 40-90 m thick. Comparison and reinterpretation of a seismic survey conducted in 1955 suggests about 100 m increase in the thickness of the ice shelf over the 32 year period. The size of the main caldera is about 20 km^2 and the elevation of the caldera floor is 1060-1200 m a.s.l. The caldera floor dips slightly from south to north and the southern and southwestern parts are believed to be covered with lava flows. In the northern and eastern parts, the lakefloor is believed to be covered with sediments. Interfaces could be seen below the lakefloor in the northern and central parts. These reflections are believed to arise from lava flows or sills within a sediment pile. It is suggested that the caldera infill is composed of a pile of lava flows and volcaniclastic sediments. Lava flows compose the greater part of the pile in the southern part but sediments are predominant in the northern part. This suggests that eruptions have been more frequent in the southern part of the caldera. The existence of the lava flows suggests that eruptions onto the lakefloor have been more voluminous than previously believed. It is suggested that the observed drop in geothermal power of the area in recent years, is caused by the reduced volcanic activity after 1940.

INTRODUCTION

In June 1987 a shallow seismic reflection survey was carried out in the ice covered Grímsvötn Caldera, Vatnajökull (Fig. 1). The purpose of the survey was to map the bottom of the subglacial lake within the caldera and in that way obtain information on the size and volume of the lake, as well as the structure of the caldera. The caldera has been highly active in the past, and numerous eruptions are known to have occurred in Grímsvötn over the last 400 years (Þórarinsson, 1974).

The area has been the focus of interest for Earth scientists since 1934 when an eruption and jökulhlaup (glacier burst) prompted the first scientific work (Áskelsson, 1936). The phenomena of volcanic and geothermal activity in the glacial environment are of great interest, as many of the present landforms observed in Iceland were formed in the subglacial eruptions during the last glaciation. The effects of the geothermal area on the glacier and the jökulhlaups are of great glaciological interest as well as being important in estimating the flood danger. As a consequence, a great deal of research

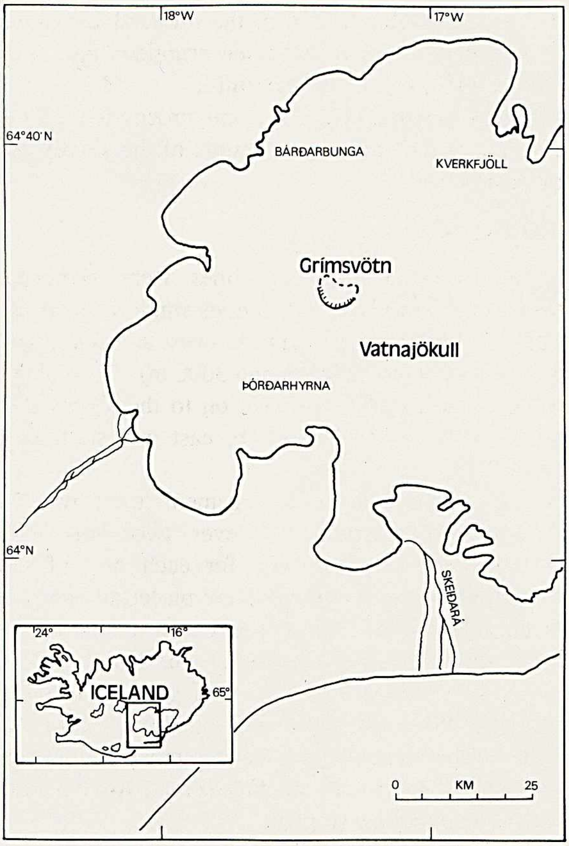

Fig. 1. Western Vatnajökull.

Mynd 1. Vesturhluti Vatnajökuls.

has been carried out on the glaciology and history of jökulhlaups and eruptions (Þórarinsson, 1953, 1965, 1974), on the glaciology and jökulhlaup mechanism and the nature of the geothermal area (Björnsson, 1974, 1983, 1988; Björnsson and others, 1982; Björnsson and Kristmannsdóttir, 1984; Steinþórsson and others, 1983) and the recent volcanic and seismic activity (Grönvold and Jóhannesson, 1984; Einarsson and Brandsdóttir, 1984; Brandsdóttir, 1984; Jóhannesson, 1983, 1984).

Grímsvötn is one of the most powerful geothermal areas in Iceland with a heat release of 4000-5000 MW thermal (Björnsson, 1974; Björnsson and Kristmannsdóttir, 1984; Björnsson, 1988). The geothermal heat melts the ice within the caldera, forming the subglacial lake which is covered by a floating ice shelf. The ice north of the caldera flows into the lake, where it is melted. The meltwater accumulates in the lake. The mechanism of the drainage of the lake is discussed by Björnsson (1974, 1988) and Nye (1976). Ice melting causes the water level in the lake to rise, and when a critical level is reached, the pressure at the bottom of the lake is sufficient to open a subglacial waterway to the edge of the glacier some 50 km to the south and the lake is drained in a jökulhlaup which lasts between one and three weeks.

Since 1934 jökulhlaups have occurred once every 4-6 years. Before that time the jökulhlaups were less frequent (approx. one every 10 years) and more voluminous (Þórarinsson, 1974, Björnsson, 1983, 1988). The total volume of water drained in each jökulhlaup before 1938 is estimated to have been 5-7 km^3 (Þórarinsson, 1974; Björnsson, 1988). Since 1938 the volume of the jökulhlaups is estimated to have ranged from 0.55 to 3.5 km^3 (Rist, 1955, 1984; Kristinsson and others, 1986).

Áskelsson (1934) was the first to suggest that Grímsvötn is a caldera. Þórarinsson (1974) estimated the size of the Grímsvötn Caldera as 35 km^2, from the glacier surface topography. He also suggested that it was composed of two calderas, the main caldera, and a second smaller caldera in the northwest part of the area. Sæmundsson (1982) suggested, also from the surface topography, that a third caldera is situated in the northeast corner of the depression. Finally, Björnsson (1988) gave a description of the subglacial topography of the area based on radio-echo soundings. He uses the term Grímsvötn Caldera for an area 6 to 10 km in diameter, bordered by the mountain ridge Grímsfjall to the south and subglacial mountain ridges to the north and east. Furthermore, he concludes that the caldera is divided into two main parts. The eastern part is 4 to 6 km in diameter and has its long axis striking NE. The western part can be divided into two elliptical areas. The southern elliptical area corresponds to Þórarinsson's main caldera and the northern area to the north caldera. The subglacial lake covers the greater part of the main caldera and at high water level it has in the past extended into the north caldera. The area covered by the lake is the area where geothermal activity is most intense.

Surveys aimed at mapping the subglacial topography in the Grímsvötn area date back to 1951, with the seismic work of the French-Icelandic expedition (Eyþórsson, 1951, 1952; Joset and Holtzscherer, 1954). More seismic reflection work was done in 1955 by the Icelandic-French expedition (Þórarinsson, 1965). However, these surveys failed to provide satisfactory estimates of the elevation of the bottom of the subglacial lake. Gravity surveys were carried out in 1960 and 1961 to obtain more information on the subglacial topography of the area (Pálmason, 1964; Sigurðsson, 1970), but interpretation of the gravity data was difficult due to lack of control points where ice thickness was known and Bouguer anomaly could be determined.

Since 1977, the subglacial topography of large parts of the Vatnajökull ice cap has been defined in considerable detail by radio-echo soundings (Björnsson, 1986, 1988). One part of this work was a detailed survey of the Grímsvötn area, carried out at the same time as the seismic survey. However, as radio waves of the frequency used in the soundings (1-5 MHz) do not penetrate water, other methods were needed in order to obtain information on the caldera floor beneath the lake.

The seismic reflection survey described in this paper complemented the radio-echo soundings by obtaining information on the subglacial lake and the main caldera in Grímsvötn. The survey objectives can be listed as follows:

1. To determine the elevation and topography of the floor of the main caldera in Grímsvötn in as much detail as possible.
2. To measure the area and the volume of the subglacial lake and in that way obtain data for the assessment of the flood potential of the area.
3. To obtain information on the structure of the caldera, in particular whether it is composed of several smaller calderas.
4. To obtain information on the material deposited on the lakefloor by volcanic eruptions and sedimentation, i.e. the caldera infill.

In this paper the fieldwork and processing of the data are described and the results of the survey are presented.

