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The Gravity Field of Northeastern Iceland

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Abstract. From 1964 to 1970, about 1,000 gravity stations have been established in parts of northeastern Iceland. The survey covered the young volcanic zone and the adjoining tertiary plateau basalts mainly in the region between 65°,5–66° N and 18°–16° W, with some extensions across the eastern basalt zone. Gravity measurements have been carried out in profiles with LaCoste-Romberg gravity meters (± 0.03 mgal), while heights have been determined by barometric levelling ($\pm 2 \ldots 3$ m). Density determinations by rock weighing and by Nettleton-profiles gave mean values between 2.0 g/cm^3 (hyaloclastite rocks) and 2.8 g/cm^3 (tertiary basalt lavas). Bouguer anomalies have been calculated with uniform density 2.6 g/cm^3 and with different density zones (2.2–2.7 g/cm^3). The accuracy of the anomalies varies between ± 0.5 and ± 4 mgal, depending on the height of the gravity station. The gravity field shows the well-known decrease from north to south (0.4...0.5 mgal/km) and a relative gravity minimum (5 mgal) in the active rift zone. The more irregular gravity behaviour in the central part of the young zone might be due to surface near mass anomalies.

Key words: Gravity anomalies – Rock densities – Icelandic rift zone.

1. Introduction

Iceland and its surroundings are of special interest for the geosciences, as here the central part of the Mid-Atlantic Ridge is rising above sea-level, thus giving the opportunity for detailed geological, geophysical and geodetic investigations at an accreting plate boundary.

A first survey of the gravity field in this region has been carried out in 1938 in northern Iceland as part of a German geological (Bernauer), geodetic (Niemczyk and Emschermann) and geophysical (Ansel and Schleusener) expedition (Niemczyk, 1943). 40 gravity stations have been established at this survey, situated mainly along the line from Akureyri to Grimsstadir (Schleusener, 1943).

A gravity survey of the whole island has been carried out by Einarsson (1954), containing about 900 stations. North of 65°30′ latitude, Einarsson occupied about 100 stations between Eyjafjördur and Vopnafjördur.

The investigation of the gravity field in northern Iceland has been taken up again in 1964 (1964–1967: Schleusener; 1970: Schleusener, Torge and Drewes). The aim of these measurements was to give the regional gravity field especially in west-east direction as well as the local gravity field of several small structures and, above all, to detect eventual gravity variations with time by repeated measurements at some monumented stations. This report deals with the result of the regional survey 1964–1970. A short discussion on the gravity field of some local structures and the general regional gravity behaviour has been given by Schleusener and Torge (1972) and by Schleusener (1974). The results of repeated gravity measurements along a west-east profile at 65°40′ N latitude have been presented by Schleusener and Torge (1971), the extension of this profile and the description of the survey 1970/71 is given by Schleusener *et al.* (1974).

The gravimetric investigations are part of the research work of German geodetic (*e.g.* Gerke and Pelzer, 1972; Gerke, 1974) and geophysical (geomagnetics: Angenheister *et al.*, 1972; microseismics: Steinwachs (1972)) groups in Iceland. This work is sponsored by the German Research Society (Deutsche Forschungsgemeinschaft).

2. Description of the Survey Area

The regional gravity survey described here, covers mainly the region between 65,5°–66° northern latitude and 18°–16° western longitude. From west to east, the survey stretches over the tertiary plateau basalt zone between Eyjafjördur and the Bárdardalur fault, and the western and central part of the young volcanic zone until the river Jökulsá á Fjöllum (Fig. 1). Some profiles extend to the eastern basalt zone, starting about 20 km east of the river Jökulsá, and end at the eastern coast of Iceland.

The tertiary basalt zone constitutes a rather homogeneous plateau-like formation of basalt lavas with intercalated sedimentary and tuff layers, strongly eroded and interrupted by deep valleys with more or less consolidated sediments. Heights of about 900 m above sea level are reached in this zone. The young volcanic zone may be divided into the flanking intermediate zones, consisting mainly of pleistocene flood basalts, and the still active zone of rifting and volcanism being the central part of this region. Here a variety of pleistocene and postglacial structures is found, as table mountains, shield volcanoes and mainly south-north directed crater rows, fissures and faults. The heterogeneous younger series filling this zone have been designated as palagonite or Móberg formation, consisting of hyaloclastite (palagonite tuff and breccia) rocks, young lava flows and sediments (Thorarinsson *et al.*, 1959; Pálmason and Saemundsson, 1974). The average height of the lowlands of the central zone is about 100...400 m, table mountains and other volcanoes reaching 800 m and more.

