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PETROLOGY AND STRUCTURE OF THE ESJA QUATERNARY VOLCANIC REGION, SOUTHWEST ICELAND

by

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ABSTRACT

The stratigraphy of Esja is described and the chronology of the succession established by means of the geomagnetic time scale. The Esja volcanic succession is about 2.4km thick, and comprises olivine tholeiites (25%), tholeiites (68%), basaltic andesites (5%), icelandites and rhyolites (2%). Volcanism was active in the Esja region for just over one million years, and during this time span, at least ten glaciations occurred in the region. The stratigraphic succession is therefore characterized by sequences of lava flows intercalated, at intervals, by thick subglacial hyaloclastite units. Two central volcanoes were active in the Esja region; the Kjalarnes volcano was active for about 0.6 million years and was succeeded after a short interval by the Stardalur volcano, which remained active for about 0.3 million years. Flood-basalt volcanism was concomitant with the central volcanism, and most of the olivine tholeiites are considered to have been erupted in fissures and shield volcanoes unrelated to the central volcanoes.

Igneous activity apparently migrated eastwards with time, reflecting the westward crustal drift away from the active volcanic zone, which is a subaerial extension of the mid-Atlantic ridge. The volcanics are tilted and downfaulted towards the east. The irregular topography created by the glaciations in Esja repeatedly prevented lavas erupted in the active volcanic zone from spreading over the tectonically less active neighbour-hood, thus producing angular unconformities in the stratigraphic

succession from which the tectonic history of the region can be read. The Esja evidence suggests that tectonic activity is chiefly restricted to the active volcanic zone, and that the crust becomes tectonically inactive soon after it has drifted away from the active zone.

Intrusive activity in Esja can be divided into three phases. The oldest dykes in the region trend N 25°E and contemporaneous sheets dip towards the Kjalarnes peninsula, where the intrusive activity culminated in the formation of a multiple dolerite sheet. This intrusion may have been preceded by a caldera collapse in the Kjalarnes area. After the intrusion of the Kjalarnes dolerites the regional trend of dykes changed to N 40°E, and a narrow dyke swarm (representing up to 20% dilation) cut across the Kjalarnes central volcano. The dyke swarm was succeeded by cone sheets focussing to the south of Leiðhamrar, and the second phase culminated in the intrusion of large dolerite sheets in Þverfell and Laugarnípa. Following a brief interval, during which flood-basalt volcanism was dominant in Esja, the Stardalur central volcano became active and, during its life span, minor intrusions were predominantly in sheet form. Caldera collapse in the Stardalur volcano was followed by the intrusion of basic cone sheets, large dolerite sheets, a sill and finally a laccolith within the caldera. Long after the caldera had been filled the caldera fault zone dominated over the regional fault pattern at depth so that basic and acid volcanics alike were erupted concentrically with and parasitically to the Stardalur caldera.

Large dolerite intrusions in Esja are found chiefly within or at the boundaries of the thick hyaloclastite units, and there is evidence of dykes cutting straight through lava successions, but spreading out laterally to form sill-like bodies once they enter the less coherent hyaloclastites. A survey of the literature shows that the majority of large basic intrusions in Iceland are accommodated in relatively soft and "structureless" host rocks, such as tuffaceous hyaloclastites, sediments, vent and caldera agglomerates, hydrothermally propylitized lavas, and "hot" and still partly liquid acid intrusive material. The majority of the large intrusions are in the form of inclined sheets, but sills and laccoliths are formed when the intrusions are emplaced at shallow levels (perhaps less than 1km) in the crust.

The coincidence of central volcanoes having a great bulk of shallow level intrusions, with positive gravity anomalies, and the sites of shallow depth to layer 3 in Iceland strongly suggests that crustal layer 3 consists mostly of basic intrusions. A comparison of the densities of primary and secondary minerals of tholeiitic rocks suggests that infilling of vesicles of porous basalt lavas by secondary minerals will not make the rock as dense as a non-porous rock of the same composition. The estimated density difference of $0.2 g/cm^3$ between crustal layers 2 and 3 can apparently not be ascribed to secondary alteration of subaerial lavas, but can readily be explained by a transition from altered lavas to non-porous intrusives. It is proposed that the sharp boundary between layers 2 and 3 results from the lavas at the base of layer 2 reaching a degree of alteration at which the rock becomes sufficiently incoherent to accommodate large basic intrusions. The "metamorphic boundary" proposed by Pálmason (1971) to explain the correlation between the thermal gradient and depth to layer 3 in Iceland is not therefore primarily a density boundary, but a boundary at which the lavas loose their strength as a result of alteration and host voluminous dense intrusives. The large scale features of crustal layer 3 in Iceland can be explained within the framework of this model.

There is a complete range in composition from olivine tholeiites to rhyolites in the Esja volcanic succession, and the majority of the rocks contain <u>some</u> phenocrysts. Crystal fractionation appears to be a feasible mechanism to explain the chemical variation within at least the basaltic rocks in Esja, but whether the intermediate and acid rocks are formed by extensive fractionation or by partial melting of crustal material cannot be answered. The apparent coincidence in time of the emplacement of large basic intrusives and the commencement of intermediate and acid volcanism in eastern Esja may suggest that the rise of voluminous basic magmas has raised the thermal gradient sufficiently to produce the intermediate and acid rocks by partial melting at the base of the crust.

Positive gravity anomalies associated with the Kjalarnes and Stardalur central volcanoes are attributed to high level intrusives in the core regions of the two centres. Specific gravity measurements of the chemically analysed rocks from Esja show a range of densities from about 2.5 (rhyolite) to about $3.15 g/cm^3$ (olivine tholeiite). Local gravity anomalies commonly found associated with central volcanoes are probably due both to local concentrations of rocks of different chemistry and to a high percentage of intrusives. A comparison of the average density of the crust in eastern Iceland and that of the crust on the Iceland-Faeroe ridge suggests that a considerable part of the negative Bouguer gravity anomaly of Iceland (Einarsson 1954) can be explained in terms of geochemical differences between the volcanics of Iceland and those of the surrounding areas. The similarity of the gravity profiles from the aseismic Iceland-Faeroe ridge and from the active Reykjanes ridge to the centre of Iceland suggests that, if a hot spot contributes to the bowl-shaped gravity anomaly of Iceland (Bott et al 1971), then it is not connected with layer 4 under the Reykjanes ridge.

A positive magnetic anomaly associated with the Stardalur caldera is explained in terms of a thick pile of normal polarity eruptives within the caldera being surrounded essentially by reverse polarity eruptives. The combined effects of a high magnetite content (which may be caused by unusually high partial pressure of oxygen in the melt) and a high palaeofield strength may cause the very high magnetic intensity of lavas, which give rise to a sharp maximum within the Stardalur magnetic anomaly. Assuming a common cause for three other strong magnetic anomalies, which, with Stardalur, lie on a straight line and are separated regularly in space and time, the possibility of a mantle controlled "high partial pressure of oxygen spot" migrating at half the spreading speed along the spreading axis is discussed.

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MAP AND SECTIONS

Geological map of Esja, scale 1:25,000, is in pocket in back cover.

Sections 1 to 7 are bound in the back.

Chapter 1 INTRODUCTION

The Esja volcanic region lies about 10km northeast of Reykjavík and about 20km west of the active zone of volcanism, which extends from the Reykjanes peninsula to the Langjökull icecap, and is the subaerial continuation of the mid-Atlantic ridge. The region is named after the mountain Esja (914m) which rises majestically from sea level forming a plateau at about 800m bounded by precipitous cliffs. The geological map covers an area of about 300km^2, and is largely confined to the mountain. Esja is a part of a continuous volcanic succession extending from the Borgarfjörður anticline to the active volcanic zone (Sæmundsson 1967a).

The primary aim of the research has been to establish the stratigraphy and structure of Esja, which is an exposed continuation of the rock formations underneath and east of Reykjavík from which natural hot water is exploited by drilling and utilized for domestic heating in the capital. The potential hot water resources to the south of Esja may be enough to supply half the Icelandic population, which inhabits the towns of Hafnarfjörður, Kópavogur and Reykjavík.

a. Previous work

Despite the short distance from Reykjavík to Esja very little detailed geological research has been carried out in the mountain. Although poets have praised its beauty, and it has been the scene of an abortive gold rush, and, more fruitfully, some calcite mining, geologists (except for Rutten 1958, see later) and travellers have made only scanty observations on the mountain, and it has not been geologically mapped in detail before.

The pioneer geologist Thoroddsen visited Esja and noted the rhyolite peaks of Móskarðshnúkar in the eastern part of the mountain, some hyaloclastite intercalations in the basalt succession, and remarked on the change in dip within the mountain, the steeper dips occurring in its lower slopes (Thoroddsen 1913). He showed Esja as the southernmost part of the Tertiary rocks of western Iceland on his geological map (Thoroddsen 1901). Pjeturss (1910) recognised glacial vestiges, and considered the top third of Esja to be of Pleistocene age. Rutten (1958) similarly drew the Tertiary-Pleistocene boundary at the base of the top plateau. Kjartansson (1960) showed the whole of Esja as the southernmost extension of the Tertiary flood basalts on his geological map (scale 1: 250,000) of southwest Iceland. Until the present study was undertaken, Esja, apart from the top plateau, had been considered to be of Tertiary age (e.g. Þ.Einarsson 1968).

Rutten (1958) included Esja in his reconnaissance geological mapping of about 600km^2 in southwest Iceland carried out in one summer. Considering the large area covered in a short time, Rutten's mapping is a considerable achievement. He divided the stratigraphic column of Esja into three series; the strongly tilted and hydrothermally altered plateau basalt series PB, the slightly tilted and fresher intermediate series IS, and the nearly horizontal and fresh "Graue Stufe" series GS. Although this threefold division is apparent in a reconnaissance study, more detailed work shows it to be untenable, as tilting occurs mainly within the active zone of volcanism and the same lava series may be steeply tilted in the active zone but horizontal in a tectonically inactive area some 20km away. Little use was made of Rutten's reconnaissance work in the present study.

The most significant contribution to the stratigraphy of Esja and indeed that of southwest Iceland was the palaeomagnetic mapping carried out by Einarsson and Sigurgeirsson with the use of field compass (Einarsson 1957a,b, Sigurgeirsson, 1957). Their palaeomagnetic map covers an area of some 900km^2, and includes the northern part of Esja from Hrútadalur eastwards and the upper slopes of Kistufell in southern Esja. They excluded the structurally more complex areas of western Esja and the Stardalur region. The more detailed palaeomagnetic mapping carried out in the present work confirms in all main respects their mapping, although the exact boundaries of the palaeomagnetic groups have in many instances been revised.

Piper (1971) interpreted and extended somewhat the palaeomagnetic maps of Einarsson (1957a,b,1962) and Sigurgeirsson (1957). His extensions in the Esja area, particularly that of the normal polarity group N3 of Einarsson and Sigurgeirsson, are, however, inaccurate. In correlating the palaeomagnetic stratigraphy column with the geomagnetic time scale of Cox (1969), Piper used evidence from the K/Ar age of Hafnarfjall (Moorbath et al 1968), intercalations of glacial tillites in the succession, and the estimated mean rate of eruption, in addition to assuming that the youngest reverse polarity rocks were of the last reverse period of the Matuyama epoch. New K/Ar age determinations and the detailed stratigraphic work of Sæmundsson in southwest Iceland broadly confirms Piper's correlation, although the short magnetic events are interpreted differently (Noll and Sæmundsson 1973, Fig.3 of Pálmason and Sæmundsson 1973). According to Piper's correlation most of Esja was formed during the lower part of the Matuyama reverse polarity epoch, and the present author agrees, except that the short normal polarity events N3 and N2 (Einarsson 1957) are considered to represent the Olduvai and Gilsá events respectively, and not the Gilsá and Jaramillo events as suggested by Piper (1971). The problems of correlation are discussed more fully on page 12.

Sigvaldason (1958) studied the Móskarðshnúkar rhyolite. Theódórsdóttir (1972) studied the geology of Grímmannsfell, which is on the southern boundary of the Esja geological map. The lowlands south of Esja have been mapped geologically by Tryggvason and Jónsson (1958), and Jónsson (1960,1972) has described glacial tillites near Reykjavík and the distribution of late interglacial (referred to as "post-erosional" in the present text) lavas south of Esja. Peacock (1926) described the geology of the island of Viðey. Einarsson (1962) has made miscellaneous geological and geomorphological studies in the areas surrounding Esja. Several groups of students from the University of Iceland, supervised by Kristján Sæmundsson, have in the summers 1971 and 1972 mapped in detail areas north of and adjacent to the Esja region.

Relatively detailed geophysical surveys have been made of the Esja region revealing strong gravity (Einarsson 1954), seismic delay time (Pálmason 1971) and magnetic (Sigurgeirsson 1970) anomalies in the region.

Short accounts on some aspects of the present study have already been published (Friðleifsson and Kristjánsson 1972, Friðleifsson 1973a,b).

b. Field work

Geological mapping was carried out during a total period of nine months in the summers of 1970, 1971 and 1972. Aerial photographs and topographical maps (scale 1: 25,000, 10 m contour lines) were supplied by the Icelandic Survey Department. The rocks were divided into mappable units on the basis of lithology and palaeomagnetic polarity directions. At the early stages of the work several detailed near vertical traverses were made from the lowlands to the top plateau series of the mountain and the polarity direction of every exposed lava flow or hyaloclastite measured with a portable fluxgate magnetometer. As the work progressed, however, and the polarity groups had been established, shorter profiles were made and the boundaries traced laterally. The extrusive rocks on the geological map are divided into olivine tholeiites, tholeiites, basaltic andesites and rhyolites; the field characteristics of Icelandic lavas of the tholeiitic series have been described by Walker (1959). An account of the classification is given on page 110. The division between olivine tholeiites and tholeiites is mostly made in the field. There is, however, complete gradation between the rock types, and borderline cases are therefore common. It appears, from a number of samples of borderline lavas examined in the laboratory, that the main shortcoming of the field identification is that fine-grained olivine tholeiite lavas can easily be misidentified as tholeiites. The estimated volume ratio of olivine tholeiite to tholeiite may therefore be slightly too low. Porphyritic rocks were not mapped separately, although they were sometimes used locally as marker horizons; they have been designated as tholeiites when they are plagioclase and/or pyroxene-phyric and as olivine tholeiites when olivine-phyric. Samples from all the basaltic andesite and rhyolite units shown on the map have been petrographically and chemically analysed. The term hyaloclastite is used loosely to cover all subaquatic (including subglacial) eruptives and sediments; a subdivision is made in the text. The bulk of the hyaloclastites shown on the geological map are subglacial eruptives, and the sediments are largely resedimented subglacial eruptives, except for the main detrital horizon in Esja, which is marked separately on the geological map. A division is not made between olivine tholeiite and tholeiite hyaloclastites on the map, but where practicable, such a division is attempted in the text. Most of the hyaloclastites are of tholeiite.

The thickness of the geological units was measured with a yardstick and/or altimeter, and the altitudes of most geological boundaries shown on the map have been checked in one or more places with an altimeter, the most notable exception to this being the inner part of Eilífsdalur valley, where the altitudes were estimated from landmarks identified on the map and aerial photographs.

A coordinate scheme with 2km grid size has been marked on the edge of the geological map in order to make it easier to locate individual geological units described in the text. The main cross section is drawn at the base of the map and its location is shown by a solid line on the map. The lines of the other seven cross sections (bound in the back) are, however, not marked on the map, but the coordinates of the end points are given on each section; all the cross sections are drawn along straight lines, except Section 7, and all but Section 7 (the Stardalur caldera section) have the same vertical and horizontal scales.

c. Outlines of the geology of Esja

The volcanic succession of Esja is about 2.4km thick. The relative proportions of the rock types is estimated to be:

Olivine tholeiite	25%
Tholeiite	68%
Basaltic andesite	5%
Rhyolite and icelandite	2%

The olivine tholeiite is thought to be erupted mainly in fissures and shield volcanoes representing the flood-basalt volcanism in the active spreading zone, whereas the tholeiite is considered to be predominantly, and the more evolved rocks entirely, the products of the two central volcanoes in Esja, which are named after Kjalarnes in the west and Stardalur in the east. The base of the Kjalarnes central volcano is not known with certainty, but unit 3 in the Esja stratigraphic column (Table 1) is assigned to the volcano. The central region of the volcano is entirely eroded, and the core lies mostly below sea level. The rocks in southwest Esja are believed to represent about 60^{o} of arc of the rim of the Kjalarnes volcano, and are considered as essentially the products of a dyke swarm associated with the centre. Interdigitated with the tholeiite volcanics are olivine tholeiites erupted outside the region of central activity. Olivine tholeiite volcanics were dominant after the Kjalarnes volcano ceased to be active and until central volcanism started in Stardalur. The Stardalur volcanic centre is relatively well exposed, and all the intermediate and acid volcanics that have been found in Esja are associated with the Stardalur centre.

Walker (1963) described Tertiary volcanism in eastern Iceland as a contest between flood-basalt and central volcanicity. A central volcano would sometimes succeed in creating a topographic cone that stood up above the surrounding plains, over which lavas from the volcano spread widely; at other times the flood-basalts would overlap, more or less burying the volcano. Glaciations, of course, alter the battlefield and tend to isolate the combatants. Subglacial volcanics tend to pile up around the eruptive orifice, and thus produce a much greater relief in the topography than do subaerial lavas. Erosion also becomes much more rapid, and less controlled by tectonic features. These general principles can be seen to have operated in the Esja region.

In southwest Iceland the earliest tillite is found in rocks about 3 million years old (Noll and Sæmundsson 1973), and from that time to the present there have been many glaciations with intermittent warmer periods. As the volcanism in southwest Iceland has been continuous from the upper Tertiary and throughout the Quaternary, the mountains give an almost unique opportunity to compare volcanic products of subaerial and subglacial eruptions. After a glaciation, hyaloclastite ridges and hills with steep slopes are the dominant features of the topography. In subsequent volcanic eruptions lava flows bank up against the hyaloclastites or flow down their slopes depending on the eruptive site. The lavas may fill the valleys between the hyaloclastites and eventually bury them.

Due to the rugged topography after a glaciation, valleys, a short distance apart, may be physically isolated from one another. Thus, an absence of volcanism in one valley may allow the development of aprons of sediments spreading out over the lava plains at the feet of the easily eroded hyaloclastite mountains, while simultaneously, active volcanism in an adjacent valley may give rise to a pile of lavas with no sedimentary intercalations. The overall effect of this, as seen in the valley sections in Esja today, is a "cedar-tree" structure with a bulky "stem" formed of mainly primary hyaloclastites with thin (tens of cm to tens of m), wedge shaped "branches" of resedimented hyaloclastite, intercalated in the lavas submerging the "stem". The number of the "branches" depends on the rate of erosion of the "stem" and the rate of eruption of lavas submerging the "stem".

*

The palaeomagnetic polarity of the rocks gives an invaluable control on stratigraphic correlations in an area of such variable geology. There were at least five and possibly seven

Table 1 **Esja stratigraphic column**

Epoch	M.Y.	Event	Unit	Rock type	Thickness (m)	Intrusions	Central volcano
MATUYAMA	1.64		26	Rhyolite hyaloclastite	100		STARDALUR CENTRAL VOLCANO
	N	Gilsá	25	Olivine tholeiite hyaloclastite (outside area mapped)			
	1.79		24	Tholeiite lavas	280		
			23	Minor rhyolite and basaltic andesite hyaloclastite	10-240		
	R		22	Tholeiite lavas	10-100		
	1.95		21	Basaltic andesite hyaloclastite (caldera lake)	($2.5 km^3$)	Stardalshnúkur,	
			20	Tholeiite hyaloclastite (caldera lake)	($0.7 km^3$)	Gráhnúkur and	
		Olduvai	19	Tholeiite and olivine tholeiite lavas	10-200	Stardalur cone sheets	
	1.98		18	Basaltic hyaloclastite	5-300		
	2.11		17	Tholeiite lavas	10-50		
	2.13	Reunion	16	Olivine tholeiite lavas	10-200	Sheets dip towards	
			15	Basaltic hyaloclastite	10-200	Stardalur.	
			14	Olivine tholeiite lavas and hyaloclastite	50-80	Þverfell and Laugarnípa	
	R		13	Tholeiite lavas	20-400	Leiðhamrar	KJALARNES CENTRAL VOLCANO
			12	Basaltic hyaloclastite	30-300	Sheets dip to Þverfell	
			11	Tholeiite lavas	10-240		
	2.43		10	Basaltic hyaloclastite	25-200	Dykes trend N 40^{o}E	
GAUSS			9	Tholeiite lavas	70-110	Kjalarnes intrusion	
			8	Tholeiite lavas	60-80	Sheets dip to Kjalarnes	
	N		7	Basaltic hyaloclastite	20-70		
			6	Olivine tholeiite lavas	10-200	Sill in Melafjall	
			5	Tholeiite lavas	30-70	Sill east of Ártún	
			4	Basaltic hyaloclastite	70-200		
	2.80		3	Tholeiite lavas	70-100	Dykes trend N 25^{o}E	
	R	Kaena	2	Olivine tholeiite lavas	125		
	2.90		1	Tholeiite lavas	200		

reversals of the Earth's magnetic field during the time the volcanic rocks of the Esja region were erupted. The correlation of the Esja stratigraphic column with the geomagnetic time scale of Cox (1969) is shown in Table 1. The correlation is based mainly on the very thick hyaloclastite units found in Esja; these could not have been produced in eruptions under local glaciers, and the thick pile of reverse polarity rocks with very much thinner normal polarity intercalations were therefore most probably erupted during the Matuyama reverse polarity epoch, which had several short normal polarity events (Watkins 1972). The base of the Esja stratigraphic succession is correlated with the top of the Kaena reverse polarity event of the Gauss epoch, as a few reverse polarity lavas are found by the sea west of Esja at the base of the thick normal polarity succession (units 1-8). Piper (1971) correlated the N3 normal polarity group (Einarsson 1957) with the Gilsá normal polarity event, and suggested that the Olduvai event may have been too short to be represented in the volcanic succession. This is considered very unlikely since the two central volcanoes are very close in space and no hiatus is found in the flood-volcanism between them. An estimate, based on the field work, of the total number of eruptions in Esja (lava flows, plugs in hyaloclastites etc.) is of the order of 400 but must be significantly higher in view of the total volume of Esja. Using this figure and the correlation with the geomagnetic time scale shown in Table 1, the Esja volcanic succession was formed in about 1 million years, giving an average rate of approximately one eruption in every 2500 years. Even using

Piper's correlation gives one eruption every 4600 years, which is still very high compared with the mean rate of eruption in southwest Iceland estimated (Piper 1971) at 27000 years.

Correlation similar to that in Table 1 has been made by Sæmundsson for the Húsafell area some 60km NNE of Esja; K/Ar dating has shown the N2 group (Einarsson 1957) east of Húsafell to be 1.8 $\pm$ 0.4 million years (Noll and Sæmundsson 1973) which correlates fairly well with the Gilsá event (Cox 1969).

Two lavas (page 43) near the top of unit 13 have normal magnetic polarity. These may represent the very short Reunion event (Table 1). If this correlation is correct then Kistufell (G4), where the two lavas are found, will be the first locality in the world where the Reunion and Olduvai events are recorded in a continuous section, and Esja will moreover be the first area where the Reunion, Olduvai and Gilsá events are all recorded in an unambiguous succession.

*

There are eight main hyaloclastite horizons in the Esja stratigraphic succession. Each of these is considered to represent a major glaciation. In the area just north of Esja two more glacial horizons* are found in Esja stratigraphic units 1 and 2; the horizon in unit 1 may be represented near

* The absence of thick subglacial volcanics representing these horizons in Esja suggests that the Kjalarnes central volcano was not yet active; unit 3 is therefore the first unit assigned to the volcano.

Bakki farm (B7) (page 17), but the horizon in unit 2 is not exposed in Esja. If these are included, then four glacial horizons have been recognized in rocks from the Gauss epoch above the Kaena event. This is the same number as are found in Húsafell (Noll and Sæmundsson 1973). There are at least six main glaciations represented in Esja in Matuyama rocks until the Gilsá event. From the Kaena event to the Gilsá event there are therefore at least ten glaciations, an average of one every 110,000 years.

*

The volcanism in Esja apparently migrated eastwards with time, reflecting the westward crustal drift away from the active volcanic zone. The rate of crustal drift in southwest Iceland since the upper Tertiary is estimated at about 1.2 cm/yr (Piper 1971). Tectonic activity is and has been chiefly restricted to the active volcanic zones in Iceland and this is very clearly demonstrated in Esja. A fairly continuous change in dip with stratigraphic level can be expected where subaerial lavas flow long distances and overlap freely the older rocks; in the Tertiary succession of eastern Iceland the dip falls from near 8^{o} at sea level to about 4^{o} at the mountain summits (Walker 1964). Angular unconformities are, however, produced when overlapping is interrupted for a period of time long enough to allow tilting, as occurs in Esja where the irregular topography created by the glaciations prevented lavas erupted in the active volcanic zone from spreading over the tectonically less active neighbourhood. It is apparent in Esja (e.g. units 16, 19 and 24) that most of the tilting occurs within less than a quarter of a million years after the rocks are erupted, i.e. within the active zone. This time span is, however, probably much shorter than average, because of the high rate of eruption and consequently subsidence in the Esja region and the tectonics associated with the central volcanoes. Depositional dip away from the active volcanic zone may also obscure or obliterate regional tilting.

Faults are numerous in the Esja region; if there were continuous exposure around the mountain there would probably nowhere be a 1km long NW-SE section that does not show evidence of faulting. The majority of the faults are normal faults and most have downthrows of less than 50m, although several are found with downthrows exceeding 100m. Faults with displacements of less than 10m are not shown on the geological map. The main fault movements are geologically short-lived (perhaps 10-100 thousand years), but the faults remain active for very much longer, as is demonstrated by the very small vertical displacement of the fairly numerous faults in the top plateau of western Esja.

Chapter 2. VOLCANIC GEOLOGY

a. Rocks from the Gauss magnetic epoch

The oldest rocks in the area mapped are found in the northwest, on the coast of Hvalfjörður. The rocks have a normal magnetic polarity direction and are believed to date from the uppermost part of the Gauss magnetic epoch.

The rock succession is formed of groups of olivine tholeiite and tholeiite lavas with two substantial basaltic hyaloclastite units intercalated. The groups of lava flows are fairly regular in thickness and can be traced over several kilometers. The regularity of the lava successions is reflected in the step-like topography with the tholeiites, harder and more resistive to erosive forces, forming platforms with prominent escarpments.

Unit 1. Tholeiite lavas, about 200m thick.

The stratigraphically lowest unit is a 200m thick group of tholeiite lavas which forms the eroded coastal plane along Hvalfjörður. The magnetic polarity of the lavas is normal, and they are believed to date from the beginning of the uppermost normal period of the Gauss magnetic epoch. This is supported by the occurrence of a few olivine tholeiite lavas with reversed polarity outcropping on the beach just north of Dalsmynni farm (C7). These are overlain by a 5m thick ill-sorted conglomerate, which underlies tholeiite with normal polarity. The reverse polarity lavas are believed to represent the Kaena event. A similar sequence is observed some 10 km further northeast along the Hvalfjörður coast (D9) (Friðriksdóttir et al

1972). The exposure on the beach is too thin to be shown on the geological map, but the reverse polarity group is indicated in Sections 1 and 2.

The tholeiites are best exposed south of Hjarðarnes farm and in Tíðaskarð pass (C8), where a few porphyritic lavas occur in the topmost part of this unit. One sedimentary horizon is found within the tholeiite unit; in a river section about 400m northeast of Bakki farm (B7) a conglomerate with rounded pebbles up to 15cm across is seen. The exposed thickness is about 2m. This may correlate with a 9m thick tillite reported within the tholeiite unit some 7km further northeast along Hvalfjörður coast (Friðriksdóttir et al 1972).

Unit 2. Olivine tholeiite lavas, about 125m thick.

Conformably on top of the tholeiite unit lie olivine tholeiite compound lava flows (Walker 1971). Typical olivine tholeiite flow units are exposed in the upper cliffs east of Tíðaskarð (C8) and in the lower cliffs further north along the main road to western Iceland. A number of dykes and small sills are weathered out of the olivine tholeiite lavas south of Tíðaskarð to show castle-ruin like features (Fig.6 in Rutten 1958).

The low dip of about 6^{o} on the lowland north of Melafjall compared with the dip values of 8^{o}-10^{o} higher up in the pile as well as further southwest along the strike may indicate a northwesterly depositional dip of 2^{o}-4^{o}. The source of the lavas may therefore have been a scutulum type (Noe-Nygaard 1968) shield volcano somewhere east of the present day exposures of the lavas.

Unit 3. Tholeiite lavas, 70-100m thick.

Conformably on top of the olivine tholeiite lavas comes a series of tholeiite lavas, forming a platform in the northern part of Melafjall (D8) (where it is about 70m thick), apparently thickening southwestwards to form a prominent platform in western Lokufjall (D8) (where it is about 100m thick), and forming low but prominent escarpments on the lowland south of Ártún (C7). In Lokufjall the lavas are vesicular and scoriaceous perhaps indicating proximity to the site of eruption.

Unit 4. Basaltic hyaloclastite, 70-200m thick.

The tholeiites are conformably overlain by a thick hyaloclastite unit. In Melafjall (D8) the total thickness is about 70m; the bottom 15m or so are of waterlain sediment which grades into pillow breccia and pillow lava (this sequence is observed in several outcrops from the vertical cliffs by the main road in the west to the low hummocks above Tindstaðir farm (E8) some 2.5km further northeast). A dolerite sill (chem.an.6)* with at least three feeder dykes (Fig.14, page 87) is intruded into this hyaloclastite in Melafjall. The hyaloclastite unit thickens considerably southwards (see Section 3). At the mouth of the Blikdalur valley (D7) the bottom of the unit is not seen, but the minimum thickness is over 200m. On the southern side of the valley this hyaloclastite is succeeded directly by hyaloclastite unit no.7.

* The sample number is given where an analysed rock body is specifically mentioned in the text. The chemical analyses are listed in stratigraphical-chronological order in Table 2, page 125.

Unit 5. Tholeiite lavas, 30-70m thick.

In Melafjall the hyaloclastite unit is overlain by a series of porphyritic tholeiite lavas and tholeiites which on the geological map are shown as one tholeiite unit. It is, however, the porphyritic character of the lavas in the lower part of this unit that makes it distinctive. The thickest of these porphyritic lava flows (about 20m thick) forms a prominent scarp which can be followed for more than 3km from just west of the Kerlingagil gorge to the western part of Melafjall. This lava flow makes the ledge of the waterfall just up the hill from the Melafjall sill.

The upper part of this unit is formed of tholeiites, some of which are fairly vesicular. In Melafjall the total thickness of the unit is about 70m, but it apparently thins southwards against the hyaloclastite unit 4. In the southern slopes of Lokufjall (D7) tholeiites corresponding to the upper part of this unit are exposed but none of the porphyritic lavas were observed. Southwest across the Blikdalur valley the group is absent with the possible exception of two thin vesicular lavas in the pillow lava pile at 250m altitude, 20m above a 30m thick sill (chem.an.7) which is shown on the geological map (C7). The two lavas thin southwards and change to pillow lava as if the lavas had flowed into water.

Unit 6. Olivine tholeiite lavas, 20-70m thick.

Conformably on top of the tholeiite unit 5 comes a series of olivine tholeiite flow units with a total thickness of over 70m in the Kerlingagil gorge (E8), some 50m thick in the northern Lokufjall cliffs (D8) (in both sections exposure is 100%), but only 20m in southern Lokufjall. The banking up of units 5 and 6 against the hyaloclastite unit 4 is shown in Section 3.

Unit 7. Basaltic hyaloclastite, 10-200m thick.

The olivine tholeiite series in Kerlingagil (E8) is overlain by 11m thick tuffaceous hyaloclastite. In southern Lokufjall pillow lavas and pillow breccias form a pile with a minimum thickness of 70m. At the top of the Lokufjall cliffs (D8) tholeiite lavas of unit 8 are seen banking up against the hyaloclastite slope from the west.

Units 5 and 6 are believed to thin out southwards against hyaloclastite unit 4, as mentioned above, and consequently unit 7 succeeds unit 4 south of the Blikdalur valley to form a hyaloclastite pile with a total thickness of about 400m. The top is not seen, but it may extend up to the reverse polarity tholeiite which caps the ridge with spot altitude 486m (D6). The next hyaloclastite unit above unit 7 has reversed magnetic polarity and is assigned to the Matuyama epoch. The hyaloclastites on the southern side of Blikdalur are intensely altered and hence a reliable magnetic polarity measurement cannot be made. However, intrusive dykes with normal polarity (almost certainly from the Gauss magnetic epoch) are found in the hyaloclastites up to 300m altitude and it is therefore concluded that the hyaloclastite pile of units 4 and 7 extends up to the reverse polarity tholeiite lavas in the westernmost part of the ridge south of Blikdalur.

Unit 8. Tholeiite lavas, 60-80m thick.

Tholeiite lavas have flowed from the west up against a hyaloclastite mountain, as is clearly demonstrated in the Lokufjall cliffs (see above). In the Kerlingagil gorge this tholeiite series is perfectly exposed and is 80m thick (17 lavas) with normal polarity in the western wall of the gorge, but only the top of the series can be seen in the eastern wall of the gorge, which has been downthrown 115m along the normal fault which runs along the gorge. This is the uppermost series showing normal magnetic polarity assigned to the Gauss normal magnetic epoch. Conformably on top of the normally magnetized tholeiites are identical tholeiite lavas which show reversed magnetic direction, (66m thick, 14 flows in the western wall of the Kerlingagil gorge). The total thickness of the tholeiite series (normally and reversely magnetized) is therefore about 146m, the average thickness of a lava flow being 4.7m.

The tholeiites can be followed to the south of Dýjadalshnúkur (E7). Outcrops are very scarce in the Blikdalur valley being confined to a few shallow gulleys. No signs were seen of the normal polarity tholeiites of unit 8 in the southern slopes of the valley, and they are considered to have thinned out southwestwards against the hyaloclastite mountain of units 4 and 7.

b. Rocks from the first reverse polarity period of the Matuyama magnetic epoch

The magnetic reversal Gauss - Matuyama occurred during an interglacial period which may have been extensive in view of the thickness of the lava pile produced and the occurrence of at least two thin red partings between lava flows (of unit 8) in the Kerlingagil gorge.

The succession consists of units of tholeiite and olivine tholeiite lavas intercalated with four substantial hyaloclastite units. The Kjalarnes volcanic centre was very active and two intrusive centres with associated dyke swarms were formed during this period. A basaltic hyaloclastite is thought to have formed within a caldera in the Kjalarnes centre. Owing to the irregularity of the topography after the glaciation, correlation of geological units over long distances becomes a difficult task, even where exposures are reasonable. In consequence, rock units formed during this earliest period of the Matuyama magnetic epoch are traceable over smaller distances than are the rocks formed during the Gauss magnetic epoch, and discussion will therefore be restricted more to individual localities. A numbering system for the geological units will be attempted, but this is more tentative than for the more regular succession from the Gauss magnetic epoch.

Unit 9. Tholeiite lavas, 70-110m thick.

As previously mentioned, this tholeiite unit is identical to unit 8, the sole difference being the magnetic polarity. This series is well exposed in the Kerlingagil gorge (E8) and forms a platform around Dýjadalshnúkur (D8). In southwest Dýjadalshnúkur an irregular circular-shaped structure in the topmost part of the lava sequence is believed to be a crater remnant. Exposures of the tholeiite lavas are not found down in the Blikdalur valley, but inferred boundaries are shown on the map (E7).

A possible member of this unit is a 80m thick group of tholeiite lavas exposed above 210m in the largely scree-covered slope northeast of Skrauthólar farm (D5). The 12 lavas exposed vary in thickness from 1.5 to 17m, with an average thickness of 6.6m. The base of these lavas is completely hidden by scree, but on top of the group comes hyaloclastite unit 10*. The group apparently thins westwards up against hyaloclastite unit 7, but exposures are very poor indeed.