FIELD PROCEDURES

Three seismic reflection lines were surveyed, using a method giving continuous single-fold coverage. Line 1 (4800 m) bears east-west and line 2 and 3 run north-south (2520 m and 3000 m). The ends of the lines reach some distance on to the slopes that border the ice shelf to north, east and southwest (Fig. 2).

The source-receiver arrangement used was in essence a split spread. However, two shots were fired in each shothole, one for each part of the spread, as all the geophones were placed to one side of the hole for the first shot and then moved to the other side for the second shot. The length of the receiving spread was 230 m, the 12 geophones being placed at 20 m intervals with a minimum offset of 10 m. The second geophone spread for each shothole was used as the first spread for the next shothole on the line (Fig. 3).

A single vertical (P-wave) geophone with a natural frequency of 8 Hz was used for each channel. During the course of the survey the weather was sunny and hot, giving rise to wet and slushy snow conditions on the surface of the glacier. Placing the geophones at about 30 cm depth in the snow seemed to give satisfactory coupling.

Small dynamite charges (150-400 g) were used as seismic sources. The charges were detonated at the bottom of 30 m deep holes, drilled by a hot water drill.

The seismic recording system used was a 12 channel Geometrics Nimbus 1210F. A 80 Hz highpass analog filter on the Nimbus was used for all recordings apart from the first two shotpoints where a 50 Hz highpass was used. Tests showed that the application of the filter produced sharper reflections and improved signal to noise ratio. Each record was 1024 ms long with a sampling interval of 1 ms.

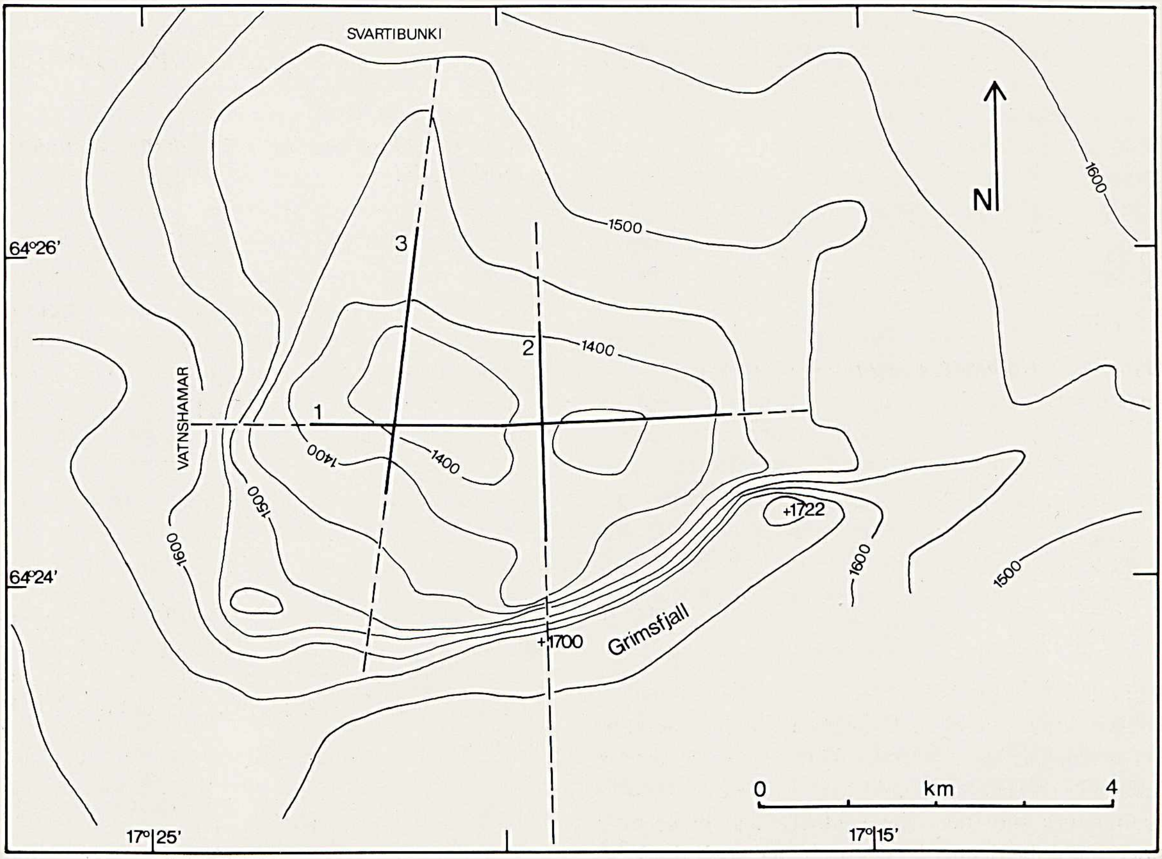

Fig. 2. A contour map of the ice surface in the Grímsvötn area (as mapped in June 1987, based on Björnsson 1988, p. 71, Fig. 5.3.). Contour interval 50 m. Bold lines mark the three seismic survey lines. The dashed lines mark the cross-sections on Figs. 6a-c.

Mynd 2. Grímsvatnasvæðið. Bil milli hæðarlína 50 m. Sveru strikin eru endurkastsmælilínurnar. Slitnu línurnar sýna legu þversniðanna á myndum 6a-c.

Hard copies of the data were obtained in the field using the printer on the Nimbus. The data were also stored on magnetic tape using a Geometrics G-724S tape recorder.

A small trailer on skis with a closed compartment was used on the survey. Inside it there was room for the seismograph, the tape recorder and the operator. The hot water drill was stored on a second trailer and a water barrel supplying the drill, was placed on the third. These trailers, when moved between shotholes, were pulled by snowmobiles or a snowcat. The number of people working on the survey ranged from 4 to 6, completing 3-5 shots per hour (360-600 m). The drilling of each shothole took only about 5 minutes.

DATA REDUCTION

The processing of the data can be divided into three main steps:

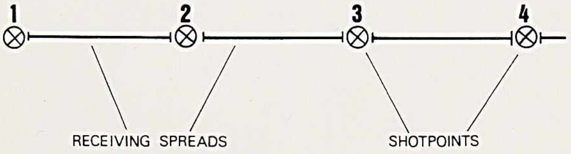

Fig. 3. Shotpoint and receiving spread arrangement on survey lines. A spread between every two shotpoints is used for recording signals from shots at both shotpoints.

Mynd 3. Fyrirkomulag skotpunkta og skjálftanema á mælilínum.

1. Velocity analysis, which is the basis for normal moveout correction of the data as well as the conversion of reflection times to depth.
2. Normal moveout correction.
3. Enhancement of the data which involved filtering and editing.

VELOCITY ANALYSIS

Two separate methods were used for velocity analysis. For the uppermost 40 m of the ice shelf a 210 m long reversed surface to surface refraction profile was shot near the southern end of seismic line 2. For the deeper layers of the ice shelf the X^2t^2 method was used (Dix, 1955; Sheriff and Geldart, 1983).

As illustrated by Joset and Holtzscherer (1953) the Wiechert-Herglotz method for calculating the velocity distribution with depth (Grant and West, 1965) is ideal for the uppermost layers of glaciers, where the transformation of snow to ice takes place. The results of applying the W-H method to the refraction profile are given in Table I and Fig. 4. The velocity increases rapidly from 700 m/s at z=0 m to 2000 m/s at 5 m. A velocity of 3000 m/s is observed at about 22 m depth and 3500 m/s at 30 m. Below 30 m the gradient is much smaller.

The main source of information on the seismic velocity at depths below 40 m (Table II and Fig. 4) consists of reflections from the ice-water interface. To invert these times into velocity the X^2t^2 method is used, which is based on the equation

Table I. Seismic velocities in the surface layers of the ice shelf as calculated by the Wiechert-Herglotz method.

Tafla I. Bylgjuhraði í yfirborðslögum íshellunnar sem fall af dýpi.