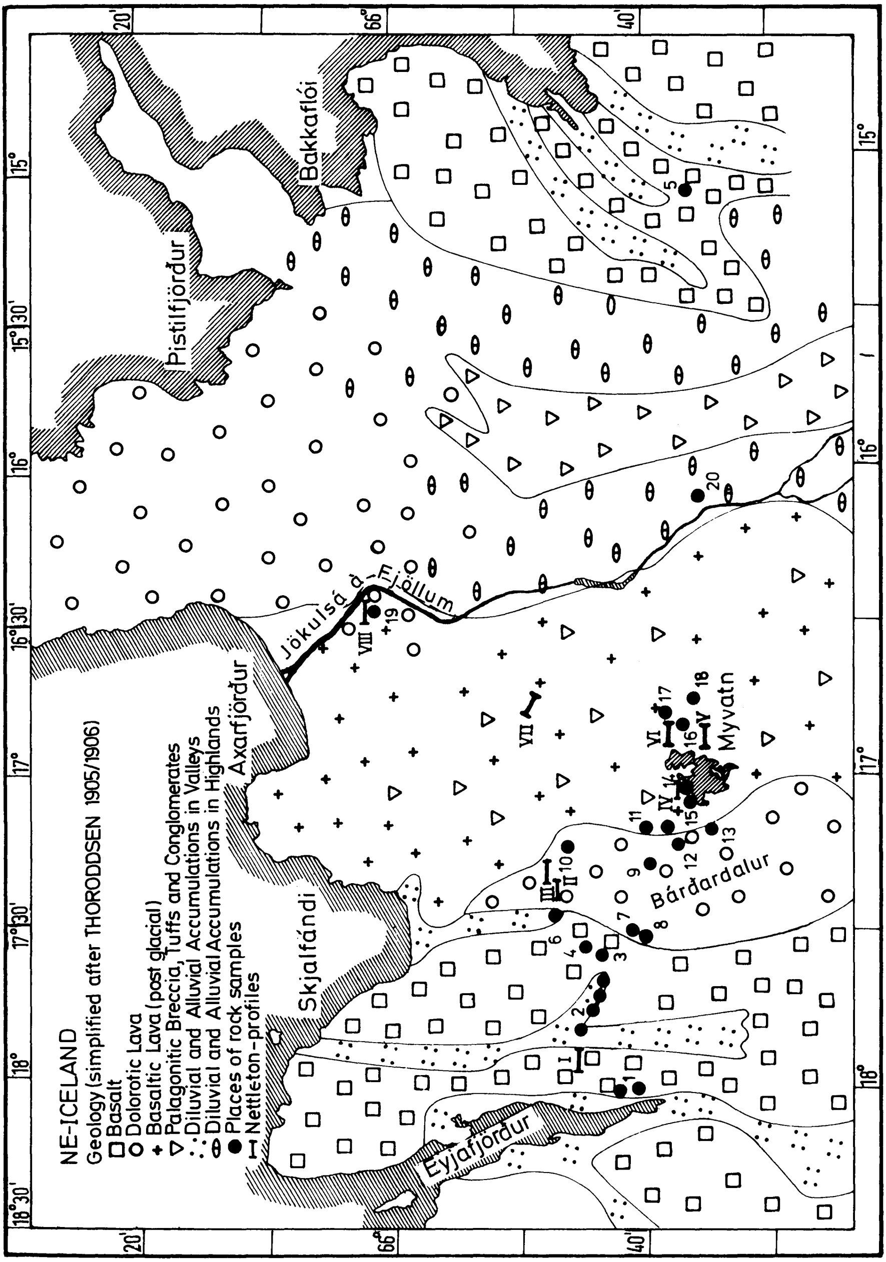

Fig. 1. Geology of the survey region, positions of rock sampling and Nettleton-profiles

3. Measurements and Data Evaluation

3.1. Survey Layout

For economical reasons the gravity stations have been established in profiles mainly along roads and tracks, so that car transport was possible (Fig. 2). The

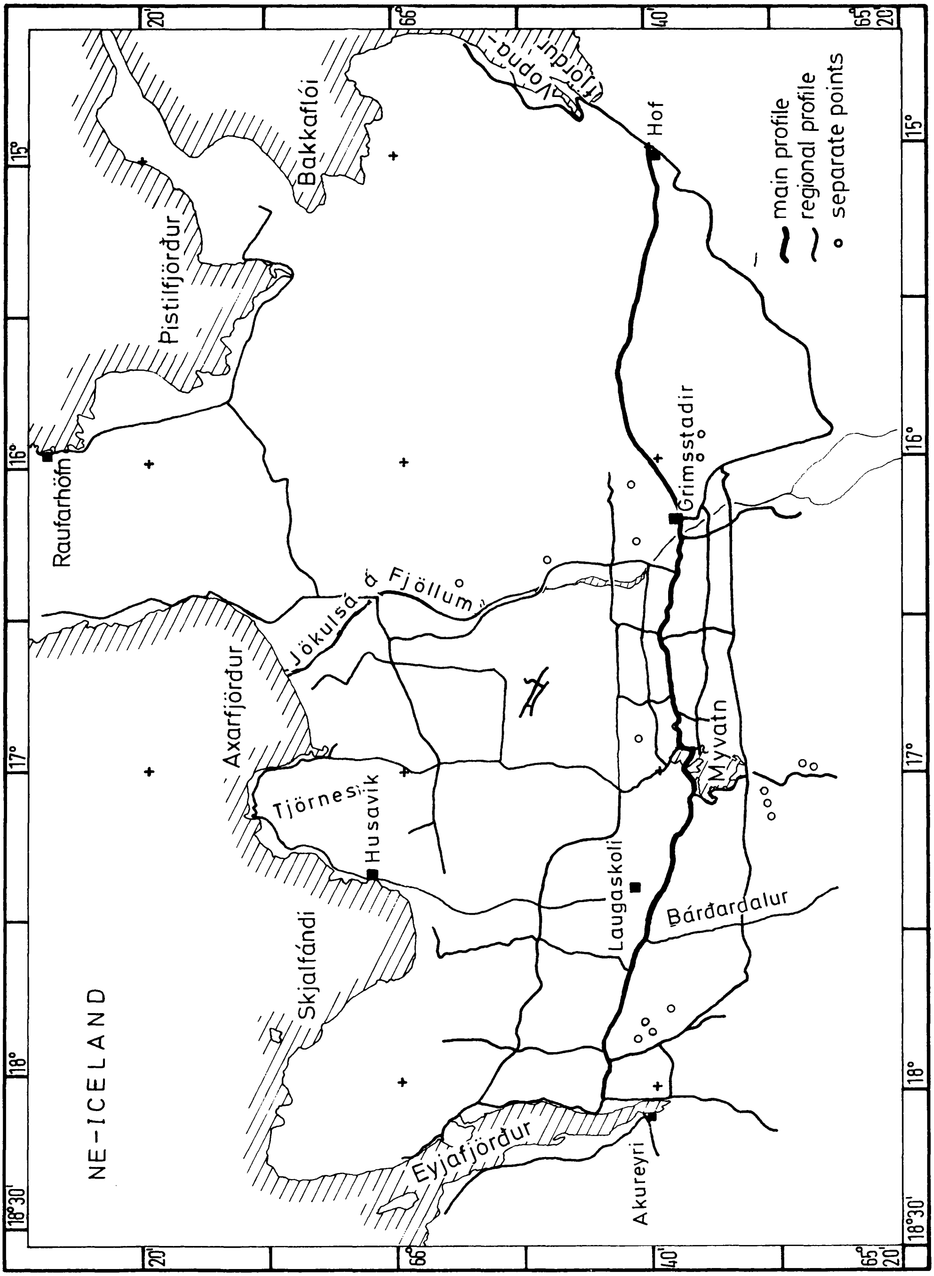

Fig. 2. Gravity profiles

west-east profile following approximately the road from Akureyri to Grimsstadir, is used for investigations about secular gravity variations (monumented main profile). In order to get a more detailed picture of the gravity field here, four parallel profiles have been observed by foot. East of the river Jökulsá á Fjöllum, the area is more or less inaccessible by car, so only a few profiles have been measured there in order to obtain an impression of the gravity of the eastern basalt zone.