The basaltic hyaloclastite found on the Kjalarnes peninsula (B5) was probably formed within a caldera simultaneously with the first tholeiite lavas of unit 9. The basaltic hyaloclastite (reversed polarity) lies directly on top of faulted tholeiite lavas with normal polarity, which cannot be younger than tholeiite unit 8. The Kjalarnes caldera will be discussed further in the chapter on intrusive activity.

* There is some evidence that the part of the hyaloclastite in actual contact with the topmost lavaflow is reworked, which might indicate that the lavas had banked up against an older hyaloclastite unit, thus making this group of lavas contemporaneous with tholeiite unit 11. This would perhaps account for the wide range in flow thickness (further discussed in unit 11), but the previous interpretation is preferred.

Fig.1. Esja from the southwest (Kjalarnes). The light coloured lower slopes are of intensely hydrothermally altered hyaloclastites. a)Lavas of unit 11. b)Volcanic plug in hyaloclastite unit 12. c)Basic cone sheets with southerly dip cut through the altered hyaloclastites and the overlying lavas. d)Hyaloclastite mountain of unit 10 (Figs. 2 and 3). e)Laugarnípa. f)Þverfell.

Unit 10. <u>Basaltic hyaloclastite, 25-200m thick.</u>

Volcanic activity was widespread during the first glacial period of the Matuyama epoch.

In the eastern wall of Kerlingagil gorge (E8) a vertical section of 200m of hyaloclastites makes a sharp contrast with the regular lava flows in the western wall. A normal fault with a downthrow to the east of some 115m runs along this 1km long gorge. The majority of the hyaloclastite is now seen on the downfaulted side of the fault which was probably active during the eruptions. In the eastern wall a thin sedimentary horizon succeeds tholeiite of unit 9 and is in turn overlain by 2-3m thick pillow lava which grades upwards into a 4m thick solid lava flow. The lava flow displays irregular jointing at the top and grades upwards into pillow lava and pillow breccia. Higher up, the hyaloclastite is mainly tuffaceous with some angular lithic fragments as well as true pillows. Large masses of crystalline tholeiite, commonly displaying irregular or fan shaped jointing, are found in the tuffs. In the western wall, immediately above the uppermost tholeiite lava of unit 9, a body of crystalline tholeiite showing irregular columnar jointing is interpreted as a plug (chem.an.15).

Massive fragments in the lowest part of the hyaloclastite on both sides of the fault are of porphyritic tholeiite with abundant, strongly zoned plagioclase phenocrysts, although phenocrysts of pyroxene are conspicuously rare. The hyaloclastite apparently thins eastwards from the Kerlingagil gorge. Two plugs in the hyaloclastite some 1.5km east of Kerlingagil are taken as representative of the upper part of the hyaloclastite unit 10. These are of porphyritic tholeiite (chem.an.16) with abundant phenocrysts of zoned plagioclase and pyroxene and rare phenocrysts of olivine. The hyaloclastite capping Dýjadalshnúkur (E8) is believed to belong to this same unit. The plug here is of porphyritic tholeiite (chem.an.17) with small phenocrysts of both plagioclase and pyroxene which commonly occur in clusters.

In the Kerhólakambur area a hyaloclastite ridge was built up at the same time as the Kerlingagil hyaloclastite. The maximum thickness is just over 200m. In the slope northeast of Skrauthólar farm (D5), the shape of this ridge can be studied in detail since the tholeiite lavas of unit 11 have banked up against it from the west and preserved it.

The main body of this hyaloclastite is formed of pillow lavas and pillow breccias. Well formed pillows can be seen less than a meter from the western contact of the tholeiite lavas of unit 11 at 400m altitude. A crudely circular-shaped area (Fig.2) of many separate, large (tens of meters), fine grained but crystalline masses of plagioclase-phyric tholeiite, many of them in the form of irregular sheets, displaying fan shaped jointing, is found within the hyaloclastite from 465m to 525m. This is almost certainly the site of eruption which has produced the upper part of the hyaloclastite.
The tholeiites of unit 11 are seen (Fig.3) banked up from the west against a slope with strike approximately 340^{o} up to an altitude of about 380m. There the strike of the old hyaloclastite mountain slope changes to about 300^{o}. This change may be the result of activity of the plug area described above.

Fig.2. Volcanic plug area in hyaloclastite unit 10 at about 500m in the slope northeast of Skrauthólar farm (D5).

Fig.3. Lavas of unit 11 banked against the slope of a hyaloclastite mountain of unit 10 northeast of Skrauthólar farm (D5). The palaeoslope of the hyaloclastite trends about 340° in the left of the photograph, but about 300° in the right of the photograph.

The hyaloclastite thins out very sharply westwards and is only about 25m thick where the lowest of the tholeiites of unit 11 have overlapped. Here the lower part of the hyaloclastite is of a tuffaceous breccia, apparently with no pillows. The uppermost 10-15m are of fine-grained laminated sediment, but the top 2m are of coarser sediment with small rounded pebbles. This is believed to represent an eroded plain at the foot of the hyaloclastite mountain.

Unit 11. Tholeiite lavas, 10-240m thick.

After the glacial episode during which hyaloclastite unit 10 was formed, the topography was dominated by steep sided hills and ridges. Lava flows in subsequent eruptions banked up against the hyaloclastites, and in some cases capped the hills. Correlation of the isolated groups of lavas as unit 11 has been based to some extent on dating them relative to fault movements.

At the mouth of the Eilífsdalur valley on the western side (G8), tholeiite lavas can be seen banked up against hyaloclastites from the east. The base of this tholeiite unit is not exposed in the Eilífsdalur valley itself, but a continuous section can be obtained in a shallow gorge from about 270m upwards. Seven tholeiite lavas with a fairly well developed flow structure and an average thickness of 4.1m are exposed under the hyaloclastite of unit 12.

Further west, in a gulley 1km southwest of Eilífsdalur farm (G8) at an altitude of about 240m, a group of 7 lavas with an aggregate thickness of about 20m occurs between two faults both downthrowing to the northwest. The lavas rest here on a conglomerate with well rounded pebbles, and they underlie pillow lava. The high dip value of 18° may indicate a depositional slope, but the block may also have been rotated by the movement of the faults. This is most probably a continuation of the Eilífsdalur tholeiite group described above.

In western Hrútadalur (F8), a group of some 10 tholeiite lavas forms a ledge in the middle of the slope of hyaloclastite units 10 and 12. The group is very heavily faulted. The lavas rest on a fairly well sorted sediment with rounded pebbles topping the hyaloclastite of unit 10. The maximum thickness is about 40m, but the group apparently becomes thinner westwards. The high dip of about 16° to the southeast may indicate a depositional slope (of perhaps 4°), and this would fit well with the thinning of hyaloclastite unit 10 from about 200m at the Kerlingagil gorge to some 120m 1.2km further east.

A prominent tholeiite lava flow forming the edge of the eastern wall of the Kerlingagil gorge (at about 470m altitude) is probably a member of unit 11, but whether it is continuous with the lavas further east in Hrútadalur is not certain. This lava flow, which has a minimum thickness of 10m (top not seen), may well postdate the main movements along the Kerlingagil fault.

Two thin and very vesicular lavas found at about 500m altitude in western Tindstaðahnúkur (E7), just east of Dýjadalshnúkur, intercalated in hyaloclastites, may well belong to unit 11. (The lavas lie at the foot of cliffs of pillow-breccia which Rutten,1958, refers to as irregular sills of basalt in fig.7, p.247).

The ridge southwest of the Blikdalur valley is capped by tholeiite lavas, which form the thickest group of lavas belonging to unit 11, reaching a maximum thickness of about 240m in the southern slopes of the ridge. No signs were seen here of the Kerlingagil fault or its closest neighbours, although the group is heavily faulted further east, much in line with the western Hrútadalur faults. The lavas have clearly banked up against an eroded hyaloclastite region from the southwest. Clear evidence of this is seen in the northwestern most part of the ridge (D6) where the tholeiite lavas pile up against hyaloclastite unit 7, and in the slopes northeast of Skrauthólar farm (D5) where the lavas have preserved splendidly the eroded landscape in hyaloclastite unit 10. Outcrops of the lavas are very scarce on the northern side of the ridge, and the downfaulted part of this unit in the Blikdalur valley southwest of Leynidalsá (F6) is inferred.

As mentioned in the description of hyaloclastite unit 10, the tholeiite lavas northeast of Skrauthólar farm flowed over an eroded plane at the foot of a hyaloclastite mountain, and piled up against a slope with a strike of approximately 340°. At an altitude of about 380m the strike of the slope changes to about 300° (Fig.3). The lavas apparently cap most of the hyaloclastite at about 450m, but have not quite submerged the top of the plug area in the east, so that hyaloclastite 12 comes there directly on top of hyaloclastite 10.

An interesting feature displayed by lavas which pile up against a steep slope is the variation in thickness of the very front or side of the lava flow which comes in actual

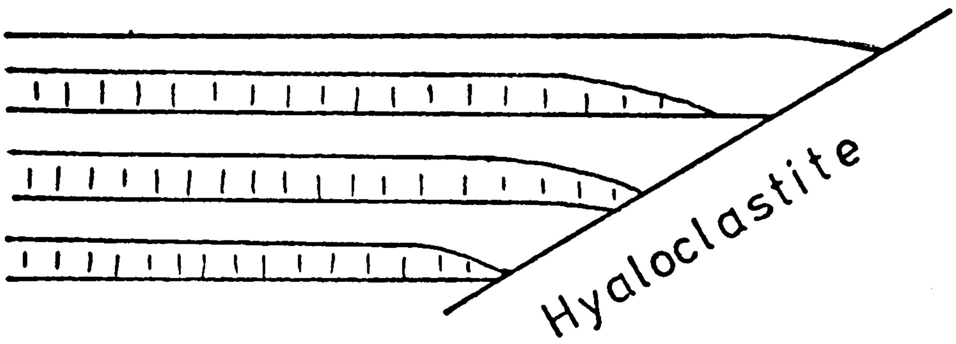


Fig.4. Lavas flowing against and/or along the side of a steep slope.

contact with the old slope. A flow starts to thin out a few meters away from the slope and the front or side that comes in direct contact with the slope has less than half the average thickness of the flow (Fig.4). This results in the formation of a shallow canal-like depression along the slope. The next lava flow in the succession fills this canal, and thus the flow front or side of this second lava is considerably thicker than the average thickness of the lava flow. This results in alternating thick and thin flow fronts (sides) in contact with the old slope.

The proximity of the old slope may be one of the reasons for the irregularity in thickness of the lava flows observed in the southern cliffs of the ridge. The measured thickness varies from 3 to 20m, the average thickness being about 7m.

The high dip value of 18^{o} for the lavas on the westernmost plateau of the ridge most probably results in part from a depositional slope. This would suggest that a stratovolcano in the Kjalarnes area (B5) produced some of these lavas. Others, however, appear to have been erupted very close to the hyaloclastite slope. An intensive dyke swarm cuts the underlying hyaloclastites and the lavas. At the base of the tholeiite lavas of unit 11 northeast of Skrauthólar farm, the dilation, represented by the total thickness of dykes, is up to 20%. The number of dykes decreases upwards in the lava pile which suggests that they are feeder dykes for the lavas, and dykes can indeed be seen ending in lava flows (e.g. the 6th lava flow from the bottom of the pile northeast of Skrauthólar farm at an altitude of 350m).

Unit 12. Basaltic hyaloclastite, 30-300m thick.

This is the thickest hyaloclastite unit in the western part of the Esja volcanic region. The main body of the hyaloclastite formed a mountain region extending from west of Kerhólakambur (D6) through Hrútadalur and beyond. It is best exposed in Hrútadalur (F7), and forms the better part of the slopes in northern Blikdalur. Only the basal part of this unit is now seen on the eroded top of the ridge southwest of Blikdalur (D6). A second hyaloclastite mountain region assigned to this unit is seen in the lower slopes of southwestern Kistufell (G4), and in the Flekkudalur valley (I8).

In western Eilífsdalur (G8), a 28m thick tuffaceous hyaloclastite with angular basaltic fragments (mostly ≤10cm) shows a northeasterly foreset bedding structure. Further west, in an exposure about 1km southwest of Eilífsdalur farm, the hyaloclastite is thickening considerably (Fig.5). Here pillow lava rests on the lavas of unit 11. Several dykes cut up into the pillow lava; one of these dykes is seen to spread out to form an irregular mass with columnar jointing.

Fig.5. Þórnýjartindur (G8) and Hrútadalur from the north. Lava units 13 and 14 bank from the east (left) against the hyaloclastite unit 12 (carved by gulleys), but the hyaloclastite is capped by unit 16 lavas. Steeply dipping lavas of unit 11 can be seen in the photograph at the mouth of the gulleys. Hyaloclastite unit 12 thickens westwards (right); the light colour of the hyaloclastites in western Hrútadalur is due to hydrothermal alteration.

In Hrútadalur, the hyaloclastite reaches its maximum thickness of over 300m. The exact thickness is not measurable, as the bottom part is not exposed. The lowest exposures are of pillow lava and pillow breccia, but the hyaloclastite becomes gradually more tuffaceous upwards, and in the uppermost part a foreset structure dipping 10°-20° northeastwards can be seen. This dip probably defines the northeasterly slope of a hyaloclastite mountain region. The very top of the unit here is of fine-grained laminated sediment which underlies a 2m thick, slightly reddish conglomerate succeeded in turn, at about 550m altitude, by two tholeiite flows (F7), which are assigned to unit 13.

At about 550m in the slopes southwest of Tindstaðahnúkur (E7) large masses (tens of meters across) of fine-grained tholeiite (chem.an.18) with columnar jointing found in the hyaloclastite are thought to indicate a site of eruption. This is a slightly porphyritic tholeiite with phenocrysts of plagioclase and microphenocrysts of augite as well as rare needles of olivine.

The hyaloclastite on the top of the ridge southwest of Blikdalur (D6) is of the same slightly porphyritic tholeiite (chem.an.19). Two plugs are eroded out of the hyaloclastite here. The larger one to the west (Fig.1) rises some 30m above the surrounding ground and is very picturesque. In its western "rim" well developed columns with diameters of about 20cm have a 30° easterly dip towards the old crater. The northeastern rim rises higher, but the columns are fan-shaped and less regular.

Further southeast along the ridge in the highest slopes west of Kerhólakambur, a 50m thick hyaloclastite, mostly pillow lava, is found on top of tholeiite lavas of unit 11. As mentioned before, the lavas of unit 11 thin out eastwards against the plug region of hyaloclastite unit 10. Hyaloclastite unit 12 comes therefore directly on top of unit 10 in the slopes above the plug (between 525 and 550m). In the northern side of the ridge (Blikdalur side), pillow lavas, corresponding to those in the southern side, are found in poor exposures in burns.

The Kerhólakambur area (D6) is believed to have been the southern flank of a hyaloclastite mountain range with a southwest-northeast trend running northwards through Hrútadalur (F7)

On the Kollafjörður shore east of Leiðhamrar (E3) a dolerite intrusion cuts hyaloclastite which is assigned to unit 12.

Assigned to hyaloclastite unit 12 is a pile of hyaloclastites, with a minimum thickness of 300m, forming the lower slopes of Kistufell north of Vellir farm (G4). The lowest exposure is of a rather ill sorted sediment, which in the eastern part, just northwest of the Kistufell fault, is seen to pass into finely laminated siltstone. A 12m thick lava flow (not shown on the geological map) rests on the sediment and forms a prominent ledge in the slope north of Vellir farm (Fig.7). The lava thins westwards*. On top of this single lava flow

* The lava flow might speculatively be connected with unit 11, thus making the underlying sediment correspond to the sedimentary top of hyaloclastite unit 10.

are pillow lava and pillow breccia in the east, and more tuffaceous hyaloclastites in the west. Lavas of unit 13 form an apron on an old southern slope of hyaloclastite 12 at and below 300m altitude north of Vellir farm, as will be described later. Underneath and north of this apron the hyaloclastite of unit 12 continues in the form of tuffaceous pillow breccia, which rises to an altitude of over 400m in the southwest spur of Kistufell, but is at an altitude of about 250m 1km further southwest where it is downfaulted on the Kistufell fault.

Also assigned to unit 12 is the hyaloclastite at the mouth of the Flekkudalur valley (I8). The Flekkudalur river dissects tholeiite pillow lava (chem.an.20) with a large proportion of tuffaceous matrix. Several tholeiite masses showing columnar jointing are seen in the river section, and irregular dykes of the same tholeiite anastomose within the hyaloclastite. Further northwest, at the northern end of the Sneiðar ridge (H9), the pillow lava is more coarsely porphyritic in character.

Unit 13. Tholeiite lavas, 20-400m thick.

At the end of the glacial episode during which unit 12 was formed, the landscape was very mountainous in the Esja region. The dominant feature was a broad hyaloclastite mountain range with a southwest-northeast trend running from the west of Kerhólakambur (D6) through Hrútadalur (F7). To the southeast of this range was a valley running from Gljúfurdalur (F5) to Eilífsdalur (G7). More hyaloclastite mountains rose to the southeast of this valley in the southwest Kistufell area (G4), and these may have extended to the northeast to form a continuous hyaloclastite mountain range as far as Flekkudalur (I8).

During the interglacial that followed, lavas ponded in local depressions, banked up against hyaloclastite slopes and flowed down the slopes, filling the valleys between the hyaloclastite ridges and submerging the hyaloclastites.

The lavas assigned to unit 13 are shown as tholeiites on the geological map. However, the unit is somewhat variable. The majority of the lavas are identified in the field as tholeiites, but some olivine tholeiites and lavas transitional between tholeiites and olivine tholeiites occur within the unit, especially in the Eilífsdalur region.

At the northern end of Sneiðar (H9), southwest of lake Meðalfellsvatn, the pillow lava of unit 12 is overlain by vesicular tholeiite lavas (total thickness about 40m) which can be traced southeastwards to the nearest fault. On top of these comes a coarsely porphyritic lava flow which is up to 35m thick and forms the lower of the two prominent ledges in the Sneiðar ridge. This lava (chem.an.21) has large plagioclase phenocrysts, showing remarkably little zoning except within very narrow rims, and smaller pyroxene phenocrysts; olivine was not detected. The lava thins southeastwards against the hyaloclastite unit 12, and is succeeded by a very prominent lava flow which reaches a thickness of 35m and shows crude columnar jointing with a column diameter of about 2.5m.

This lava forms the cliff above the summerhouse settlement in the western slope of Sneiðar ridge, and forms prominent cliffs on the eastern side of the ridge inwards along the Flekkudalur valley, where it is seen thinning against and also capping the hyaloclastite; it forms the waterfall in Flekkudalur river at 150m altitude and forms a prominent crag in the northwestern slope of Sandsfjall (I8). The lava is easily identified by its olivine and plagioclase phenocrysts (chem.an.22 and 23). These two porphyritic lavas have probably ponded in a local depression in the hyaloclastite. Tholeiite lavas come on top of the porphyritic lavas, but their exact thickness in Flekkudalur is not known.

Lavas assigned to unit 13 are exposed in burns in the lowest slopes of eastern Eilífsdalur (H7), but they are mostly covered by scree. The lavas are much better exposed in northwestern Eilífsdalur (G8), where nineteen tholeiite flows with an average thickness of 4m are seen. An intercalated 6m thick sedimentary layer (rounded grains mostly $<$1cm in the lower part, but more tuffaceous in the upper part) is correlated with a 8m thick sediment in eastern Eilífsdalur. The tholeiite lavas are thought to have banked up against (Fig.5) the unit 12 hyaloclastite mountain (in Hrútadalur) from the east. No signs of the lavas are seen in eastern Hrútadalur. Two tholeiite lavas (20m total thickness) occurring at about 550m altitude in western Hrútadalur (F7), and some ten tholeiite lavas occurring above 500m in the hyaloclastite slope northeast of Leynidalsá in Blikdalur (F7) are tentatively assigned to unit 13, although it has not proved possible to correlate them with the Eilífsdalur group.

In the southwestern part of the hypothetical Gljúfurdalur-Eilífsdalur valley, lavas banked up against the hyaloclastite, and some of these were indeed erupted on the slope of the hyaloclastite in the Kerhólakambur area (E5). The crater region has a northeast-southwest extension. Three of the most obvious crater remnants are shown on the geological map. All three are of slightly porphyritic tholeiite (chem.an.24 and 25), with plagioclase phenocrysts much larger than the groundmass plagioclase. Slightly porphyritic lavas can clearly be seen thickening southeastwards away from the scoria heaps associated with the largest crater ruin furthest southwest (Fig.6).

Fig.6. The tholeiite lavas of unit 13 thicken away from the scoria cone associated with the parent crater ruins in south-west Kerhólakambur; note the difference in the sketch in Fig.4 where the lavas have piled against an old slope.

Immediately southeast of this crater, 38 lavas were counted (binocular geology) in a 170m high cliff (average thickness 4.5m per flow). The total thickness of flows of unit 13 in the Lauganípa-Mógilsá area (E5-F4) is over 400m. This clearly reflects the vigorous activity of the Kjalarnes central volcano (dyke swarm) at this time. The lavas are perfectly exposed in precipitous cliffs carved by narrow gorges in the Lauganípa-Kerhólakambur area. No attempt was made, however, to do detailed traverses in these practically unscaleable cliffs. All the lavas measured in the Kerhólakambur cliffs and in the gulleys running north from the Gljúfurdalur river showed reversed magnetic polarity, with the exeption of two lavas at about 630m altitude in the western cliff edge of Lauganípa (E5) which showed weak normal polarity. The lavas in the Gljúfurdalur-Mógilsá area have a northwesterly dip which is in a sharp contrast with the regional southeasterly dip. This may be due partly to a depositional dip of the lavas, but is thought to be mainly the result of uplift associated with the Þverfell intrusions (E4). All the lavas in the Mógilsá river traverse (F4) from sea level to the top of the ridge between Mógilsá and Gljúfurdalur have reversed magnetism.

East of the Mógilsá landslip (G4), exposures of tholeiite lavas assigned to unit 13 are found at two levels in the southwestern Kistufell slopes. As mentioned previously, tholeiite lavas form an apron on the southern slope of hyaloclastite 12. The lavas are thought to have been erupted on the hyaloclastite slope; a dyke is seen feeding a lava flow at 250m altitude in the western of the two deep gulleys north of Vellir farm (G4), and a second possible site of eruption is seen some 700m further east, at about 270m altitude, where two prominent tholeiite hummocks, with a very irregular flow structure, appear along the strike of two tholeiite cone sheets (each about 4m thick and having a 50° southerly dip) which may be their feeders.

The single lava flow forming a ledge within hyaloclastite 12 north of Vellir farm has a southeasterly dip of 10°, and can be used as a reference in the study of the depositional dip of the lavas of unit 13 (Fig.7). The lavas of unit 13 furthest west show a 20° southwesterly dip. A little higher up on the platform at about 300m the dip is very low. As the lavas are traced to the east the dips increase progressively to 20° southeasterly. This is interpreted as being due to the hyaloclastite, on which the lavas were erupted, having a southwesterly slope of some 25°, levelling off eastwards, and having a southeasterly slope of up to 10°. This was referred to by Rutten (1958) as "...local strong tilting ... reminiscent of a small anticline..." (p.245).

A thin red parting is seen between the two lowest tholeiite lavas of unit 13 between the two deep gulleys north of Vellir farm. Both these lavas have reversed magnetic polarity direction. The third lava has normal polarity, but the fourth and fifth lavas have reversed polarity. Some 300m further northwest, in a gulley bordering the lava apron, the lowest lava, just below 300m altitude, has reversed polarity; the

Fig.7. Kistufell from the south. The numbers are unit numbers in the stratigraphic column and the dip of the lava apron of unit 13 is indicated. The single lava flow dipping 10° (see text) is marked A. The downthrow on the Kistufell fault is 110m to the southeast.

second lava, which forms a platform east of the gulley, is normally magnetised; the third lava shows both normal and reversed magnetic directions. These lavas can be seen thinning northwards against hyaloclastite unit 12 in the gulley at 320m altitude. The hyaloclastite apparently underlies the reversely magnetised tholeiite lavas at about 450m, but no exposures are found between 400 and 450m in this southwestern corner of Kistufell. This group of tholeiite lavas which is assigned to unit 13, can be followed eastwards along the foot of the main Kistufell slopes. In a section some 500m east of the southwest Kistufell spur, a sequence of lavas is observed similar to the sequence between the two gulleys north of Vellir farm (described on p.41). A 50cm thick red parting is found between the reversely magnetised second and third lavas, the fourth and fifth lavas show normal magnetic polarity, the sixth lava shows both normal and reversed polarity, the subsequent six lavas show reversed polarity whereas the uppermost tholeiite lava of unit 13 shows weak normal magnetism. (The average thickness of all twelve lavas is 5m). It is concluded that the lavas forming the apron are contemporaneous with the lavas at the foot of the main Kistufell slopes. Exposures of the unit are terminated southeastwards by the Kistufell fault.

The normally magnetised lavas must represent a very short normal magnetic event within the first reverse polarity period of the Matuyama epoch and may possibly correlate with the Reunion event. Perhaps the two normal polarity lavas observed on Lauganípa also correspond to this event.

Unit 14. <u>Olivine tholeiite hyaloclastite and lavas, 50-80m thick.</u>

The unit, comprised of a hyaloclastite and lavas, is conformable with, and has largely the same distribution as unit 13, except that no signs were seen of the hyaloclastite in the Gljúfurdalur area (E5). The scale of the geological map precluded subdivision of the unit into hyaloclastite and lavas; the maximum thickness of the hyaloclastite is 25m. The hyaloclastite is thought to represent either a short temporary glacial or lacustrine environment. A lake on the lava plain east of Kerhólakambur-Hrútadalur hyaloclastite mountain range (by now extending into the Kerhólakambur-Mógilsá area) would perhaps better explain the relatively uniform thickness of the olivine tholeiite hyaloclastite. In subglacial eruptions, the hyaloclastite is locally thicker around the eruptive site.

The unit was not mapped in detail in Flekkudalur (I7) and is not separated from units 15 and 16 there on the geological map. However, the boundary of units 13 and 14 is thought to lie in the upper part of a series of some 20 thin lavas which are seen (binocular geology) on top of the prominent olivine-plagioclase porphyritic flow of unit 13 in the northern Sandsfjall cliffs (I8). These lavas are overlain by a hyaloclastite which thins out eastwards, and which in turn is succeeded by eight thin lava flows which are thought to form the top of unit 14 here. These are capped by a substantial hyaloclastite which is assigned to unit 15. There are no exposures of the unit in the Flekkudalur river traverse, but a hyaloclastite found there at about 270m altitude is assigned to unit 15.

The unit is well exposed in the eastern slopes of Elífs-dalur valley (H7). Here the bottom of the unit consists of olivine tholeiite flow units, with a total thickness of 12m (each unit less than 1m). This is capped by a 15m thick hyaloclastite which is tuffaceous in the lower half but grades upwards into olivine tholeiite pillow lava. On top of this come olivine tholeiite lavas and two tholeiite lavas. The total thickness of unit 14 here is about 60m.

On the western side of Eilífsdalur valley, the bottom 15m are of olivine tholeiite lavas which are capped by an olivine tholeiite pillow lava about 20m thick. The pillow lava thins southwards and a sedimentary hyaloclastite with a southerly foreset structure takes its place. This is capped by one tholeiite lava followed by a 10m thick conglomerate (boulders up to 20cm across). The conglomerate is overlain by a 10m thick tholeiite lava which has normal magnetic polarity, and is the only normal polarity lava found in Eilífsdalur valley below the normally magnetised unit 19. The tholeiite is succeeded by olivine tholeiite lavas, 8m thick. The total thickness of unit 14 here is about 70m. The unit thins against the Hrútadalur hyaloclastite.

Unit 14 is represented in Kerhólakambur (E5) by an approximately 20m thick olivine tholeiite cap on a hill at 750m altitude. The unit is downfaulted about 200m to the east; the olivine tholeiite lavas form an approximately 80m thick cap on the tholeiites of unit 13 and can be traced eastwards to Þverfellshorn. No signs were seen of the hyaloclastite within the unit here. The Mógilsá landslip covers all exposures of the unit east of Þverfellshorn.

In the southern slopes of Kistufell, the unit consists of a 25m thick olivine tholeiite pillow lava which is overlain by 25m thick olivine tholeiite lavas (flow units) which can be followed eastwards to the Kistufell fault.

Unit 15. Basaltic hyaloclastite, 10-200m thick.

Volcanic activity was widespread during the glaciation which followed the extrusion of the lavas of unit 14. A hyaloclastite blanket was formed in the western Esja mountain range, and hyaloclastite ridges were formed in the central south Kistufell area (H4) and in the southeastern Eilífsdalur area (H6).

A characteristic feature of the hyaloclastite blanket of this unit in the western Esja region is its light colour. The light brown layer in the highest slopes of the northern Blikdalur valley wall can be seen from afar, and the hyaloclastite reaching from Kerhólakambur to the east of Þverfellshorn can clearly be seen across the Kollafjörður bay from Reykjavík. The light colour is due at least partly to hydrothermal alteration. In northeastern Blikdalur, west of Gunnlaugsskarð (F6), the hyaloclastite is mainly of pillow lava and pillow breccia; olivine tholeiite pillows are seen at about 600m, but higher up the hyaloclastite is dominantly of small tholeiite pillows set in a glassy, tuffaceous matrix. The hyaloclastite is here approximately 200m thick but apparently thins westwards. In Kerhólakambur-Þverfellshorn area the hyaloclastite is mainly fine-grained, tuffaceous material with small (10-20cm) tholeiite pillows and pillow fragments. This southern part of the hyaloclastite is thickest (approximately 170m) in the Þverfellshorn area.

A narrow hyaloclastite ridge in the central south Kistufell area (H4) consisting mostly of a tuffaceous olivine tholeiite pillow lava and pillow breccia, reaches a maximum thickness of about 110m, but thins to the west and east where lavas of unit 16 and units 16 and 17 respectively bank up against the ridge. The hyaloclastite is downfaulted on the Kistufell fault and is seen to be mainly of pillow lava and pillow breccia in the gulleys north of Norður-Gröf farm (H4).

A second hyaloclastite ridge is postulated in the southeastern Elífsdalur valley, but only its northwestern side is exposed. The hyaloclastite reaches a maximum thickness of some 130m and can best be inspected in the deep and very narrow gorge of the Fossá river (H6). The lowest part is of olivine tholeiite pillow lava, but the great bulk of the hyaloclastite is of tuffaceous material with only rare lithic fragments. The hyaloclastite apparently thins gradually northwestwards and is represented by some 30m of pillow lava capped by a sediment northwest of Þverá (G7), and by a 8m thick tuff in northwestern Eilífsdalur. In eastern Eilífsdalur, the hyaloclastite is mainly tuffaceous (20m) and is capped by a 1m thick sediment. The unit is not shown on the geological map in Flekkudalur, but is believed to occur at about 270m altitude in the Flekkudalur river, and in the upper northern-Sandsfjall cliffs as previously stated.

Unit 16. <u>Olivine tholeiite lavas, 10-200m thick.</u>

Erosion took place at the end of the unit 15 glacial episode and the land was apparently peneplaned in the Eilífsdalur and Flekkudalur regions. There were hyaloclastite mountains in the Kerhólakambur (E5) - Hrútadalur (F7) region and in the eastern Eilífsdalur (H6) and Kistufell (H4) regions. The main volcanic activity was by now on a line from Grafardalur (H4) to northern Eyjadalur (J8). Tilting had almost ceased in the region northwest of this line so that an unconformity is seen there between the steeply dipping lavas of units 13 and 14 and the relatively flat lying lavas of unit 16.

Fine-grained olivine tholeiite lavas spread over the peneplain in the Eilífsdalur region. In Þórnýjartindur (G7) the individual lavas vary greatly in thickness; one fine-grained olivine tholeiite lava is 34m thick (chem.an.33). The aggregate thickness of lavas in western Þórnýjartindur is about 100m, and in eastern Þórnýjartindur about 50m, the unit apparently thinning southwards. It consists of three lavas (about 25m thick) in the slope northwest of Þverá (G7) (a thin, bright-red parting is seen between the two upper flows) and only two lavas southwest of Þverá. Three lavas are found southwest of Fossá (H6), but they thin eastwards against hyaloclastite unit 15.

In southwestern Kistufell (G4) olivine tholeiite lavas (total thickness about 20m) bank up eastwards against hyaloclastite unit 15.

In eastern Eilífsdalur, in the western slope of Skálatindur (H7) a single lava flow (chem.an.34) represents unit 16. This prominent lava flow is clearly seen disintegrating southeastwards into pillow breccia (overlying waterlain sediment of unit 15) as if it had flowed into water. In a gulley only 10m south of where the flow "ends", a continuous profile reveals a tuffaceous hyaloclastite (unit 15) with a thin sedimentary cap, succeeded by a pillow breccia which in turn is capped by a sediment (unit 18).

The unit thickens considerably eastwards and is about 100m thick in the Flekkudalur river (the second lowest lava forms the waterfall at about 280m altitude). Most of the lavas here are of olivine tholeiite. The thickness is of the order of 150-200m in the Eyjadalur area (J8). The uppermost three thin lavas in western Eyjadalur below hyaloclastite unit 18 are of tholeiite and are thought to correspond to the tholeiite lavas of unit 17 which cap the olivine tholeiite lavas in Möðruvallaháls (K8).

In the southern slopes of Kistufell (H4), olivine tholeiite lavas can be seen banking up against hyaloclastite unit 15 from the east. The lava pile thickens considerably eastwards and has a minimum thickness of 80m (the base is not exposed) in the Grafardalur valley (H4). The unit forms the southwestern end of the Þverárkotsháls ridge (I4) and is also found in Þverárdalur (I5).

Unit 17. <u>Tholeiite lavas, 10-50m thick.</u>

Along and to the south of the Grafardalur (H4) - Eyjadalur (J8) line the olivine tholeiites of unit 16 are capped by a few tholeiite lavas which, because of their higher resistance to erosion, form escarpments.

Unit 17 is represented by two lava flows thinning westwards against hyaloclastite unit 15 in southeastern Kistufell, and occurring on both sides of the Kistufell fault. The unit thickens slightly eastwards and is more than 30m thick in the eastern slope of Grafardalur valley (I5). The unit was not detected in the poor exposures in Þverárdalur valley (I5). The tholeiite lavas form prominent escarpments at the mouth of Svínadalur valley (K8), where their aggregate thickness is about 50m, but the unit thins westwards in Eyjadalur (J8).

Unit 18. Basaltic hyaloclastite, 5-300m thick.

During the glaciation that followed the extrusion of lava units 16 and 17, the volcanism was most active along a line from Þverárdalur (I4) to Trönudalur (L7) and gave rise to a rugged hyaloclastite mountain region. Sediments were formed to the northwest of the active volcanoes; the retreat of the glaciers is probably represented by the conglomerates commonly found at the base of the sedimentary sequence, the more tuffaceous re-sedimented hyaloclastites in the upper part of the unit corresponding to the less violent erosion of the next interglacial period.

In the extreme west of the Esja area the identity of the unit is not entirely clear, and the sediments at the first two localities described here are only tentatively assigned to unit 18. At the head of the Blikdalur valley, southwest of Gunnlaugsskarð (G6), a 10m thick sediment is found on top of the unit 15 hyaloclastite and below the normally magnetised lavas of unit 19. The bottom 1m is of conglomerate but the sediment above is of fairly tuffaceous laminated material. In western Hrútadalur (F7), on top of the lavas tentatively assigned to unit 13 at 550m altitude, a 20m thick, fine-grained, laminated sediment lies underneath the lavas of unit 19. The sediment becomes slightly coarser at the top. The main part of this bed perhaps belongs to unit 15, and the coarser top part to unit 18.