Z (m)	V (m/s)	Z (m)	V (m/s)
1.3	1131	15.7	2629
2.6	1548	25.3	3118
4.1	1833	29.8	3478
4.8	2000	36.1	3557
9.5	2250	43.3	3723
11.7	2500		

$$t^2 = t_0^2 + \frac{X^2}{V_{rms}^2}\,, \qquad (1)$$

where t is the reflection time measured at the surface at a distance X from the shotpoint and t_0 is the reflection time as measured at the shotpoint. The root-mean-square velocity, V_{rms}, is defined by the equation

$$V_{rms} = \left[(\sum_{i=1}^{n} V_i^2 t_i) / (\sum_{i=1}^{n} t_i) \right]^{1/2} . \qquad (2)$$

The model of the subsurface is of n horizontal layers where V_i and t_i are respectively the velocity in and the traveltime through layer i. In order to calculate the velocities of individual layers, the interval velocities, the Dix formula is used

$$V_n = \left[(V_{rms,n}^2 \sum_{i=1}^{n} t_i - V_{rms,n-1}^2 \sum_{i=1}^{n-1} t_i) / t_n \right]^{1/2} . \qquad (3)$$

The Dix formula follows directly from (2). It is important to note that the formula is not valid for large offsets, as it is assumed that the only difference in raypaths for waves reflected from interfaces n-1 and n is the additional travel between the two interfaces. If the two interfaces are not parallel large errors can occur and the Dix formula gives results that are meaningless. This however, is not a problem in the present survey, as areas where the ice-water interface is dipping can easily be identified

Table II. Seismic velocities as calculated by the X^2t^2 method. V_{rms} is the root-mean-square velocity for the ice-water reflection and V_2 is the interval velocity of layer 2. T_0 is the reflection time at zero offset.

Tafla II. Bylgjuhraði reiknaður samkvæmt X^2t^2 *aðferð.* V_{rms} *er veginn meðalhraði fyrir alla íshelluna en* V_2 *er hraði í ísnum neðan 30 m dýpis.*

Shotpoint	V_{rms} m/s	T_0 ms	V_2 m/s
4	3690	143.3	3800
5	3670	138.0	3780
12	3820	136.2	3940
13	3530	133.3	3630
26	3360	135.6	3450
27	3490	134.1	3590
28	3560	133.5	3660
29	3440	127.5	3540
39	3440	139.3	3530
40	3670	137.0	3780
mean	3570±100		3670±110

The mean of the two velocity values are given with 95% confidence intervals.

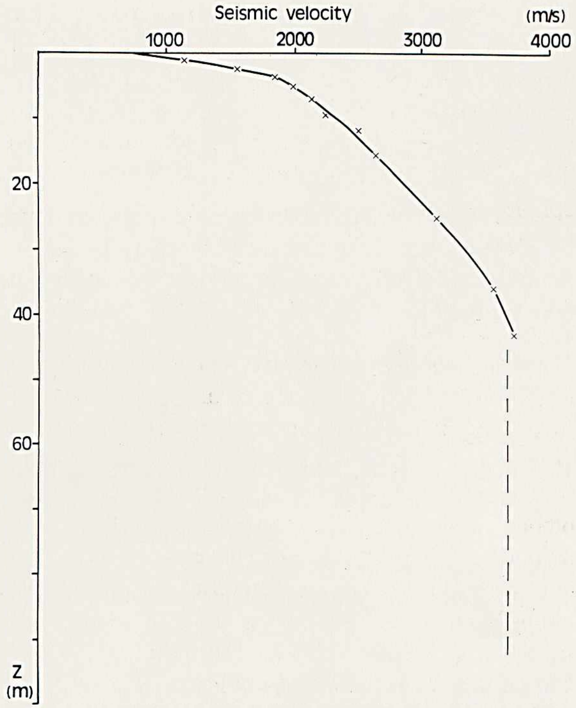

Fig. 4. A velocity profile of the ice shelf. The solid line shows the velocity profile of the uppermost 40 m as calculated by the W-H method (Table I). The broken line shows the interval velocity of layer 2 as calculated by the X^2t^2-method.

Mynd 4. Bylgjuhraði í íshellunni sem fall af dýpi.

and lateral changes in velocity are very small, if present at all.

For the purposes of velocity analysis it was decided to regard the ice shelf as consisting of two layers, above and below 30 m below the surface. The thickness of the upper layer is equal to the depth of the shotholes and corresponds roughly to the layer in which the compaction of snow and its transition to ice takes place.

Out of 40 split spread shotpoints in the Grímsvötn survey, 12 were either at the edges of the ice shelf or on the slopes surrounding it. Out of the 28 on the ice shelf proper, 10 were chosen for X^2t^2 analysis. The criteria were that there should be little or no dip, that the ice water reflection was free from irregularities and that the reflection times could be read with good accuracy (±1 ms). Shotpoints 4-13 are situated on line 1, 26-29 on line 2 and 39 and 40 on line 3 (Fig. 5). A least squares routine was used to calculate the RMS-velocities for each shotpoint.

The mean RMS velocity for the 10 shotpoints for the reflection from the ice-water interface is 3570 m/s. It is perhaps of interest to note that there are indications of higher velocities in line 1 (bearing east-west) than lines 2 and 3 (bearing north-south). The mean RMS velocity for shotpoints 4-13 is 3680 m/s, but for shotpoints 26-40 the mean is 3490 m/s. A more detailed study of the data is needed to see whether this phenomena is real or not. However, for the purposes of the present survey it is not important and the mean value of 3570 m/s is used for the data processing.

In order to calculate the thickness of the ice-layer from the reflection times, the interval velocity of layer 2 is needed. The RMS velocity for an imagined reflection from the bottom of layer 1 was calculated numerically using the velocity model from the refraction profile, giving V_{rms}=2390 m/s. The uphole time (13 ms) was used as the interval time for layer 1. Using the observed V_{rms} for the bottom of layer 2, Dix formula was used to calculate the interval velocity of layer 2. The results are given in Table II and Fig. 4.

No analysis was carried out to extract a velocity for the water layer from the data. The velocity of fresh water is a slowly varying function of temperature and the value used here is 1420 m/s, which corresponds to a temperature of 4 °C (Zemansky and others, 1966, p. 258). No measurements of the temperature of the subglacial lake have been carried out nor have there been any theoretical studies of its temperature profile (Björnsson and Kristmannsdóttir, 1984; Björnsson, 1988). The value of 4 °C is chosen here as it is the temperature at which a body of water is most dense and gravitationally stable.

PROCESSING

The processing carried out on the data included frequency filtering, normal moveout correction, spatial filtering and summing of adjacent traces. The frequency filtering was carried out in the time domain and the passband used was 80-150 Hz on the largest part of the data. On some parts of the data two passbands (50-125 Hz and 160-200 Hz) were used. The purpose of the frequency filtering was to reduce random noise and cut off spurious resonance frequencies present on some traces due to an improperly placed or slightly defective geophone.

Normal moveout correction (NMO) was carried out using a velocity function based on the results of the velocity analysis. The velocity values used as input for the function were V=3570 m/s for the ice and V=1420 m/s for the water.

In addition to frequency filtering, spatial filtering (mixing) was applied to the seismic sections after moveout to further reduce the noise content. A lowpass filter with a cutoff wavelength of 30 m turned out to give the best results. As the distance between reflection points along the seismic lines is 10 m (half the geophone spacing) the filter cuts off only the highest spatial frequencies present. In this set of data such spatial frequencies are only caused by random noise and perhaps by diffractions.

Despite the loss in resolution and a slight danger of aliasing, horizontal summing of adjacent traces turned out to be advantageous, as the signal to noise ratio was improved. It was therefore applied to all the data after frequency filtering, NMO and spatial filtering.

Parts of the data were contaminated by random noise caused by a faulty tape recorder. In addition to the processing steps described above, considerable editing of noisy traces had to be carried out. A large part of this process involved treating the noisy parts of the data with a gain function which reduced the amplitude of those parts of the traces which evidently only contained random noise. The original hard copies from the survey were used to define these areas on the seismic traces.

RESULTS

The processed seismic lines (Figs. 5a-c) show clearly the reflections from the ice-water interface ('a') and the lake bottom ('b'). Deeper reflections appear to be present in the northern part of line 2 ('c'-'e') and in places on lines 1 ('c') and 3 ('c'-'e'). Multiples are mostly absent from the data, but a ghost trails the ice-water reflection by about 25 ms in places. Some parts of the sections are seriously downgraded by noise. These parts are 3.5-3.7 km in line 1 and 2.0-2.1 km, 3.3-3.4 km and 4.2-4.3 km in line 3.