The average spacing of points on the profiles is about 1 km. Altogether about 1,000 stations have been observed in the measuring periods 1964, 1965, 1967 and 1970.

Gravity, position and height have been determined simultaneously, a survey crew generally consisting of two observers. One observer was responsible for the gravity measurements, the other one for positioning and height determination. If possible the stations have been marked by color, furthermore a point sketch has been prepared. So the station could be found again for connecting surveys.

With the exception of the authors, the following persons participated at the field work and the data evaluation: H. Lehrke, A. Berger, W. Brosche, H. Hahn, B. Köhler, U. Meyer, M. Mimus, H. Mittendorf, B. Stache, H.-J. Voss, H. Zimmermann and the Icelandic and German helpers B. Benediktsson, T. Keil, V. Sauer, H. Tomasson. The Icelandic geologist Thomas Tryggvason † has been a valuable guide to Icelandic geological and geographical conditions and a never tired active aid in difficult field situations, form 1938 to 1965.

3.2. Gravity Measurements

The gravity measurements have been carried out with the LaCoste-Romberg model *G* gravity meters no. 79 and no. 85. Worden gravity meter no. 530 has been used in measuring some Nettleton-profiles for determination of rock density.

The profile measurements always started and ceased at a station of the "main profile", the gravity of these stations being known with $\pm 0.01 \ldots 0.02$ mgal accuracy. These loop measurements allowed the detection of gross errors and a drift control. One loop generally contained 15 . . . 25 stations, which could be observed within 8 . . . 10 hrs. At each station, three readings have been made for control, the mean value being introduced as observed quantity.

Evaluation started with the transformation of the mean reading to mgal-scale, using the manufacturers conversion tables. Tidal corrections have been applied from the tables calculated by Goguel (1954) and published annually in "Geophysical Prospecting". Residual loop misclosures were supposed to result from instrumental drift and distributed proportional with time.

Gravity values refer to the Potsdam gravity system, according to the connections measured from 1964 to 1970 between Reykjavik and the continent (Torge, 1971). The gravity value of Reykjavik, University, station no. 21941 A (IAG-SSG5-Catalogue) has been found with

$$g_{\text{Reykjavik}} = 982{,}279.93 \pm 0.05 \text{ mgal.}$$

This value refers to the conventional Bad Harzburg (station no. 21510 A) value $g = 981{,}180.40$ mgal. As the IGSN 71 (Morelli *et al.*, 1974) value for Reykjavik

A is

$g_{\text{Reykjavik}} = 982{,}264.96 \pm 0.02$ mgal,

a correction of -14.97 mgal must be applied to the gravity values of the present survey, if transformation to IGSN 71 is desired.

Accuracy of the gravity values may be estimated from the root mean square errors (r.m.s.e.) of the main profile reference stations and of the observed gravity differences. For the main profile stations, r.m.s.e. of $\pm 0.01 \ldots 0.02$ mgal have been found, referring to the station Akureyri being the base station for the survey. The station Akureyri has been connected to Reykjavik with an accuracy of about $\pm$ 0.01 mgal (Schleusener *et al.*, 1974). Small gravity differences can be measured with the same order of accuracy (Schleusener and Torge, 1971). So the r.m.s.e. of the stations is about $\pm$ 0.03 mgal.

In 1970, 31 field stations established at earlier measuring periods, have been reoccupied. From the differences, a r.m.s.e. of $\pm$ 0.07 mgal has been calculated. As this value includes also errors in position and height identification as well as gravity variations caused by local mass shifts, the result is consistent with the value estimated above.

3.3. Positioning

Position of the gravity stations has to be determined for mapping and (geographical latitude) for calculating normal gravity values. Due to economical considerations, positioning was performed by identifying the gravity stations in the topographical maps 1:100,000, published by "Landmaelingar Islands". Identification was carried out in the field, using local structures and distances to identified points, determined by the car kilometre recorder (100 m-reading) or by stepping. The accuracy of the geographical coordinates obtained by this method, has been evaluated in the eastern part of the main profile between Grimsstadir and Hof, where positions were determined by map identification and by tacheometric methods. R.m.s.e. of $\pm 0\overset{\prime}{.}1$ in latitude and $\pm 0\overset{\prime}{.}2$ in longitude, corresponding to approximately ± 150 m, have been found (Schleusener *et al.*, 1974).

3.4. Height Determination

Due to the high accuracy of gravimeter measurements, the accuracy of gravity anomalies depends mainly on the reduction quantities, especially on the station height. At regional surveys, barometric levelling is the most economical method for height determination, as it is analogue to the gravimetric process. Reading of barometer, thermometer and hygrometer takes about the same time as gravity meter readings. If the horizontal pressure gradient, time variations of pressure and instrumental drift are sufficiently taken into account, height differences at extensive surveys can be determined barometrically with r.m.s.e. of $\pm 2 \ldots 5$ m, depending on the difference and on spacing of control points (Bachem *et al.*, 1972).

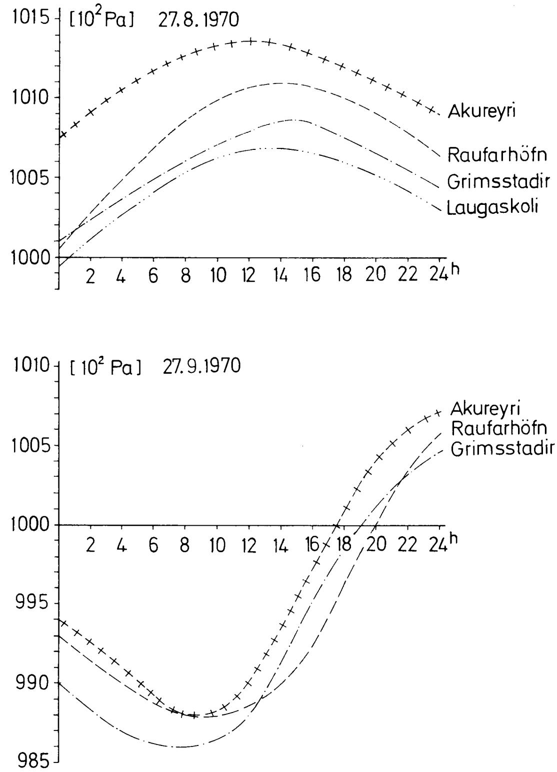

Fig. 3. Time variations of air pressure (example)

In 1965 and 1967, one Thommen and three Paulin altimeters were used for height determination. In 1970, a group of 3 Thommmen altimeters type 3B4 has been operated, in order to control readings and to improve accuracy. Pressure variations with time have been recorded by a Lambrecht microbarograph installed at a central point (Laugaskoli). In addition, pressure records of the meteorological stations at Akureyri, Grimsstadir and Raufarhöfn were available. Similar to the gravity measurements, barometric levelling started and ceased at a reference station with known height.