In the northwestern spur of Þórnýjartindur (G7), a hyaloclastite assigned to unit 18 is found on top of the lavas of unit 16. Furthest west the hyaloclastite reaches a maximum thickness of about 30m; here it consists of pillow breccia, which may be a primary volcanic feature, but it thins eastwards changing progressively into a conglomerate. In the northeastern spur of Þórnýjartindur the conglomerate (about 20m thick) is polygenetic with rounded pebbles and graded bedding. It apparently thickens southwards, and, north of Þverá river (G7) overlying lavas of unit 16, the sediment is 50m thick, with rounded pebbles and repeated fining-upwards cycles. Southwest of Þverá river on top of the lavas of unit 16 come pillow lava and pillow breccia which grade upwards into a tuffaceous hyaloclastite. Only the uppermost meter or so, which is greyish in colour, has well rounded pebbles. The hyaloclastite here has a total thickness of about 30m. In the Fossá river (H6) traverse, a 20-25m thick hyaloclastite occurs between the lavas of unit 16 and those of unit 19, but no good exposure was found to establish whether it is a primary volcanic or re-sedimented hyaloclastite. In western Skálatindur (H7), the unit is represented by a 20m thick sediment with rounded pebbles at the base but it becomes more tuffaceous upwards and has a reddish colouration at the top. The unit apparently thins out northwards in the slopes of Nónbunga (H8), and where it is found again in the eastern slopes of Nónbunga in Flekkudalur it is in the form of a 1m thick sediment with small rounded pebbles. The unit thickens southwards in Flekkudalur and is seen as 15-20m thick pillow lava and pillow breccia at the head of the valley. The unit again thins northeastwards in eastern Flekkudalur and Sandsfjall (J8), but is about 45m thick on the western side at the mouth of Eyjadalur and thickens southeastwards from there to form the main hyaloclastite region in Svínadalur (L8), southern Eyjadalur (J6) and Þverárdalur (I4).

In southwestern Kistufell (G4), the unit it represented by a conglomerate horizon which can be traced along the western slopes of Kistufell. The pebbles are well rounded, range in size up to 15cm and sit in a matrix which is greyish in the lower half of the unit, and brown (more glassy) in the upper half where the pebbles are also more vesicular. The thickness is variable, but commonly about 5m. The unit thickens eastwards to the hyaloclastite ridge of unit 15, and that ridge has clearly served as a source region for the upper part of the sedimentary unit 18.

To the east of the hyaloclastite ridge, the unit has a fairly uniform thickness of about 45m and can be studied both north and south of the Kistufell fault. The lowest 20m or so have rounded pebbles (mostly smaller than 10cm) but the upper half is almost entirely of tuffaceous hyaloclastite. In eastern Grafardalur the bottom and top parts of the unit consist of tuffaceous hyaloclastite, separated by a reworked hyaloclastite with small, rounded pebbles. A similar sequence is found at the base of the unit at the head of Þverárdalur (I5). Further south in Þverárdalur, however, the unit is composed entirely of pillow lavas, pillow breccias and tuffs which have built up a ridge reaching from Þverárkotsháls (I4), through eastern Þverárdalur, to southern Eyjadalur and Trönudalur (K7) and beyond. The thickness at the ridge crest is of the order of 200m in Þverárdalur, but is more than 300m in the Eyjadalur-Trönudalur region. The hyaloclastite ridge is mostly composed of olivine tholeiite.

c. <u>Rocks from the Olduvai normal magnetic polarity event</u>

The change from reverse to normal magnetic polarity direction occurred a short time after the retreat of the glaciers. The first lava succeeding the hyaloclastite unit 18 in northern and western Esja commonly shows both normal (more common) and reverse polarity, and in eastern Esja the first few lavas in the succession at some localities are reversely magnetised. The magnetic field intensity during this normal event was high and the volcanic products are easily mapped since their strong normal polarity is in striking contrast to the reversely magnetised rocks above and below in the succession. The volcanics erupted during this event provide a good marker horizon.

Unit 19. <u>Tholeiite and olivine tholeiite lavas, 10-200m thick.</u>

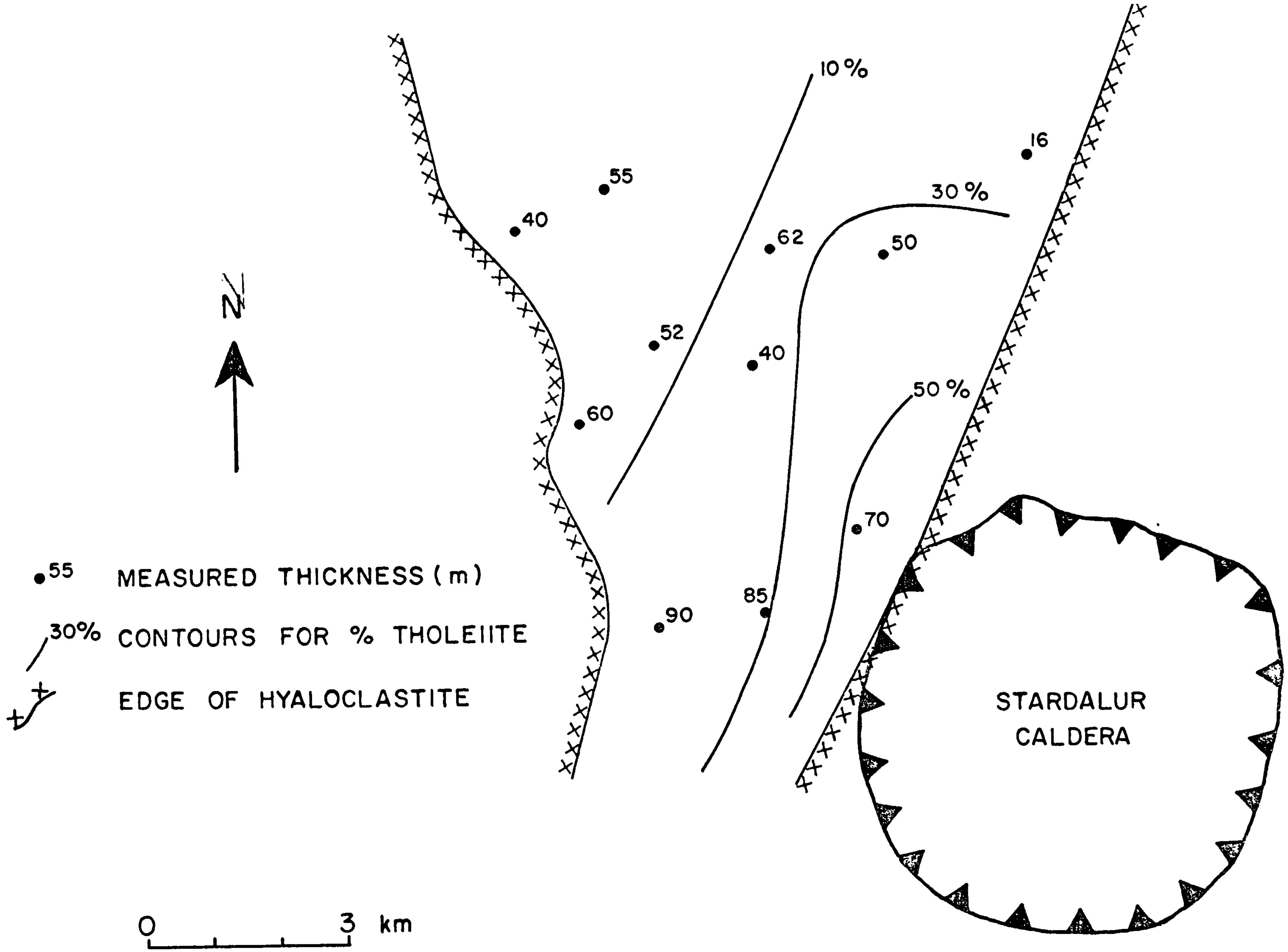
The main features of the landscape after the glaciation were an eroded hyaloclastite mountain region in the Kerhóla-kambur (E5) - Hrútadalur (F7) area and the newly formed hyaloclastite ridge running from Þverárkotsháls (I4) through Trönu-dalur (K7). The latter divided into two the area accessible to the normally magnetised subaerial volcanics. To the northwest of this ridge the volcanism was mixed, with shield volcanoes giving rise to olivine tholeiite compound lavas, and much less voluminous tholeiite lavas probably being erupted from fissures. To the southeast of the ridge, on the other hand, the lavas were almost exclusively of tholeiite.

In western Esja, the hyaloclastite mountain region apparently sloped down to the northeast which may reflect the high productivity of the Kjalarnes central volcano, and/or may be due to the uplift associated with the Þverfell and Lauganípa intrusions (E4). The lowermost lava in the unit is of tholeiite which is followed by two olivine tholeiite series separated by one and three slightly porphyritic tholeiite lavas in Skálatindur (H7) and Kistufell (H4) respectively. Further southeast there is a higher proportion of tholeiite lavas in the pile. Fig.8 shows the measured thicknesses of the unit and isopachytes for the approximate % of tholeiite lavas in the pile.

The rugged nature of the hyaloclastite ridge running from Þverárdalur through Trönudalur exerted a strong control on the distribution of unit 19 lavas. Those erupted on the ridge flowed along local depressions in the hyaloclastite; in northeastern Þverárdalur (I5) six lavas (or flow units), all less than 2m thick, are found in a small valley in the hyaloclastite; in southwestern Eyjadalur (J6) a few tholeiite lavas can be seen banking against hyaloclastite 18 from the south. The two lowest lavas exposed have reversed magnetism, but the upper lavas have normal magnetism. Despite very limited exposure, it is thought likely that the lavas have flowed in an isolated valley within the hyaloclastite ridge.

East of the hyaloclastite ridge, tholeiite lavas banked against the ridge but did not submerge it. The thinning of the unit is well demonstrated in eastern Þverárdalur (J5) where the lava pile thickens steadily eastwards from the hyaloclastite

Fig. 8



10 %
16
55
30 %
40
62
50
N
52
40
50 %
60
70
55 MEASURED THICKNESS (m)
85
90
30% CONTOURS FOR % THOLEIITE
EDGE OF HYALOCLASTITE
STARDALUR
CALDERA
0
3 km

ridge both within the Stardalur caldera and in Svínadalur (L6). The maximum thickness of the lavas can nowhere be measured. Drillhole H-1 (Friðleifsson and Tómasson 1972) sunk at Stardalur farm (L4) penetrated a 155m thick pile of tholeiite lavas but the bottom of the unit was not reached. Resistivity measurements failed to indicate the thickness of the unit.

The lavas east of the hyaloclastite ridge were erupted at the beginning of the normal polarity event, and a large number of lavas were erupted in a short period of time. This is suggested indirectly by the lack of erosive horizons between the lavas, and (probably more convincingly) by the measurements of the magnetic inclination of lavas from drillhole H-1 in Stardalur (Friðleifsson and Kristjánsson 1972). In 103 specimens from the drillcore between 41 and 143m depths, the inclination values averaged 81°, with a standard deviation of 4°, a deviation much smaller than would be expected from the long-term (about 2000 years) dispersion of geomagnetic field directions in Upper Cenozoic times; this long-term dispersion is about 22° in Tertiary basalts in NW Iceland (Kristjánsson 1970) and 10° in the present field. The lavas in Stardalur may therefore have been emplaced in a time interval much shorter than the major secular variation time scales (about 2000 years).

The lava production was brought to a sudden end by a caldera collapse. The caldera (Fig.9; Section 7) is roughly circular in shape and has a diameter of about 6.5km. Its northern margin is clearly marked by a zone of intensive faulting and brecciation. A gradual change from slightly dislocated lava blocks to intensely brecciated and mylonitized rocks can be seen in the caldera fault zone. The width of the zone varies from a few tens to about 300m. The caldera faulting is shown by a single line on the geological map, but this line represents a number of closely spaced parallel faults, most of which are normal with dips of 50^{o}-70^{o} towards the centre of the caldera. There has been relatively little overall vertical displacement across the fault zone; the faults have mainly served to adjust to a change in dip.

In the exposed parts of the caldera this marked change in dip is clearly seen; the rocks within the caldera dip steeply towards its centre in a markedly regular fashion, whereas the rocks outside the caldera rim dip away from it or show the regional southeasterly dip. The dip change is sharp and coincides with the caldera fault zone both east of Stardalur farm (L4) and in northeastern Þverárkotsháls (I5), where a few tholeiite lavas form a steeply dipping apron on hyaloclastite unit 18. In a gulley carved out along the caldera fault in eastern Þverárdalur (J5), the dip of the lavas is northerly closest to the fault, but is seen to change gradually to a southeasterly dip along the ridge towards Gráhnúkur (J5). Similarly in Þverfell (J5), the dip is seen to change inside the main caldera rim.

The southern half of the caldera is mostly covered by the post-erosional olivine tholeiite compound flow which was probably erupted in the Mosfellsheiði shield volcano after erosion had produced the present day landforms in the area. Here the inferred caldera boundary is based on geophysical evidence, which will be discussed on page 158.

At the head of the Stardalur valley (L4), just east of the edge of the geological map, an icelandite lava (chem.an.43) is seen above the tholeiites of unit 19 but below the basaltic andesite hyaloclastite of unit 21. The exposed thickness is about 15m, but the lava is probably considerably thicker. The lava has well developed flow alignment, and dips about 30° to the southeast.

The lavas within the central 3km diameter portion of the caldera, e.g. the tholeiite lavas southeast of Skeggjastaðir farm (J3), may at least partly have acquired their centripetal dip through uplift associated with the intrusion of the basic cone sheets. But the overall picture indicates that the caldera collapse was brought on by a sudden drainage of magma from under the central part of the caldera. Taking the base of normally magnetised lavas (unit 19) as having been at the same altitude of 400m south of Móskarðshnúkar (J5) and north of Grímmannsfell (J2) prior to the caldera collapse, the downwarp in the centre of the caldera is found (by projection of observed dips) to be about 900m. A right circular cone of radius 6.5km and height 0.9km has a volume of about $10km^3$. An almost identical shape and volume of the cone is found by taking the present day base of the normally magnetised lavas (unit 19) from northwest of Þverárdalur (I5) and south of Múli (L3) as being the pre-caldera base.

Fig.9. Stardalur caldera from the northwest. Compare with Section 7. a)Móskarðshnúkar rhyolite; note the light rhyolite dyke on right hand side of the peak. b)Bláhnúkur basaltic andesite plug. c)Þverfell. d)Stardalshnúkur dolerite intrusion; note the saucer shape. In the left foreground: e)caldera fault zone capped by f)lavas of unit 24. In the central foreground: g)Gráhnúkur sill. In the background (right): h)Borgarhólar crater from which the post-erosional olivine tholeiite lava came.

Unit 20. Tholeiite hyaloclastite in caldera.

The caldera basin was filled with water. Tholeiite magma erupted within the caldera formed pillow lavas and pillow breccias in the caldera lake. The pillows tend to be large (50cm long axis). The pillow lavas are best exposed at the mouth of the Leirvogsá river gulley (J4) northeast of Skeggjastaðir and in gulleys in the southern slopes of Stardalshnúkur (K4). Especially at the latter locality large masses of the same rather coarse-grained vesicular tholeiite can be seen displaying irregular columnar jointing within the pillow lava pile. These masses are intrusive, but there is no evidence of chilling against the pillows, and they are thought to have intruded the pillow lavas before the latter had cooled, and may indeed have served as feeders.

In a 240m deep drillhole H-2 (Friðleifsson and Tómasson 1972) by Tröllafoss (J4), the bottom of the hyaloclastite was not reached. By extrapolation of measured lava dips the maximum thickness of this hyaloclastite unit in the centre of the caldera is estimated to be about 400m. From this and the present day exposures the total volume of the unit is calculated to be approximately 0.7km^3.

Unit 21. Basaltic andesite hyaloclastite in caldera.

After the eruption of unit 20, there was a pause in the volcanism sufficiently long to allow a 10m thick layer of fine-grained, tuffaceous, laminated sediment to settle at the bottom of the caldera lake. The sediment is best exposed in a subsidiary gulley from the Leirvogsá river between Haukafjöll and Tröllafoss (J4). The sediment dips less steeply here than the tholeiite lavas further west, indicating that some of the intra-caldera tilting had occurred before the sediment was formed. The sediment can also be seen in the main Leirvogsá gulley some 300m southwest of Tröllafoss, where it is thoroughly baked adjacent to the dolerite sheets. The 8m thick tuffaceous sediment found at 60m depth in drillhole H-2 by Tröllafoss (Friðleifsson and Tómasson 1972) is probably the same bed slightly downfaulted.

After this period of quiescence, volcanism became very active on the caldera rim and several volcanic vents, below or at the water level of the caldera lake, were erupting at the same time, giving rise to hyaloclastites. Picturesque feeder plugs of basaltic andesite and with very similar chemical compositions have been eroded out of this hyaloclastite in Bláhnúkur (J5), in southwestern Svínaskarð (K5), and in Múli (L4). It is not known whether there were any vents near the centre of the caldera lake, but the fine-grained hyaloclastites, which succeed the tuffaceous sediment in the Tröllafoss region, are also of basaltic andesite and it seems likely that at least the central, northern and eastern parts of the caldera lake were filled with basaltic andesite hyaloclastite. In Múli (L4) and in the northern part of the caldera the basaltic andesite hyaloclastite was laid directly on top of the down-warped tholeiite lavas (unit 19), whereas in the central region the coarser grained tholeiite hyaloclastite (unit 20) forms the base of the caldera filling and is separated from the basaltic andesite hyaloclastite by the tuffaceous sediment. It is feasible that the tuffaceous laminated sediment represents tephra falling into the caldera lake at the beginning of the Bláhnúkur and Múli eruptions, before the slower moving basaltic andesite pillow breccia had reached the bottom of the caldera lake. This would shorten the time interval between the eruptions of the tholeiite and the basaltic andesite hyaloclastite to days or weeks from years, if the sediment had been produced by accumulation of sediments from streams running down the caldera sides.

The basaltic andesite is everywhere in the form of pillow breccia, but the proportion of the glassy tuffaceous matrix to pillows and lithic fragments is variable. The only remnant of tephra build up in the crater regions is found northeast of Bláhnúkur where a light-green coloured, highly altered tuff with rare basaltic andesite fragments forms a ridge parallel to the caldera rim. Tholeiite lavas of unit 24 are seen to have piled against the steeply eroded, northern side of this ridge. The volume of the basaltic andesite hyaloclastite is estimated to be about 2.5km^3.

Events in the southern part of the caldera at this time are uncertain. There are several fine-grained plugs within normally magnetised hyaloclastites in northern Grímmannsfell. The author thought at first that these were of identical composition to, and contemporaneous with the Bláhnúkur and Múli plugs. But two of these plugs (chem.an.73 and 74) proved to be of olivine tholeiite and very similar to a small normally magnetised plug (chem.an.72) on the caldera rim at about 490m altitude in southwestern Skálafell (K5), which could, in fact, belong to the Gilsá event basaltic volcanism that produced the hyaloclastite higher up in Skálafell. Theódórsdóttir (1972) suggested that the normally magnetised hyaloclastites in northern Grímmannsfell belong to the same normal magnetic event as the rhyolite in Grímmannsfell, which the present author believes to be contemporaneous with the Móskarðshnúkar rhyolite, and of Gilsá event age.

While the Stardalur caldera was being formed and filled in eastern Esja, little or no volcanic activity characterized the area to the west of the Þverárkotsháls-Trönudalur hyaloclastite ridge of unit 18. A thin blanket of sediments was deposited over the lavas of unit 19. The thickness is less than 5m in the Kistufell (H4), Eilífsdalur (H7) and Flekkudalur (I8) regions, but increases southwards towards the hyaloclastite ridge and Stardalur, being about 20m thick in eastern Grafardalur*. The base of the horizon (e.g. in Eilífsdalur) often has small well-rounded pebbles set in a brownish matrix, but the characteristic feature of the horizon is its light colour. The main constituent is commonly yellow tuff which shows lamination and contains both vesicular basic and glassy intermediate fragments. The top of the bed is usually red. In eastern Grafardalur there is some stratification, with the

* This horizon is only shown on the geological map where it exceeds 10m in steep topography, or less in the gentle slopes in Flekkudalur. The horizon is omitted in the cross section on the geological map.

lowest part consisting mainly of basalt scoria, the central part mainly of light coloured tuff with both basic and intermediate fragments, and the top part mainly of basalt scoria. The light-coloured tuff probably originated in the basaltic andesite volcanoes on the caldera rim.

d. Rocks from the second reverse polarity period of the Matuyama magnetic epoch

Volcanic activity continued unabated after the caldera had been filled, and during the time interval between the Olduvai and Gilsá magnetic events (about 160000 years, Cox 1969) a variety of basic, intermediate and acid volcanics were erupted.

Unit 22. Tholeiite lavas, 10-100m thick.

After the formation of the Stardalur caldera, eruptions of tholeiite lavas continued until the next glaciation. Unit 22 is (in the present day exposures) confined to the region northwest of the Þverárkotsháls-Trönudalur hyaloclastite ridge. The magnetic reversal from normal to reverse polarity is recorded within this unit, only the first lava on top of the tuffaceous sediment of unit 21 in northwest Kistufell (G5), and the lowest four or so lavas of this unit in northern Möðruvallaháls (K8) having normal polarity. The polarity transition can also be observed in northern Nónbunga (H8).

A thin horizon of reversely magnetised pillow breccia (chem,an.59) is found at the base of the unit in western Eyjadalur (J8). It overlies waterlain sediment of unit 21, and could represent a lava which flowed into water.

The total thickness of the lava unit ranges from about 40m in southern Esja to about 100m in northern Esja, and the unit is seen to thin southeastwards against the Þverárkotsháls-Trönudalur hyaloclastite mountain region. The tholeiite lavas are fresh, and range in thickness from 2 to 20m (average 9m). The scoria between several of the lavas of this unit in western Eyjadalur (J7) is bright red, which may indicate proximity to the eruptive site.

Light-coloured tephra is found in the scoriaceous lava tops in several localities. Up to three such horizons have been found at a single locality. These are assumed to be airborne representatives of intermediate or acid volcanism, but the source volcanoes have not been traced.

Unit 23. <u>A minor rhyolite, and a basaltic andesite hyaloclastite, 10-240m thick.</u>

The rocks of unit 23 were produced during a major glacial episode and on a land surface that was relatively flat except for (by now) a minor hyaloclastite ridge reaching from Þverárkotsháls through Trönudalur and a slightly elevated caldera rim. The Stardalur central volcano was very active during this glaciation, but the central activity had by now migrated northeastwards to the Svínadalur-Trönudalur area.

The majority of the volcanic products were of basaltic andesite composition, but rhyolite was also erupted. At the end of the glaciation the whole of Esja was covered by a thick blanket of conglomerates.

In northeastern Múli (L7) a large composite sheet, with a southeasterly dip, is seen cutting through the normally magnetised lavas of unit 19. The sheet is reversely magnetised. It is probably a volcanic feeder and either preceded unit 23, or fed the first volcanic eruption that can be attributed to the unit. The basic part, which is of icelandite composition (chem.an.63), is seen only at the foot of the scree-covered northern end of the hill. The major part of the sheet is of rhyolite (chem.an.61 and 62) which appears as a light-coloured, slightly vesicular, felsitic rock, except on the highest part of the ridge, where it grades into glassy pitchstone with thin, drawn out vesicles, and frothy glass with elongated pitchstone wedges showing good flow alignment. No other exposures have been found of this rhyolite. There is no indication as to whether its eruption was subglacial or subaerial, but, if the latter, then it may well have been a source for one of the light coloured tephra horizons in unit 22.

A small exposure of an intermediate, perhaps icelandite, hyaloclastite is found at the head of the Svínadalur valley in a gulley that climbs westwards into the pass between Trana and Móskarðshnúkar (K6). In a part of this glassy material there is a suggestion of columnar jointing resembling that found in large pillow-like masses in icelandite and rhyolitic hyaloclastites. The rock is far too altered and crumbly for sampling and the magnetic polarity direction could not be established. This body, which is too small to be shown on the geological map, seems to underlie the basaltic andesite hyaloclastite.

Voluminous subglacial eruptions of basaltic andesite gave rise to a hyaloclastite pile which is thickest in Trana (K7) (about 240m), but apparently much thinner between Trana and Bláhnúkur (J5), so that at the end of the glaciation there was a valley between the northern rim of the Stardalur caldera and Trana with an east-west extension parallel to the caldera rim. This may be an erosive feature (a glacial valley), but, since the strike of cone sheets in the vicinity is nearly parallel to the valley, it is thought likely that the basaltic andesite hyaloclastite formed an arc-segment concentric with the caldera.

The hyaloclastite pile resulted from a number of eruptions, that show a slight chemical variation. A small plug (chem.an.64) outcrops in the gulley where the "icelandite" hyaloclastite is found (page 67), and a very much larger plug (chem.an.66) is seen in the western slopes of Trönudalur. There is a range from pillow lavas with the typical small (10cm diameter) basaltic andesite pillows through pillow breccias to entirely tuffaceous hyaloclastite in the slopes of Trana. The tuff is light-brown with small, black, glassy fragments. In the gulleys running westwards from the mouth of Trönudalur (L7), large, solid masses of crystalline rock (chem.an.65) are found amongst the pillows and pillow breccias.

The only relict of the basaltic andesite volcanism on the western side of Eyjadalur is found at about 540m altitude (J7). A 40m thick pile of basaltic andesite pillow lava (chem.an.67) is sandwiched between conglomerate beds, about 20m thick below and about 70m thick above. The pillow lavas thin southwards, but have not been traced to the northwest.

The retreat of the glaciers is marked by a thick blanket of fluvioglacial sediments which spread over the whole of Esja. Rutten (1958) frequently referred to this horizon as "the morainic horizon at the base of the Graue Stufe", and drew the Tertiary-Pleistocene boundary at the base of those sediments.

The sediments are thickest (more than 100m) furthest east, closest to the basaltic andesite area, but thin westwards. Southeast of the Þverárkotsháls-Trönudalur hyaloclastite ridge, against which the lavas of unit 22 banked from the northwest, the sediments succeed directly hyaloclastite unit 18 in Þverárdalur (I5) and basaltic andesite hyaloclastite unit 21 on the rim of the Stardalur caldera in Svínaskarð (K5) and northwest of the Bláhnúkur plug (J5). On both sides of Þverárdalur, sediments overlap the caldera fault zone.

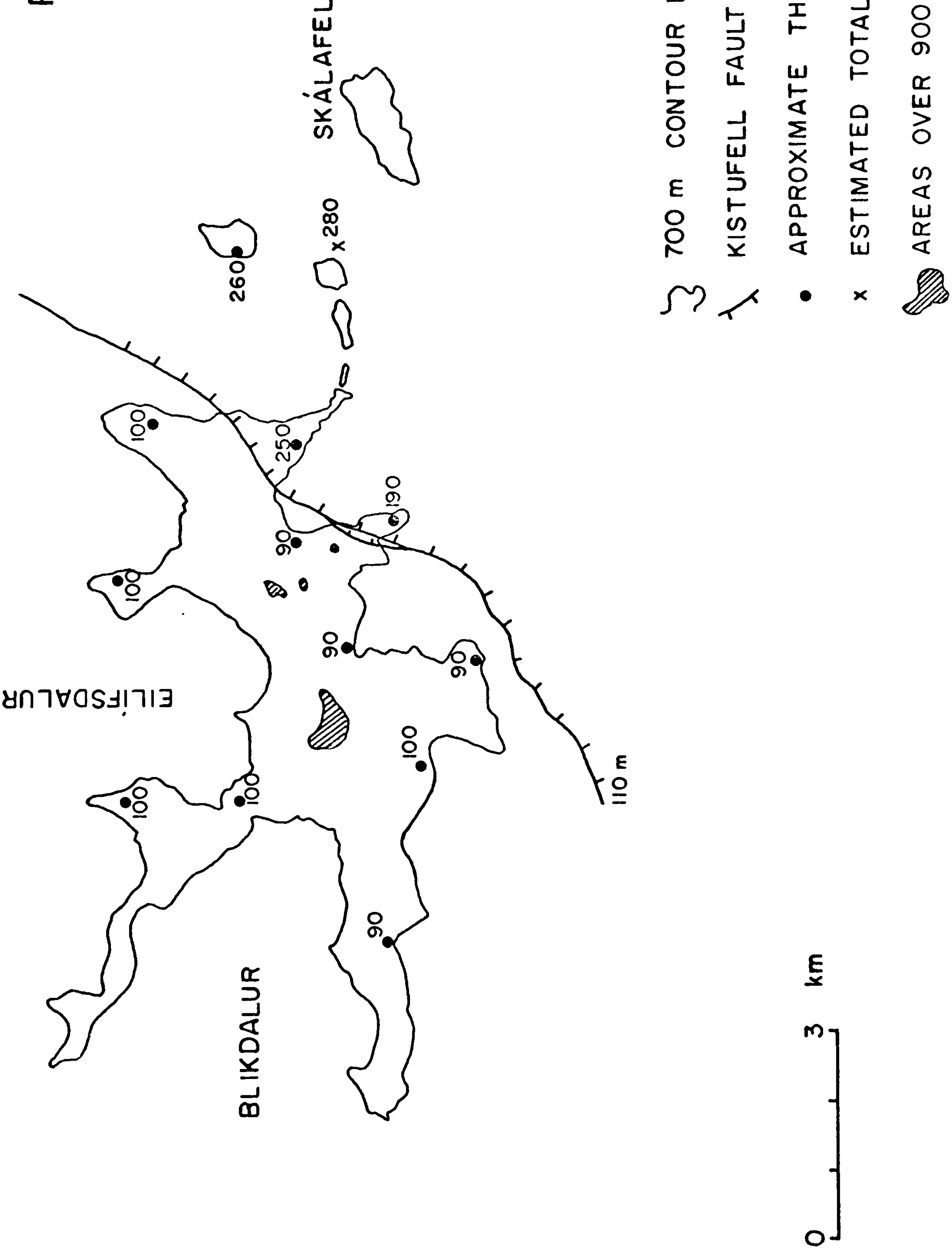
With the exception of a 3m thick lava flow in the middle of the unit at 670m in the slopes west of Hátindur (I5) and the basaltic andesite pillow lava mentioned previously (in Eyjadalur), the unit seems to be entirely sedimentary in origin to the west of Eyjadalur. The thickness is about 40m in Þverárdalur, Grafardalur and southern Kistufell, but is only 15m thick in northwest Kistufell (G5), and about 20m in northeast Gunnlaugsskarð (G6). The base of the unit usually consists of greyish, ill-sorted, coarse conglomerate with boulders often 20-30cm in diameter but occasionally up to 1m across. This is succeeded by a brownish zone with smaller and less numerous but well-rounded pebbles. The middle of the unit is often of well stratified conglomerate with rounded pebbles in a sandy matrix. The top part is more tuffaceous with fewer, usually well-rounded pebbles, and shows some stratification.

Unit 24. Tholeiite lavas, about 280m thick.

After the conglomerate blanket had been desposited, most of Esja was a flat plain. In the northeast, however, the basaltic andesite hyaloclastite mountain of unit 23 rose above the plain, and there was a valley trending east-west between Trana and the northern rim of the Stardalur caldera, which was still slightly elevated. The lavas of unit 24 filled this valley, submerged the Stardalur caldera and the Trönudalur hyaloclastite, and spread over the whole of the Esja plateau. These lavas form the present top of Esja.

The estimated total thickness of the unit in Svínaskarð (K5) is 280m. The approximate thickness of the unit at the highest altitude spot of Esja today is 270m. This suggests that the valley floor between Trana and the Stardalur caldera rim was level with the Esja plain to the west at the commencement of the lava eruptions of unit 24, and that the unit has suffered little subsequent erosion at the highest altitudes of Esja,

Fig.10


SKÁLAFELL
EILÍFSDALUR
BLIKDALUR
x 280
260
250
190
100
100
100
100
100
90
90
90
90
110 m
0
3 km
700 m CONTOUR LINE
KISTUFELL FAULT
APPROXIMATE THICKNESS AT 740 m
ESTIMATED TOTAL THICKNESS PRE-EROSION
AREAS OVER 900 m ALTITUDE

assuming a near uniform thickness of the unit at the end of the tholeiite lava volcanism. Fig.10 shows the approximate thickness of the unit at 740m altitude (chosen because of the spot altitude of Trana) and demonstrates the flatness of the pre-unit 24 plain as well as the tectonic inactivity (little overall vertical displacement on faults) of the area to the northwest of the Kistufell fault.

The crater region(s) from which the tholeiite lavas were erupted cannot be located with certainty. But thick accumulations of scoria and volcanic cinders intercalated with the lavas in eastern Svínaskarð (K5) strongly suggest the proximity of a crater region.

Some of the first tholeiite lavas of the unit are slightly porphyritic with plagioclase and less abundant pyroxene phenocrysts occurring in glomeroporphyritic groups; e.g. a lava resting on the basaltic andesite hyaloclastite in the eastern slope of Trana (chem.an.69) resting on top of unit 23 in southeastern Kistufell above Karl (H4) at 650m. At this latter locality, evidence is found for the lava having flowed into water, probably a shallow lake. The underlying sediment has small rounded pebbles, but becomes progressively finer-grained upwards and is a varved siltstone at the top. At the base of the lava is a bed (about 50cm thick) formed of small pillows, some of which are surrounded by the silty sediment. The pillows are cogenetic with the lava and are succeeded by 2-3m high columns (40-50cm in diameter) which grade upwards into smaller columns succeeded by hackly jointed kubbaberg type entablature (Sæmundsson 1970) which is 10-20m thick.

The majority of the lavas are of nearly aphyric tholeiite (chem.an.70 and 71). An average thickness of 5.1m was calculated from 137 lavas measured in the field and counted (binocular geology) in cliffs of known height. The lower half of the unit is near perfectly exposed in steep cliffs all around Esja, but the upper half has been carved and eroded to various extents during subsequent glaciations and can best be studied in Hátindur (I5) and to the east of the Kistufell fault.

Up to three sedimentary horizons have been found intercalated within the unit at any one locality, but these have not been traced or correlated in order to establish whether the sedimentary episodes were widespread or localised. In Kistufell all three horizons are conglomeratic, but the lowest one has a thick (8-19m where measured) resedimented tuff succeeding the conglomerate. The highest occurrence of a conglomerate is at 900m in Hátindur (I5) about 240m above the base of the unit.

The unit is reversely magnetised. The magnetic intensity is fairly high in the lowest 100m of the unit, becomes very low near the middle of the unit but increases again upwards. This must reflect variations in the Earth's magnetic field strength at the time as the petrology of the lavas is apparently (from field inspection) fairly uniform. The occasional, low intensity lava near the middle of the unit gave both reverse and normal polarity readings (samples from the same lava) in Kistufell, Hátindur and Móskarðshnúkar. Whether these ambiguous polarity measurements (made with a portable fluxgate magnetometer) represent a true polarity reversal(s) or should be attributed to "noise", can only be answered by laboratory treatment of the samples.

At the very top of the unit in western Skálafell (K5) and southern Móskarðshnúkar lavas have a weak normal polarity magnetism, which is assigned to the Gilsá normal magnetic event.

e. Rocks from the Gilsá normal magnetic polarity event

Unit 25. Olivine tholeiite hyaloclastites.

At the end of the tholeiite lava volcanism of unit 24 the whole of the Esja region was covered by a thick pile of lavas. Glaciation set in on the flat lava plain, and olivine tholeiite hyaloclastites were erupted just outside the area under investigation, i.e. in western Skálafell (K5) and probably simultaneously in northern Grímmannsfell (J2). Feeder plugs for the hyaloclastites are found in western Skálafell and in northern Grímmannsfell just outside the observed and inferred caldera rim respectively; the magma has apparently found an easy path in the caldera fault zone. The normally magnetised hyaloclastite underlies the rhyolite in Grímmannsfell (Theódórsdóttir 1972), and the rhyolite volcanism in other parts of the Esja region may well have started slightly later than the basaltic volcanism.