The ice-water interface reflection is everywhere strong, especially in the interior of the shelf where it is regular and continuous. Furthermore, the reflection from the lakefloor can be seen on all three seismic lines. Only in one place do the two reflections merge, at 2.4 km in line 3 (Fig. 5c). In other places the lakefloor reflection disappears near the edges of the ice shelf. The absence of reflections near the end of the lines indicates that the slopes bordering the relatively flat caldera floor are fairly

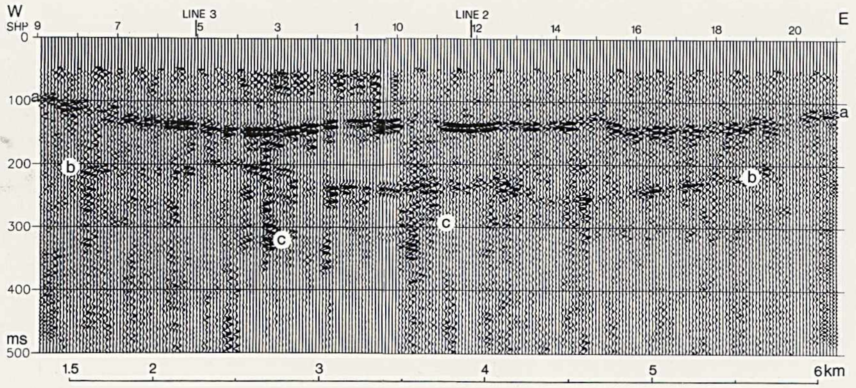

Fig. 5. a) Seismic line 1, time section. The spacing between traces is 20 m. 'a' marks the ice-water reflection, 'b' is the lake bottom reflection and 'c' is a reflection from below the lakefloor. Zero on horizontal distance scale the same as on Fig. 6a.

Mynd 5. a) Endurkastsnið, mælilína 1. 'a' er endurkastið frá botni íshellunnar og 'b' frá botni vatnanna. 'c' er endurkast frá lagmótum undir vatnsbotninum. Núllpunktur á láréttum skala sá sami og á mynd 6a.

steep. The critical angle at the ice-water interface for an upward travelling wave is only 23 degrees. A wave reflected from a dipping interface below the shotpoint is likely to have a larger angle of incidence and would consequently be totally reflected from the interface back into the water layer.

Cross sections through the main caldera in Grímsvötn are shown in Figs. 6a-c. The sections highlight the absence of any major topographic features in the interior areas of the caldera. The ice shelf is in most places 240-260 m thick and the water layer 40-90 m thick.

The data make it possible to measure the volume of the subglacial lake. The area of the ice shelf and the lake at the time of the survey was about 10 km^2. The volume was calculated by estimating the thickness of the water layer in all parts of the lake and then summing. The result is a total volume of 0.5±0.1 km^3.

The survey was conducted 9 months after the last jökulhlaup from Grímsvötn, which occurred in September 1986. As a consequence, the water level at the time of the survey was relatively low, or 1374 m a.s.l. which can be compared with about 1350 m at the end of the 1986 jökulhlaup and 1430 m just before its start (Björnsson, 1988). Fig. 7 shows the subglacial topography of the main caldera. The topography in the area to the north and east of the caldera is based on Björnsson (1988). The caldera has a long axis at N80°E, and is slightly eggshaped with the narrow end towards east. Its size is about 20 km^2. The deepest areas are in the northern part of the caldera where the lowest points have an elevation of about 1060 m a.s.l. The depth of the caldera from the highest rims (Eystri and Vestari Svíahnúkur) is therefore about 650 m. The caldera floor dips gently down towards north from the foot of the southern caldera wall. The total height of the southern wall is 500-550 m but the slopes bordering the caldera towards north are 250-300 m high. The height of the slopes on the western margins is about 400-450 m. There are two subglacial openings through the caldera walls into the caldera. A broad pass with an elevation of about 1100 m connects the main caldera to the north caldera. A second opening, slightly higher, is towards northeast. It is through the latter that many of the jökulhlaups are believed to force their way out of the subglacial lake (Björnsson, 1988).

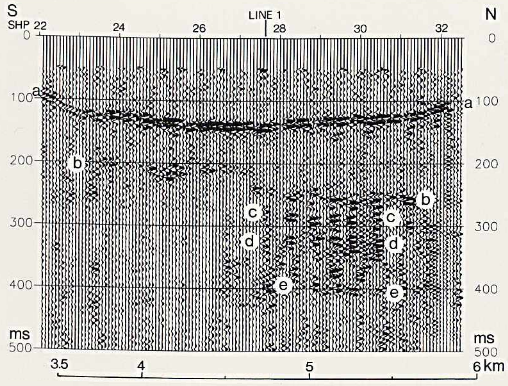

Fig. 5. b) Seismic line 2, time section. 'a' and 'b' as in Fig. 5a and 'c', 'd' and 'e' are deeper reflections.

Mynd 5. b) Endurkastsnið, mælilína 2. 'a' og 'b' þau sömu og á mynd 5a og 'c'-'e' eru endurköst frá lagmótum undir vatnsbotninum.

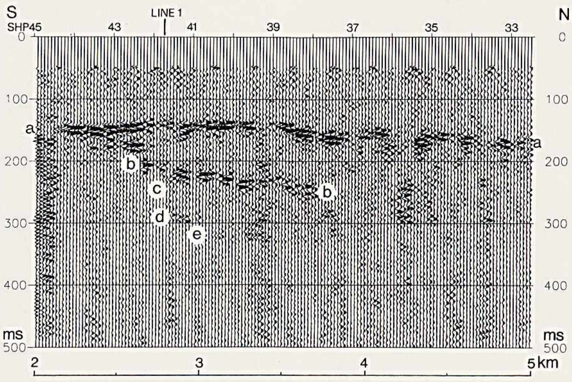

Fig. 5. c) Seismic line 3, time section. 'a' and 'b' as in Fig. 5a and 'c', 'd', and 'e' are deeper reflections.

Mynd 5. c) Endurkastsnið, mælilína 3. 'a'-'e' eins og á mynd 5b.

MORPHOLOGY OF CALDERA FLOOR

The strength of the lake bottom reflection varies along the seismic lines. Strong signals were obtained in the northern half of line 2, the area were the lake is present on line 3 and some central parts of line 1. Steep or vertical steps are present at 4.7 km in line 2 and 2.6 km in line 3. A step is present at 2.9 km in line 1 but is somewhat exaggerated on the time section, as the ice shelf is thickening above the step. The fourth step can be seen at 4.2 km in line 1. The throw of the steps in lines 2 and 3 is similar, 25-30 ms, which corresponds to about 20 m. At the western step on line 1 the bottom reflection splits in two and the continuation of the lower reflection can be detected for a further 300 m to the west. A possible explanation is that the step cuts the line at a low angle.

The data suggest that the steps seen in the lake bottom mark the divide between two seismic provinces. The areas on the downthrown side of the steps give in general a stronger, more coherent signal than do the highstanding areas. This can be seen clearly on the northern part of line 2, where the signal is very strong, and where deeper reflections appear to be present. On line 1 the bottom reflection is weak in the westernmost part of the caldera and in the area where lines 1 and 2 cross. In these areas steep steps with a bearing similar to line 1 are known to be present. The focusing effects at the ice-water interface would help to suppress reflections and diffractions from the edges and the northwardly dipping lakefloor north of the steps, explaining the weak signals in these areas.

Magnetic profiles measured along the seismic lines in 1987 and 1988 are shown above the cross sections in Figs. 6a-c. As can be seen on Fig. 6b, a clear anomaly is associated with the southern part of the caldera floor. The decrease in the magnetic field takes place directly above the step, clearly indicating that the rocks on the southern side of the step are magnetic. In Fig. 6c, a high is also associated with the southern part of the caldera floor but the main anomaly is somewhat to the south of the step in line 3. No clear anomalies can be seen over the steps on line 1 (Fig. 6a). However, the general east-west magnetic field gradient along line 1 decreases directly above the step at 2.8 km (Fig. 6a). A weak positive anomaly could therefore be superimposed on the general decrease in the magnetic field towards west. This would suggest that the rocks