Absolute heights were available for the bench marks of a high precision geometric levelling between Akureyri and Grimsstadir, carried out by Spickernagel (1966).

Before reading the altimeters at each station, the instruments were kept unclamped for 10 ... 15 min to adapt the local pressure.

At evaluation, at first the instrumental pressure correction determined at the laboratory, has been applied. A correction for time variations of pressure was determined from the pressure registrations of the meteorological stations next to the measured profile, using linear interpolation between the stations. From Fig. 3, showing some extreme pressure changes, we see that pressure behaves similarly at points 50 ... 150 km distant from each other. The different pressure level of the stations results from the different station heights. Finally a temperature and humidity reduction was applied to the height differences (Möller, 1962). Comparing the sum of the height differences in one loop with the known difference between the control points gave the misclosure, which was distributed proportional with time.

Accuracy of barometric heights depends mainly on the procedure employed for taking into account horizontal pressure gradients, time variations of pressure and instrumental drift. From analysis of more than 300 stations in about 15 profiles, r.m.s.e. of $\pm 1.3 \ldots 2.5$ m have been found for height differences of $0 \ldots 50$ m between adjacent points, observed with one altimeter. Averaging the results of three altimeters and regarding the correlation produced by identical meteorological effects, the r.m.s.e. reduces to $\pm 1 \ldots 1.5$ m. A comparison of barometric with tacheometric heights in the eastern part of the main profile resulted in a height r.m.s.e. of ± 1.2 m. Finally, repeated altimeter measurements at 21 stations, carried out in 1965 and 1970, gave a r.m.s. difference of ± 4.4 m, this value including identification errors. As a result, the height r.m.s.e. will probably not exceed ± 3 m (one altimeter) resp. ± 2 m (three altimeters).

4. Density Determinations

In order to obtain a smoothed gravity field being free from the irregular influence of the topographic masses, the observed gravity values are reduced by the Bouguer and the terrain reduction. One important error source in the topographic reduction is the uncertainty in estimating the mean density of the topographic masses.

The surface boundaries of the geological formations in the survey area are given in the geological map 1:500,000 by Thoroddsen (1905/06). After checking the existing data about rock densities in Iceland, especially in the northern part, it has been felt necessary to obtain some more information by taking rock samples and observing Nettleton-profiles. The places of observation are distributed over the survey region in order to get values for the different geological formations (Fig. 1).

4.1. Rock Weighing

Rock samples have been taken from 5 places in the tertiary basalt zone and from 15 places in the young volcanic zone. At most places more than one probe has been chosen, especially if larger differences in type occured. The samples were placed under water for 24 hrs and then weighed in water and in air by means of a spring balance. From the difference of the two weighings, the density of the rock sample was calculated. Altogether 179 samples of weight $0.1 \ldots 0.5$ kg have been measured, the mean results for the different places are given in Table 1. Within one place, density of the samples of one type varied in the range of 0.3 g/cm^3 for solid and 0.5 g/cm^3 for porous basalt lava, 1.1 g/cm^3 for tuffs and 0.4 g/cm^3 for tuff breccias. The r.m.s.e. of weighing is estimated to be $\pm$ 0.1 g/cm^3.

Table 1. Mean densities from rock weighing

No.	Place	Material	Number of samples	Density g/cm^3
Tertiary basalt zone				
1	Vadlaheidi	basalt (solid)	3	2.97
2	Ljósavatndalur	basalt (solid)	6	2.95
		basalt (porous)	3	2.64
		tuff	4	2.31
3	Krossôxel	basalt	1	2.91
		tuff	5	2.14
4	Krossoxel	basalt (solid)	1	2.93
		basalt (porous)	2	2.50
		tuff	7	2.22
5	Urdafell	basalt	11	2.89
		tuff	3	2.63
Young volcanic zone				
6	Kinnarfell	tuff (unconsolid.)	2	1.91
7	Ingjaldistadir	basalt (solid)	2	2.90
		basalt (porous)	2	2.51
8	Godafoss	basalt (solid)	2	2.88
		basalt (porous)	3	2.52
9	Reykjadalur	basalt	1	2.79
		breccia	8	2.21
10	Thorgerdharfjall	basalt (porous)	4	2.66
11	Laxárdalur	basalt (solid)	2	2.95
		tuff	2	2.10
		tuff breccia	3	1.98
12	Laxárdalsheidi	basalt (solid)	5	2.89
13	Hofstadir	basalt (porous)	2	2.42
14	Mývatn	basalt (porous)	5	2.28
15	Vindbelgjarfjall	basalt (solid)	1	2.86
		basalt (porous)	10	2.13
		hyaloclastite	8	1.72
16	Námaskard	basalt	1	2.93
		hyaloclastite	5	2.31
17	Halaskógarfjall	basalt (porous)	13	2.63
		tuff	5	2.12
		hyaloclastite (consolid.)	7	2.47
18	Búrfellsrhaun	basalt (porous)	1	2.42
19	Ásbyrgi	basalt (porous)	33	2.50
20	Tungufjöll	basalt (solid)	1	2.91
		basalt (porous)	1	2.26

As there is a large scattering of the results found at different places, it was decided to apply mean densities for the calculation of the Bouguer anomaly, using as well one common density for the survey area as a few density zones. Taking the simple mean values from Table 1 and calculating the r.m.s.e., we obtain the results given in Table 2.