Unit 26. <u>Rhyolite hyaloclastite, 100m thick.</u>

Although the main remnants of the rhyolite volcanism today are restricted to Móskarðshnúkar (J5) and Grímmannsfell (I1), the distribution of rhyolite dykes shows that it has been widespread. The rhyolite dykes are conspicuously concentric with, and most of them dip (up to 45^{o}) towards the Stardalur caldera. Some are composite with thin basaltic margins and thick rhyolite centres. They range in thickness from 2 to 12m. Northwest of the caldera rim, the dykes are seen to cut right through the tholeiite unit 24, and there is little doubt that they erupted, perhaps producing domes similar to those seen in Móskarðshnúkar. The rhyolite volcanism either postdates or was partly contemporaneous with movements on the Kistufell fault, as a rhyolite dyke follows the fault path in eastern Grafardalur (J5).

The two westernmost rhyolite peaks in Móskarðshnúkar can be seen extending from their feeder dykes, and the third peak may also have a seperate feeder. The westernmost feeder (on the north side, Fig.11) began to expand within the topmost part of tholeiite unit 24. The feeder to the central peak seems to have cut right through the tholeiite lavas (Fig.12) before expanding. The rhyolite (slightly crystalline grey pitchstone) displays fairly regular columnar jointing (columns 10-50cm in diameter) forming a cliff about 10m high, but higher up and further north, on the strike of the feeder dyke, it is mainly in the form of regular sheets of black pitchstone (with the same dip as the feeder). The rhyolite screes are coarsest

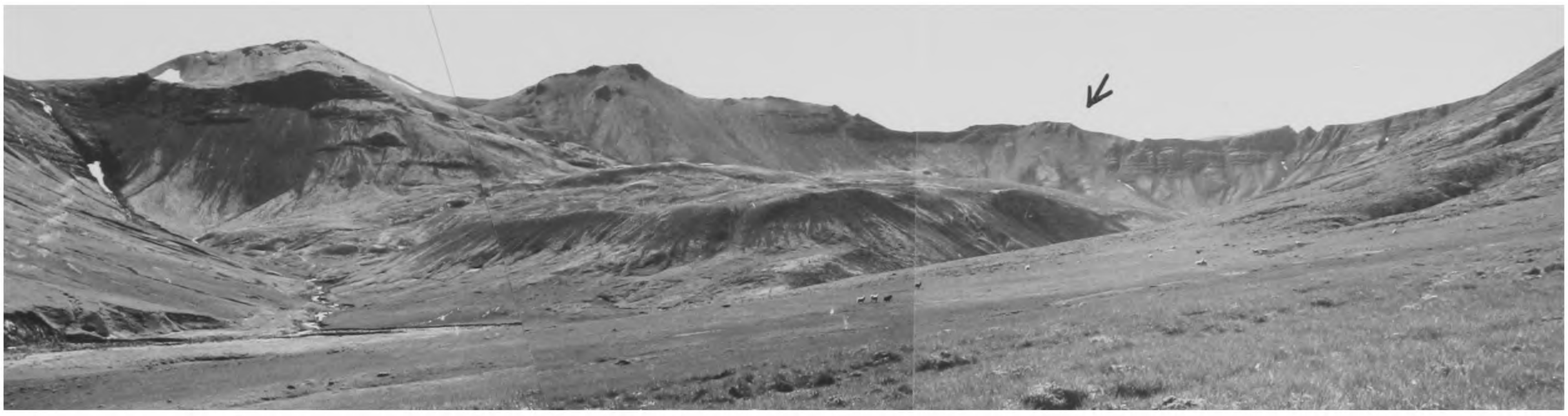


Fig.11. Móskarðshnúkar rhyolite peaks from the north (Eyjadalur). Note how the rhyolite feeder has expanded within the lavas of unit 24 in the westernmost (right) peak and compare with Fig.12.

Fig.12. The central rhyolite peak of Móskarðshnúkar from the west. Note how rhyolite feeder (light coloured scree on the right) has cut through the lavas of unit 24 (right and foreground) before expanding to form columnar jointed pitchstone cliffs (top right). The landslip in the easternmost peak is in the left of the photograph and further away is the top of Trana (unit 24).

closest to the exposed rhyolite columns where large pieces have broken from the columns. These large pieces subsequently break into smaller fragments and form thick screes. Contacts of the easternmost peak of Móskarðshnúkar with the underlying tholeiite lavas are completely covered by screes. An eroded platform southwest of the pinnacle at about 700m altitude is thought to represent the top of the tholeiite unit; the rhyolite reaches its maximum thickness of about 100m in this easternmost peak.

The core of the rhyolite body in Grímmannsfell is of dark grey, very slightly crystalline pitchstone (chem.an.78), which displays regular vertical columnar jointing. Surrounding this columnar unit are screes of thoroughly altered tuff (former granulated glass?) and rhyolite fragments similar to those in the screes of Móskarðshnúkar.

Spherulites are very common in the rhyolites and are of two types: albite spherulites and albite-quartz spherulites (Sigvaldason 1958). Their abundance, which is particularly noticeable macroscopically in the pitchstone, prompted Rutten (1958) to call it sago-obsidian.

The mode of extrusion of the Móskarðshnúkar rhyolite is by no means certain. Sigvaldason (1958), after studying flow structures in the rhyolite, refers to it as a quellkuppe (dome). Rutten (1958) rejects the quellkuppe idea and suggests that the peaks are formed by multiple intrusion of rhyolite sheets. No outcrop of rhyolite lava has been found in Esja. The only flat lying rhyolite bodies found are a normally magnetised, and therefore intrusive, felsitic body (chem.an.76) within the reversely magnetised lavas of unit 16 in the Grafardalur valley (I4), and a very small lenticular body (visible thickness 1m) of rhyolite within the tholeiite lavas of unit 24 on the ridge crest just north of Hátindur (I5). The slopes of the ridge are scree covered on both sides, so a possible feeder would be obscured.

Subglacial rhyolites consist of rock varieties ranging from granulated glass to columnar lithic lobes encrusted with pitchstone (Sæmundsson 1972). The size of the lobes, which are the equivalents of pillows in basaltic subaquatic volcanics, may vary from 1 to 100m, and their shape can be spherical, ellipsoidal or flattened and almost dyke-like. The crystallinity of the rhyolite varies with the size of the lobes; in a large lobe a gradual change from glassy pitchstone to fairly lithic rhyolite can be observed. Small differences in texture and vesicularity change the appearance of the rock markedly. The grey granulated rhyolite glass, which is usually the most abundant material in subglacial rhyolites, is not consolidated like the glass of the basaltic hyaloclastites (Grönvold 1972), and hence easily eroded.

Aspects of the Móskarðshnúkar and Grímmannsfell rocks such as the black pitchstone, the grey slightly crystalline rhyolite columns, and the platy, often vesicular, fragments of the screes, which range from dark, glassy material to light-coloured, more lithic rhyolite (disintegrated lithic lobes), compare well with descriptions of subaquatic rhyolite.

The rhyolite bodies in Móskarðshnúkar and Grímmannsfell are considered the remnants of large lithic lobes formed immediately above the feeder dykes. They probably formed the cores of much larger rhyolite hyaloclastite mountains, which consisted largely of granulated glass and small lithic lobes, and were easily eroded away.

f. Alteration and erosion

Secondary alteration is less intense in the northeastern part of Esja than in the vicinity of the central volcanoes in the south and west. The high local thermal gradients associated with the centres are clearly demonstrated by the distribution of the secondary minerals and by colour changes in the rocks. The most intense effects are seen in southwest Esja where the rocks have suffered intense hydrothermal alteration; small crystals of pyrite and quartz, and large crystals of platy calcite are abundant, but epidote is rare. The areal extent of the propylitization is indicated on the geological map. The hyaloclastites in the propylitized area are light coloured, due to the alteration, and light coloured narrow, vertical channels extend from the hyaloclastites up into, and in some instances up through, the much darker (less altered) tholeiite lavas on the ridge (D6) west of Kerhólakambur; these are probably funnels through which steam and hot water penetrated towards and, in some cases, reached the surface. The eastward extension of the propylitized area in Blikdalur (E6) is unknown owing to lack of exposures. In the thick hyaloclastites in Hrútadalur (F6), stilbite,* heulandite and apophyllite are found as well as platy calcite. This is probably the northeastern limit of the high temperature area associated with the Kjalarnes central volcano. Both in Hrútadalur and in Kerlingagil (E8) intensely altered patches, reminiscent of those in present day hydrothermal areas, are seen to be associated with faults, and the crumbly soil still smells of sulphur. In the lavas north of Kerlingagil, chabazite, thomsonite, analcime, levyne, stilbite and scolecite are found as well as platy calcite, chalcedony** and quartz, and in the whole area westwards to the sea and southwards to Kjalarnes, quartz, opal and chalcedony are abundant, reflecting the mixing of thermal waters with cold ground water (Arnórsson 1970) on the outskirts of the high temperature area. The southeastern limit of the propylitized area is on the eastern margin of the Þverfell dolerite sheet where calcite veins up to 40cm thick are found in the intensely altered lavas; pyrite is abundant.

A maximum value can be estimated for the life span of the Kjalarnes hydrothermal system. Chabazite crystals are found at the base of the tholeiite lavas (post-Olduvai) capping the strongly altered units 14 and 15 in Kerhólakambur (E5) indicating

* Field identification of secondary minerals was supplemented by about 100 XRD powder photographs.

** Chalcedony amygdales have nearly horizontal onyx-type banding in the steeply dipping lavas indicating that thermal activity, probably below 100^{o}C (Arnórsson 1970), persisted for some time after the tilting occurred.

a "normal" thermal gradient there some 200,000 years after the Kjalarnes volcano became inactive.

Intense hydrothermal alteration in the Stardalur area is restricted to the (exposed parts of the) caldera fault zone and to the parts within the caldera where cone sheets are most abundant, such as west of Haukafjöll (J4) and south of Skeggjastaðir (J3). Platy calcite, pyrite and quartz are common, but epidote (identifiable with a hand lens) is rare. North of the caldera rim, large crystals of heulandite are fairly common, and west of the caldera an analcime zone is fairly well developed at about 400m altitude in Grafardalur and in the southern slopes of Kistufell; stilbite, scolecite and heulandite are found in the lower slopes of Kistufell.

In northern Esja the secondary mineralization is less affected by the central volcanoes. Stilbite, scolecite and mesolite are found up to about 300m altitude in western and southeastern Eilífsdalur, but the upper limit of the mesolite-scolecite zone (Walker 1960) apparently drops to a lower level northeastwards from Skálatindur (H7). Chabazite and thomsonite are commonly found below and in the lowest part of the top-plateau series of Esja, but rarely found above 700m altitude.

Judging from the level of the analcime zone (at about 400m) in Kistufell and from the rare occurrence of zeolites above 700m altitude it is considered unlikely that more than 200m have been eroded off the present highest top of Esja. It is suggested that the last major eruptives were deposited on Esja during the glaciation at the beginning of the Gilsá normal polarity event. The hyaloclastites (basic and acid) probably sheltered the Esja region from lava flows advancing from the active volcanic zone further east during the next interglacial, but were eroded in subsequent glaciations.

Erosional features of the Esja region will not be discussed here. Attention is, however, drawn to the landslips north of Móskarðshnúkar (J6) and the Mógilsá landslip (F4), which are shown on the geological map. A much smaller, but interesting landslip is seen in Kvensöðlar (D5) where large dislocated blocks of lavas have slid down the mountain slope and come to rest showing very steep and most irregular dips.

g. Post-erosional volcanism

The last volcanic eruption known to have occurred within the Esja region produced the Mosfell (H3) olivine tholeiite hyaloclastite mountain, which is situated just outside the inferred caldera rim of Stardalur. The eruption occurred after the Esja region had been eroded essentially to its present level, and possibly as late as during the third last glaciation. During the second last interglacial (Th. Einarsson 1968) olivine tholeiite compound lavas were erupted some 10km east of the Esja region. The lava, which spreads over the lowlands south of Esja and all the way to the sea, where its base is in the form of pillow deltas, was probably erupted in the Borgarhólar crater in Mosfellsheiði. The distribution of the lava suggests that the landscape in the Esja region at the time was essentially the same as it is today. Chemical analyses were made of samples from Mosfell (chem.an.79) and from the lava (chem.an.80) in the course of the present work. Jónsson (1972) has briefly discussed the distribution and petrography of post-erosional olivine tholeiite compound lavas near Reykjavík.

Chapter 3 INTRUSIVE ACTIVITY

Igneous intrusions in Esja are most heavily concentrated in the two volcanic centres, Kjalarnes and Stardalur, although within every unit of the Esja stratigraphic column intrusions of one kind or another can be found. There is considerable variety in the shapes of intrusions and they range from a few centimeters to hundreds of meters in size. In Esja there was a steady "eastward" migration with time of the intrusive as well as the extrusive activity. This reflects the crustal drift away from the active volcanic zone of southwest Iceland, which may have remained stationary for the last several million years.

In the following account the main points of interest are a) the migratory aspect of the intrusive activity, and b) the interrelation of the size and shape of the intrusive bodies with the lithologies of the host rocks. The magnetic polarity direction of the intrusives is indicated on the geological map where these have opposite polarities to the host rock, and where there is a cluster of dykes of opposite polarities.

a. Intrusives older than the Kjalarnes dolerites

The oldest intrusives in Esja are normally magnetised and date from the Gauss magnetic epoch. The number of normal polarity dykes decreases upwards within the pile of normal polarity units 1 to 8. The dykes correspond petrologically to the lavas, and there is little doubt that many of the dykes are feeders. The actual percentage dyke dilation can rarely be measured (because of lack of continuous exposures), but the figures of 3% and 6% dilation obtained within unit 2 north of Melafjall (E9) and at the mouth of Blikdalur (C7) respectively suggest more active volcanism in the southern part of the region. The dilation at the base of unit 1 west of Hjarðarnes (C8) is over 10%. The dykes range in thickness from a few centimeters to 5m, and have an average thickness of 2.2m which compares well with the average dyke thickness for the whole of the Esja volcanic region (Fig.13). The dykes trend N25°E.

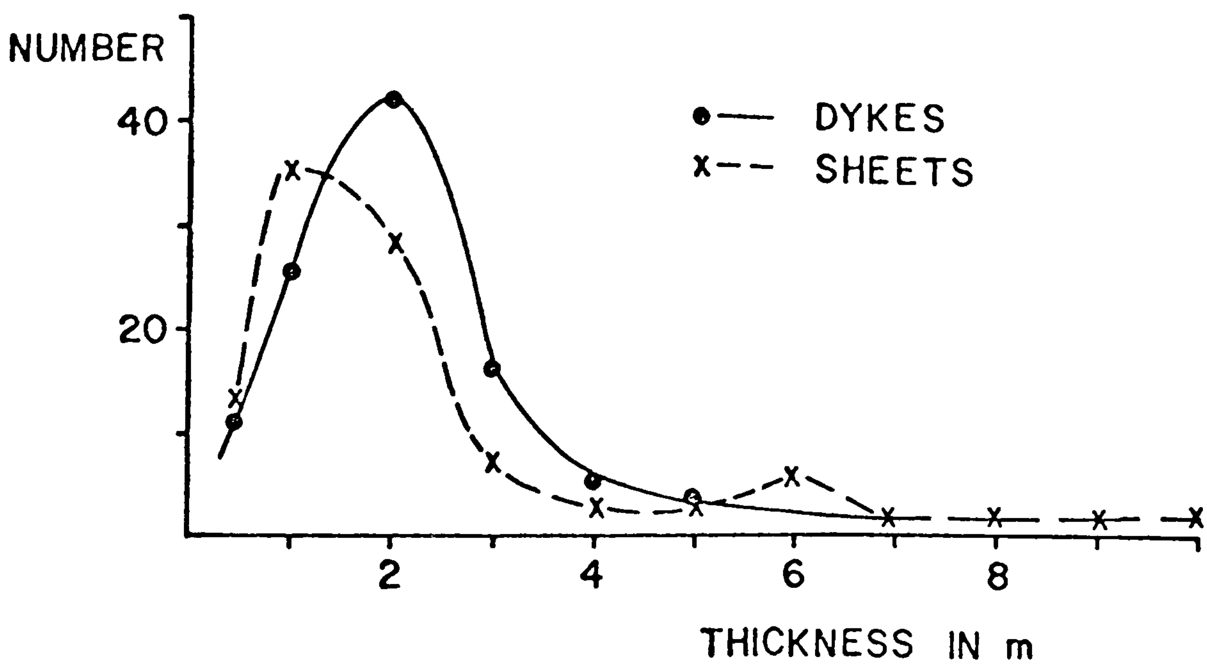


Fig.13. Average thickness of dykes and sheets ($\leq$ 10m) measured in Esja.

A large number of low angle sheets and sills are found, especially within units 1 and 2. The sheets dip towards the Kjalarnes peninsula and indicate that high level magmatic activity had started in the Kjalarnes central volcano. There is a range in thickness from thin veins to 6m thick sills. The large majority of the sheets and sills are of very fine-grained tholeiite. The sills are more abundant within the softer olivine tholeiite lava piles than within the hard and brittle tholeiites.

There is a striking contrast between the intrusives within the lavas and those within the hyaloclastites. The soft and "structureless" hyaloclastites break irregularly when subjected to stress, and the intrusive magma thus forms anastomosing masses. There is a tendency for the magma to spread out laterally to form either irregular sheets or even sill-like bodies within the hyaloclastite, rather than to penetrate upwards into succeeding beds. In the brittle lavas, on the other hand, parallel cracks are formed, giving rise to more regular, uniformly thick and parallel sided intrusives.

A good example of intrusive behaviour in hyaloclastite is seen in unit 4 in Melafjall (D8). In a gulley trending NW-SE, three dykes can be seen cutting through waterlain sediment at the base of the hyaloclastite unit and feeding a sill above (Fig.14) which is at least 10m thick (top not seen). The sill (chem.an.6) is of coarse-grained plagioclase-pyroxene-phyric tholeiite. Petrographically similar dykes are found cutting the underlying lava units; e.g. a 4m thick dyke in unit 2 northeast of Tíðaskarð (chem.an.5).

Fig.14. Melafjall sill with three feeder dykes (1,2,3) cutting through laminated sediment 4, which grades into pillow breccia 5.

A 30m thick sill of fine-grained plagioclase-phyric tholeiite (chem.an.7) is found in the same hyaloclastite unit 4 at 200m altitude east of Ártún (C7). This sill has a similar chemical composition to the porphyritic tholeiite lavas (chem.an.8) at the base of unit 5 in Melafjall (E8), and could well be contemporaneous. A feeder was not found for the sill, but standing out of the screes some 10m further down the slope is a 10m thick mass with subvertical columns and a chilled margin against the hyaloclastite (upslope), which suggests that the intrusion extended further to the west.

Intrusive activity continued in much the same manner at the beginning of the Matuyama as it had been during the last part of the Gauss epoch. Many reversely magnetised dykes with trends near N25^{o}E are found up to unit 9, and the reverse polarity sheets exposed on the Hvalfjörður coast dip towards Kjalarnes as did the normal polarity ones. This phase of the intrusive activity reached its culmination in the intrusion of the large dolerite sheets on the Kjalarnes peninsula (B5).

b. The Kjalarnes multiple dolerite sheets

It is thought possible that the intrusion of the multiple sheets on Kjalarnes was preceded by a caldera collapse in the Kjalarnes area. Only a small arc of the hypothetical caldera fault zone is seen north of Brautarholt farm (B6). The brecciated basalt in the fault zone has clearly been intensely heated as a thoroughly baked and hardened breccia forms a 5-6m broad promontory which extends several tens of meters out into the sea. A reason for suggesting that this represents a caldera fault zone is that the reversely magnetised, very fine-grained, tholeiite pillow lava and pillow breccia which forms the tip of the Kjalarnes peninsula (chem.an.9) and is also found forming a hummock (chem.an.10) in the fault zone north of Brautarholt farm, succeeds normally magnetised tholeiite lavas in Borgarvík (B5) which cannot be younger than unit 8 (the uppermost normal polarity unit). The pillow breccia must have formed in a subaquatic eruption. One possibility* is a caldera lake.

The Kjalarnes dolerite intrusion is formed of multiple sheets of very coarse-grained dolerite intruded along the boundary between the fault zone and the pillow breccia. Individual sheets are up to 30m thick and together measure about 350m. Identical, isolated sheets, each up to 6m thick, are found both north of the fault zone and within the fine-grained tholeiite pillow breccia on the tip of the Kjalarnes peninsula.

The dolerite sheets are composed of large crystals of plagioclase (30-50% modally), pyroxene (25-40%) and euhedral to skeletal ilmenite (10-15%) and have an ophitic texture. The grain size of the mesostasis (10-20%) minerals decreases towards the margins of the sheets. The larger sheets show fairly regular layering (10-30cm layers), which arises from slight variations in the proportions of pyroxene and plagioclase crystals, and of mesostasis; the lightest coloured rock has the highest percentage of mesostasis.

The sheets appear to have been intruded as crystal-mushes, the crystals having suffered strain when the individual mush-sheet

* An alternative possibility, neither favourable nor unfavourable to the caldera hypothesis, is that the pillow breccia formed during the unit 10 glaciation. This would imply a sheltering of the Kjalarnes area while the reverse polarity lavas of unit 9 flowed. Support for the latter possibility is perhaps found at the foot of the cliff north of Presthúsatangar (B5) where a stratified conglomerate can be seen underlying the tholeiite pillow breccia. The conglomerate indicates an interval of time between the formation of the normal polarity lavas in Borgarvík and the pillow breccia. The length of the interval is, however, difficult to evaluate.

was emplaced and/or when the neighbouring sheets in the multiple intrusion were intruded. The mechanical deformation of the crystal mush is spectacularly demonstrated by the elongated augite crystals which are bent through up to 90° without breaking (Fig.15). Similar arcuate augites occur within multiple intrusives on the island of Viðey (Peacock 1926) which lies 6km to the south of the Kjalarnes peninsula (B2), and which may also lie within the Kjalarnes caldera. Additional evidence for sill formation in hyaloclastites is found on Viðey where Peacock (op cit) mapped a large multiple sill (at least 37m thick) and several dolerite dykes intruding a hyaloclastite series, the oldest formation on the island. It is not known whether the Viðey sill was intruded at the same time as the Kjalarnes sheets or whether it is closer in time to the Þverfell sheets.

Fig.15. Bent augite crystal in a Kjalarnes dolerite sheet believed to have been intruded as a crystal mush. Note that in crossed polars both ends are in extinction demonstrating bending through 90°. The crystal is 3mm long.

c. Reverse polarity dykes younger than the Kjalarnes dolerites

After the intrusion of the Kjalarnes dolerite sheets the trend of the dykes changed to N40°E (from N25°E). The first eruptions which can reasonably be correlated with dykes having this new trend are those of hyaloclastite unit 10 in Kerlingagil (E8). There is a range in thickness from thin veins to 10m thick dykes, the latter occurring southeast of Ártún (C7). The large majority of the dykes is of tholeiite, but one 3m thick dyke trending N40°E in the cliffs east of Tíðaskarð (C8) consists of icelandite (chem.an.12). This is the most evolved rock found in western Esja.

The minimum width of the volcanically active belt at any given time was apparently 5-10km. The greatest dyke dilation, about 20%, is found in the slopes west of Kerhólakambur (D5). The dykes in this swarm are feeders for units 11, 12 and 13. Relatively few dykes have been found in northern Esja. This is certainly largely due to the scarcity of outcrops, but may also reflect the greater number of eruptions in the central volcanic region.

d. The Leiðhamrar, Þverfell and Lauganípa intrusions

For the second time in the geological history of Esja the igneous activity culminated in the formation of large dolerite intrusions, this time in the area between Skrauthólar (D5) and Mógilsá (F4). The oldest of these intrusions is probably the Leiðhamrar dolerite (E3), which intrudes a hyaloclastite assigned to unit 12. Columnar jointing in this very coarse-grained, almost gabbroic, plagioclase-phyric tholeiite (chem.an.27) body is nearly vertical, which indicates a horizontal sill-like form. It is tempting to suggest a correlation between the feldspar-phyric lavas of unit 13 in Sneiðar (H9) and Kerhólakambur (E5) with this intrusion, the lavas being (expectedly) considerably poorer in plagioclase phenocrysts than the dolerite.

The Leiðhamrar intrusion was accompanied or followed by a large number of intrusions, consisting mostly of sheets of olivine-free tholeiite, which can best be inspected in the Gljúfurdalur gulley by Esjuberg farm (E4). The sheets are commonly of the order of 20m thick, but much thinner where they cut the lavas of unit 13 in the Mógilsá river traverse (F4). The transition from nearly pure intrusives* to mostly host rocks (lavas) is fairly sharp (about 100m), and it is very likely that the intrusive sheets are largely accommodated on the boundary of the hyaloclastites (units 10 and 12) and the lavas (unit 13).

* The inferred boundary between the dolerite and the lavas in Gljúfurdalur (E4) is drawn on the geological map where the sheets are estimated to form less than 50% of the outcrop.

Associated with the dolerite sheets in the Þverfell-Lauganípa area were a swarm of tholeiite cone sheets (range in thickness from veins to 2m) which can be seen in the slopes and on the beach west of Kerhólakambur, and both east and west of the Mógilsá landslip. The sheets are most spectacular where they cross cut the hyaloclastite and lavas of units 10 and 11 west of Kerhólakambur, but are also found to intrude unit 13. The sheets dip towards the Kollafjörður bay and apparently have a centre of origin about 7km east of the centre for the Kjalarnes sheets.

The climax of the sheet intrusive activity was reached with the large olivine tholeiite dolerite sheets in Þverfell (E4) (chem.an.31) and Lauganípa (D5) (chem.an. 29 and 30). These intrusions certainly postdate unit 13. The lavas of that unit are thought to have been uplifted when the Þverfell sheets were intruded, and thus aquired their northerly dip in Gljúfurdalur. The Lauganípa sheet may represent the magma chamber which fed the olivine tholeiite unit 14.

The Þverfell sheet (Fig.16) is probably over 200m thick. The estimated total thickness of the dolerite intrusives in the Esjuberg - Leiðhamrar area is about 1km. The area between the Þverfell and the Lauganípa sheets is heavily faulted and has not been studied in any detail. Whether there was a physical contact between the two sheets has not been established, but it is thought very probable that the sheets were intruded more or less simultaneously.

Fig.16. Þverfell (E4) dolerite sheet from the west. Note the steep northerly dip of the lavas (unit 13) in Gljúfurdalur (left); this may have been caused by uplift associated with the intrusions in the Leiðhamrar-Esjuberg area.

The Lauganípa sheet has apparently been intruded into the lavas near the hyaloclastite slope against which lava unit 13 banked. The sheet has a maximum thickness of about 200m, and there are a number of smaller dolerite sheets associated with it, especially in the slopes southeast of Kvensöðlar (D5). The sheet terminates at about 600m altitude in western Kerhólakambur. Some of the lavas have bent at the contact with the dolerite and apophyses of the dolerite intrude the lava pile.

e. Discussion on the intrusives of the Kjalarnes central volcano

If the correlation of the Esja stratigraphic column with the palaeomagnetic time scale of Cox (1969) is correct and units 1 and 2 are taken to represent about 200,000 years, then the life span of the Kjalarnes central volcano was about 0.6 million years. For the first 0.3 million years the main centre was to the south of the Kjalarnes peninsula, but intrusive activity migrated eastwards and a secondary centre was formed in the Kollafjörður region (Fig.17). The two intrusive centres, as defined by the cone sheets, do form a largely continuous intrusive core over 10km broad, and are therefore taken to represent one central volcanic complex with a "siamese twin" intrusive centre.

The majority of the core of the Kjalarnes central volcano is now either under sea level or covered by the post-erosional olivine tholeiite lava, but its dimensions can to some extent be deduced from the geophysical anomalies (page 146), which have been recorded within the region. Both the positive gravity anomaly (Einarsson 1954) and the negative magnetic anomaly (Sigurgeirsson 1970) are elliptical with the long axis aligned northwest-southeast (Figs. 28 and 29), and extend from the Kjalarnes peninsula through the island of Viðey and south of Geldinganes. The anomalies are about 16km long, but only 5-7km broad. An unpublished aeromagnetic map of the Reykjavík area made by a Canadian team in 1959 (Sigurgeirsson 1967) shows within the main magnetic anomaly several local steep gradient negative anomalies, the strongest of which is located southwest of the island of Þerney. One of these local anomalies coincides with the outcrop of the multiple dolerite sill in Viðey, and it is thought likely that all of them are connected with reversely magnetised intrusives within the altered hyaloclastites. Several coarse-grained intrusives have been penetrated by drilling near the southeastern margin of the anomalies (Tómasson, pers. comm.).

The regional fault trend in central and eastern Esja is N30°E - N40°E. There are, however, several deviations from this trend in the Kjalarnes central volcanic region. Geological mapping of the area east of Geldinganes (D1) (Thors, pers. comm.) has revealed that the faults with the greatest displacement have a trend almost N-S. These faults are cut by younger NE-SW faults. The downthrow of the earlier faults is almost invariably on the western side and the greatest displacement (about 100m)

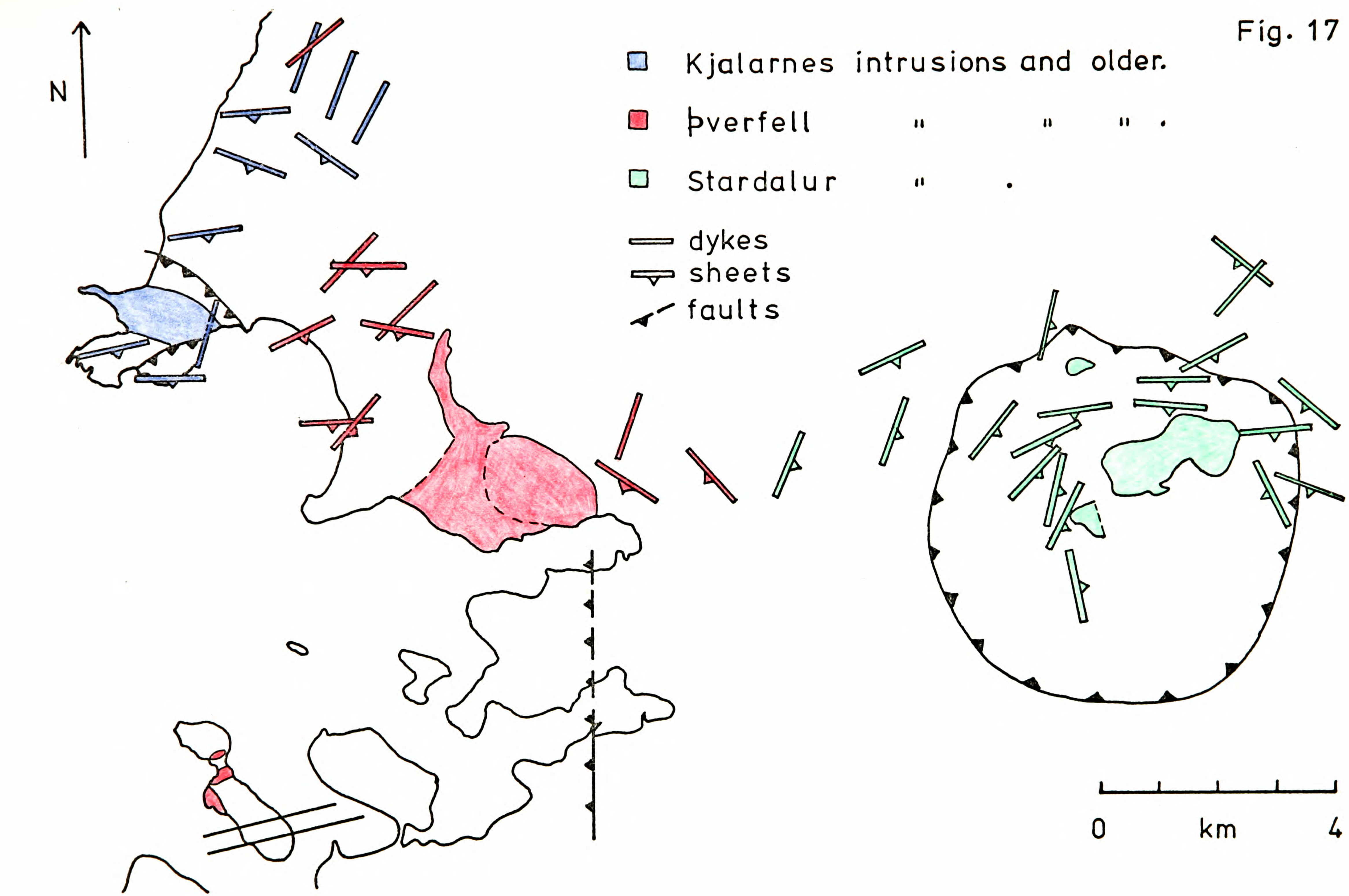

N
Kjalarnes intrusions and older.
þverfell " " " .
Stardalur " .
dykes
sheets
faults
0 km 4

Fig. 17

is observed in the westernmost of these faults. The northern extension of this fault and the fault trend on Viðey is shown in Fig. 17. The disposition of these faults and of the fault zone in the Kjalarnes peninsula makes it possible to visualize the curved northern and eastern margins of the Kjalarnes complex, the geophysical data providing a picture of its overall size and shape (Figs. 28 and 29).

f. Pre-caldera intrusives in eastern Esja

The oldest known sheets dipping towards the Stardalur centre are the tholeiite sheets with 50° southeasterly dip which are suggested as possible feeders for the uppermost lavas of unit 13 in Kistufell (page 41). It is not, however, until in hyaloclastite unit 15 that Stardalur-orientated sheets are found frequently. Within and stratigraphically above this unit, sheets are much more abundant than dykes.

The magnetic polarity measurements of the intrusives demonstrate clearly the increment in the intrusive activity just prior to, during, and just after the caldera formation since normal polarity sheets, intruded during the short Olduvai event, are far more numerous than reverse polarity sheets pre- and post-dating that event.

g. The Stardalur cone sheets

The bulk of the Stardalur cone sheets were probably intruded just after the caldera collapse, as the sheets often have similar or only slightly higher dips than the tilted lavas within the caldera. The sheets are most commonly between 1 and 3m thick where they intrude lavas, but they expand to form 10-20m thick sheets where they intrude the tholeiite hyaloclastite (unit 20) formed in the caldera lake. The hyaloclastite had been very recently formed and was presumably scarcely consolidated when the sheets were intruded.

As can be seen on the geological map the cone sheets are most abundant in outcrops which lie roughly on a circle with a diameter of about 4km. These outcrops are probably not fortuitous, and may represent a zone of greatest sheet intensity. In some of these outcrops the sheets form over 50% of the exposed rock. In the 240m deep drillhole H-2 by Tröllafoss, which is located about 0.9km west of the centre of the hypothetical circle mentioned above and well inside of the zone of highest sheet intensity, a log made from drill chips (Friðleifsson and Tómasson 1972) shows coarse-grained dolerite intrusives forming 18% of the rock. In the 90m deep drillhole H-3 (the uppermost 29m are of post-erosional formations) which is located 3.2km west of the centre of the circle, a coarse-grained intrusive was found only at the bottom of the hole and forms 16% of the "old rock" penetrated. It should be noted, however, that these logs are made from drill chips collected at 2m intervals and fine-grained intrusives may have been misidentified as lavas and lithic masses in the hyaloclastite. The percentages of intrusives given above are therefore minimum values.