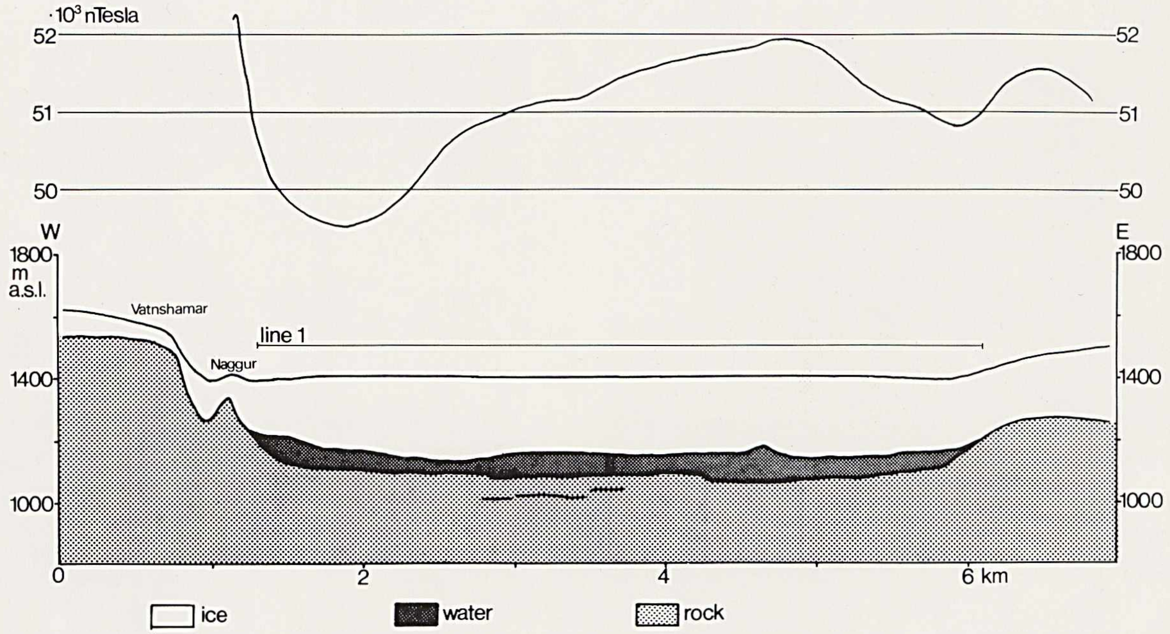

Fig. 6. a) East-west cross-section through the main caldera in Grímsvötn. Vertical exaggeration 2:1. The interface giving rise to reflection 'c' in 5a is shown. Above, the magnetic field strength in nanoteslas.

Mynd 6. a) Þversnið frá vestri til austurs gegnum meginöskju Grímsvatna. Fyrir ofan er segulsviðið eftir sniðinu.

forming the step are magnetic.

It is therefore proposed here that the higher areas represent a lava flow, or flows, covering a portion of the southern and western caldera floor. The steps are believed to be the edges of these lavas. In the lower areas in the northern and eastern parts of the lake, the lakefloor is believed to be covered with sediments, at least in the areas where reflections from horizons beneath the lakefloor are present.

DEEPER REFLECTIONS

Reflections from interfaces below the lakefloor are believed to be present in the northern half of line 2 (Fig. 5b). In the area 4.9-5.6 km, the water bottom reflection ('b') is at about 250 ms. At around 290 ms, the first deeper event can be seen ('c'). A second event ('d') is present at 310-320 ms and a third reflection ('e') can be seen at 390-400 ms. The possibility that these three deeper events are multiples can be ruled out as their arrival times are very different from those for the double reflections that might be present. Events, not as clear as the other three, are present between 'd' and 'e' (Figs. 5b and 6b).

In the central part of line 1, a reflection ('c') is present at 280-300 ms. It is not a strong event on most traces, but it is continuous and its arrival time is not consistent with that of a multiple. It is also in an area of a low lying lake bottom, considered to be covered with sediments. This event must therefore be considered to be a genuine reflection. The lake bottom in line 3 between 2.7 and 3.7 km is considered to be sedimentary in nature. There are indications of weak deeper reflections in places between 250-350 ms ('c', 'd' and 'e'), but individual reflections are continuous for only 200-300 m.

It is unclear whether the same deeper horizons are seen on all lines, but the limited extent of each

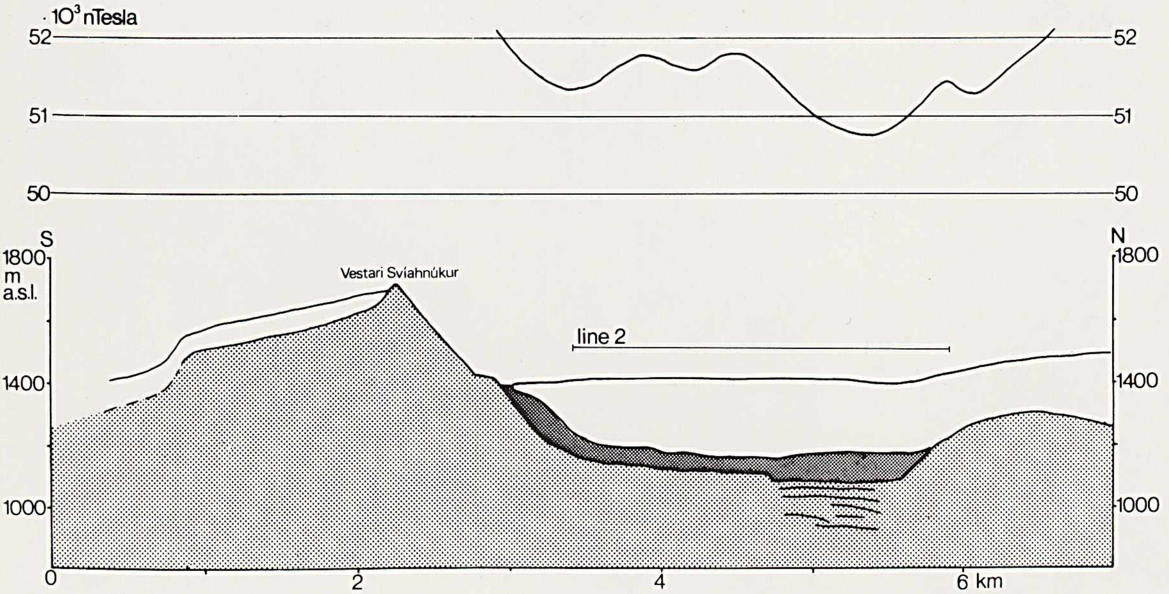

Fig. 6. b) North-south cross-section along seismic line 2. Reflections 'c' and 'd' on Fig. 5b correspond to the two uppermost interfaces below the lakefloor. Reflection 'e' corresponds to the deepest interface. The open water undir the southern caldera wall is at the site of the 1983 eruption.

Mynd 6. b) Þversnið frá suðri til norðurs gegnum meginöskjuna eftir endurkastsmælilínu 2. Vatnið undir Vestari-Svíahnúk er á gosstöðvunum frá 1983.

reflection suggests, that each horizon only exists over a limited area. The reflections are only seen in sediment covered parts of the lake bottom. The reflections are cut off fairly abruptly over the step which marks the edge of the lava flow in line 2. The roughness of the lava surface could be partly responsible for masking the predominantly high frequency signal from the underlying layers.

The deeper reflections are fragmentary and weak everywhere except in the northern half of line 2. However, as far as can be seen from the data, the general trend in these reflections is of dip towards north. Furthermore, there are indications that steps, similar to the ones observed in the present caldera floor, can be seen in places (line 1, 3.5 km; line 2, 5.0 km, below 350 ms.; line 3, 3.0 km). It is therefore believed that these reflections arise from buried lava flows and possibly sills intruded at shallow depths into sediments.

The existence of these reflections shows that substantial amounts of material have accumulated within the caldera since its formation. The caldera infill is made up of lavas and sediments. The sediments are probably mostly hyaloclastites from eruptions within the caldera, but sediments from glacial erosion may also be deposited on the caldera floor. The deepest reflections are 100-150 ms below the lakefloor which corresponds to a 75-150 m thick pile of sediments and lavas.