These results agree sufficiently with the density values found by other authors from rock weighing. In the same area Bernauer (Schleusener, 1943, p. 144/145)

Table 2. Mean densities for rock formation from weighing

Rock formation	Number of places	Density g/cm^3	Range g/cm^3
tertiary basalt lavas	7	2.83 ± 0.07	2.5–3.0
tertiary tuffs	4	2.33 ± 0.11	2.1–2.6
quarternary basalt lavas	19	2.65 ± 0.06	2.1–3.0
hyaloclastite rocks	8	2.10 ± 0.08	1.7–2.5

obtained for

tertiary basalts: density 2.77 g/cm^3 (2.3–3.2), $n=10$
palagonite tuffs: density 2.20 g/cm^3 (2.1–2.5), $n=8$.

A very detailed study of Hospers (1952) carried out in parts of the main profile, gave for

basalt lava: density 2.84 g/cm^3, $n=48$
palagonite tuff: density 2.24 g/cm^3, $n=1$
moraine matrix: density 2.24 g/cm^3, $n=8$
sandstone: density 2.13 g/cm^3, $n=9$
unconsolidated sand: density 1.80 g/cm^3, $n=1$.

Einarsson (1954) used variable densities at the calculation of Bouguer anomalies for Iceland, the values varying between 2.0 and 2.9 g/cm^3. In the central parts of Iceland mainly the density 2.6 g/m^3 has been applied. For basalts he gives the values 2.7 ... 3.0 g/cm^3, for tuffs 2.2 g/cm^3.

Taking these results into account and regarding, that a precision of 0.01 g/cm^3 in density would be unrealistic for a mean value, we may round off the values given in Table 2. Thus we obtain as mean values from rock weighing:

tertiary basalt lavas: density 2.8 ± 0.1 g/cm^3
tertiary tuffs: density 2.3 ± 0.1 g/cm^3
quarternary basalt lavas: density 2.6 ± 0.1 g/cm^3
hyaloclastite rocks: density 2.1 ± 0.1 g/cm^3

4.2. Nettleton-Profiles

Nettleton-profiles (Nettleton, 1939) have been observed and calculated at some places with sufficient height differences. Clear results were obtained at 8 places, they are given in Table 3. Some examples are shown in Fig. 4.

Taking simple means, we obtain the following Bouguer densities for different structures from Nettleton-profiles:

basalt mountains (basalts and tuffs):
tertiary basalt zone: density 2.7 g/cm^3
young volcanic zone: density 2.6 g/cm^3,
tuff volcanoes (hyaloclastite rocks): density 2.0 g/cm^3.

Table 3. Density values from Nettleton-profiles

No.	Place	Material	Density g/cm^3
Tertiary basalt zone			
I	Vadlaheidi (basalt mountain)	basalt, tuff	2.7
Young volcanic zone			
II	Múlaheidi (basalt mountain)	basalt, tuff	2.8
III	Thorgerdharfjall (basalt mountain)	basalt, tuff	2.5
IV	Vindbeljarfjall (tuff volcano)	hyaloclastite	1.8
V	Hverfjall (tuff volcano)	hyaloclastite	1.8–2.0 (6 profiles)
VI	Námaskard (tuff volcano)	hyaloclastite	2.05
VII	Hrutafjöll (tuff volcano)	hyaloclastite	2.15
VIII	Ásbyrgi (basalt mountain)	basalt, tuff	2.53 (2 profiles)

4.3. Final Density Assumptions

The results found from rock weighing and from the Nettleton profiles have been compared with the values given by Pálmason (1963, 1971) from *P*-velocities obtained at refraction seismic work in Iceland.

For the near surface rocks (recent lava flows, hyaloclastite tuffs and breccia) of the young volcanic zone (layer o) the density range 2.1 ... 2.5 g/cm^3 (mean 2.3 g/cm^3) is proposed by Pálmason. For the density of layer 1, consisting of tertiary and quarternary basalt lavas with intercalated sediments and tuffs, a value of 2.6 g/cm^3 is given. Layer 2 and 3 do not rise above sea level, so they are of no interest for gravity reduction. The mean values for layer o and layer 1 agree sufficiently with the results found in the survey area.

It was decided to carry out one reduction process with uniform density 2.6 g/cm^3, and a second one using three density zones:

tertiary basalt zone: density 2.7 g/cm^3
intermediate zone: density 2.5 g/cm^3
central volcanic zone: density 2.3 g/cm^3.

The boundaries of zones have been determined from the geological map 1:600,000 and from Pálmason and Saemundsson (1974). The gravity stations situated in the eroded and sediment filled valleys of the coastal region have been reduced with the lower density 2.2 g/cm^3, assuming that the unconsolidated material has a considerable thickness (Fig. 1). This procedure has been performed also by Hospers (1952).