The large majority of the cone sheets have dips between 20^{o} and 40^{o}. By projecting accurate dip measurements, poles plotted for the sheets at various depths (500m, 1000m etc.) were found to cluster most closely at depths of 600-700m under the present surface. This is thought to indicate the upper margin of a more or less continuous layer of intrusions under the central and eastern part of the caldera. Seismic measurements (Pálmason 1971) indicate crustal layer 3 at a depth of 500-600m in this area, as will be discussed further on page 169.

The Stardalur cone sheets vary in grain size from fine-grained tholeiite sheets, which are usually less than 2m thick, to coarse-grained (sometimes "sugary") tholeiite dolerite sheets, which range in thickness from less than 1m to over 20m. Despite the "picritic" look of the coarse-grained sheets in the field, olivine crystals have rarely been found in thin sections of the samples collected in the course of this study. The coarse-grained sheets are found to contain approximately equal modal amounts of pyroxene and plagioclase, and the ore is commonly in the form of skeletal ilmenite. In the fine-grained sheets, on the other hand, there is a much higher percentage of plagioclase than of pyroxene. Basaltic andesite sheets have been found only in the caldera fault zone east of Þverfell (K5).

The commencement of intrusion of coarse-grained sheets may well have coincided with the caldera collapse. It is certain, however, that they were being intruded until after the caldera was filled with basaltic andesite hyaloclastite unit 21, which they intrude both at Tröllafoss (J4) and in Þverfell (J5).

As mentioned on page 99, the sheets expanded in volume where they intruded the tholeiite hyaloclastite in the caldera filling. The best examples of this are seen in the gorge of the Leirvogsá river west of Tröllafoss.Some of the sheets here expand to form sills with weakly developed, nearly horizontal layering, the largest of these occurring near the contact of the tilted lavas (unit 19) and the hyaloclastite. Further east, within the basaltic andesite hyaloclastite, the sheets are thinner and more regular. The thickest intrusion (16m) penetrated by drillhole H-2 at Tröllafoss was found to underlie the tuffaceous sediment which is thought to seperate tholeiite hyaloclastite unit 20 and the basaltic andesite hyaloclastite unit 21. The three intrusives penetrated at greater depth in the tholeiite hyaloclastite are each 8-9m thick. It therefore appears that the intrusives expand where they enter the softer hyaloclastite, as demonstrated by the sills at the contact of the lavas (unit 19) and the hyaloclastite (unit 20), and also where they meet a more coherent rock such as the basaltic andesite hyaloclastite pillow breccia (unit 21).

h. <u>The Gráhnúkur and Stardalshnúkur intrusions</u>

There are two large dolerite intrusives within the caldera, the Gráhnúkur sill and the Stardalshnúkur laccolith. Both have normal magnetic polarity and could have been intruded either at the end of the Olduvai event or during the Gilsá normal polarity event.

The Gráhnúkur tholeiite dolerite (chem.an.57) sill (J5) intrudes the brecciated lavas of the caldera fault zone. It covers an area of about 0.1km^2, but may have been considerably larger before it was eroded. The exposed thickness of the sill is about 90m; columnar jointing (about 20m high columns) is well developed at the lowest exposed level of the sill, and weak horizontal layers are weathered out in the lower part of the columns.

The Stardalshnúkur olivine tholeiite dolerite (chem.an.58) intrusion (Fig.18) is saucer shaped and intrudes the tholeiite hyaloclastite of unit 20. It covers an area of about 2.3km^2. The base of the eastern part of the intrusion is reflected in the topography on the southern side where gulleys are carved in the easily eroded hyaloclastite, but the dolerite intrusion higher up forms steep columnar cliffs up to 70m high. The columns are up to 2m in diameter, and their regular fan-shaped disposition in the whole of the intrusion indicates cooling perpendicular to the surface of a saucer shaped body. In the northeast the intrusion is probably accommodated along the contact of the hyaloclastite and the underlying steeply dipping lavas of unit 19; this is reflected on the northern side of

Fig.18. The eastern half of the Stardalshnúkur intrusion from the southwest. Deep gulleys (right) are carved in hyaloclastite unit 20.

Fig.19. Layering in northern Stardalshnúkur (from the west) which may represent the northern edge of the roof of the intrusion. The layering may be due to variation partly in mineral proportions and partly in grain size.

the intrusion where its base appears to rise in altitude eastwards as it gets closer to the contact of the hyaloclastite and the dipping lavas. The base of the western part of the intrusion is not exposed. Its roof is completely eroded, but regular layering (Fig.19) dipping 24° northwards in the northernmost part of the intrusion may be the northern edge of the roof. The maximum exposed thickness is about 100m; the total thickness may have been 150-200m in the eastern part, but possibly considerably more in the western part. The dolerite is coarse-grained with euhedral olivine crystals (0.1-2.0mm in size) sitting in a patchwork of poikilitic augite (commonly up to 5mm in diameter) and plagioclase laths (mostly 0.1-1.0mm long). Olivine accumulation has apparently occurred, but this has not been studied in detail.

i. Late intrusives in eastern Esja

Very few reverse polarity intrusives post-dating Olduvai have been found in eastern Esja, and those found are mostly sheets assigned to unit 23. It is considered possible that feeders of unit 24 lavas have been concentrated on arc fractures concentric with but just outside the caldera rim, probably mainly on the northern side.

Several normal polarity dykes and sheets have been found post-dating unit 24; these are assigned to the Gilsá normal polarity event. It has already been mentioned (page 74) how the olivine tholeiite plugs (unit 25) in Skálafell and Grímmannsfell are apparently associated with the caldera fault zone, and how the rhyolite sheets of unit 26 are conspicuously concentric with and commonly dip towards the Stardalur caldera. A few normal polarity basic dykes have been found cutting unit 24 lavas, and irregular normal polarity basic veins and thin sheets (up to 1m thick, generally striking northeast and dipping southeast) have been found as far west as western Kistufell (G5). These have not been mapped in detail, but it appears that volcanism was more widespread during the Gilsá event in the Esja region than during units 23 and 24 times. Whether the normal polarity dykes and sheets were contemporaneous with unit 25 or whether they post-date the rhyolite volcanism of unit 26 is not known, as the eruptives have been completely eroded and no basic intrusives have been found cutting the rhyolites.

Chapter 4 PETROLOGY

a. Petrochemical setting

In the last decade a flood of new chemical analyses has revealed the main pattern of the chemical characteristics of volcanic rocks in Iceland. Jakobsson (1972) has outlined the distribution pattern of recent basaltic rocks in Iceland. He divides the postglacial volcanic zones into petrological regions within which the petrochemistry has remained fairly constant throughout postglacial and possibly late Pleistocene time. He recognized three main groups: tholeiites, transitional alkali basalts and alkali olivine basalts. The tholeiite regions, which are very much larger than the alkaline ones, are further divided into an olivine tholeiite region, which is a direct continuation of the Reykjanes ridge, and a tholeiite region (quartz normative tholeiites) in middle and northern Iceland. The intermediate and acid rocks show, as far as is known, the same regional pattern as the basalts.

Jakobsson draws attention to the connection between the petrochemistry and the depth to crustal layer 4 (uppermost mantle). The depth is less than 9km below the olivine tholeiite region in the Reykjanes volcanic zone, but increases from 9 to 14km below the alkali basaltic regions.

It is still premature to generalize much about the petrochemical distribution pattern of Tertiary and early Pleistocene rocks in Iceland, as there are yet large areas with many central volcanoes for which very few or no chemical analyses exist.

It is of interest, however, to look at the structural history of Iceland keeping in mind the possible correlation of the petrochemistry and the crustal structure. According to Sæmundsson's hypothesis (Sæmundsson 1973), until about 4 M.y. ago, there was only one main active volcanic zone which extended right across Iceland from the Reykjanes ridge in the southwest to the Kolbeinsey ridge in the north. Most of the Tertiary rocks in Iceland were formed in this zone; the rocks being progressively older away from this zone to the east and west, as a consequence of plate spreading. About 4 M.y. ago volcanism died out north of Langjökull in central Iceland, and a new volcanic zone was formed in eastern Iceland. This was preceded by a trough that reached right through the country. The eastern volcanic zone was thus superimposed on the Tertiary rocks of eastern Iceland. Transitional alkali and alkaline volcanics are erupted at the southern end of this volcanic zone, and coincide with and lie south of a transform fault system (Ward et al 1969) that extends from the Reykjanes volcanic zone to the eastern volcanic zone. The only other known alkaline region in Iceland is the Snæfellsnes peninsula, where Sigurðsson (1970a) has described three central volcanic series which are tholeiitic, transitional and alkaline, and date from late Tertiary early Quaternary and late Quaternary times respectively. The alkaline and transitional volcanism can be attributed to magma generation at a deeper level in an area having a relatively low thermal gradient away from the main active volcanic zone, and may be associated with transcurrent faulting through the Snæfellsnes peninsula (Sigurðsson 1970b).

Alkaline volcanism in Iceland thus appears to be associated with anomalous tectonic conditions. The Tertiary volcanics, if formed in a single volcanic zone extending from the Reykjanes ridge to the Kolbeinsey ridge would therefore be expected to be entirely tholeiitic in character.

Fairly extensive chemical analyses are now available from fourteen central volcanic regions in Iceland:

The Tertiary centres are: Þingmúli (Carmichael 1964) in eastern Iceland; the Hornafjörður region (Newman 1967), and the Vesturhorn intrusive centre (Roobol 1969) in southeast Iceland; Króksfjörður (Hald et al 1971), and Setberg Centre I (Sigurðsson 1970a) in western Iceland. All these centres are tholeiitic and chemically very similar to the Þingmúli series with the exception of the Vesturhorn intrusive centre, which is significantly more alkaline than Þingmúli (the acid rocks are slightly lower in total alkalis than Setberg Centre II). It is of considerable interest to note that the only one of these Tertiary centres showing slight transitional tendencies is the intrusive complex, which may have been formed outside the main volcanic zone at the time. According to Roobol (1969) the intrusions rose to a high level in the crust but did not produce a large central volcano. Acid magmas predate, accompanied and postdate the emplacement of the basic intrusions, and most of the rocks of intermediate composition in Vesturhorn are xenocryst-bearing hybrids. Petrographic evidence (some of which is supported by chemical analyses) from other Tertiary centres that have been investigated indicates that they are tholeiitic in character (e.g. Walker 1959, Walker 1963, Blake 1964, Beswick 1965, Gibson et al 1966, Annells 1968).

<u>The early Quaternary centres are:</u> Setberg Centre II (Sigurðsson 1970a) in western Iceland, which is transitional; Húsafell (Grönvold 1972), and the two central volcanoes in Esja, Kjalarnes and Stardalur, in southwest Iceland, all of which are tholeiitic and similar to Þingmúli. - A few analyses are also available from the Fitjaskógar (Sigurðsson 1970a) and Stóra-Laxá (Friðleifsson 1970) centres in southern Iceland (located on the eastern side of the Reykjanes volcanic zone) which appear to be tholeiitic, although the analysed basalts in both centres are unusually alkali rich for tholeiites.

<u>The late Quaternary and recent centres are:</u> The Snæfellsnes alkaline province (Sigurðsson 1970a), which cannot be referred to as a central volcano; Kerlingafjöll (Grönvold 1972) in central Iceland which is tholeiitic; Hekla (Thorarinsson 1967, Sigvaldason 1973) and Torfajökull* (Grönvold 1972) in southern Iceland, which are tholeiitic and transitional respectively; Námafjall* (Grönvold 1972) in northern Iceland, which is tholeiitic, but markedly lower in total alkalis than Þingmúli.

At the present state of knowledge the volcanic rocks in Iceland seem to have been of a fairly uniform tholeiitic character during the Tertiary. The marked diversity in chemical characteristics found in Quaternary and recent rocks may be tectonically controlled.

* The basaltic end of these series has not been extensively sampled so comparisons have to be made mainly from the intermediate and acid rocks.

The petrochemical setting of Esja is simple. Recent studies of late Tertiary and early Quaternary volcanic rocks erupted within and spreading westwards from the Reykjanes volcanic zone in southwest Iceland indicate undiversified tholeiitic characteristics.

b. Rock nomenclature of the tholeiitic series

In the following sections dealing with the petrochemistry of Esja, it will be convenient to refer to the rocks using clearly-defined petrological names. This section therefore comes first, although the rock terminology adopted partly makes use of chemical criteria, and, to this extent, anticipates data to be presented in detail subsequently.

A multilingual rock nomenclature was in circulation in the literature on Icelandic geology until the early 1960s. It was the second generation of British geologists working in Iceland that established a classification scheme for the tholeiitic series on which now there is a broad consensus amongst students of Icelandic geology. Walker (1959) divided the basalt lavas of Reyðarfjörður into olivine basalts, porphyritic basalts and tholeiites; he referred to the intermediate lavas as andesites, and the acid lavas as rhyolites, but added in a footnote that many of the latter are, in fact, dacitic. This classification is based on field characteristics, and works very well. The term olivine basalt is, however, unsatisfactory, because of the possibility of confusion with the same term used as shorthand for alkali olivine basalts.

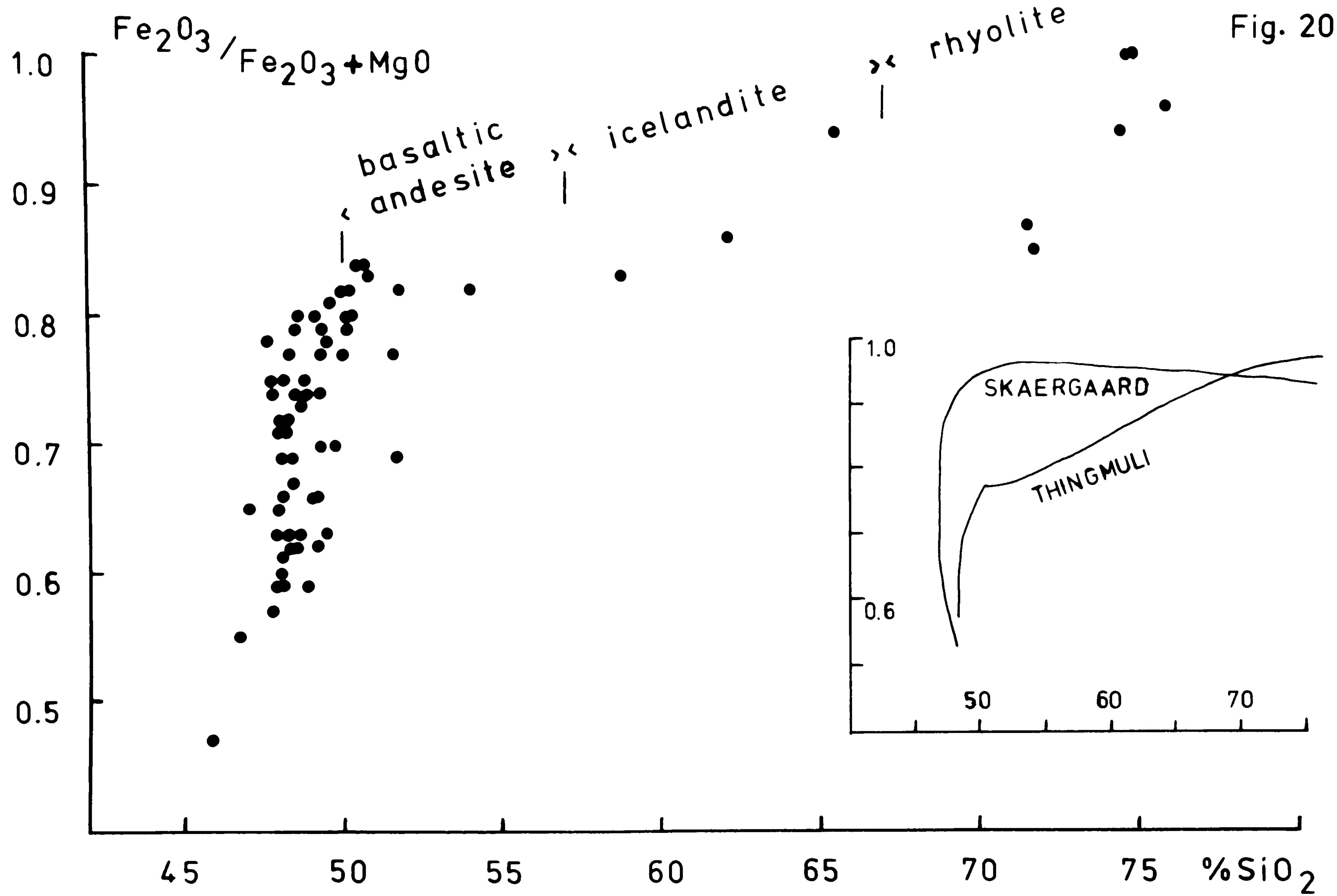
Carmichael (1964) improved the scheme by referring to the olivine-bearing basalts as olivine tholeiites, adding the term basaltic andesite, and suggesting that, in order to avoid confusion of the very fine-grained andesites in Þingmúli with the typical porphyritic orogenic andesites, the Icelandic andesites could be called icelandites. Grönvold (1972) modified this term to icelandesites. There is a complete gradation in the field character (Walker 1959), and in the mineralogy and chemistry (Carmichael 1964) from olivine tholeiites, through (olivine-free) tholeiites, basaltic andesites, and andesites (icelandites), to rhyolites.

Both Sigurðsson (1970a) and Grönvold (1972) adopted Carmichael's classification of the intermediate and acid rocks, except that the former added the term rhyodacite (66-70% SiO_2) and the latter the term dacite (63-69% SiO_2) to the scheme.

Carmichael's (1964) division between olivine tholeiites and tholeiites was based on whether the rock had more than accessory amounts of olivine or not. Sigurðsson (1970a) used the CIPW-normative classification to divide the basalts into olivine tholeiites and quartz tholeiites, calling those basalts tholeiites which have Di, Hy and Pl as the principal normative components and have very little normative Qz or Ol. He pointed out, however, the shortcomings of a normative classification when dealing with altered (oxidized) rocks. The FeO/Fe_2O_3 ratio must be adjusted in altered rocks in order to obtain a meaningful normative composition. Normative classification is therefore not suitable in regions having a high thermal gradient, and consequent secondary alteration.

The classification of Carmichael is adopted in the present study unchanged except that the boundary between tholeiites and basaltic andesites is drawn at a lower silica level (50% SiO_2 as opposed to 52.7% in Þingmúli). It is proposed that this boundary should be drawn at the point where there is a marked change in slope on the $FeO+Fe_2O_3/FeO+Fe_2O_3+MgO$ versus SiO_2 diagram (Fig.20). Where this method leads to ambiguity (e.g. a rock plots away from the trend of Fig.20, or has suffered secondary silica enrichment) such rocks are called tholeiites or basaltic andesites according to whether MgO is greater or less than 4% respectively.

Fig. 20


Fe2O3/Fe2O3+MgO
1.0
0.9
0.8
0.7
0.6
0.5
basaltic andesite
icelandite
rhyolite
45
50
55
60
65
70
75
%SiO2
SKAERGAARD
THINGMULI
1.0
0.6
50
60
70

c. Chemical analyses

Major and minor element analyses were carried out by XRF methods (analytical methods are described in Appendix 2) on samples from most of the geological units in the Esja stratigraphic column. Selection of samples for analysis was limited by the prevalence in the Esja rocks of hydrothermal alteration effects. Larger numbers of the generally fresher intrusives and fine-grained tholeiite lavas than of coarser-grained and more altered olivine tholeiite lavas were selected for analysis. However, since those olivine tholeiite units, which would unfortunately not provide meaningful analyses, were probably erupted from shield volcanoes outside the areas of central volcanism, this limitation should not seriously affect the study of the petrochemical development of the centres. Samples representing the hyaloclastite units were taken from plugs and large lithic masses within the hyaloclastites. An attempt was made to distinguish chemically between the Kjalarnes and Stardalur centres, but they were found to be very similar. The chemical analyses are listed in stratigraphical-chronological order in Table 2. Sample localities and short petrographic descriptions are given in Appendix 1.

Nearly all the rock samples collected in Esja contain <u>some</u> phenocrysts; the abundance ranges from one phenocryst in a thin section to thoroughly porphyritic rocks. The majority of the rocks, however, contain only accessory amounts of phenocrysts. The most abundant phenocryst is plagioclase, which is found in all the rock types, and it is sometimes accompanied by one or more of olivine, pyroxene (augite) and ore. When considering chemical variations within the rock suite it is important to keep in mind that one, two or possibly three phenocryst phases may be present in some of the rocks. Crystal accumulation and liquid fractionation appears to be a feasible mechanism to explain the chemical variation within at least the basaltic rocks in Esja.

Variation diagrams of the major and minor elements versus silica are shown in Figs. 21, 22 and 23. The trends in the Esja rocks correspond very closely to those of the Þingmúli series (Carmichael, 1964). Analyses from Þingmúli, the best documented tholeiitic series in Iceland, are plotted on the total alkalis versus silica diagram (Fig.21) for comparison.

The most outstanding feature shown by the major element versus silica diagrams is the steady decrease in Al_2O_3, MgO and CaO and the corresponding increase in Fe_2O_3 (total iron) and TiO_2 for nearly constant but slightly rising silica values until, at about 50% SiO_2, the trends change direction more or less abruptly and strong silica enrichment is thereafter accompanied by gentle decreases in MgO, CaO, Fe_2O_3 (total iron) and TiO_2, the Al_2O_3 content remaining fairly constant (Figs. 22 and 23).

Al_2O_3 contents can be related to petrographic features of the basalts, in particular their feldspar mineralogy. The highest alumina values are found in the olivine tholeiites, which (with their more basic plagioclase) are always rich in alumina, and in the plagioclase-phyric tholeiites. Nearly all the olivine-free tholeiites with Al_2O_3 higher than 14% are plagioclase-phyric, and most of those with Al_2O_3 higher than 15% can be referred to as porphyritic (phenocrysts more than 10% modally). The only apparent exceptions to this are the tholeiite lavas in unit 22 (chem.an.60) and especially those in unit 24 (chem.an.68, 70 and 71) which are exceptionally alumina rich. The trend (Fig.22) is fairly smooth from the basaltic andesites to the rhyolites, and the icelandites do not show the enrichment in alumina characterising the Þingmúli series (Carmichael 1964).

The apparent plagioclase control on the chemical trend of the basalts can be further elucidated by variation diagrams showing alumina versus magnesia (representing plagioclase versus pyroxene and olivine), and alumina versus lime (representing plagioclase versus plagioclase and pyroxene).

In the Al_2O_3/MgO diagram (Fig.24) there is a broad, but linear, trend from the basaltic andesites to the olivine tholeiites. "Above" this trend are mainly plagioclase porphyritic tholeiitesand some MgO poor olivine tholeiites; "below" the trend are mainly pyroxene-rich dolerite intrusions. In the Al_2O_3/CaO diagram (Fig.24) there is less scatter. There is a linear trend from the basaltic andesites to the plagioclase porphyritic basalts. "Below" this trend are the pyroxene-rich dolerite intrusives; "above" the trend are olivine tholeiites.

It is also possible to relate, in a general way, MgO and CaO contents to modal features, but this is more difficult since these oxides occur in, respectively, two and three major mineral phases, as opposed to Al_2O_3, which is largely confined to the plagioclase only. In the CaO/MgO variation diagram (Fig.25) the linear trend is fairly narrow in the basaltic andesites and the aphyric tholeiites, but in the more basic rocks the olivine-phyric rocks and the pyroxene-rich dolerites fall "below" the trend, and some of the plagioclase-phyric rocks are "above" it.

Tha sharpest "breaks" in the chemical trends are seen in the Fe_2O_3 (total iron)/SiO_2 and the TiO_2/SiO_2 diagrams (Fig.23). These two diagrams have a remarkably similar appearance, with a broad scatter but an overall increase in both metallic oxides from the olivine tholeiites to the basaltic andesites, an abrupt decrease in each at about 51% SiO_2, and a gentle decrease in each, commencing at about 52% SiO_2, through to the rhyolites. The samples having between 15.5-17.0% Fe_2O_3 at silica level 47.0-48.6% are from units associated with the Kjalarnes centre. The very high iron and titanium content of these rocks is reflected in their high density, which is believed to contribute to the gravity anomaly in the Kjalarnes centre (page 161). Samples with less than 13.5% Fe_2O_3 at SiO_2 level of 48.5-52.0% are mostly dolerite intrusives.

The iron and titanium enrichment is represented modally by high abundances of opaque oxide minerals, as noted by Carmichael (1964). The basaltic andesites and the tholeiites contain the iron ore in evenly distributed, equant, euhedral grains of magnetite (magnetite-ulvöspinel solid solution) which may be larger than the groundmass pyroxene. In the olivine tholeiites, on the other hand, the magnetite is much less abundant and occurs mainly as irregular grains in interstices between the pyroxene and plagioclase in the groundmass. Independent acicular ilmenite is present in the olivine tholeiites, the tholeiites, and in the basaltic andesites. Carmichael drew attention to the size distribution of magnetite in the basaltic andesites of Þingmúli; the large fraction (microphenocrysts) being larger than the groundmass minerals, whereas the smaller interstitial size fraction is smaller than the other groundmass minerals. In the icelandites and rhyolites the large majority of magnetite occurs in the groundmass as tiny grains. These features apply also to the Esja rocks.

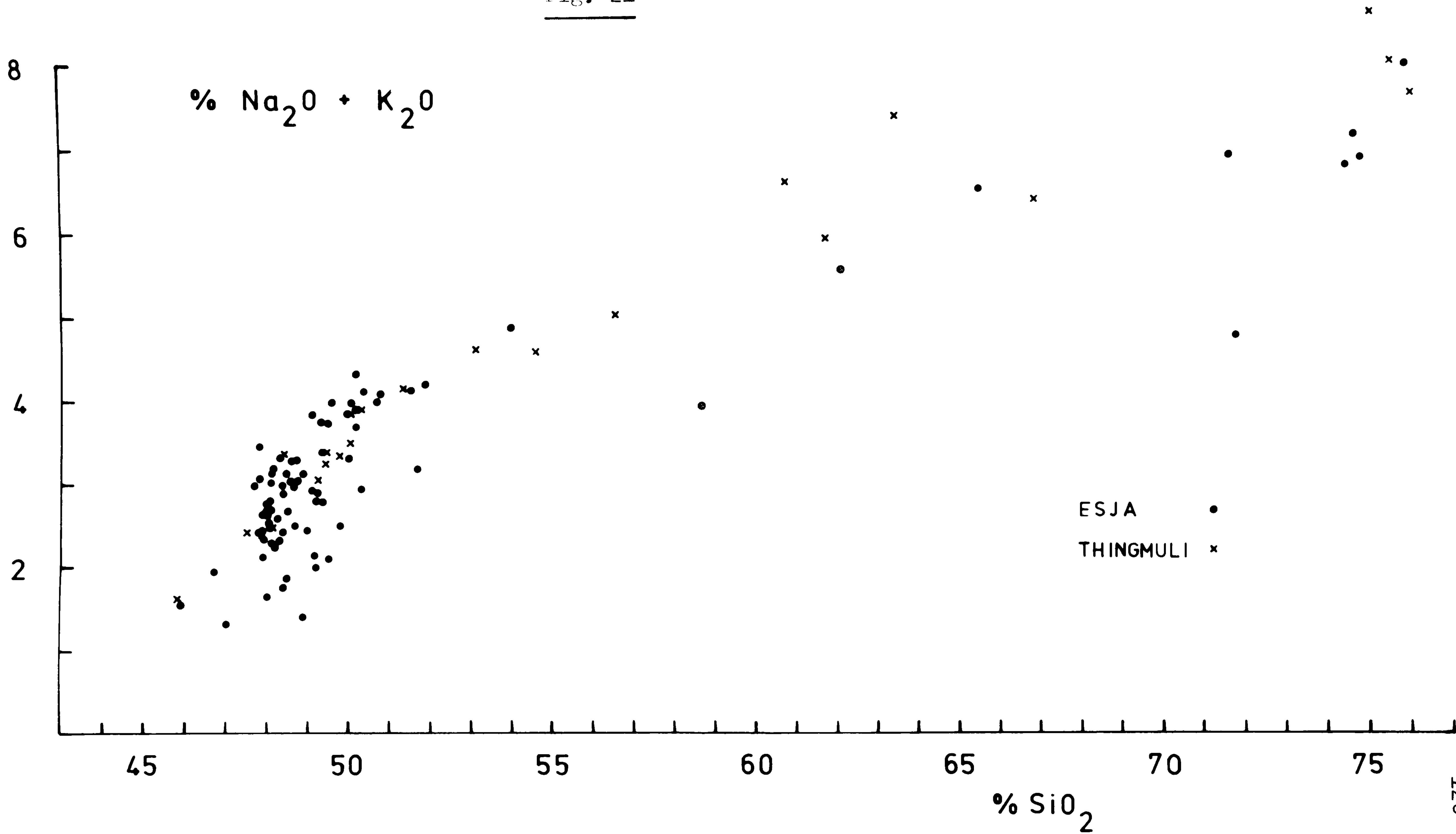
MnO increases with increasing silica until about 50% SiO_2, and thereafter decreases gradually to the rhyolites (Fig.23). The MnO pattern corresponds closely to those showed by Fe_2O_3 (total iron) and TiO_2; the points falling above the main trend on the MnO/SiO_2 variation diagram for silica values less than 48.6% represent the iron- and titanium-rich rocks of the Kjalarnes centre.

P_2O_5 increases from the olivine tholeiites through the tholeiites and basaltic andesites until a silica value of about 54% is reached; sharp linear decrease proceeds to about 65% SiO_2; thereafter the decrease is less marked through the rhyolites.

P_2O_5 is the only element showing, on a silica variation diagram, a marked change in slope occurring elswhere than at about 50% SiO_2. The P_2O_5 trend probably reflects fractionation of apatite, although this phase, if present, must be in the groundmass and has not been detected in thin-section.

There is close correspondence in the patterns of Na_2O and K_2O, and the sums of these two oxides are plotted against silica in Fig. 21. There is a slight change in slope at about 50% SiO_2 when Na_2O is plotted against silica, but this change in slope is less marked in the K_2O/SiO_2 plot. The rhyolites show enrichment in K_2O relative to Na_2O, which is probably due to secondary alteration of the rhyolites. The two points falling below the main trend on the alkali silica diagram are from the same composite body (chem.an.61 and 63) and are particularly low in Na_2O. A less altered rhyolite sample (chem.an.62) from this body conforms to the main pattern.

Fig. 21



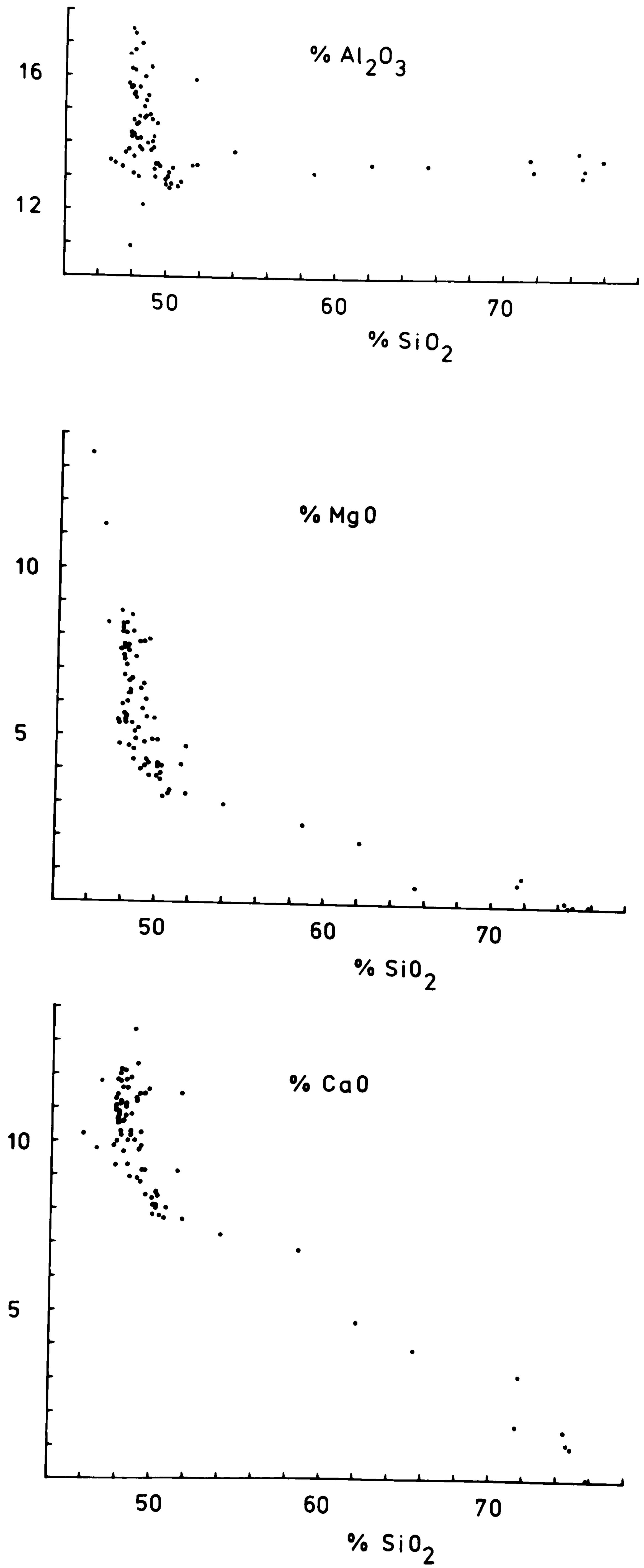
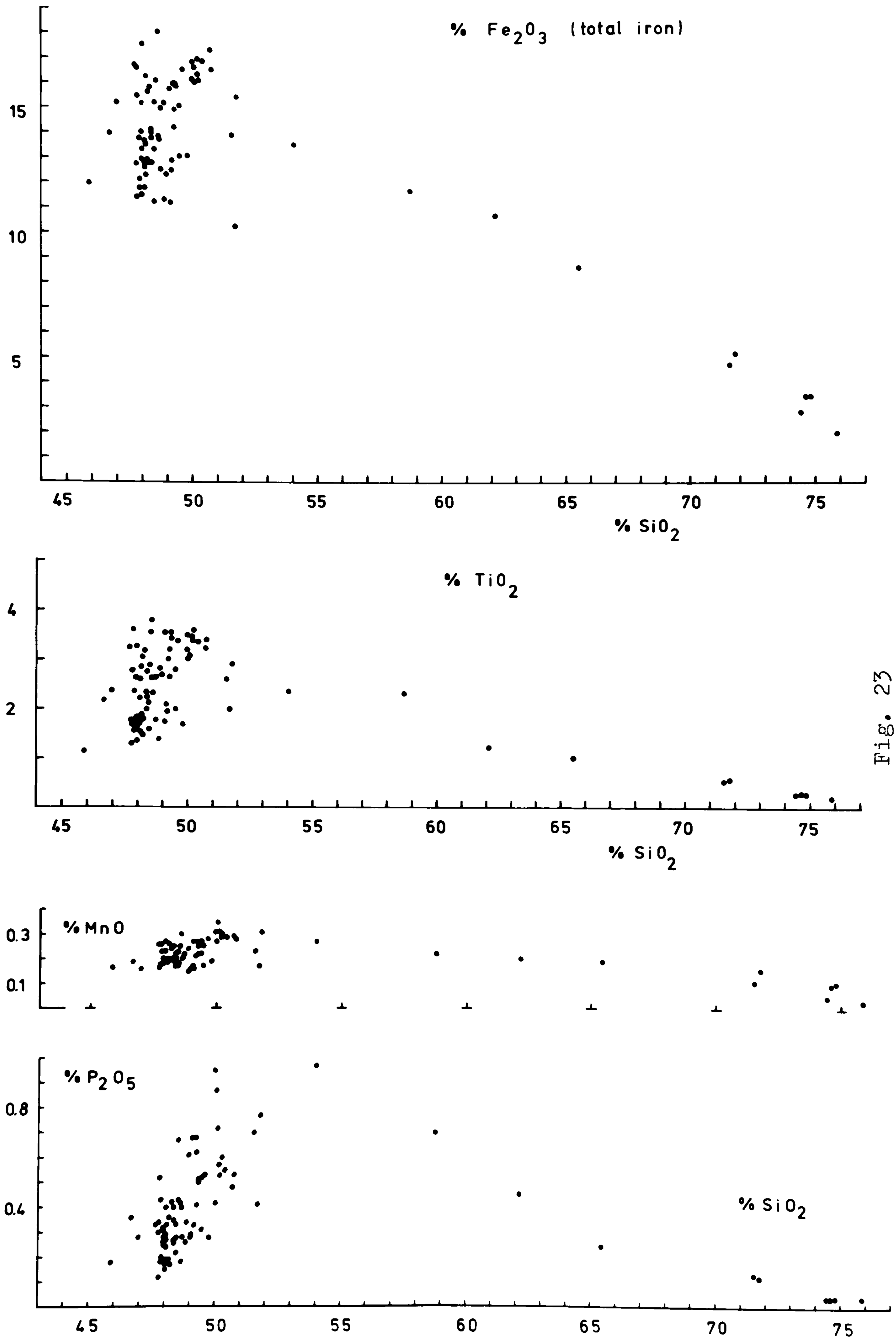