COMPARISON WITH PREVIOUS SURVEYS

In 1951 and 1955, seismic reflection surveys were carried out in the Grímsvötn Caldera (Joset and Holtzscherer, 1954; Þórarinsson, 1965). These surveys were essentially point soundings with one shotpoint and 3-12 geophones at an offset of 300-

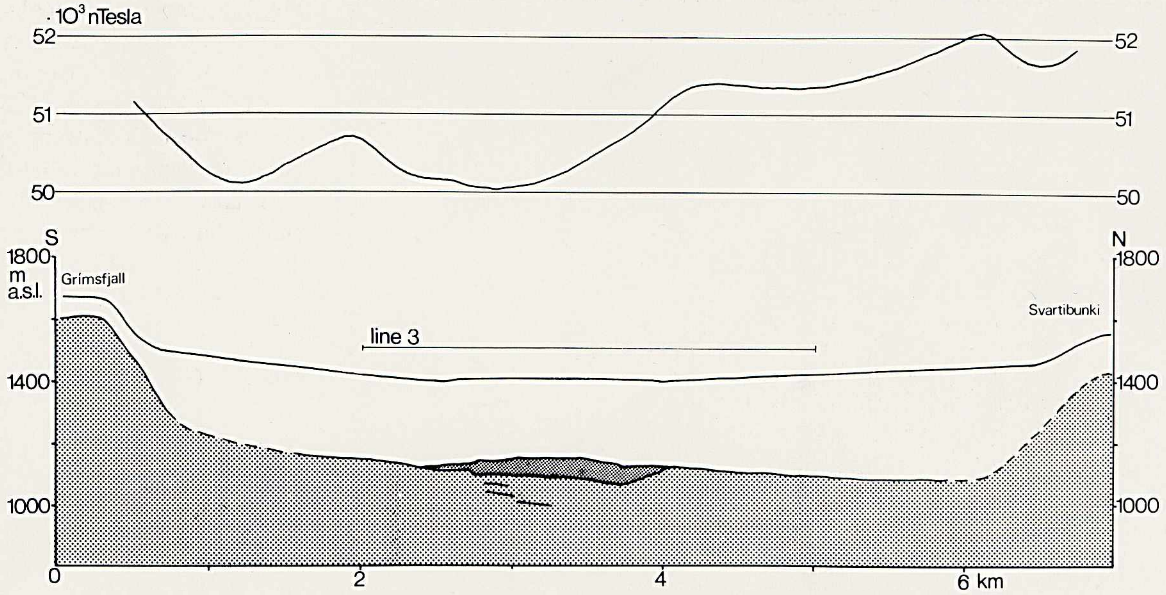

Fig. 6. c) Cross-section along seismic line 3. The broad ridge in the bedrock at 4 km is the divide between the main caldera and the north caldera.

Mynd 6. c) Þversnið frá suðri til norðurs gegnum megin- og norðuröskjuna eftir endurkastsmælilínu 3.

1000 m. One successful sounding was made on the ice shelf in 1951 by the French-Icelandic expedition and there were four successful soundings during the Icelandic-French expedition in 1955. Details of the latter survey have never been published but the results were summarized by Þórarinsson (1965).

These surveys were successful, in the sense that a reflection from the lakefloor was observed. However, no reflection from the ice-water interface was detected. The fact that only one reflection was observed led the French seismologists to interpret their data as if no water layer were present. They therefore calculated the depth to bedrock using ice velocities (V_m=3550 m/s, Joset and Holtzscherer, 1954). The elevation of the caldera floor which results from these calculations is 800-900 m a.s.l., contrasting with 1060-1100 m a.s.l. calculated from the present survey.

The existence of the subglacial lake over the past decades is well documented and beyond dispute (Wadell, 1920; Þórarinsson and Sigurðsson, 1947; Þórarinsson, 1953, 1965, 1974; Björnsson, 1974, 1988). It can therefore be taken as a fact that a water layer was present at the time of the previous surveys. The failure to detect the ice-water interface was probably caused by the field methods used. The long offsets result in a relatively short time interval between the arrival of the P-wave and that of a shallow reflection. Furthermore, a surface dynamite source appears to produce a high amplitude P-wave with a duration of about 100 ms (5-6 complete cycles, Joset and Holtszcherer, 1954, p. 20, Fig. 18) which masks any other arrivals which might occur in that time interval.

It is possible to reinterpret the results of the earlier surveys converting the ice thickness values given by Joset and Holtzscherer (1954) and Þórarinsson (1965) back to traveltime. To do this, information given by Joset and Holtzscherer (1954, p. 20) on the method of calculation is used.

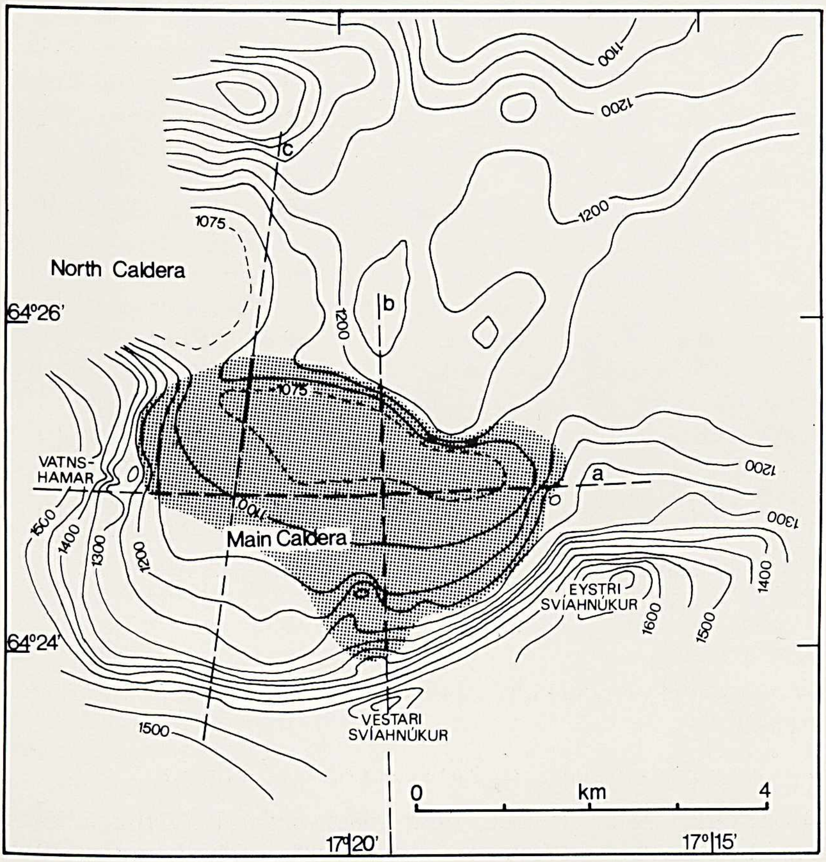

Fig. 7. Grímsvötn, subglacial topography. Contour interval 50 m. The topography of the area north and east of the main caldera is based on Björnsson (1988). The area covered by the subglacial lake in June 1987 is shaded. Dashed lines mark the cross-sections on Figs. 6a-c.

Mynd 7. Grímsvötn, botnkort. Bil milli hæðarlína 50 m. Skyggingin afmarkar það svæði sem vatnið náði yfir í júní 1987. Slitnu línurnar sýna legu þversniðanna á myndum 6a-c.

Furthermore, comparison of the velocity data of the earlier surveys with that of the present survey suggests that no significant changes in the velocity structure of the ice shelf have taken place. If it is assumed that the elevation of the lakefloor has not changed, the thickness of the ice shelf and the water layer can be calculated.

The present survey covers the areas where stations III and IV of the 1955 survey were located. The lakefloor elevation at these places is 1070-1100 m and 1100 m respectively, but 812-847 m and 922 m according to Martin's calculations. An ice shelf thickness of 120 m at station III and 150 m at station IV at the time of Martin's survey would be in agreement with unchanged elevation of the lakefloor.

Björnsson (1988) presented data on the thickness of the ice shelf in Grímsvötn for the period 1958-1987, based on measurements of the ice surface elevation above the water level. The results show that the ice thickness was 200-230 m in 1958-1966, but 250 m in 1980 and 1987. Furthermore, local melting reduces the ice thickness to 150 m in 1958 and 1964. Even though the figures from the reinterpretation of the 1955 survey must be treated with caution, they clearly suggest a substantial thickening of the ice shelf, in agreement with Björnsson's results.

DISCUSSION

The evidence from the seismic lines suggests that parts of the present caldera floor are covered with lava flows, and also that similar lava flows or sills are present, buried by sediments and other lava flows. Direct evidence from observations since 1919 as well as historical records of jökulhlaups, extending back to the sixteenth century, suggests that the caldera lake has existed for at least 400 years. At the same time numerous eruptions are known to have occurred in Grímsvötn (Þórarinsson, 1974). The lavas must therefore have been extruded subaqueously. The edges of the lava flows are 15-20 m high and their surface is relatively flat apart from a slight dip towards north.

The lava flows observed on the present caldera floor were formed in eruptions in the southern part of the caldera. The dip of the deeper reflections is similiar to the general dip of the caldera floor. This suggests that the lavas, that are believed to give rise to these reflections, also originated in the southern part. Apparently, eruptions have been more frequent under the southern caldera wall than elsewhere within the caldera.