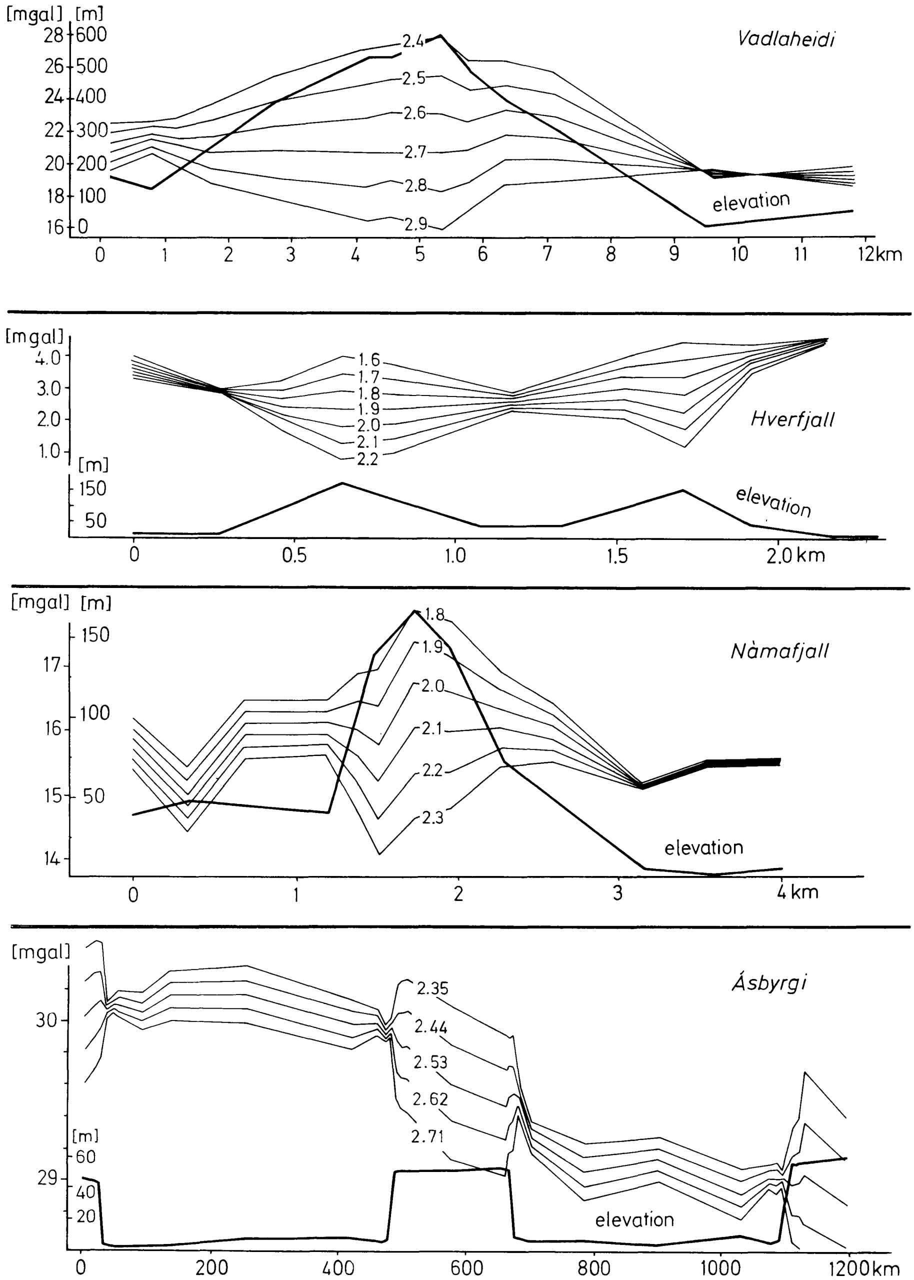

Fig. 4. Nettleton-profiles (examples)

5. Calculation and Accuracy of Bouguer Anomalies

5.1. Reference Systems

The Bouguer anomaly is defined by

$$\Delta g'' = g - \gamma_o + \delta g_F - \delta g_{\mathrm{Top}},$$

where g is the observed gravity, γ_o the normal gravity at sea level, δg_F the free air reduction and δg_{Top} the topographical reduction, taking into account all the masses above sea level. g and γ_o depend on the gravity reference system and on the reference ellipsoid. Reference for the gravity values of the survey is the Potsdam gravity system (see section 3.2). For normal gravity, the international gravity formula of 1930 has been used:

$$\gamma_o = \gamma_a(1 + \beta \sin^2 \phi + \beta_1 \sin^2 2\phi)$$

with

$$\gamma_a = 978{,}049 \text{ mgal}, \qquad \beta = 0.0052884, \qquad \beta_1 = -0.0000059,$$

ϕ being the geographical latitude.

5.2. Gravity Reductions

The free air reduction is given by

$$\delta g_F = -\frac{\partial g}{\partial H} H$$

(H=height above sea level). As actual vertical gravity gradients $\partial g/\partial H$ have not been determined, a mean normal vertical gravity gradient of -0.3086 mgal/m has been used. The topographical reduction has been split up—as usual—into the simple Bouguer reduction δg_B and the terrain correction δg_T:

$$\delta g_{\mathrm{Top}} = \delta g_B - \delta g_T.$$

The Bouguer reduction has been calculated from gravitation of an infinite plane slab:

$$\delta g_B = 2\pi G \rho H$$

with $G = 6.67 \cdot 10^{-8}$ cm^3 g^{-1} s^{-2} being the gravitational constant and ρ the Bouguer density. The terrain correction was calculated from planar formulas up to a distance of 50 km, using the templates of concentric circles and radii given by Schleusener (1940). The error produced by this procedure will probably not exceed 0.1 mgal (Drewes, 1974). For the inner zone (0 ... 100 m) a two-dimensional mass distribution was assumed, the effect of gravitation being estimated in the field. Mean heights were estimated from the maps 1:750,000, 1:250,000, 1:100,000, 1:50,000, published by "Landmaelinger Islands".

From repeated determinations, the r.m.s.e. of the terrain correction is estimated to be $\pm$ 0.2 mgal. This estimation contains only the errors of the volume integration, and not the density errors.

5.3. Accuracy of Bouguer Anomalies

The r.m.s.e. of the Bouguer anomalies depends on the errors of the quantities gravity g, geographical latitude ϕ, height H, vertical gravity gradient $\partial g/\partial H$, Bouguer density ρ, and terrain correction δg_T. Introducing the reduction terms into the formula for the Bouguer anomaly $\Delta g''$, differentiating this expression and applying the law of error propagation, gives the m.s.e. of $\Delta g''$:

$$m^2_{\Delta g''} = m^2_g + (\gamma_a \beta \sin 2\phi)^2 m^2_\phi + \left(\frac{\partial g}{\partial H} + 2\pi G\rho\right)^2 m^2_H + H^2 m^2_{\partial g/\partial H} + (2\pi GH)^2 m^2_\rho + m^2_{\delta g_T}.$$

From the Bouguer anomalies calculated with uniform density, we obtain a summed up information about all mass anomalies underneath the observation points. The Bouguer density introduced may be regarded as an errorless reduction quantity. The same is valid for the vertical gravity gradient. The r.m.s.e. of the other quantities mentioned above, has been estimated in the previous sections:

$$m_g = \pm 0.03 \text{ mgal}, \quad m_\phi = \pm 0\rlap{.}'1, \quad m_H = \pm 2m, \quad m_{\delta g_T} = \pm 0{,}2 \text{ mgal}.$$

With these values, we obtain the height independent r.m.s.e. of the Bouguer anomalies

$$m_{\Delta g''} = \pm 0.5 \text{ mgal},$$

the main influence resulting from the height errors.