% Al_2O_3
16
12
50
60
70
% SiO_2
% MgO
10
5
50
60
70
% SiO_2
% CaO
10
5
50
60
70
% SiO_2

Fig. 22

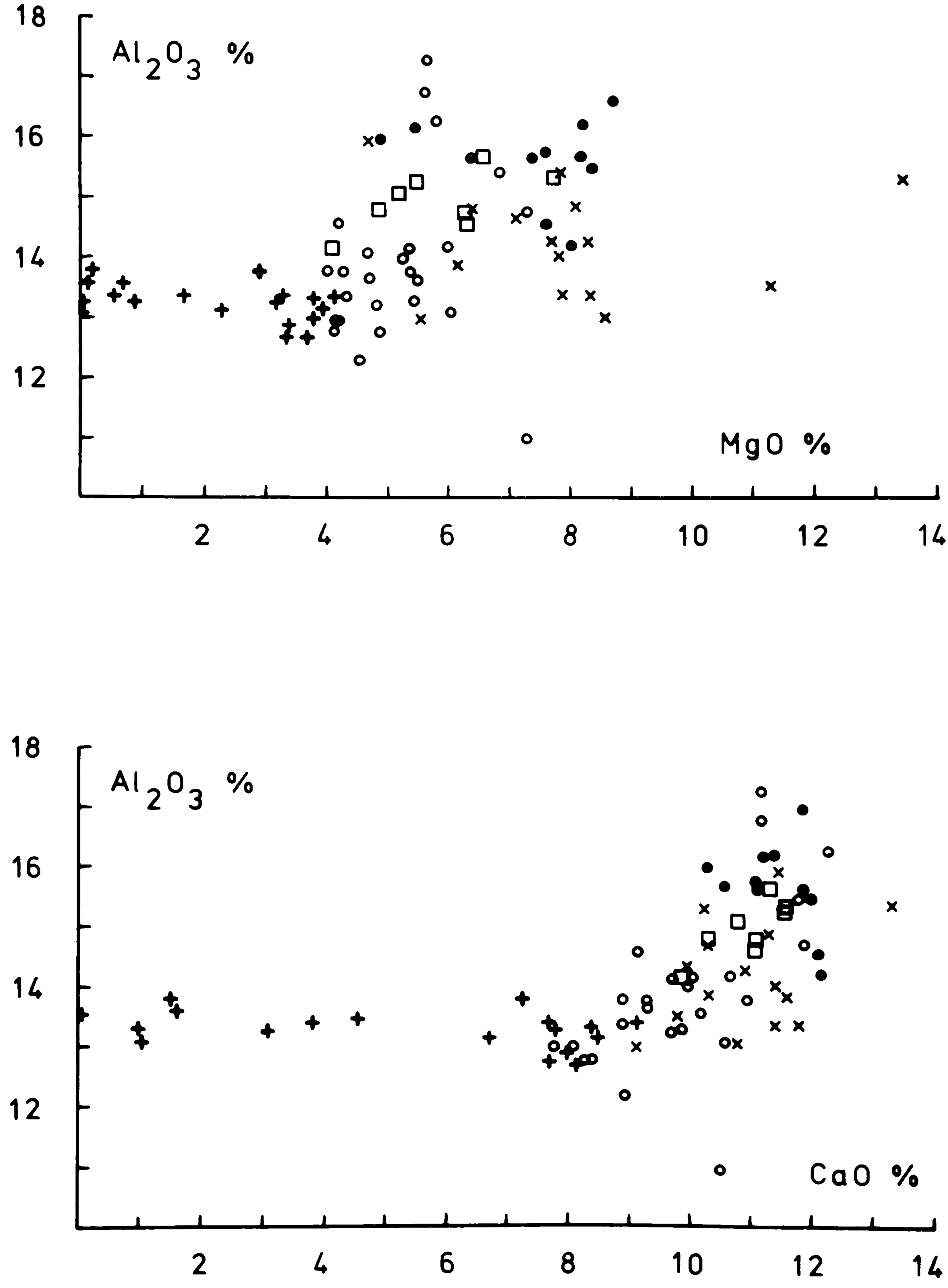


% Fe_2O_3 (total iron)
15
10
5
45
50
55
60
65
70
75
% SiO_2
% TiO_2
4
2
45
50
55
60
65
70
75
% SiO_2
% MnO
0.3
0.1
% P_2O_5
0.8
0.4
% SiO_2
45
50
55
60
65
70
75

Fig. 23

Fig. 24

• ol. tholeiite
∘ tholeiite
□ plagioclase phyric (>10%) tholeiite
+ bas.andesite, icelandite, rhyolite
× dolerite intrusive



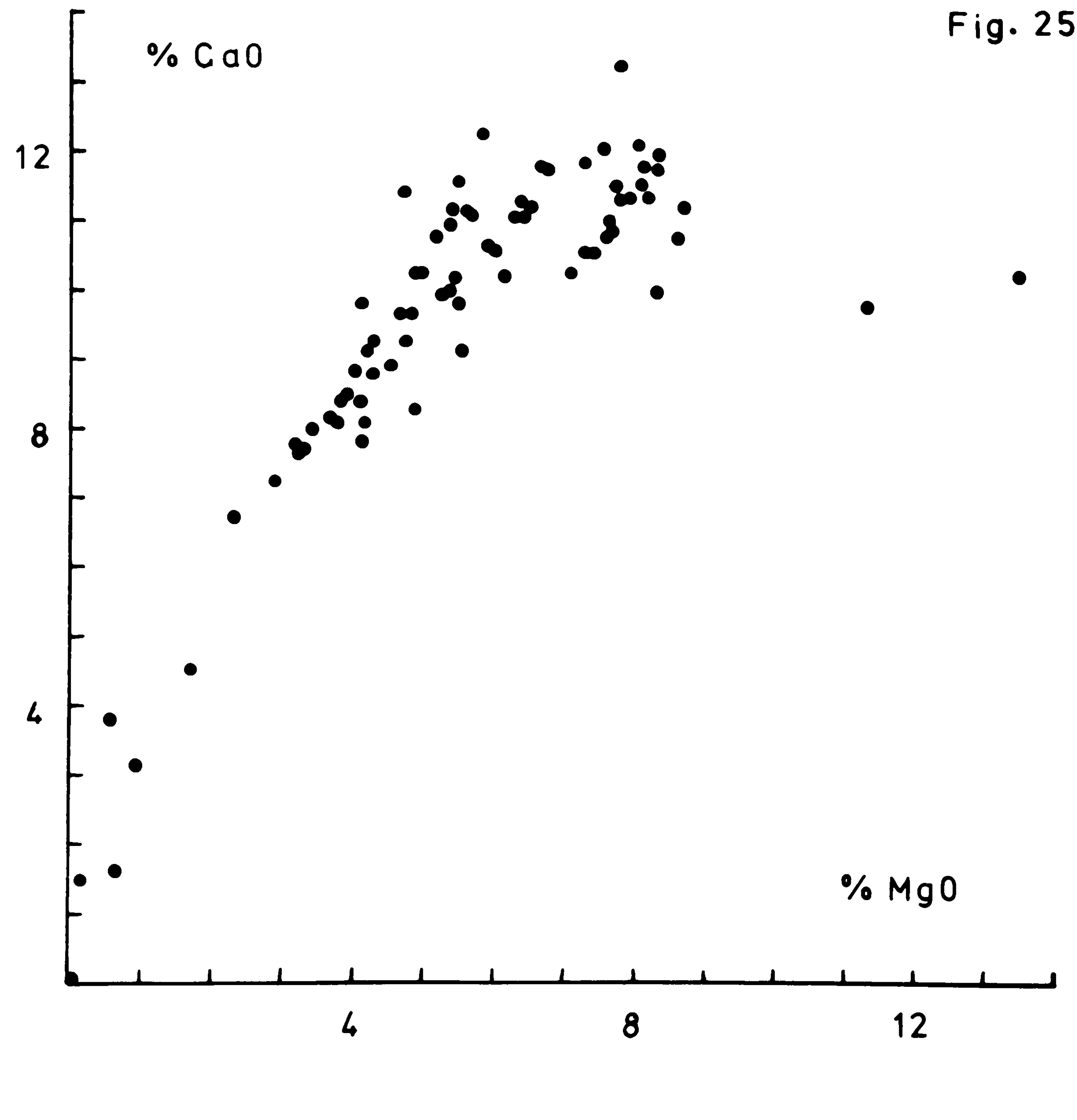


Fig. 25

Table 2 Chemical analyses

Sample localities are listed in Appendix I and analytical methods are described in Appendix II.
Total iron is listed as Fe_2O_3.

Sample	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	FeO	Density g/cm^3
1	49.26	13.25	16.03	4.84	9.72	2.61	0.31	3.03	0.27	0.68	9.73	3.08
2	49.09	13.82	15.84	4.04	8.86	3.28	0.57	3.55	0.27	0.68	12.79	3.13
3	48.56	13.83	16.13	4.32	9.33	2.92	0.41	3.57	0.25	0.67	13.04	3.20
4	47.83	13.69	16.66	4.75	9.30	2.89	0.48	3.62	0.26	0.52	12.92	3.15
5	49.28	13.89	14.32	6.14	10.27	2.49	0.33	2.65	0.22	0.41	9.71	3.03
6	48.38	13.04	14.13	8.58	10.79	1.70	0.07	2.76	0.17	0.40	7.55	2.97
7	48.05	15.47	13.04	6.84	11.81	2.36	0.18	1.83	0.20	0.24	8.10	3.15
8	48.20	15.36	12.99	7.74	11.57	2.14	0.12	1.50	0.19	0.19	8.83	3.09
9	47.71	13.32	16.77	5.46	9.87	2.65	0.37	3.26	0.26	0.33	11.80	3.11
10	49.35	13.43	16.01	4.33	8.79	3.15	0.62	3.55	0.27	0.50	11.82	3.13
11	51.72	15.95	10.37	4.73	11.45	2.67	0.53	2.00	0.17	0.41	6.81	2.97
12	62.15	13.44	10.71	1.73	4.53	3.98	1.61	1.20	0.20	0.45	5.47	2.73
13	48.24	13.12	15.73	6.02	10.61	2.46	0.14	3.08	0.24	0.36	9.52	3.09
14	48.48	14.17	15.31	5.37	10.05	2.87	0.27	2.90	0.23	0.33	11.68	3.13
15	49.30	14.20	15.02	4.11	9.86	2.92	0.52	3.20	0.25	0.62	11.54	3.02
16	48.40	14.61	13.93	6.31	11.12	2.71	0.20	2.25	0.22	0.27	10.46	3.15

Sample	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	FeO	Density g/cm^3
17	47.82	13.83	15.56	5.44	10.96	2.82	0.26	2.78	0.23	0.30	11.37	3.12
18	48.29	14.14	15.93	4.68	9.74	2.88	0.46	3.22	0.25	0.42	12.56	3.15
19	48.89	14.04	15.25	5.26	9.99	2.79	0.35	2.83	0.24	0.34	11.31	3.09
20	48.14	13.64	16.36	5.48	10.21	2.53	0.18	2.87	0.26	0.33	11.24	3.05
21	49.80	15.27	13.14	5.52	11.59	2.41	0.11	1.69	0.19	0.28	8.47	3.09
22	47.99	15.69	13.44	7.39	10.55	2.57	0.20	1.65	0.20	0.32	10.06	3.13
23	47.81	15.84	12.87	7.64	11.06	2.22	0.21	1.82	0.20	0.34	9.08	3.12
24	47.97	14.21	15.26	5.96	10.68	2.62	0.15	2.64	0.23	0.28	10.40	3.12
25	48.38	14.80	14.16	6.31	11.10	2.31	0.12	2.35	0.20	0.26	8.88	3.04
26	48.66	15.10	13.93	5.21	10.80	2.74	0.25	2.66	0.21	0.42	8.84	3.08
27	48.86	15.43	11.38	7.81	13.34	1.25	0.15	1.38	0.15	0.26	6.32	2.96
28	49.01	14.86	12.47	6.44	11.30	2.10	0.35	2.70	0.16	0.61	8.05	3.01
29	48.10	14.73	13.75	7.13	10.30	2.45	0.34	2.62	0.19	0.40	10.97	3.15
30	47.93	14.26	13.84	8.29	10.03	2.41	0.24	2.38	0.20	0.43	11.45	3.17
31	46.69	13.48	14.07	11.31	9.79	1.71	0.24	2.18	0.19	0.36	11.15	3.19
32	48.01	10.96	17.59	7.32	10.62	1.23	0.43	3.27	0.27	0.31	12.13	3.07
33	48.48	16.99	11.34	6.69	11.85	2.45	0.24	1.58	0.17	0.22	9.82	3.10
34	47.93	15.66	12.20	8.16	11.86	1.94	0.20	1.70	0.18	0.18	9.57	3.08
35	47.88	16.25	11.90	8.21	11.43	2.19	0.18	1.59	0.18	0.20	9.00	3.09
36	48.37	15.71	12.87	6.38	11.11	2.80	0.21	2.00	0.19	0.35	11.70	3.04
37	48.12	16.20	13.27	5.45	11.21	2.82	0.21	2.23	0.19	0.29	11.06	2.91

Sample	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	FeO	Density g/cm^3
38	48.70	16.00	13.82	4.92	10.31	3.04	0.27	2.34	0.20	0.40	11.94	2.99
39	50.01	12.80	16.23	4.90	8.33	2.64	0.67	3.21	0.27	0.95	9.90	3.05
40	50.28	12.82	16.19	4.11	8.40	3.19	0.49	3.63	0.29	0.60	12.33	2.96
41	50.09	13.00	16.13	4.19	8.12	3.24	1.11	3.09	0.31	0.72	10.87	3.03
42	50.06	13.00	16.72	4.16	7.84	3.06	0.93	3.01	0.35	0.87	9.54	2.92
43	65.48	13.41	8.69	0.58	3.80	5.17	1.40	1.03	0.19	0.24	6.37	2.69
44	50.73	12.75	17.41	3.36	7.74	3.25	0.76	3.23	0.29	0.48	14.04	3.04
45	50.21	12.71	17.03	3.69	8.16	3.13	0.77	3.47	0.30	0.53	11.38	3.05
46	49.64	13.35	16.61	3.81	8.42	3.36	0.64	3.38	0.28	0.53	13.13	3.08
47	50.03	12.98	16.92	3.83	8.14	3.14	0.72	3.51	0.31	0.42	14.11	3.08
48	50.42	13.29	16.95	3.20	7.83	3.42	0.70	3.36	0.28	0.55	13.70	3.14
49	50.77	12.88	16.60	3.42	8.03	3.42	0.67	3.40	0.28	0.53	13.14	3.12
50	49.16	15.73	12.59	6.55	11.27	2.09	0.06	2.10	0.16	0.28	7.33	3.00
51	47.00	13.40	15.26	8.36	11.82	1.33	0.00	2.38	0.16	0.28	6.21	3.00
52	47.99	14.33	14.09	7.68	10.91	2.12	0.59	1.86	0.18	0.25	7.36	2.98
53	49.22	14.05	12.98	7.83	11.41	1.86	0.15	1.96	0.21	0.33	6.37	3.09
54	48.49	13.86	13.42	8.12	11.62	1.71	0.16	2.14	0.18	0.28	6.24	2.99
55	49.54	13.43	13.14	7.89	11.41	1.91	0.20	2.00	0.17	0.31	6.93	3.05
56	48.72	14.76	12.63	7.34	11.89	2.42	0.08	1.79	0.20	0.18	10.04	3.15
57	49.35	13.03	15.96	5.55	9.16	2.34	0.44	3.44	0.22	0.51	10.23	3.02
58	45.92	15.31	12.01	13.45	10.25	1.38	0.17	1.15	0.17	0.18	8.62	3.09

Sample	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	FeO	Density g/cm^3
59	48.60	12.18	18.12	4.56	8.95	2.60	0.46	3.80	0.30	0.43	12.95	3.04
60	49.53	14.63	15.15	4.21	9.14	3.17	0.60	2.81	0.25	0.52	13.24	3.02
61	71.78	13.25	5.23	0.94	3.13	2.59	2.22	0.58	0.16	0.12	1.56	2.41
62	71.56	13.58	4.74	0.70	1.63	4.28	2.71	0.55	0.11	0.13	1.34	2.45
63	58.78	13.16	11.82	2.34	6.74	2.85	1.08	2.29	0.22	0.70	6.70	2.73
64	54.04	13.81	13.57	2.93	7.26	3.79	0.98	2.38	0.27	0.97	10.66	3.00
65	50.22	13.15	16.21	3.93	8.52	3.00	0.70	3.40	0.29	0.57	12.22	2.97
66	51.55	13.38	14.03	4.17	9.14	3.45	0.69	2.66	0.23	0.70	11.82	3.12
67	51.78	13.41	15.52	3.33	7.73	3.33	0.89	2.93	0.31	0.77	11.54	2.91
68	49.11	16.31	11.27	5.83	12.33	2.77	0.17	1.75	0.17	0.29	8.24	3.04
69	48.75	14.77	15.07	4.94	10.27	2.74	0.31	2.64	0.22	0.28	11.56	3.08
70	48.13	16.81	12.72	5.59	11.21	2.95	0.19	1.92	0.20	0.27	10.51	3.08
71	48.14	17.28	12.42	5.66	11.17	3.03	0.18	1.73	0.19	0.19	9.40	3.07
72	48.07	14.25	12.86	8.05	12.17	2.40	0.11	1.73	0.20	0.15	8.85	3.13
73	48.27	14.61	12.94	7.55	12.11	2.25	0.09	1.80	0.20	0.17	8.83	3.14
74	48.09	15.46	11.89	8.35	12.03	2.14	0.16	1.54	0.18	0.17	8.30	3.10
75	75.91	13.63	2.01	0.08	0.00	4.05	4.03	0.23	0.03	0.04	0.72	2.46
76	74.45	13.79	2.87	0.20	1.48	3.92	2.93	0.27	0.05	0.04	0.16	2.44
77	74.78	13.34	3.49	0.00	1.02	3.79	3.14	0.28	0.11	0.04	1.68	2.58
78	74.64	13.08	3.56	0.00	1.07	4.29	2.94	0.29	0.10	0.04	2.34	2.60
79	47.77	16.66	11.52	8.70	11.28	2.37	0.08	1.33	0.17	0.12	8.95	2.73
80	48.00	17.45	11.60	7.64	10.84	2.48	0.19	1.35	0.18	0.26	9.27	2.98

d. Petrochemical development of Esja

The chemically analysed samples are arranged in stratigraphical-chronological order in Table 2. The relative ages of the lavas and hyaloclastites are certain, but the placing of the intrusives into the table is, of course, more tentative. As a rule an intrusive is placed as low in the chronological scale as its field relations, such as the age of the rocks it intrudes, the magnetic polarity direction, and the strike (in case of dykes and sheets), allow.

As more than one sample has been analysed from some of the geological units, and no analyses are available from several of the units, the petrochemical pattern demonstrated by the analyses is incomplete, and certainly biased towards the finer-grained and hence fresher rocks. However, a clear pattern emerges for the petrochemical development of Esja (Fig.26). If a representative sample of every geological unit had been analysed, the olivine tholeiites would have been more prominent in the pattern.

During the time before the intrusion of the Kjalarnes dolerite sheets, tholeiite volcanism was twice interrupted by olivine tholeiite volcanism (units 2 and 6). Despite the thickness of the olivine tholeiite lava piles, each of these units may represent a fairly short period of time, as they were probably erupted from shield volcanoes, which characteristically eject large volumes of lava in single eruptions; e.g. Skjaldbreiður $17km^3$, Kjalhraun $4km^3$,

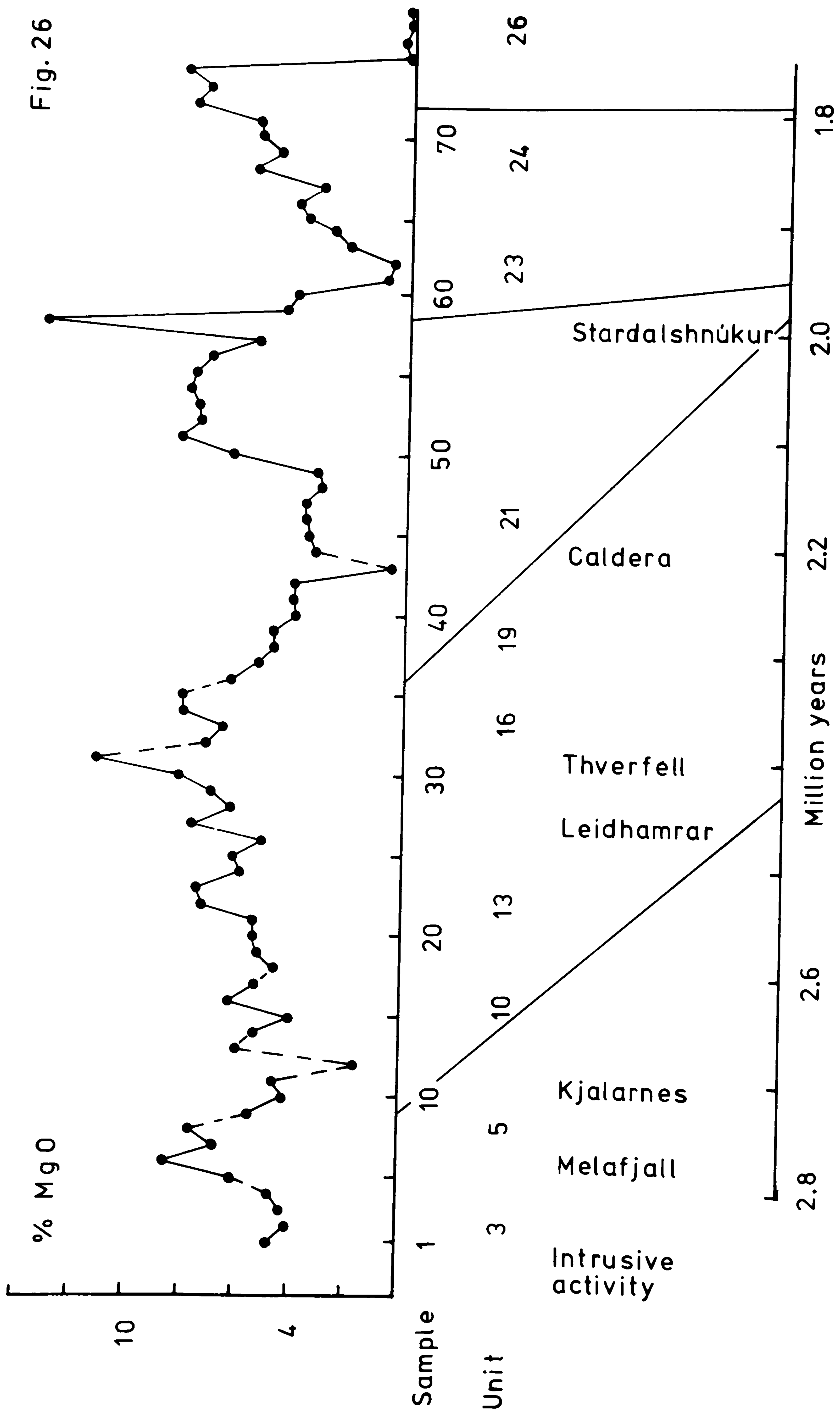

% MgO
10
4
Sample
Unit
1
10
20
30
40
50
60
70
3
5
10
13
16
19
21
23
24
26
Intrusive activity
Melafjall
Kjalarnes
Leidhamrar
Thverfell
Caldera
Stardalshnúkur
Million years
2.8
2.6
2.2
2.0
1.8

Fig. 26

Leggjabrjótur 4km^3, Baldheiði 2km^3, Selvogsheiði 1km^3 (Kjartansson 1967). The tholeiites (unit 3 and above), which are thought to be products of the Kjalarnes central volcano, are characterized by very high iron and titanium values. The bulk of the dykes, sheets and sills are of very fine-grained tholeiite and are chemically very similar to the lavas. A small proportion of the tholeiite lavas are plagioclase porphyritic. Low iron and titanium values suggest that they may be unrelated to the Kjalarnes centre.

After the intrusion of the Kjalarnes dolerite sheets the trend of dykes changed to N 40°E (from N 25°E). One of the dykes with this trend is of icelandite. When it was emplaced is not known. Its polarity direction is reverse, but a possible eruptive counterpart has not been found. It is considered likely that its emplacement, if not its generation, may be related to the tectonic upheaval associated with the Kjalarnes intrusions. Surprisingly, considering the length of the life span of the Kjalarnes central volcano, this is the only rock of intermediate composition that has been found in western Esja. However, one has to consider the fact that the rocks exposed at present all lie outside the main centre of activity and that only a 60° arc of a hypothetical peripheral rim of the centre is exposed. Substantial amounts of intermediate and acid magma may have been erupted within the central region or parasitically on other sides of the Kjalarnes central volcano.

After the Kjalarnes intrusives had been emplaced and before the intrusion of the Þverfell and Lauganípa dolerite sheets, all the volcanics erupted in western Esja (so far as is known) were of tholeiite. Most of these tholeiites contain some phenocrysts, which must indicate a delay in the passage of the magma from its source to the surface. However, despite the phenocrysts, all the analysed rocks, except the porphyritic lavas in Flekkudalur (chem.an. 21, 22 and 23), are characterised by high iron an titania contents, and are considered a part of the Kjalarnes central activity. The high productivity of the volcano at this stage can be judged from the thickness of the units shown in Table 1, page 11 . Most of the volcanics were probably erupted from a dyke swarm which cut across the central volcano.

Coincidental with the halt in tholeiite volcanism were the intrusions of the olivine tholeiite dolerite sheets in Þverfell and Lauganípa. Gravitational crystal settling of olivine has probably occurred in these sheets (samples 29 and 30, taken from the top and the lowest exposed part of a sheet east of Kvensöðlar).

Possibly during, and certainly after the emplacement of these sheets, the volcanics were mainly of olivine tholeiite composition, until the Stardalur central volcano became active. The bulk of these olivine tholeiites are probably not related to the central volcanoes, and some of the lava sequences are in the form of flow units characteristic of shield volcanoes, whereas the hyaloclastite ridges suggest fissure eruptions. There may have been periods of fissure and shield volcano eruptions similar to postglacial volcanism in the Reykjanes peninsula (Jónsson 1967).

The tholeiites of unit 17 may be related to the Stardalur central volcano, but the oldest analysed samples from that centre are from unit 19. As mentioned on page 54, the Esja region was divided by a hyaloclastite ridge at this stage. The three analysed olivine tholeiite samples (36,37,38) from the area to the northwest of the ridge may all be products of the same eruptive episode of a single shield volcano; if so, they show a systematic chemical variation with the early products of the eruption (chem.an.36) being more basic than the late products (chem.an.38). To the southeast of the ridge, in the vicinity of Stardalur, tholeiite lavas were erupted. The analysed samples of these lavas are, in fact, on the borderline of tholeiites and basaltic andesites in the adopted classification. Whether the eruption of the icelandite lava (chem.an.43), which marks the top of this lava sequence and which is found only to the east of the caldera fault, was connected with the caldera collapse, is not known. Following a brief spell of tholeiite volcanism, a more voluminous episode of basaltic andesite volcanism filled, and may have been confined to, the caldera.

The bulk of the dolerite sheets and sills were probably intruded after the caldera had been filled with basaltic andesite; the virtually olivine-free but pyroxene-rich dolerites indicate a marked change in the chemistry of the products of the Stardalur centre in a short interval of time after the caldera collapse. There is a considerable chemical variation in the intra-caldera intrusives, the plagioclase-phyric dyke in Múli (chem.an.50), and the Gráhnúkur sill (chem.an.57) contrasting with each other and with the more typical dolerite intrusions. The last major intrusive to be emplaced within the caldera was the olivine-rich Stardalshnúkur dolerite. This intrusion, and also some of the other normally magnetized dolerites, may have formed later than the Olduvai event, e.g. during the Gilsá event. The dolerite cone sheets, however, do not appear to postdate unit 21 (top of Olduvai event).

The spell of tholeiite volcanism that followed the Stardalur activity, but preceded the intermediate volcanism in Svínadalur, may possibly be unrelated to the Stardalur centre. An analysis from unit 22 is significantly richer in Al_2O_3 than other Esja tholeiites with similar CaO and MgO contents. A hypothetical olivine or pyroxene settling in the pre-eruption magma to give this composition must have been very thorough, as the analysed lava (chem.an.60) is aphyric.

After the caldera episode, central activity of the Stardalur volcano migrated northeastwards and a period of acid and intermediate volcanism commenced. The oldest analysed samples from this period are the icelandite and rhyolite parts of the composite sheet in Múli (page 67). This was followed by basaltic andesite eruptions; the first basaltic andesite eruptives (chem.an.64) are characterized by accessory amounts of plagioclase and pyroxene phenocrysts, whereas the subsequent basaltic andesites (chem.an. 65, 66 and 67) contain only rare plagioclase phenocrysts, rare microphenocrysts of pyroxene, but have accessory amounts of plagioclase microphenocrysts.

A period of tholeiite and olivine tholeiite volcanism followed the acid and intermediate activity and separates it from the final acid eruptive phase in the Stardalur volcano. The first tholeiite lavas to be erupted are characterized by their plagioclase phenocrysts (chem.an. 68 and 69), but higher up in the lava series these are less conspicuous. Both the samples below the middle and the one from near the top of the series do contain rare plagioclase phenocrysts, but, in addition, carry accessory amounts of olivine microphenocrysts, and are on the boundary of tholeiites and olivine tholeiites in the classification. These samples (chem.an. 70 and 71) are unusually rich in Al_2O_3 compared with other tholeiites in Esja. Fractionation of olivine from an olivine tholeiite parent magma may have produced this rock composition.

A period of olivine tholeiite volcanism followed. The location of the eruptive sites of this unit (unit 25) certainly suggest a tectonic if not a magmatic connection with the Stardalur central volcano. The very fine-grained olivine tholeiites, which have been analysed, contain rare, large phenocrysts of olivine and plagioclase.

In the final eruptive phase of the Stardalur central volcano rhyolites were erupted outside but concentrically with the Stardalur caldera. There is a considerable variation in the chemical composition of the rhyolites (chem.an. 75, 76, 77 and 78). Three published analyses (Sigvaldason 1958) of the Móskarðshnúkar rhyolite vary to a similar degree, but compare reasonably well with the analyses listed here. The analysed samples vary slightly petrographically in that sample 77 contains accessory amounts of plagioclase phenocrysts, sample 76 (the least altered of the rhyolite samples) contains rare micro-phenocrysts of pyroxene and olivine, whereas the other two samples are virtually aphyric. Despite these petrographic differences, however, the chemical variability of the samples is probably largely the result of secondary alteration.

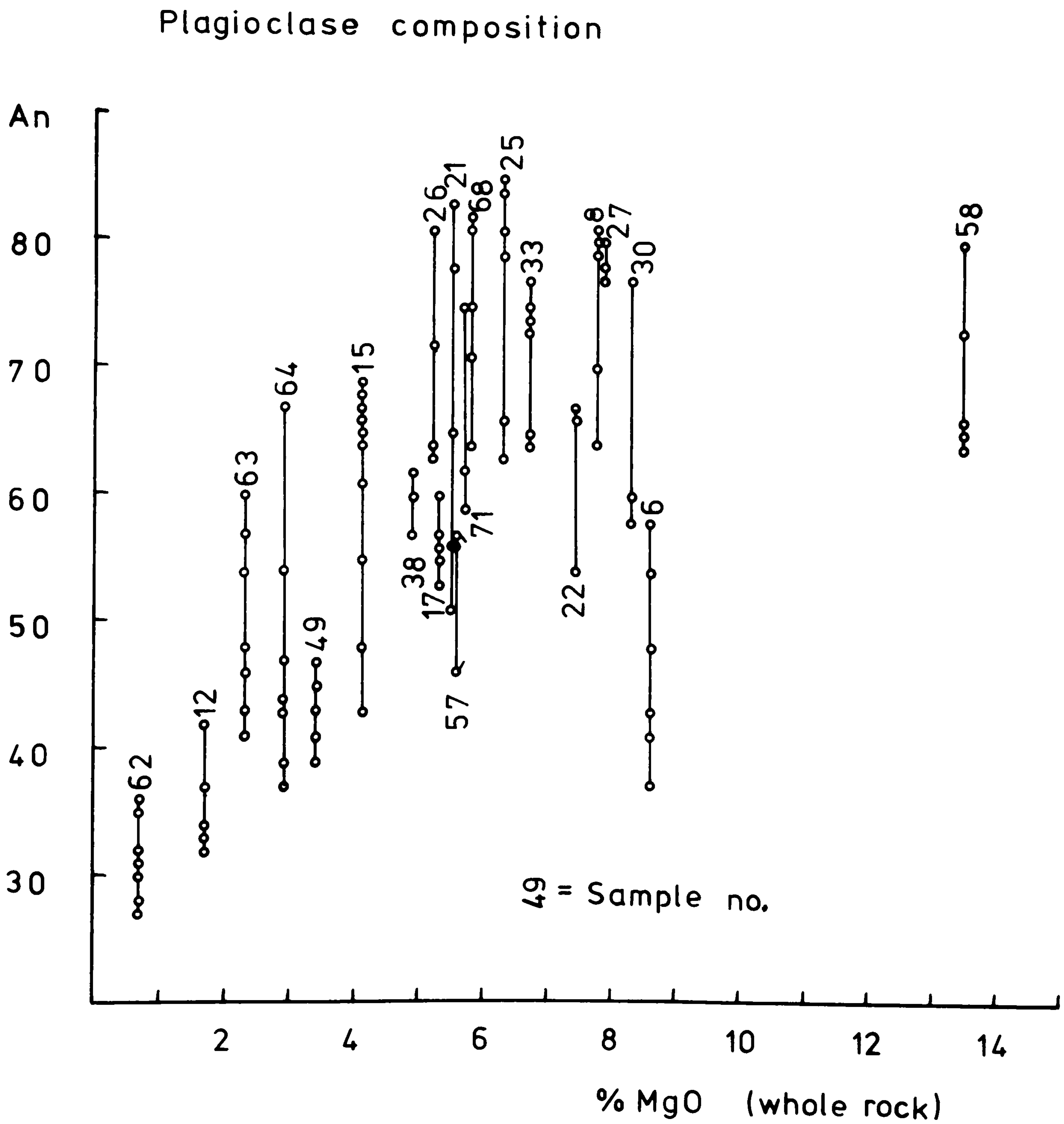
e. Problems of crystal fractionation versus partial melting in Iceland

Detailed mineralogical and petrochemical studies have been carried out on two tholeiitic central volcanic series in Iceland, the Þingmúli series (Carmichael 1964, 1967) and the Setberg Centre I series (Sigurðsson 1970a). Both these workers came to the conclusion that crystal fractionation from a basaltic parent magma (olivine tholeiite) could explain the continuous chemical and mineralogical variation from the basalts to the rhyolites.