Most of the lava flows are believed to be small. Each flow covers an area of only a few square kilometres. This implies that the volume of each flow is probably less than 0.1 km^3. Larger flows would cross the caldera floor to the northern caldera wall. The deeper reflections in line 2 do indeed suggest that larger lava flows have been erupted occasionally. Alternatively, these deeper units could be sills, intruded into the caldera infill. Further, the data suggest that the caldera infill is mostly made up of lavas in the southern and southwestern part of the caldera but that sediments constitute a greater proportion in the northern part (Fig. 8).

The contribution of volcanic eruptions within the caldera to the melting of ice has been considered negligible, as the eruptions have been considered small, producing mostly tephra (Þórarinsson, 1974). The existence of lava flows within the caldera suggests, that these ideas need to be reconsidered. The flows could be produced by the eruptions that have been observed at the end of some jökulhlaups. They could also be produced by entirely subaqueous eruptions which were not detected, probably occurring at high water level. If eruptions producing lava flows have been frequent, their effect on the thermal heat flux of the area could be substantial.

Björnsson (1983, 1988) estimated variations in subglacial melting at Grímsvötn over the period 1860-1986. These variations reflect variations in the thermal power of the area. His results show a base heat transfer rate of 4500-5500 MW. Peaks in the heat flux when the total power reaches 10000-15000 MW, are superimposed on the base heat flow. Björnsson suggests that these peaks are caused by volcanic eruptions at the glacier base, but that the base heat flux is drawn from a deeper magma body by hydrothermal convection. A decline in the base rate from 5000 MW to 4500 MW is detected over the period up to 1976. Since 1976, a sudden drop in the thermal power to 2500-3000 MW has been observed. This decreased geothermal output of Grímsvötn is further supported by the measurements mentioned earlier, on the increased thickness of the ice shelf. Also, maps compiled at various times over the last 50 years show that the area of the lake has decreased in this period (Björnsson, 1988).

As pointed out by Björnsson (1974, 1988), an eruption within the lake is unlikely to spark off a jökulhlaup immediately, as the water level would be raised only slightly. This can be explained by the fact that melting of the ice shelf does not change the water level, and rising of the water level would be controlled by the flow of ice into the lake. On the other hand, a subglacial eruption north of the lake could melt large volumes of ice and the meltwater would be drained into the lake causing a sudden rise in water level. The jökulhlaup in 1938 was an example of such an event (Björnsson, 1988). Therefore, a difference exists between the effects of volcanic eruptions on the behaviour of the jökulhlaups: frequent eruptions within the caldera are likely to sustain a high level of melting but raising of the water level would be gradual. On the other hand, an eruption north of the caldera would most likely spark off a large unexpected jökulhlaup. Consequently, a sharp peak would be detected in the observed thermal power of the area.

The thermal effects of eruptions onto the lakefloor can be estimated in the following way: Consider a period of high eruption frequency, with an eruption occurring once every 10 years. In each event, a total

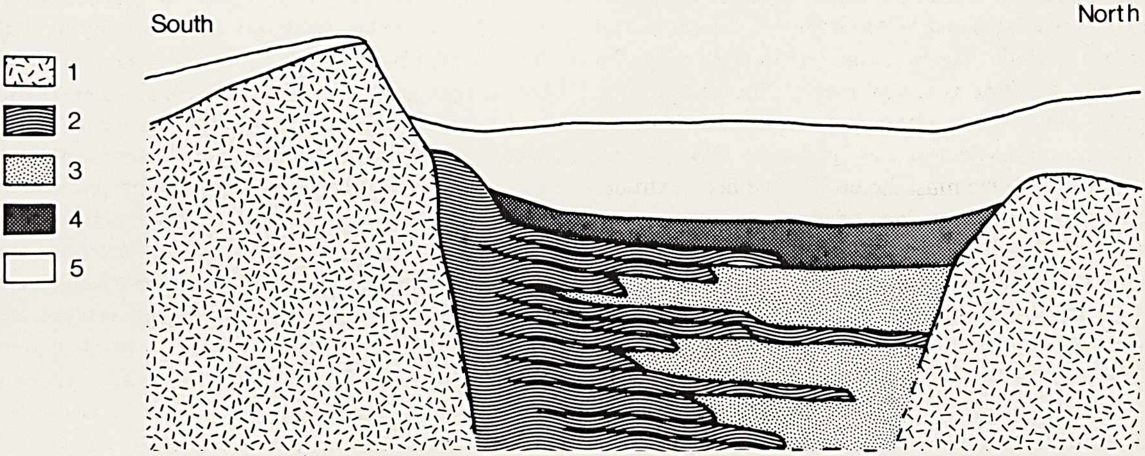

Fig. 8. Schematic cross-section of the infill of the main caldera. Figure is not to scale. Lava flows and sills are believed to constitute the major part of the infill in the southern part of the caldera but sediments have accumulated in the northern part.

Explanations / *Skýringar*:

1 Hyaloclastite rocks/ *Móberg*
2 Lava flows and shallow intrusions/ *Hraunlög og grunn innskot*
3 Sediments/ *Setlög*
4 Water/ *Vatn*
5 Ice/ *Ís*

Mynd 8. Einfaldað líkan af meginöskju Grímsvatna.

volume of 0.1 km^3 of magma is erupted onto the caldera floor and intruded at shallow depths. This amounts to 10^7 m^3/year of magma. According to the calculations of Björnsson (1988, p. 104), the heat released by magma which solidifies and cools from 1300 °C down to 200 °C is $4.2 \cdot 10^9$ J/m^3. If the volume of magma is completely solidified and cooled down over the 10 year period the contribution to the power of the geothermal area is 1300 MW. Evidence from the Heimaey eruption shows that lava is cooled very rapidly when in direct contact with water (Björnsson, 1987). It is therefore likely that lavaflows on the lakefloor are solidified and cooled down within few years from the time of eruption.

Heat release from intrusions is probably more gradual, as access of water is not as free as on the lakefloor. However, it is unclear how much role penetration of water into an intrusion plays in extracting its heat, which makes it difficult to quantify the process. A likely upper limit on the time it takes a sill-like intrusion to cool down can however be estimated by assuming that conduction is the only means of heat transport within the intrusion and below it. In the following model the upper surface of the intrusion is kept at a constant temperature which equals that of the surroundings before the magma was intruded. This corresponds to the situation where all heat that is conducted through the upper surface is carried away by hydrothermal convection. No convection is supposed to take place below the intrusion. Buntebarth (1980) presents a solution to the heat-flow equation with the above boundary conditions

$$T(z,t) = T_2 + \frac{(T_1-T_2)}{2}\left[\mathrm{erf}\left(\frac{h-z}{\sqrt{2kt}}\right) + 2\mathrm{erf}\left(\frac{z}{\sqrt{2kt}}\right) - \mathrm{erf}\left(\frac{h+z}{\sqrt{2kt}}\right)\right]. \quad (4)$$

Here T is temperature at time t and depth z measured down from the upper surface of the intrusion. h is the thickness of the intrusion and T_1 and T_2 are respectively the initial temperature values of the intrusion and the surroundings. erf is Gauss error function and k is the coefficient of thermal diffusivity. A value of $k=20\ m^2/year$ for basalt is used (Cermák and Rybach, 1982).

The heat left in the sill-like body and the rocks below it at any given time t can be estimated by integrating T(z,t) with respect to z from z=0 down to where $T(z,t) \approx T_2$. These calculations indicate that a 10 m thick sill has lost 80% of its heat through its upper surface after about 5 years from its formation and 90% after 15-20 years. For a 20 m thick sill 80% is lost after 20 years and 90% after 50 years. These results of the model must by treated with caution, as it may be unrealistic that no warming up of the rocks above the sill takes place while it is still molten or very hot. Furthermore, the role of convection in heat transport may be underestimated once the intrusion is solidified. However, the results clearly indicate that the contribution of shallow intrusions to the power of a geothermal area becomes negligible within few decades from their formation.

Eruptions are known to have occurred in Grímsvötn in 1867, 1873, 1883, 1922, 1934, 1938 and 1983, and are considered likely in 1861, 1892 and 1903 (Þórarinsson, 1974; Jóhannesson, 1983; Björnsson, 1988). The 1983 eruption was considerably smaller than the other eruptions and would probably not have been detected without the local seismic monitoring network (Einarsson and Brandsdóttir, 1984). It is possible that minor eruptions occurred in 1945 and 1954 (Áskelsson, 1959; Jóhannesson, 1983). In the period 1860-1940, 6-9 eruptions occur in the area or one eruption every 9-13 years. Since 1940, 1-3 very minor eruptions have occurred over a 50 years interval. It is therefore possible that part of the base heat flux observed in the period 1860-1976 was derived from shallow intrusions and lava flows erupted onto the caldera floor. The observed drop in thermal power of the area could be caused by the reduced eruption frequency since 1940, as the heat from lava flows and intrusions has been exhausted. If this is true, the geothermal power observed at present reflects the energy drawn from more deeply seated magma bodies by hydrothermal convection.