When we intend to represent mass anomalies underneath a reference level, usually "mean sea level", we have at first to carry out a geological preinterpretation or, at least, to introduce some generalized density assumptions for the formations above sea level. Density and vertical gravity gradient have then to be introduced as quantities with errors. For density we may assume $m_\rho = \pm 0.1$ g $\cdot$ cm^{-3}, while for the gradient no estimate is available in the survey region. From the values given below, we find that the height dependent influence of density errors in the r.m.s.e. of the Bouguer anomaly, surpasses the height independent terms already at heights of about 200 m:

H	0	100	200	500	1,000	m
$m_{\Delta g''}$	$\pm$0.5	$\pm$0.6	$\pm$1.0	$\pm$2.1	$\pm$4.2	mgal

For error analysis between adjacent points, the correlation behaviour of the densities and the vertical gradient must be known, which generally is not available.

From the values given above, we see that for the survey discussed here, a contour interval of 1 mgal for the isoanomalies in the Bouguer maps along the profiles seems to be reasonable. Using an interval < 1 mgal will produce gravity structures, which are not significant.

6. Description of the Bouguer Anomalies Field

Bouguer gravity anomalies have been mapped in scale 1:250000, the contour interval of the isoanomalies being 1 mgal. Reduced scale maps with 2 mgal-interval (for readability) are given in Figs. 5 and 6. In regions with large profile spacing, interpolation errors will influence the position of the contour lines, and produce errors of the isoanomalies being greater than the point errors. At first, a uniform Bouguer density 2.6 g cm^{-3} has been used, corresponding to the main density value of Einarsson (1954) and representing an average density value in the survey region (see Section 4.3).

The main feature of the gravity field is the well-known gravity decrease to the centre of Iceland, producing a north-south gravity gradient of $-$ 0.4 ... 0.5 mgal/km in the survey area. The gravity bowl of 70 ... 80 mgal from the coasts to the islands centre found and discussed by Einarsson (1954), is generally explained now by inhomogenities in the density of the upper mantle, only a minor part of the variation being contributed by variations in crustal thickness (Pálmason and Saemundsson, 1974). This regional trend covers and distorts smaller gravity structures in the survey region, thus making their detection and interpretation more difficult.

Passing from west to east, we find relative gravity minima at the Eyjafjördur (a few mgal) and in the tertiary flood basalts southeast of Akureyri (about 5 mgal). East of the Bardardalur, at latitude 65°40′ to 65°50′ N, a west-east inclined gravity plain is to be seen, which stretches approximately to the river Jökulsá á Fjöllum (22 ... 16 mgal). Along the main profile (65°40′ N) there is a negative structure of about 5 mgal approximately 15 ... 25 km east of lake Myvatn, at the central part of the young volcanic zone. To the east, the Bouguer gravity anomalies increase to about 40 mgal at Vopnafjördur. At the southern part of the Tjörnes peninsula, a rather high gravity gradient of about 1.3 mgal/km is found. From the seismic refraction profile across northern Iceland, given by Pálmason (1963, 1967), we may suppose that the relative minimum in the young volcanic zone is produced by the thickening of the light layer 0 (quarternary volcanic rocks) and the corresponding dipping in of layer 1 (upper part of flood basalts). Smaller gravity structures might be due to local near surface mass anomalies, but as the resolution of the survey is of the order of 1 mgal (see section 5.3) they should be regarded with caution.

As the investigations about rock density have shown, regional density differences are probable (see Section 4.3). Therefore another Bouguer anomaly map has been drawn, using the density zones 2.2, 2.3, 2.5, 2.7 g cm^{-3} (Fig. 6). While the regional behaviour of the gravity field does not vary much, local structures are partly changed. For instance, the relative minimum at Eyjafjördur is strongly reduced, whereas the minimum southeast of Akureyri is more pronounced. Altogether, the gravity picture seems to be more quiet now, thus justifying the introduction of density zones, for reducing the influence of mass anomalies above sea level.