So far as the Esja series is concerned, the presence and distribution of phenocrysts also favours fractional crystallization (see page 115). Partial electron-microprobe analyses were made of plagioclase (Ca), pyroxene (Ca, Mg, Fe) and olivine (Mg, Fe) in 21 samples from Esja. The results were largely in agreement with the mineralogical trends of Þingmúli, but in view of the incompleteness of the analyses they are not listed in detail in this thesis. The composition of plagioclase* is plotted versus the MgO content (whole rock) of the samples in Fig. 27. Each point represents an average of three analyses in a single grain, but some of the points represent several grains. No attempt is made in Fig. 27 to distinguish between phenocryst and groundmass crystal analyses, nor between core and rim analyses in zoned crystals. The most interesting feature to note

* An approximate composition of plagioclase was obtained by analysing for Ca and the An/Ab ratio read from a calibration curve made from synthetic "plagioclase" glass standards.

Fig. 27


Plagioclase composition
An
80
70
60
50
40
30
62
12
63
64
49
15
38
17
26
21
68
71
57
25
33
22
8
27
30
6
58
49 = Sample no.
2
4
6
8
10
12
14
% MgO (whole rock)

is the wide range in plagioclase composition in basaltic andesite sample 64 and icelandite sample 63. It would be difficult to obtain such a basic plagioclase if these intermediate rocks were formed in single batches by partial melting, whereas it could be explained by incomplete fractionation of a tholeiite. The icelandite is, however, from the basic part of a composite body, and both rocks (samples 63 and 64) could possibly be hybrids of a tholeiite and rhyolite (which itself might be produced by partial melting). The other basaltic andesite (sample 49) and icelandite (sample 12) in Fig. 27 do not show this abnormally wide range in plagioclase compositions.

Mount Hekla has commonly been regarded as a "living" example of fractional crystallization occurring within a central volcano. Thorarinsson (1967) presented data "that up to a certain limit" suggested that the extent of differentiation of Hekla's magma was related linearly to time; the longer the repose period before an eruption the higher the silica content of the initial (tephra) phase of the eruption. The eruption of Hekla in 1970, after a repose period of only 23 years, provided a welcome opportunity to test the differentiation theory. But instead of Hekla producing only basaltic andesite, as could be expected from Thorarinsson's curve, acid tephra with a silica content of up to 72% was erupted initially (Thorarinsson and Sigvaldason 1973). New whole rock analyses, mineral analyses (major, minor and trace elements) as well as petrographic data has shown (Sigvaldason 1973) that two types of magmatic liquid are produced beneath Hekla, a dacite (65-74% SiO_2) and a basaltic andesite (about 54% SiO_2). On the basis of the new data, he proposes that the dacites are produced by partial melting of olivine tholeiite basalts, which were originally erupted in the western (Reykjanes) rift zone (Sæmundsson 1973) and drifted eastwards (to the eastern volcanic zone) and subsided due to isostatic equilibrium to deep crustal levels (Pálmason 1973); the basaltic andesites could be formed by further partial melting of the depleted basalts, although the chemical data indicate some admixture of mantle material, and the intermediate products of Helka may be the result of mixing of the two liquids (Sigvaldason 1973).

Recent strontium isotope measurements (O'Nions et al 1973, O'Nions and Grönvold 1973) indicate that late and postglacial basalts in Iceland have been generated from a source region which is essentially homogenous with respect to Sr^{87}/Sr^{86}; the mean ratio for the basalts analysed is 0.70328 and the range is from 0.70317 ± 6 to 0.70334 ± 5 (2 sigma). Acid rocks from the Kerlingafjöll and Námafjall central volcanoes (Grönvold 1972) have Sr^{87}/Sr^{86} ratios which are indistinguishable from the basalts, but intermediate and acid rocks from the Torfajökull central volcanic region have significantly higher Sr^{87}/Sr^{86} ratios (0.70369 ± 10), and could not have been derived by fractional crystallization from basaltic magmas similar to those in the region (O'Nions and Grönvold 1973). A sample from the Hekla 1970 eruption has a ratio of 0.70337 ± 6, which is higher, but not statistically distinguishable from the average basalt composition. O'Nions and Grönvold (1973) proposed that the intermediate and acid rocks in the Torfajökull region were generated by partial melting of the base of the crust (layer 3) under the area, but this is a "cold palaeocrust", as will be discussed on page 183. Late glacial and recent rhyolitic volcanics cover an area of about 450km^2 making the Torfajökull region the largest rhyolite area in Iceland (Sæmundsson 1972). O'Nions and Pankhurst (1973) have demonstrated a secular variation in the Sr-isotope composition of Icelandic volcanics with the initial Sr^{87}/Sr^{86} ratios progressively decreasing from 0.7036 15 million years ago to 0.70328 at the present day. Extrapolation from the secular variation diagram may suggest that the intermediate and acid rocks in the Torfajökull region are formed by partial melting of 15-16 million years old basalts, whereas the source rock for Hekla may be 4-5 million years old. This is not altogether unrealistic with regard to the known crustal structure of southern Iceland.

O'Nions and Pankhurst (1973) measured the Sr^{87}/Sr^{86} ratios of five samples from Esja, which are listed in Table 3, sample descriptions and the localities can be obtained in Appendix 1. The age is inferred from the palaeomagnetic stratigraphy.

Table 3

Sample	Type	Unit	Age (m.y.)	Sr^{87}/Sr^{86}
2	Tholeiite	3	2.6	0.70331 ± 6
19	Tholeiite	12	2.3	0.70329 ± 6
38	Olivine tholeiite	19	2.0	0.70319 ± 5
58	Olivine tholeiite	21	2.0	0.70331 ± 6
63	Icelandite	23	1.9	0.70332 ± 5

All the analyses fall within the range found in late and postglacial basalts in Iceland, and are, with the exception of sample 38, virtually identical. This is as expected irrespective of whether one assumes that the icelandite is produced by crystal fractionation of a basaltic magma or that it is produced by partial melting of "young" lower crustal material.

O'Nions and Grönvold (1973) showed by comparison of observed rare earth element patterns and model calculations that both the tholeiitic and alkali basalt compositions in Iceland could be generated from the same source material (mantle peridotite) by different degrees of partial melting. They suggested that intermediate and possibly acid rocks might be produced by fractional crystallization from upper mantle derived basalt liquid, but might equally be produced by remelting of crustal material (layer 3); acid rocks by small degrees of melting and intermediate rocks by more extensive melting. The time interval between the emplacement of layer 3 and its subsequent partial remelting would decide whether the initial strontium ratios of the remolten products would be the same (short interval) as mantle derived contemporaneous basalts or different (long interval), assuming a secular variation of strontium ratios in the mantle source.

There is little doubt that crystal fractionation produces some of the chemical variations observed in the basaltic rocks of Esja. But whether the intermediate and acid rocks are formed by extensive fractionation or by partial melting of crustal material cannot be answered. It is apparent that the commencement of intermediate and acid volcanism in eastern Esja coincides with and follows the emplacement of the large intrusive bodies in the Stardalur caldera area which may (page 171) be a continuous extension of layer 3. The "upward surge" of layer 3 has certainly raised the thermal gradient in the crust and partial melting near the base of the crust may have produced the intermediate and acid rocks.

Chapter 5 GEOPHYSICAL ASPECTS OF ESJA AND THE CRUSTAL STRUCTURE OF ICELAND

a. Magnetic anomalies

The aeromagnetic survey of Iceland by Sigurgeirsson (1970) has revealed several distinctly localized magnetic anomalies. Many of these anomalies are apparently associated with central volcanoes, but a detailed survey of the magnetic properties of the rocks and the geological features affecting the shape of the anomalies has only been carried out on the strongest of these localized anomalies, the Stardalur magnetic anomaly (Friðleifsson and Kristjánsson 1972). A very much weaker anomaly is associated with the Kjalarnes central volcano. An aeromagnetic map of the Esja region (Fig.29) shows the total magnetic field intensity at 900m above sea level (reproduced from the map of Sigurgeirsson 1970).

The Kjalarnes centre magnetic anomaly is negative with a maximum amplitude of about -1800γ at 900m altitude (the mean regional field intensity is about 51700γ). The anomaly is roughly elliptical with a long axis of about 16km aligned northwest-southeast; the width varies from approximately 3 to 7km (as measured from the 50500γ contour line in Fig.29). The sharp boundary of the anomaly northeast of the Kjalarnes peninsula coincides with the boundary of the normal polarity (Gauss epoch) rocks northeast of and the reverse polarity (Matuyama epoch) rocks southwest of the Kjalarnes fault zone. In other places the boundary underlies either the sea or rock formations younger than the Kjalarnes centre. The total field intensity contour lines (Fig.29) north of the anomaly follow approximately the northeast-southwest strike of the country rocks and the intensity values correspond to the normal polarity rocks in western Esja and the reverse polarity rocks in central and eastern Esja. The anomaly is, however, elongated perpendicular to the strike, and it coincides roughly with the positive gravity anomaly of the Kjalarnes centre (Fig.28). This strongly suggests that the magnetic anomaly is caused largely by the reversely magnetised intrusives of the Kjalarnes central volcano.

Further evidence for this hypothesis is found in the lithology of the country rocks. The rocks outcropping within the boundaries of the anomaly are reversely magnetised (with the exception of the post-erosional olivine tholeiite lavas), and reversely magnetised rocks are "known to exist under Reykjavík to a depth of at least 650m" (Sigurgeirsson 1967), and extend downwards to more than 1km below sea level in the eastern end of the anomaly judging from the strike and dip of the rocks in Esja. These rocks will contribute to the magnetic anomaly. Extending the Esja stratigraphy into the anomaly region, however, suggests that the bulk of the reverse polarity rocks are hyaloclastites. These have suffered intense hydrothermal alteration and have therefore a low net magnetization, whereas the less permeable intrusives are fresher and have retained their magnetization. Kristjánsson (1972) measured the susceptibility and artificial TRM (thermal remanent magnetic) intensity of 12 samples of drill chips between 800 and 2200m depth

Fig. 28

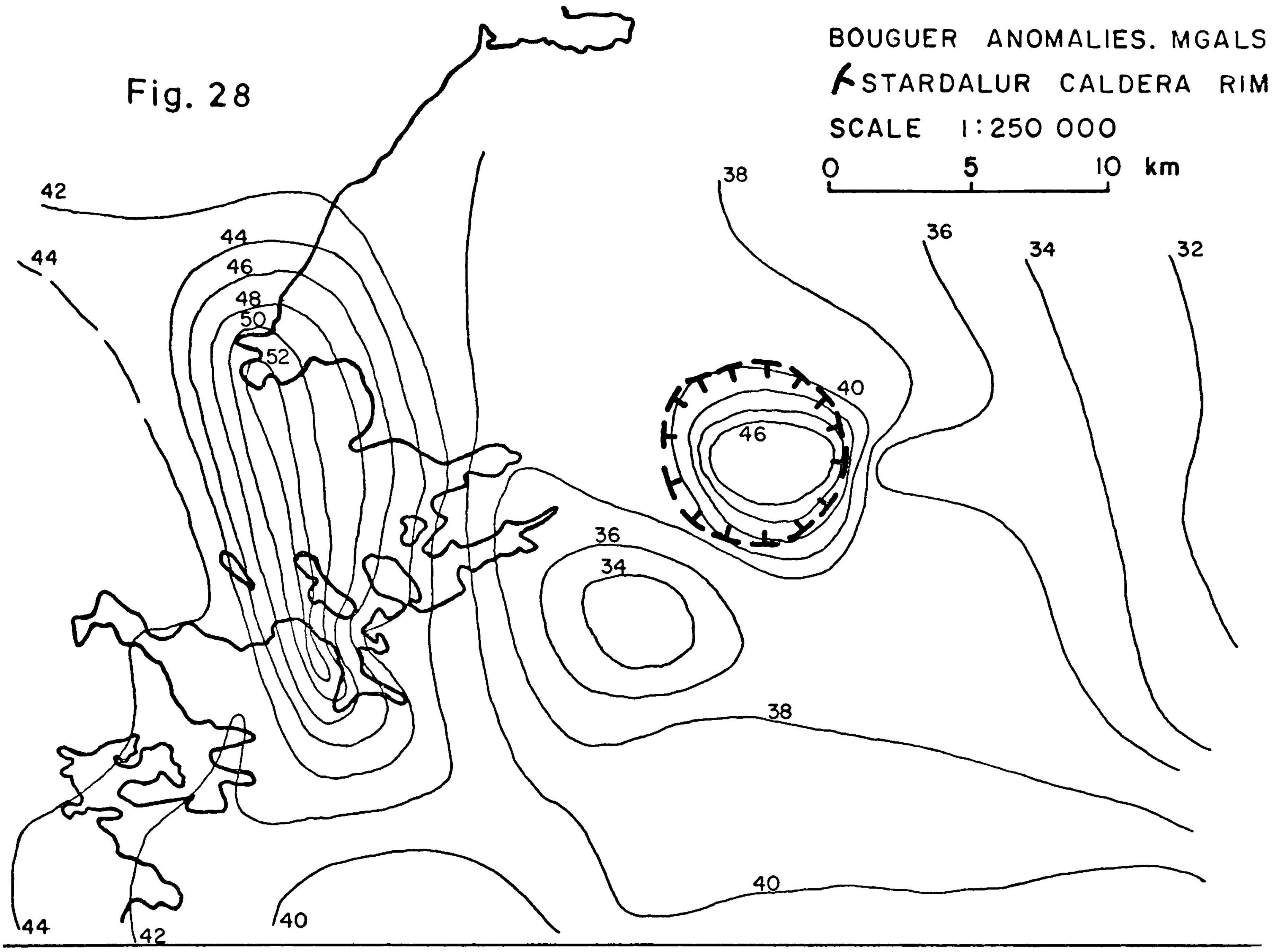
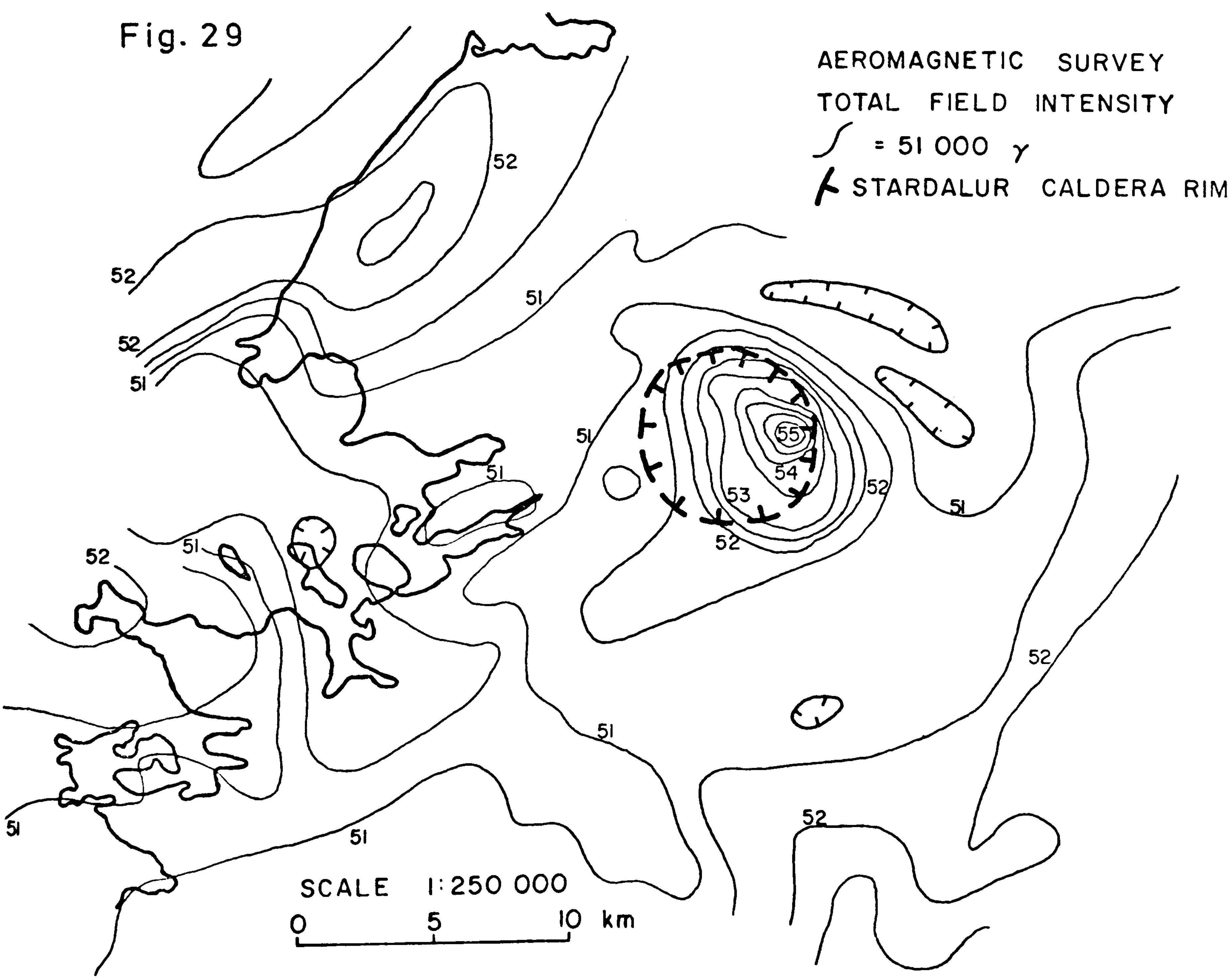

BOUGUER ANOMALIES. MGALS
STARDALUR CALDERA RIM.
SCALE 1:250 000
0 5 10 km
42
44
44
46
48
50
52
38
36
34
32
40
46
36
34
38
40
44
42
40

Fig. 29


AEROMAGNETIC SURVEY
TOTAL FIELD INTENSITY
= 51 000 γ
STARDALUR CALDERA RIM.
52
52
51
52
51
51
55
54
53
52
52
51
51
52
51
52
51
51
51
52
SCALE 1:250 000
0 5 10 km

below Reykjavík and found indications that, in this depth interval (which is probably below the main hyaloclastite units), the amount and composition of magnetite is similar to that in oxidised surface basalts.

The Stardalur magnetic anomaly is limited to a slightly oval shaped area 8-10km across. It is positive with an amplitude of 4500γ at 700m altitude. Ground surveys indicate (Kristjánsson 1970, Búason 1971) that the anomaly is composed of a broad anomaly of dimensions 7km by 5000γ, on which are superimposed a few sharp local maxima, the highest one reaching 79000γ at ground level (the regional field is about 51700γ).

On the simplified geological map (Fig.30, taken from Friðleifsson and Kristjánsson 1972), the palaeomagnetic polarities are shown and the polarity epochs and events numbered* as in Einarsson (1957). As mentioned on page 54 the normal polarity lavas of unit 19 succeed the reverse polarity hyaloclastite unit 18. In the caldera collapse the normal polarity lava unit, which thickens southeastwards away from the hyaloclastite ridge of unit 18, was faulted and tilted within the caldera. The subsequent intra-caldera filling and the intrusives within the caldera are also normally magnetised. Within the caldera is therefore a pile up to 700m thick of normally magnetised eruptives (lavas and hyaloclastites) and a very high concentration of normal polarity intrusives at

* It should be noted that the basic hyaloclastites south of the inferred caldera rim are now assigned to N2 (Gilsá event) and not N3 (Olduvai event) as indicated on the map. The overlying basic lavas are similarly assigned to R1 and not R2.

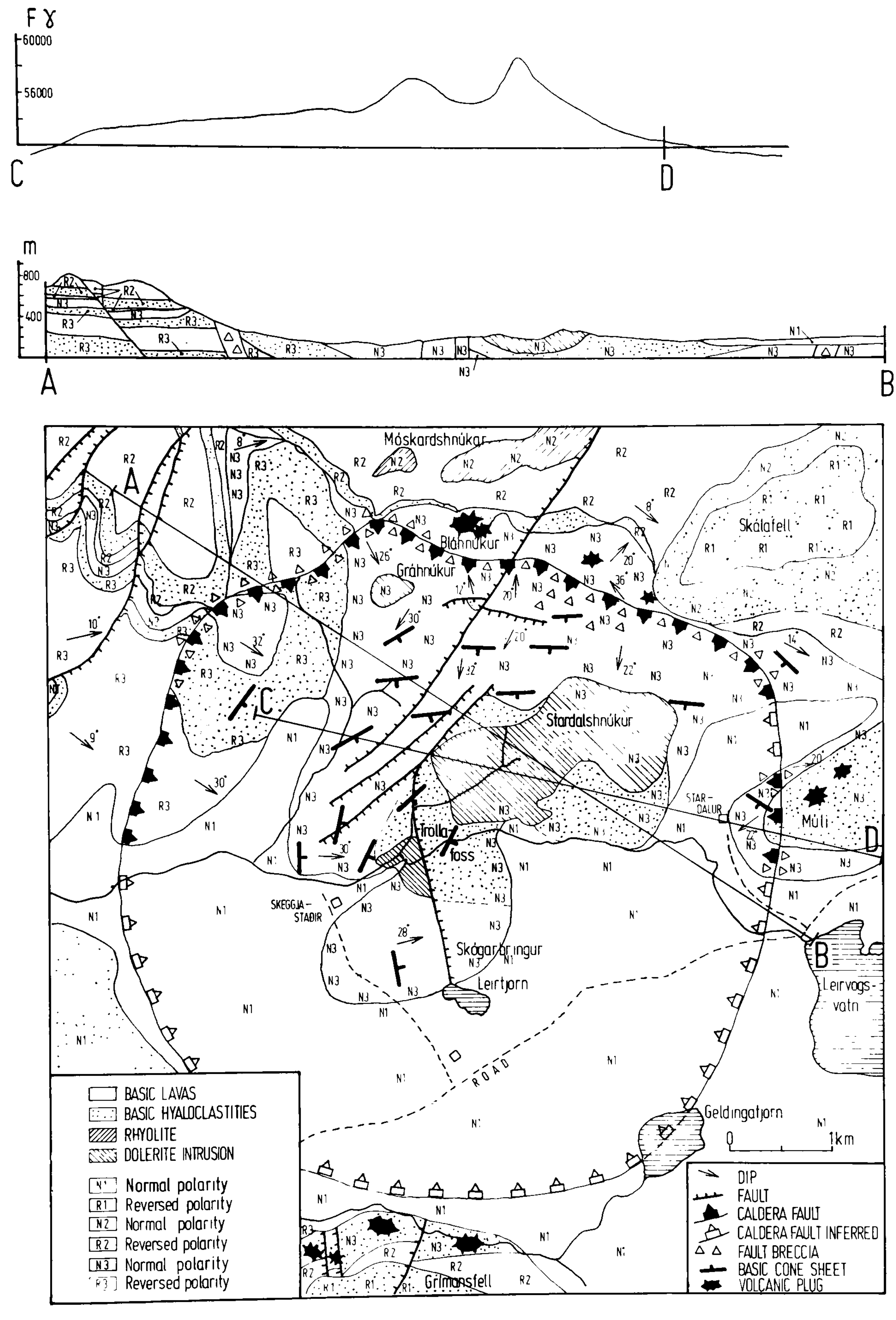


Fig.30. Geological-palaeomagnetic map of the Stardalur caldera. The aeromagnetic profile was kindly provided by Professor Þ. Sigurgeirsson. (From Friðleifsson and Kristjánsson 1972).

greater depth. This is in sharp contrast with the rocks to the north and west of the caldera where a less than 100m thick layer of normal polarity rocks is sandwiched between about 2km and 300m of reverse polarity rocks below and above respectively. The aeromagnetic intensity profile C-D in Fig.30 corresponds very closely to the geological features, and the shape of the anomaly is clearly related to the caldera structure. The intensely brecciated and hydrothermally altered (i.e. having a low net magnetization) rocks of the caldera fault zone probably help to sharpen the boundaries of the anomaly.

In order to elucidate further the cause of the broad magnetic anomaly, the remanence and susceptibility values of 160 samples from surface formations in the Esja region were measured by Kristjánsson (Friðleifsson and Kristjánsson 1972). The results of these measurements are highly scattered, but logarithmic averages of total magnetization (remanent plus induced, regardless of sign) are given in Table 4 for the main categories.

It is evident from this table that the fine-grained, near surface intrusives (dykes, sheets, plugs) make an important contribution to the the broad magnetic anomaly in Stardalur. The magnetization values depend very much on the chemical composition of the rocks, with the iron-rich varieties (tholeiite and basaltic andesite) having higher magnetization than the rocks with lower iron contents (olivine tholeiites). The average total iron values of analysed samples from some of

Table 4

Total magnetization of samples from surface formations in the Esja region

No. of Samples	Type	Locality	Average magnetization in 10^{-3}G.
23	olivine tholeiite lavas	Stardalur, Reykjavík	2-3
9	tholeiite lavas	Stardalur	6-8
30	tholeiite lavas	inside Stardalur caldera; Úlfarsfell	1-3
20	basaltic andesite plugs	Bláhnúkur, Tröllafoss, Múli	10-15
7	dolerite (olivine tholeiite) intrusions	Stardalshnúkur, Leiðhamrar, Kvensöðlar	1-2
16	dolerite (tholeiite) intrusions	Músarnes, Gráhnúkur	6-8
12	tholeiite dykes	Stardalur, Skálafell, Skeggjastaðir	10-20
14	rhyolite	Esja, Grímmannsfell	0-1

the rock bodies listed in Table 4 are plotted against the average magnetization values in Fig.31. It should be noted, however, that the chemical analyses and the magnetic measurements were not made on the same samples. From this preliminary comparison it seems clear that the grain size of the rock and the order of crystallization (i.e. whether magnetite crystallises early as a phenocryst phase) are also very important factors; the fine-grained basaltic andesite plugs have much higher average magnetization than for example the coarse-grained, iron-rich, tholeiite dolerite of Gráhnúkur (chem.an.57). It should be stressed, however, that important factors such as

Fig. 31

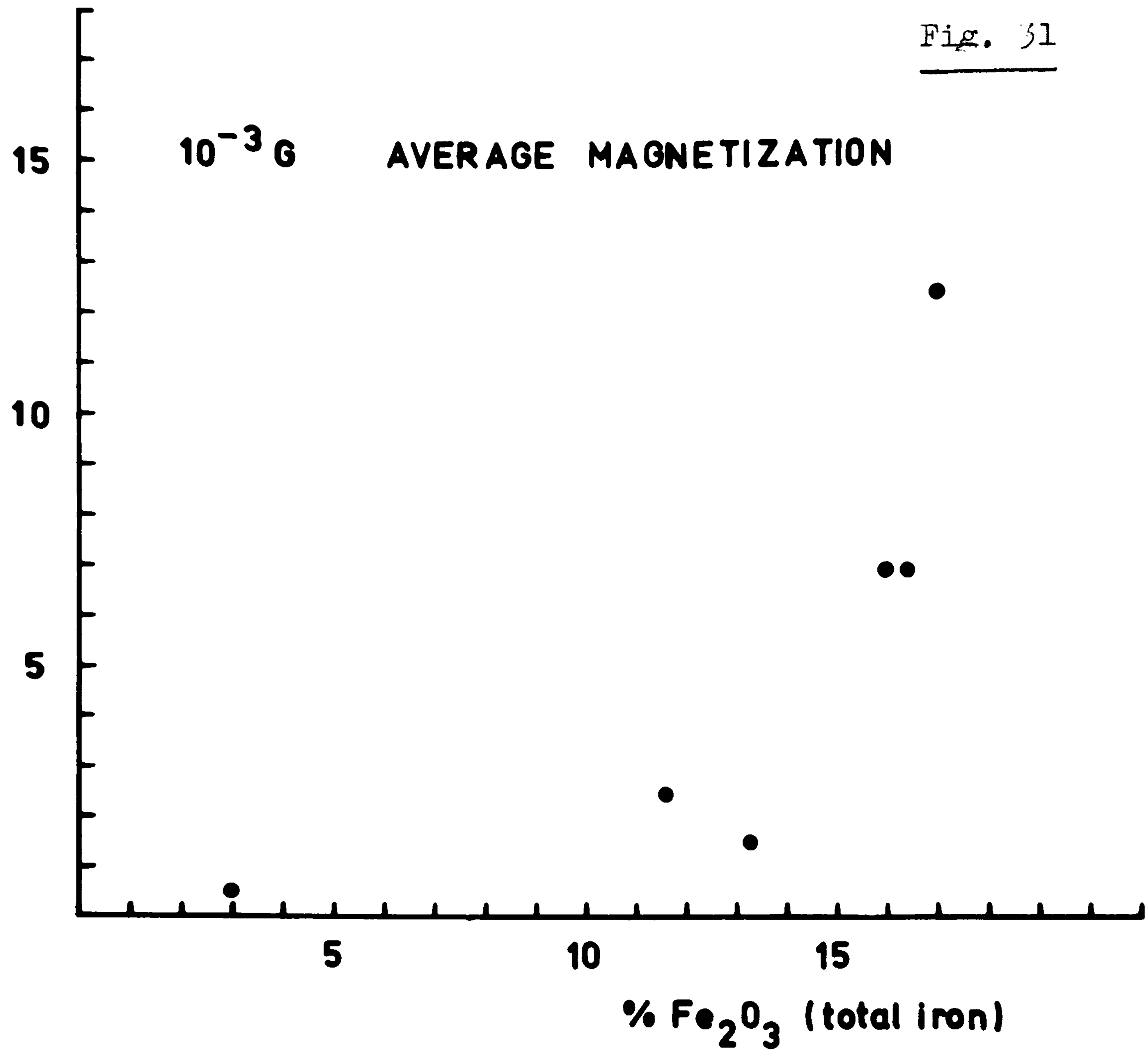
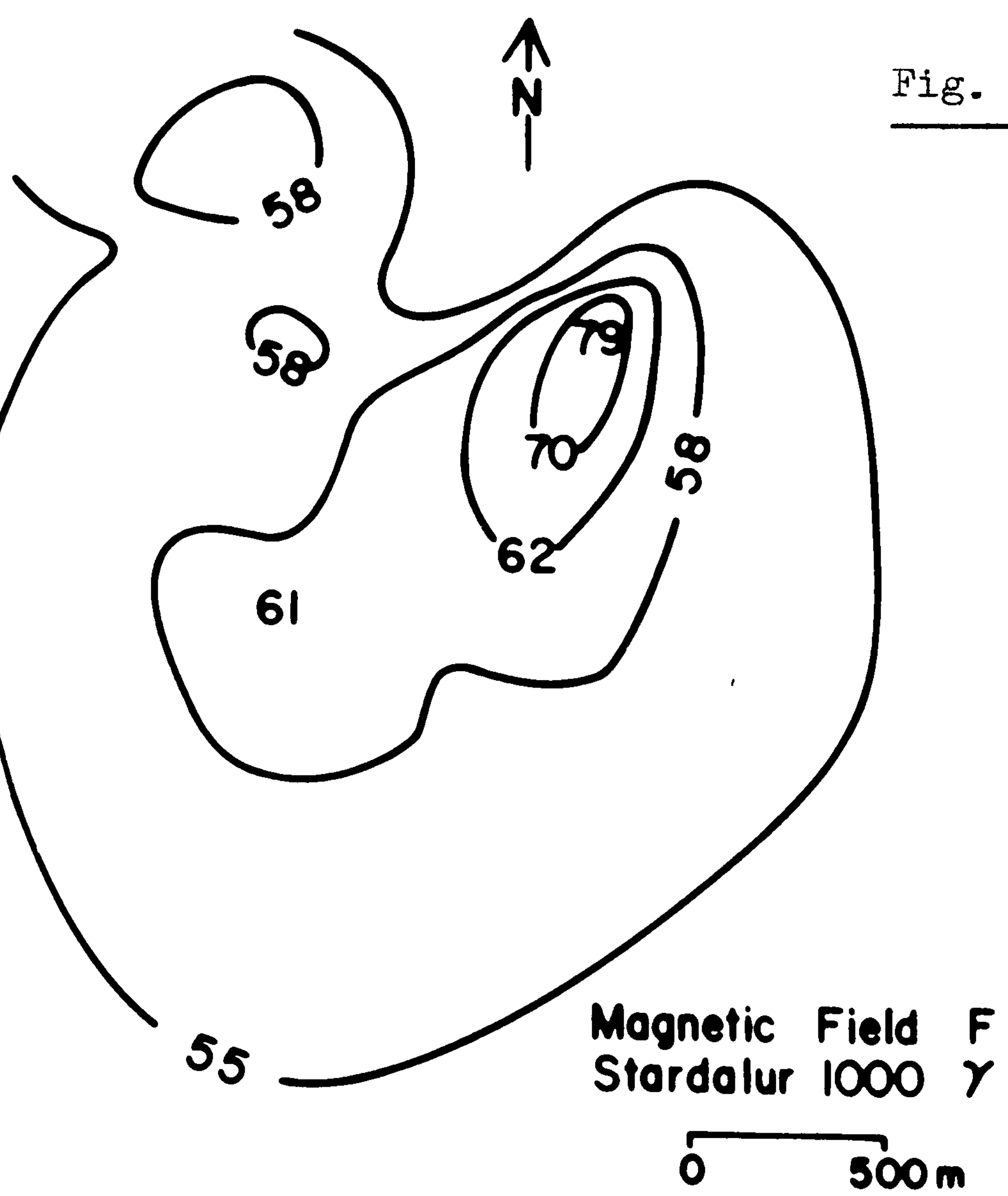

15
10
5
10^{-3} G AVERAGE MAGNETIZATION
5
10
15
% Fe_2O_3 (total iron)

Fig. 32


N
58
58
79
70
62
58
61
55
Magnetic Field F
Stardalur 1000 γ
0 500m

the partial pressure of oxygen in the crystallizing melt, the Earth's magnetic field strength at the time of solidification, and secondary alteration, are omitted in this comparison.

Sigurgeirsson's discovery of the Stardalur magnetic anomaly, which is considerably stronger than any other recorded magnetic anomaly in Iceland, aroused much interest among scientists and laymen, since strong magnetic anomalies may indicate ore bodies. Fig. 32 (from Kristjánsson 1970) shows the simplified features of a total-field ground magnetic survey of the highest local maximum, located southwest of Stardalur farm. From preliminary ground survey data, Kristjánsson (1970) computed the dimensions of a possible single body causing this highest maximum as 200 by 600m, striking NE, with an estimated upper surface at a depth of 50-60m, and a total magnetic intensity (remanent plus induced) of 0.05-0.06 Gauss. A more detailed survey and analytical treatment by Búason (1971) yielded a total magnetization of 0.08 Gauss and slightly different dimensions.

A 200m deep drillhole (H-1 of Friðleifsson and Tómasson 1972) was sunk at the centre of the highest maximum with the dual purpose of investigating the cause of the magnetic anomaly and to measure the thermal gradient. Below the post-erosional olivine tholeiite lavas the drill entered the tholeiite lavas of unit 19 (in the Esja stratigraphic column) and did not reach the base of that unit. Búason (1971) measured the total magnetization values of 103 specimens of the drillcore between 41 and 143m depths and found a range from 0.021 to 0.119 Gauss, which is almost an order of magnitude higher than the average value of Tertiary and Quaternary basalts in Iceland.

Steinþórsson and Sigvaldason (1971) examined petrographically many of the samples measured by Búason, and analysed 17 samples for Fe, Sr, Rb, Zr and Y. The average iron content of the samples was 11% Fe (or 15.6% Fe_2O_3 as total iron), but there was a range from 8.75 to 14.3% Fe. They found that the fresher samples showed higher magnetic intensity than the altered ones with similar iron contents.