CONCLUSIONS

The results of the 1987 seismic reflection survey have been presented and an interpretation presented and discussed. It is believed that the following conclusions are supported by the data:

1. The ice shelf covering the subglacial lake is 240-260 m thick and the thickness of the water layer is 40-90 m.
2. The area of the subglacial lake in June 1987 was 10 km^2 and the volume of water stored in the lake was 0.5 km^3.
3. The area of the main caldera in Grímsvötn is about 20 km^2. The elevation of the caldera floor is 1060-1200 m a.s.l. The caldera floor dips slightly towards north and the deepest part is under the northern caldera wall.
4. The caldera floor can be divided into two separate areas. The deeper northern and eastern parts are covered with sediments but the southern part is covered with lava flows.
5. Comparison with earlier seismic surveys (1955) suggests, that the ice shelf was 120-150 m thick at that time, i.e. about 100 m less than at present.
6. The caldera infill is believed to be made up of a pile of lava flows and sediments, with a minimum thickness of 100-150 m. The lava flows seem to have been erupted in the southern part of the caldera, but the sediments have accumulated in the northern part.
7. The existence of the lava flows suggests that the thermal effects of eruptions within the caldera

have been substantial. It is suggested that the observed drop in heat flux from the area in recent years can be explained by the decrease in volcanic activity since 1940. The heat stored in lava flows and shallow intrusions is now exhausted.

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ÁGRIP

GRÍMSVATNAASKJAN, LANDSLAG OG GERÐ JARÐLAGA

Vitneskja um stærð og rúmtak Grímsvatna hefur verið af skornum skammti fram til þessa, þar sem dýpi þeirra hefur verið óþekkt. Gerðar voru tilraunir til að mæla dýpið vorið 1951 og aftur 1955 með endurkastsmælingum með skjálftabylgjum, en niðurstöður voru frekar óvissar. Íssjá er ekki unnt að beita til að finna botn Grímsvatna þar sem rafsegulbylgjur, eins og hún notar, sjá ekki gegnum vatn. Sumarið 1987 voru því enn á ný gerðar endurkastsmælingar með skjálftabylgjum á íshellu Grímsvatna. Tilgangur mælinganna var að kortleggja botn vatnanna og afla þannig upplýsinga um stærð þeirra og rúmtak. Einnig var ætlunin að fá vitneskju um gerð jarðlaga í Grímsvatnaöskjunni, en talið er að henni megi skipta í a.m.k. 2 smærri öskjur. Sú syðri er stærri um sig og er hér nefnd meginaskjan. Norðvestur af henni og samtengd er svo nyrðri askjan. Vötnin sjálf eru innan meginöskjunnar en teygja sig inn í þá nyrðri við háa vatnsstöðu.

Mæld voru þrjú endurkastsnið samtals um 10 km að lengd (mynd 2). Myndir 5a-c sýna endurkastsniðin og sjást víðast hvar 2 endurköst og sumstaðar koma fram fleiri. Sterkasta endurkastið ('a') kemur frá botni íshellunnar en botn vatnanna ('b') sést í öllum þremur sniðunum. Á öllum sniðunum koma fram einhver endurköst frá jarðlögum undir vatnsbotninum ('c'-'e') en sterkust og greinilegust eru þau í norðurhluta sniðs 2.

Íshellan reyndist víðast hvar vera 240-260 m þykk en vatnslagið þar undir var 40-90 m (myndir 6a-c). Stærð þess svæðis þar sem vatn var að finna undir ísnum var um 10 km^2 og rúmmál vatnsins var um 0.5 km^3. Mælingarnar voru gerðar við tiltölulega lága vatnsstöðu þar sem aðeins 9 mánuðir voru liðnir frá hlaupinu 1986. Setur það mark sitt á niðurstöður hvað varðar stærð vatnanna og rúmtak. Þrátt fyrir það staðfesta þessar niðurstöður að vötnin hafa mjög skroppið saman á síðustu árum. Hefur það haldist í hendur við minnkandi Skeiðarárhlaup. Ennfremur bendir samanburður við endurkastsmælingarnar sem gerðar voru 1955 til þess að íshellan hafi þykknað um allt að 100 m síðan þá. Er það í samræmi við niðurstöður Helga Björnssonar (1988) sem byggðar eru á hæð yfirborðs íshellunnar yfir vatnsborðinu eins og hún var mæld í leiðöngrum Jöklarannsóknafélagsins 1958-1966, 1978, 1980 og 1987.

Meginaskjan er um 20 km^2 að flatarmáli og liggur botn hennar dýpst í um 1060 m y.s. norðantil (mynd 7). Askjan er því 600-650 m djúp frá hæstu brúnum Grímsfjalls. Megin- og norðuraskjan virðast aðskildar með breiðum en lágum hrygg. Botn meginöskjunnar hækkar til suðurs og benda segulmælingar til þess að stallar sem fram koma í endurkastsniðunum séu jaðrar hrauna sem runnið hafa frá upptökum undir suðurbörmum öskjunnar (Grímsfjalli). Hraunbrúnir þessar eru 15-20 m háar en talið er að hvert hraunlag hafi takmarkaða útbreiðslu. Líklegt er að hvert hraun sé aðeins fáir ferkílómetrar að stærð. Endurköst frá jarðlögum undir botni vatnanna eru talin koma frá samskonar hraunlögum sem grafist hafa í set, en einnig getur verið að endurköstin komi frá yfirborði grunnra innskota (silla) sem troðist hafa inn í jarðlagastaflann á botni öskjunnar Svo virðist, að jarðlögum innan meginöskjunnar halli heldur til norðurs (mynd 8). Er talið að gosvirkni hafi verið öflugri undir Grímsfjalli en annars staðar innan öskjunnar. Hraun og innskot mynda þar því stærri hluta jarðlagastaflans en í norðurhlutanum.

Sveiflur í afli jarðhitasvæðis Grímsvatna hafa verið metnar út frá stærð Skeiðarárhlaupa af Helga Björnssyni (1988). Kemur þar meðal annars fram að heldur hefur dregið úr aflinu síðustu 120 árin. Sérstaklega virðist hafa dregið niður í Grímsvötnum eftir 1976. Hugsanlegra skýringa á þessu gæti verið að leita í því, að gosvirkni í Grímsvötnum hefur minnkað mjög eftir 1940. Hraunin og innskotslögin, sem talin eru sjást í endurkastsgögnunum, hljóta að hafa aukið mjög á varmaflæðið á svæðinu meðan þau voru að storkna og kólna. Á tímabilinu 1860-1940 varð gos að meðaltali á um 10 ára fresti en síðan hefur aðeins einu sinni gosið svo öruggt sé. Meðan gos voru tíð er líklegt að stöðugt hafi bæst við ný grunn innskot og hraun. Varmaflæðið mestan hluta tímabilsins gæti því hafa átt sér tvær uppsprettur. Í fyrsta lagi er sá varmi sem fólgin var í hraunum og grunnum innskotum og í öðru lagi varmi hins reglulega jarðhitasvæðis sem talinn er kominn frá kvikuhólfi undir Grímsvötnum. Nú þegar mjög hefur dregið úr gosvirkni er líklegt að varmi hrauna og grunnra innskota sé að mestu uppurinn. Sé þessi tilgáta rétt er það varmaflæði sem nú mælist jafnt grunnvarmaflæði hins eiginlega jarðhitasvæðis.

Grímsvötn from the north on 27 February 1986 5 months before a jökulhlaup. Flight altitude 5 km a.s.l. Photo Oddur Sigurðsson.

Grímsvötn úr norðri 27. febrúar 1986 5 mánuðum fyrir hlaup. Flughæð 5 km y.s. Mynd Oddur Sigurðsson.

Grímsvötn from the east on 10 September 1986 at the end of a jökulhlaup. Flight altitude 3 km a.s.l. Photo Oddur Sigurðsson.

Grímsvötn úr austri 10. september 1986 í lok jökulhlaups. Flughæð 3 km y.s. Mynd Oddur Sigurðsson.