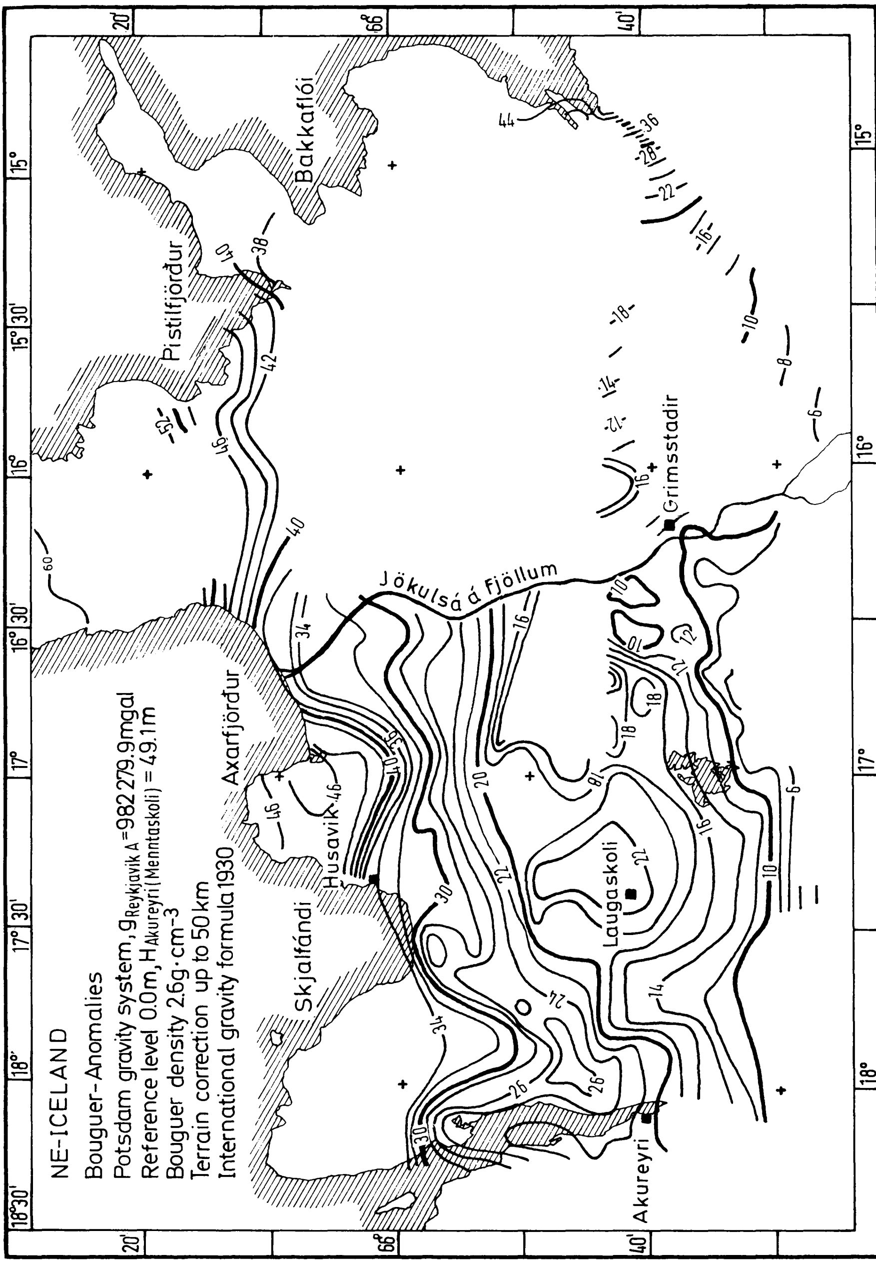

Fig. 5. Bouguer gravity anomalies, uniform density

7. Conclusion

In the present report, the results of a regional gravity survey in northeastern Iceland are given. Special attention has been turned to accuracy investigations of gravity, position and height determination and to evaluation of Bouguer

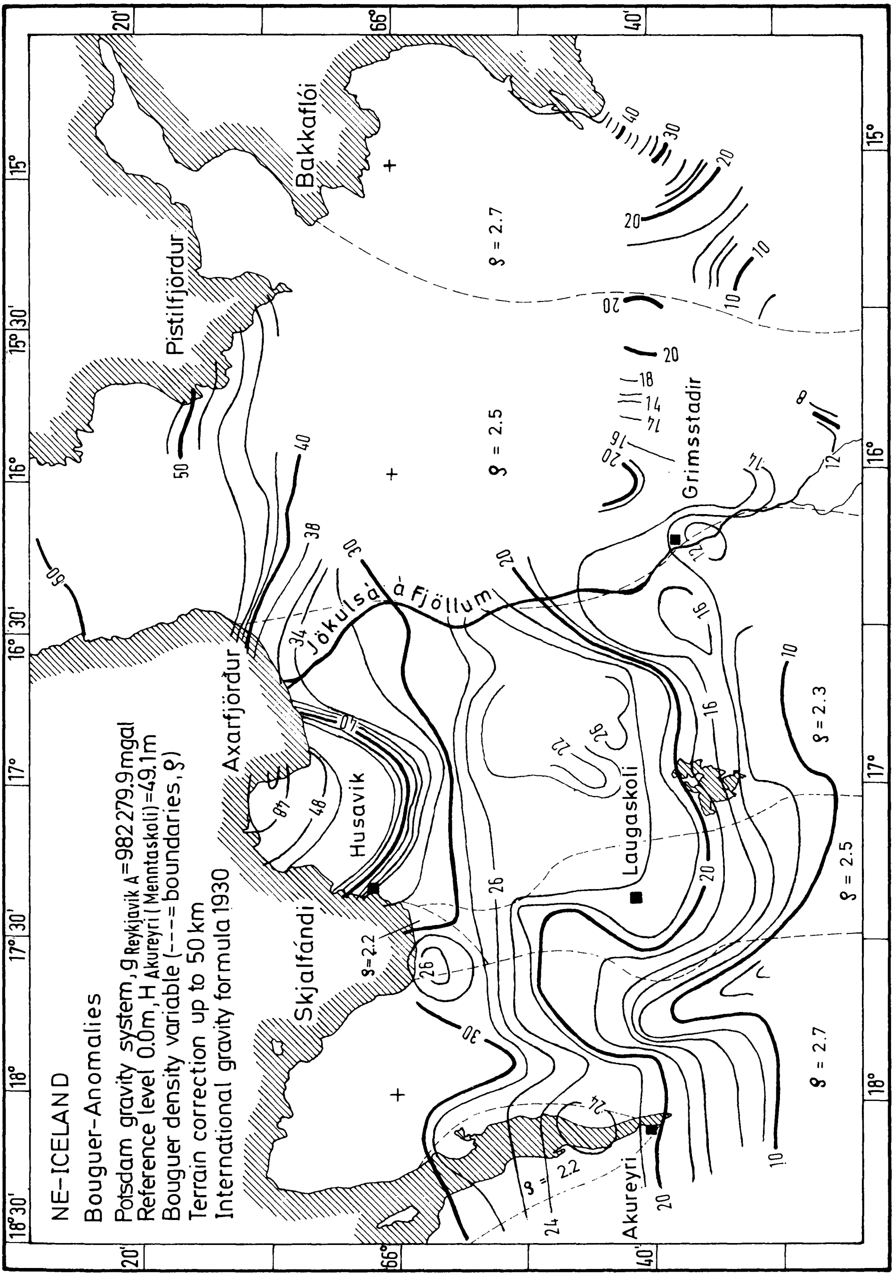

Fig. 6. Bouguer gravity anomalies, density zones

densities. The accuracy of the Bouguer anomalies derived is about ± 1 mgal. While rock densities show a rather large scattering, Bouguer densities may be comprehended to density zones. The Bouguer gravity field is characterized by the well-known decrease towards the island's centre, superposed by some smaller structures of a few mgal amplitude which are partly correlated with the young volcanic zone.

The results of this survey especially give a detailed picture of the gravity field along the western and the central part of the monumented gravity profile Akureyri-Grimsstadir-Hof, established for investigations of gravity variations with time. These results contribute with some details to the gravity map of Iceland, based on a regional gravity survey with about 10 km average point spacing, which started in 1968 by the National Energy Authority, Reykjavik (Pálmason *et al.,* 1973). As one of the seismic refraction profiles measured by Pálmason (1967) crosses the region of the gravity survey, the results may be used for combined investigations about the crustal structure. Finally, regional type calculations in gravimetric geodesy could be supported by the observed data.

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