Various magnetic properties of four randomly selected Stardalur drill core specimens were investigated by Kristjánsson (summarized in Friðleifsson and Kristjánsson 1972). He found the Curie points typical for high-temperature oxidized titanomagnetites, where the magnetic constituent is magnetite (T_c= 580°C), possibly somewhat oxidized towards maghemite composition. The conclusion that the Stardalur magnetic mineral is magnetite was supported by Kristjánsson's microscope observations of polished thin sections and hysteresis curve shapes, as well as by x-ray work (Steinþórsson and Sigvaldason 1971). From susceptibility data the mean magnetite content in the Stardalur core is estimated to be about 2.5 volume % in comparison with the average for Icelandic basalts of 1.0 $\pm$ 0.2 volume %.

One possible cause of the high remanence intensities in the Stardalur tholeiite lavas, the average of which is of the order of ten times higher than those in other Icelandic Lower Quaternary basalts (Kristjánsson 1970), is that the NRM (natural remanent magnetization) resides in magnetite grains of single domain size. However, no significant difference was found between the Koenigsberger ratio (a good criterion for indicating this) of the Stardalur tholeiites and those of other Icelandic basalts (Kristjánsson 1972).

A contributing cause to the high natural remanent magnetization intensity at Stardalur may be a high geomagnetic field strength at the time of emplacement. This may be investigated by comparing the NRM of specimens with the TRM acquired in a field of known strength. Such considerations indicate that the field strength may have been 2.5-3 times higher than the average magnetic field strength in Iceland in Upper Cenozoic times (Friðleifsson and Kristjánsson 1972).

Steinþórsson and Sigvaldason (1971) found the partial pressure of oxygen of the Stardalur tholeiite melt to have been unusually high in comparison with other Icelandic basalts. This provides a plausible explanation of the unusually high volume percentage of magnetite in this iron-rich, but not unusually iron-rich, tholeiite. It would be of interest to apply similar tests to the other three magnetic anomalies in southwest Iceland recorded by Sigurgeirsson (1970) and noted by him to lie on a single line, which includes Stardalur (Fig.33). The four magnetic anomalies are, in order of increasing age: Skálafell (late Quaternary, within the present active volcanic zone); 25km further NNW is Stardalur (about 2 m.y. old); 23km further NNW is Hvalfjörður (about 3 m.y. old); and 20km further NNW is Hvanneyri (about 4 m.y. old). The Hvanneyri and Hvalfjörður anomalies are both associated with central volcanoes, but Skálafell is formed of subglacial hyaloclastites and mostly covered by a late interglacial shield volcano lava, and is not known to be associated with a central volcano.

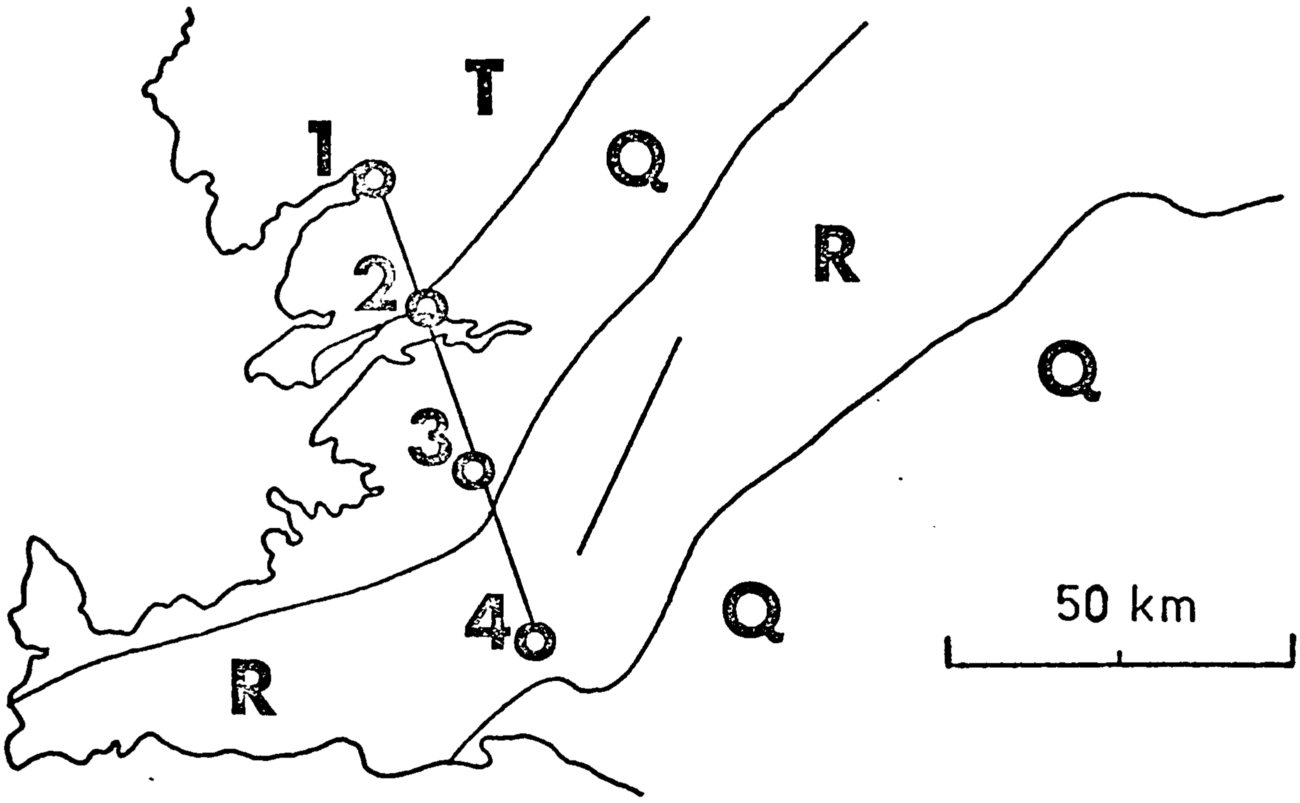


Fig.33. The four strong magnetic anomalies in southwest Iceland. 1 Hvanneyri, 2 Hvalfjörður, 3 Stardalur, 4 Skálafell. The geological boundaries between T Tertiary, Q Quaternary and R recent volcanics are shown and fault direction within the recent volcanic zone is indicated. The "hi-p-p-o" spot may have migrated about 1cm/yr southwestwards along the spreading axis.

Assuming a common cause for the four magnetic anomalies, then of the two proposed main causes for the unusually high remanent intensity of the Stardalur lavas, i.e. the unusually high magnetite volume percentage of the tholeiite and the high palaeo-field strength (Friðleifsson and Kristjánsson 1972), only the former is likely to be applicable to all four anomalies. An unusually high partial pressure of oxygen could therefore be expected at all four anomaly sites. If this is the case the questions arise: why do these conditions occur at such regular intervals in time (about 1 m.y. between the three older ones) or is this fortuitous, and why does the phenomenon follow a fairly straight line in space, or is this fortuitous?

Vogt (1971) suggested that V-shaped volcanic basement features symmetrical about the spreading axis on the sea floor southwest of Iceland were due to asthenosphere flow away from the Iceland "hot spot" with a flow rate an order of magnitude greater than the spreading rate.

A line drawn through the sites of the four magnetic anomalies makes an approximately 45° angle with the eruptive fissures in the Reykjanes neo-volcanic zone. A hypothetical "high partial pressure of oxygen spot" (hi-p-p-o spot!), if confined within the main spreading zone, would therefore migrate southwestwards along the axial spreading zone at a speed equivalent to half the overall spreading rate.

It is of considerable interest to note that of the five samples analysed for Sr^{87}/Sr^{86} (Table 3, page141) four had virtually identical Sr isotopic ratios, but one had a slightly lower ratio; this rock was probably erupted contemporaneously with the tholeiite lavas in Stardalur, which reflect unusually high partial pressure of oxygen conditions. Both the high partial pressure of oxygen and the low Sr^{87}/Sr^{86} ratio are likely to be "mantle controlled", but are the two phenomena related?

b. Gravity anomalies

The gravity survey of Iceland by Einarsson (1954) showed a bowl-shaped Bouguer anomaly with a minimum of about -35 mgals in the central part of the island and an average of about 40 mgals along the coast line. Superimposed on this main anomaly are several local gravity anomalies, ranging in diameter from a few kilometers to a few tens of kilometers, and up to 15 mgals in strength relative to the background. These are seen particularly well on Einarsson's gravity map of southwest Iceland where the density of gravity stations is larger than in other parts of the country. Most of the positive, local gravity anomalies are associated with extinct central volcanoes.

Einarsson (1954) discovered two positive gravity anomalies in the Esja region (Fig.28, page 146), and associated these with the intrusives in the Kjalarnes centre region and in the Stardalur region.

The Kjalarnes centre Bouguer anomaly has a relative height of 15 mgals; it is elongated NW-SE with a long axis of about 19km and a short axis of up to 8km (as measured from the 44 mgal contour line). The shape may, however, be distorted by the location of the gravity stations along the coast line. The Lauganípa-Þverfell-Leiðhamrar intrusives lie on the eastern rim of the anomaly.

The Stardalur Bouguer anomaly has a relative height of 10 mgals. It is roughly circular in shape with a diameter of about 7km (as measured from the 40 mgal contour line) and coincides essentially with the exposed parts of the Stardalur caldera rim (Fig.28). The non-exposed southern half of the caldera rim is inferred on the basis of the gravity anomaly, but this is supported by electrical resistivity measurements (Friðleifsson and Tómasson 1972). Twenty resistivity profiles, up to 1500m deep, were measured within, and east and west of the caldera. Low resistivity values, probably indicating the presence of hot water, were found on the eastern and southern parts of the caldera rim supporting field evidence (hot water aquifers and drill hole data) suggesting that the caldera fault zone provides a path for hot water flowing from depth. Relatively high resistivity values were encountered within the central 4km diameter of the caldera. By comparing the thermal gradients measured in the three drillholes within the caldera, the resistivity values at the drill sites, and similar data from other areas outside central volcanoes, it is thought likely, that the dense intrusives within the caldera (which would be fresher and contain less pore water than the country rocks) are the cause of the high resistivity, and hence the resistivity profiles may indicate their presence and the location of the caldera fault zone.

It seemed likely that large dolerite intrusives are the main contributors to the two gravity anomalies. However, in order to establish the extent to which the smaller, fine-grained intrusives and the fine-grained extrusives (which are much more glassy and vesicular) may contribute, in addition to studying the variation of density with composition, the specific gravity of all the chemically analysed samples of the Esja collection were measured*. The density values are listed with the chemical analyses in Table 2, page 125. The density values are plotted against MgO (the most variable oxide in the basalts) in Fig. 34, and separate symbols are used for coarse-grained intrusives, fine-grained intrusives and lavas.

The porosity of the samples was not measured. Pálsson (1972) found a linear relationship between the dry density and the porosity of rocks from the Tungnaá area in Iceland, which agrees with the results of Manghnani and Woolard (1968) for Hawaiian rocks.

In Fig. 34 there is a considerable scatter in the density values for the lavas; this probably reflects the porosity of the samples. There is much less scatter in the data for the fine-grained intrusives, which are almost free of vesicles, and thus give a much clearer picture of the true rock density. The coarse-grained intrusives have a fairly wide scatter in density values; this is probably due to a combination of porosity, alteration and absorbed water. There seems to be a fairly linear increase in densities from the rhyolites to the tholeiites where the increment decreases and the values approach 3.15 g/cm^3 for the olivine tholeiites. There is, however, a fairly wide scatter in the density values for the basaltic andesites and the tholeiites; the samples having the

* The specific gravity measurements were made on a Walker's steel yard using deionised water and a "wetting agent" to reduce the surface tension. Five cubes (1-4cm^3) were measured from each sample; measurements deviating more than ± 0.04 g/cm^3 were rejected and more cubes measured except where only one measurement was rejected in which case an average of four was taken.

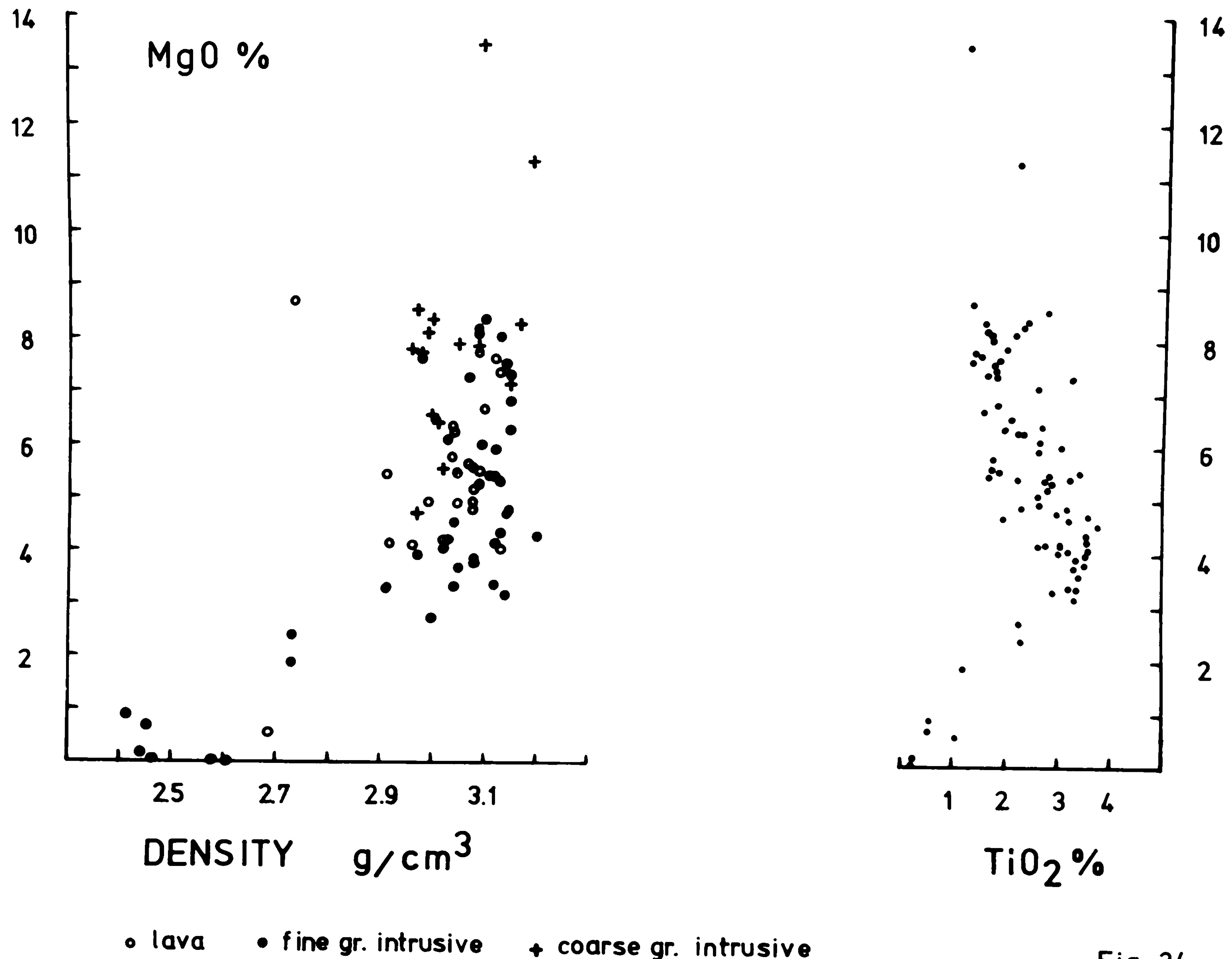

MgO %
14
12
10
8
6
4
2
2.5
2.7
2.9
3.1
DENSITY g/cm3
1
2
3
4
TiO2 %
o lava
• fine gr. intrusive
+ coarse gr. intrusive

Fig. 34

highest densities are the very iron- and titanium-rich rocks associated with the two central volcanoes. TiO_2 is plotted against MgO in Fig. 34 for comparison with the density pattern.

Within the complete compositional range from rhyolites to olivine tholeiites there is a density difference from about 2.5 to about 3.15, or about 0.65 g/cm^3. Clearly this range of density values can give rise to local gravity anomalies in the central volcanic regions, where there may be local concentrations of rocks of different chemistry. In the case of the Kjalarnes centre the iron- and titanium-rich tholeiites (e.g. the lavas of unit 3 and the pillow breccia on the tip of the Kjalarnes peninsula) must be considerably denser than the average tholeiites of similar lithology. If a shallow level chamber with a magma having a similar composition to these tholeiites ever existed, the early forming ilmenomagnetite crystals may have fractionated out of the magma and the cumulate would, of course, have a much higher density than any of the normal rock types. No examples of magnetite cumulates have, however, been found in this or any other part of Iceland. In the case of the Stardalur centre the tholeiites and basaltic andesites are similarly contributing factors. The fine-grained intrusives in both centres are likely to be denser than their coarse grained equivalents due to the preferential alteration of the coarser rocks.

c. Effects of petrochemistry on rock density in Iceland

Einarsson's (1954) bowl-shaped gravity anomaly reaches a minimum value of -35 mgals near the centre of Iceland, 75 mgals lower than the average value around the coastline (40 mgals). The Bouguer anomaly is a mirror image of the average topography, and the island is apparently in isostatic equilibrium.

Einarsson (1954) suggested that the gravity bowl could be explained by a thin sialic crust under Iceland. No indications have, however, been found of such a layer in seismic work (Pálmason 1971). The seismic refraction data have not as yet given an indication of the cause of the main bowl-shaped anomaly.

Bott et al (1971) have studied the deep structure of the Iceland-Faeroe ridge. They showed that passing from the centre of Iceland onto the ridge, the Bouguer anomaly increases from -35 mgals to about 110 mgals. The gravity gradient is gentle and the steep but small bathymetric scarp separating the Iceland block from the ridge does not show up as more than a local feature on the Bouguer anomaly profiles. A crust thinning away from Iceland could account for the 145 mgals change, but the seismic work of Pálmason (1971) and Bott et al (1971) demonstrated that the reverse is the case, the crust under the Iceland-Faeroe ridge being slightly thicker than that under Iceland, and there is every reason to believe that the ridge was formed subaerially. Bott et al (1971) also considered the possibility of the mean density of the Iceland crust being lower because of higher temperature, but found that even with a temperature difference of several hundred degrees the density difference of such a thin crust would be too small to explain the Bouguer anomaly. They came to the conclusion that the most likely explanation of the gravity anomaly is lateral variation in the density of the upper mantle (i.e. not the crust) which in turn could be attributed to a hot spot under Iceland.

One possibility, not considered by Bott et al (1971), is that the mean density of the Iceland crust is lower than the crust on the ridge as a consequence of petrochemical differences between the two regions. The estimated proportions of different rock types in an area of 500-600km^2 in Reyðarfjörður, eastern Iceland (Walker 1959), are listed in Table 5 together with density values estimated from Fig. 34. The range of densities in Table 5 is probably a minimum.

Table 5

Rock type	Estimated %	Estimated density (g/cm^3)
Olivine tholeiite	23	3.15
Tholeiite	48	3.11
Porphyritic tholeiite	12	3.07
Andesite (icelandite)	3	2.75
Rhyolite	8	2.60
Detrital beds	6	2.30

The average density of the Reyðarfjörður pile, if it were non-porous, would be about 3.01g/cm^3. Taking the crustal thickness of eastern Iceland as about 15km and assuming that the Iceland-Faeroe ridge is composed entirely of olivine tholeiite* (density=3.15g/cm^3), the gravity anomaly resulting from this density difference would be about 88 mgals.

No account has been taken in these calculations of the effect of vesicularity (the density decreases linearly with increasing porosity). All field volcanologists know that olivine tholeiite lavas tend to be more vesicular than tholeiite lavas, and this may cancel out some of the effect of the density difference rising from the chemistry in the lava pile. But the relationship would hold true in the intrusive layer 3, which is thicker than the lava layers. It appears that a considerable part of the negative gravity anomaly of Iceland can be explained by the geochemical difference of the volcanics in Iceland in comparison with the surrounding areas.

Talwani et al (1971) found a Bouguer minimum on the Reykjanes ridge with a gravity gradient comparable to that found further south (at 30°N) on the mid-Atlantic ridge. They explained the isostatic equilibrium of the ridge in terms of alteration of the top part of the mantle, and the Bouguer low over the Reykjanes ridge crest (average about 60 mgals lower than that over the mid-Atlantic ridge crest) in terms of a thicker

* To the author's knowledge no analysed dredge samples exist from the ridge. Bott et al (1971) show a slight thickening of the crust under the Faeroes and a lighter crust there than under the ridge. The petrochemistry of the lavas in the Faeroes (Noe-Nygaard and Rasmussen 1968) is exclusively tholeiitic basalts with a lower aphyric tholeiite series (30%), a middle porphyritic series (40%), and an upper olivine tholeiite series (30%). It can therefore be reasonably assumed that the denser material on the ridge is to a large extent olivine tholeiite.

anomalous mantle layer under the Reykjanes ridge. They found that the Bouguer anomaly on the Reykjanes ridge crest, some 300km southwest of the tip of the Reykjanes peninsula, had an absolute value of about 110 mgals, and explained the absolute difference of 145 mgals between the Bouguer anomalies on the ridge crest and in the centre of Iceland in terms of the anomalous mantle being thicker under Iceland. North of Iceland the Bouguer anomaly is about 120 mgals at 68°N (Meyer et al 1972).

It is of interest to note that the Bouguer anomaly difference between the volcanically active Reykjanes ridge and Iceland, ant that between the aseismic Iceland-Faeroe ridge and Iceland are of similar magnitudes. Schilling (1973) has demonstrated a progressive increase in the incompatible and minor elements along the Reykjanes ridge towards Iceland indicating a gradual change from the olivine tholeiites on the ridge towards more evolved rocks in Iceland.

The gravity data suggests that, if the negative Bouguer anomaly is due to a hot spot under Iceland, the hot spot must be very well focussed under the centre of the country. The gravity profiles from the centre of Iceland to the Iceland-Faeroe ridge and to the Reykjanes ridge are fairly similar, suggesting that the hot spot is not connected with the anomalous mantle under the Reykjanes ridge.

d. Seismic structure

Detailed seismic refraction measurements have been carried out in Iceland to study the seismic velocity structure of the crust and its thickness (Pálmason 1971). A characteristic seismic layering has been found, resembling the oceanic crust in velocity values, but the Icelandic crust is thicker. The lowest seismic velocities, Layer 0 (Table 6, from Pálmason 1971), are found in the surface layer in the active volcanic zones. This layer consists of recent lava flows, hyaloclastites and sediments, and reaches a maximum thickness of 1km. Layer 1, which consists of Tertiary and Quaternary lavas with some hyaloclastite intercalations and is a surface layer outside the active volcanic zones, has distinctly higher velocity values and correspondingly higher average density. Its thickness is usually 0.5-2.0km with an average value of about 1km. Layer 2, which consists of lavas which have been buried fairly deeply in the volcanic pile (and have hence suffered burial metamorphism) and usually contains several percent of dyke rock, is only found at the surface where the lava pile has been deeply eroded, mainly in southeast Iceland. Its thickness is usually in the range 1-3km with an average value

Table 6 (from Pálmason 1971)

Layer	Average P-velocity(km/sec)	Average density(g/cm^3)
0	2.8	2.1-2.5
1	4.2	2.6
2	5.1	2.65
3	6.5	2.9
4	7.2	3.1

close to 2.1km. Layer 3 is found beneath the whole of Iceland but is nowhere at the surface. The depth to its upper boundary has been mapped in some detail (Fig.35, from Pálmason 1971) and is quite variable, usually in the range 1 to 5km but increasing conspicuously southeastwards in southern Iceland. The thickness of the layer is usually in the range 4 to 5km, but a greater thickness is indicated in northern Iceland (Pálmason 1971). Layer 3 in Iceland, with an average P-wave velocity of 6.5km/sec, is probably equivalent to the oceanic layer, although the average P-wave velocity of the latter is commonly given as 6.7-6.8km/sec. The lower velocity in the Icelandic crust cannot be explained wholly by higher temperatures (Pálmason and Sæmundsson 1973) but may be explicable in terms of the Icelandic crust having a lower average density than the crust surrounding Iceland because of petrochemical differences (see page 164). Layer 3 probably consists of metamorphosed basalt lavas with a large volume percentage of intrusives (Pálmason and Sæmundsson 1973). Layer 4 is found at a depth of 8 to 9km in southwest Iceland, but this increases to 14 to 15km east of the Reykjanes-Langjökull volcanic zone in southern Iceland, probably in northern Iceland and also west of the Snæfellsnes peninsula in western Iceland. The depth to its upper surface is, however, unknown in large areas of northern and eastern Iceland. This layer is thought to represent the upper mantle.

The very large variations in the depth to the upper surface of layer 3 in Iceland (Fig.35) pose in many ways the most interesting problem in the interpretation of the seismic refraction data (Pálmason 1971). Pálmason found a close

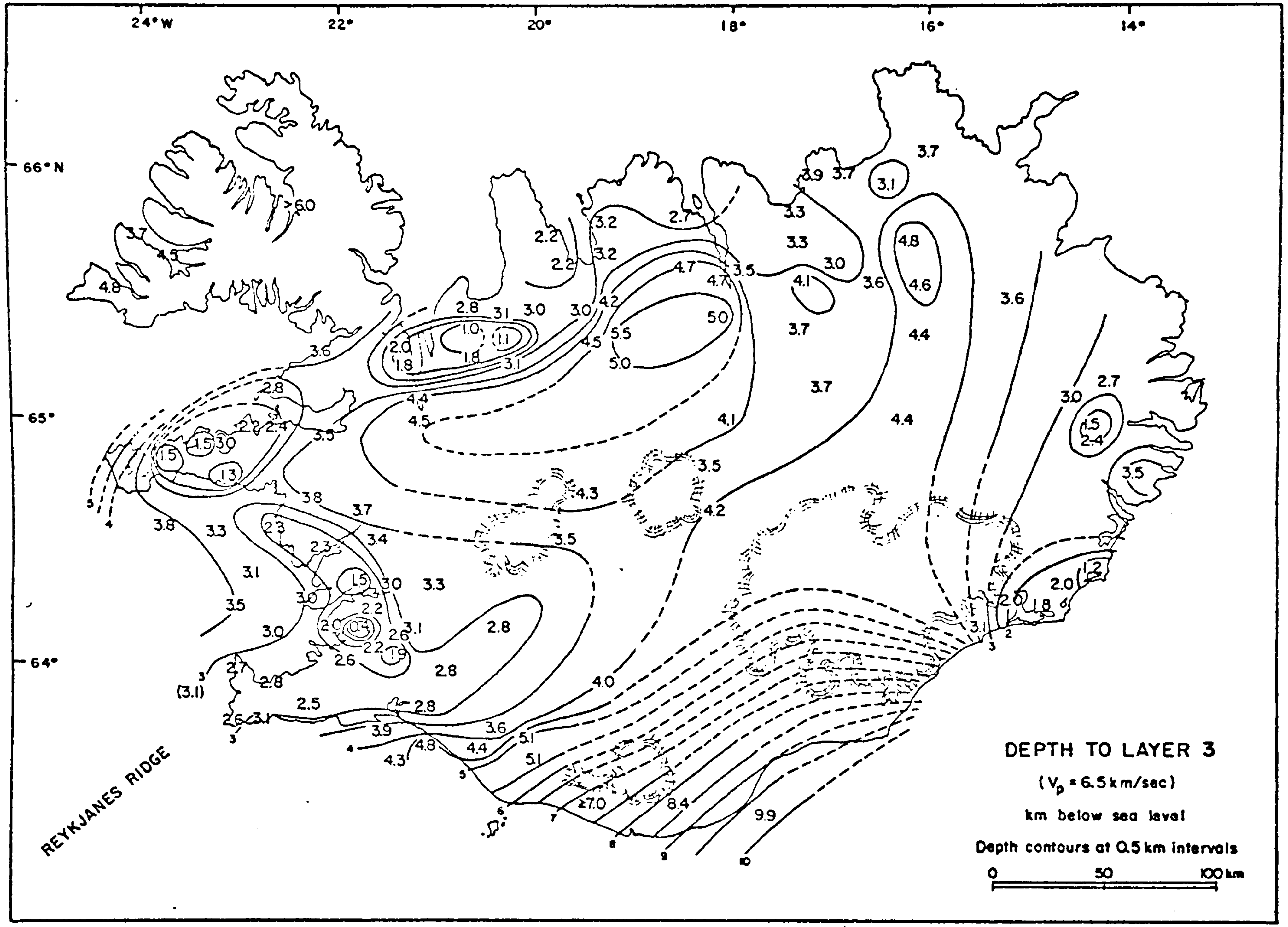


Fig. 35 (From Pálmason 1971)

correlation between extinct central volcanoes and shallow depths to layer 3, and furthermore he found a clear correlation between small scale features of the gravity field and the depth to layer 3. Strong positive gravity anomalies coincide with some of the central volcanoes where shallow depths to layer 3 have been recorded.

An unreversed 17km lond E-W profile across the Stardalur caldera, with a shot point just west of Leiðhamrar (E3 on the geological map) indicated layer 3 at a depth of 500-600m below the ground surface in the Stardalur area. "This estimate is relatively inaccurate due to the uncertainty in the shot point delay time, but it is nevertheless clear that this is the shallowest depth to layer 3 found so far in Iceland" (Pálmason 1971). The extent of the seismic delay time anomaly agrees fairly well with the extent (6-7km) of the Stardalur gravity anomaly discovered by Einarsson (1954), and coincides with the Stardalur caldera. The estimated upper surface of layer 3 is in good agreement with the clustering of the Stardalur cone sheets at a depth of 600-700m which is thought to indicate the upper margin of a more or less continuous layer of intrusions under the central and eastern part of the caldera (see page 100). It is concluded that the gravity anomaly and the seismic delay time anomaly are caused by the basaltic intrusives of the Stardalur central volcano.

An unreversed offshore profile westward from the Kjalarnes area into the bay of Faxaflói with a recording station just west of Leiðhamrar (E3), close to the shot point of the profile discussed above, indicated some irregularities in the structure of layers 1 and 2 along the first 10 to 15km of the profile, which can be correlated with the Kjalarnes central volcano. On the basis of travel times of the two profiles the depth to layer 3 under the Kjalarnes area is estimated as 1.8 to 2.2km depending on the relative thicknesses of layers 1 and 2 (Pálmason 1971). This depth increases to 2.7-3.2km beyond a distance of 15km west of Leiðhamrar. Although the seismic structure of the Kjalarnes centre is not as well known as that of the Stardalur centre, as the two profiles have end points within the former centre, it seems clear that the depth to layer 3 is significantly greater in the Kjalarnes centre than in Stardalur. This may seem to contradict the fact that the Bouguer gravity anomaly in the Kjalarnes centre is considerably stronger than the one in the Stardalur centre. The difference in depth to layer 3 may perhaps be explained by a thicker pile of pre-intrusion hyaloclastites under the Stardalur region than below the Kjalarnes region, which would accommodate larger volumes of coarse-grained intrusions in the former centre. Extending the Esja stratigraphy across the two centres supports this (four hyaloclastite horizons in Kjalarnes but six plus the intra-caldera hyaloclastites in Stardalur; the respective total thicknesses of the hyaloclastites in each centre can, however, not be estimated), and thick hyaloclastite units have been found down to a depth of 1100m in drillholes southwest of the Stardalur area (Tómasson, personal communication). Furthermore the lava pile through which the magma had to penetrate before spreading out laterally to form large intrusive bodies within the hyaloclastites is likely to have been considerably thicker under the Kjalarnes centre. It is, indeed, quite likely that the intrusions in the Kjalarnes centre are separated from the main layer 3 by a thick pile of subaerial lavas, whereas the Stardalur intrusions could be a direct continuation of layer 3 as, in fact, the seismic data suggests. The stronger Bouguer gravity anomaly of the Kjalarnes centre could perhaps be explained by the intrusives of that centre having a higher average density than the Stardalur intrusives. The characteristically iron and titanium rich eruptives of the Kjalarnes centre support this suggestion. A lava pile chemically equivalent to unit 3, for example, will contribute significantly to the average density of the crust, and intrusives of that composition even more so.

Another major seismic travel time anomaly is found associated with the Hvalfjörður central volcano where one of the four strong magnetic anomalies (page 155) was discovered by Sigurgeirsson (1970) and where Einarsson (1954) recorded a 5 mgals positive Bouguer anomaly. The estimated depth to layer 3 at this locality is 1.2-1.9km, depending on the relative thicknesses of layers 1 and 2 (Pálmason 1971). A reversed profile with end points south of the Hvalfjörður and east of the Stardalur seismic travel time anomalies indicates that the two anomalies are not connected.

A short account of other sites reported by Pálmason (1971) where depths to layer 3 are less than 2km are listed below:

Hraunsmúli, depth to layer 3 about 1.3km. This locality coincides with the Setberg Centre II central volcano where Sigurðsson (1970a) has described a cone sheet swarm, gabbroic ring features and a major granophyric cone sheet. A positive gravity anomaly is associated with this site (Pálmason 1971).

Kolgrafarfjörður, depth to layer 3 about 1.5km. This is the central volcano Setberg Centre I where Sigurðsson (1966, 1970a) described a caldera, a coarse-grained gabbro intrusion, and found an apical focus for thick acid and thin basic cone sheets at about 2.6km depth. A positive gravity anomaly is associated with this site (Einarsson 1954).

Fróðá, depth to layer 3 about 1.5km. This locality is about 20km west of Setberg Centre I. Kjartansson (1968) records a granophyre intrusion and rhyolite (indicating central volcanic activity) at this locality, but the geology has not been mapped in detail. A positive gravity anomaly is found at this site (Pálmason 1971).

Borgarvirki, depth to layer 3 about 1km. This locality is on the western margin of the Víðidalsfjall intrusive centre where Annells (1968) has described large acid and basic intrusions and two sets of basic cone sheets; the early set with an apical focus at about 5km, the late set at about 2km depth.

Hvammur, depth to layer 3 about 1.1km. This locality is on the eastern side of the Víðidalsfjall intrusive centre. There are several dolerite intrusions at this locality, and the profile lies just inside a subsidence trough which may have formed because of withdrawal of magma at depth (Annells 1968). A positive gravity anomaly of about 10 mgals is found at this site (Pálmason 1971).

Þingmúli, depth to layer 3 about 1.5km. This is the site of the Þingmúli central volcano where Carmichael (1964) described an intensive swarm of acid and basic dykes, acid intrusions and a set of acid cone sheets.

Austurhorn, depth to layer 3 about 1.2km. The profile is in the vicinity of the largest intrusions in Iceland. The shot point of the profile is by the Austurhorn acid-basic intrusion (Blake 1966), the profile lies southeast of the Lón central volcano and ends close to the Slaufrudalur acid intrusion (Beswick 1965). The depth to layer 3 increases to 2km towards Slaufrudalur. The depth of emplacement of the Austurhorn intrusion is estimated 1.7km (Blake 1966).

Hornafjörður, depth to layer 3 about 1.8km. This profile is also in the vicinity of the largest intrusions in Iceland. It extends from the Vesturhorn acid-basic intrusion (Roobol 1969, 1972) to north of the Ketillaugarfjall acid (and small basic) intrusion (Jónsson 1954, Annels 1967). The profile indicates a fairly constant depth to layer 3 in this area. The estimated depth of emplacement at the Vesturhorn intrusion (Roobol 1969) is 1.7km.

Hengill, depth to layer 3 about 1.9km. The shallowest depth to layer 3 is recorded in the northern part of the late Quaternary to recent Hengill central volcano (Sæmundsson 1967b), which is inside the Reykjanes-Langjökull volcanic zone.

All the sites where layer 3 has been recorded at depths less than 2km are associated with central volcanoes. All these sites, except for the relatively uneroded Hengill area, have large coarse-grained intrusives exposed on the surface, and many of the sites have cone sheet swarms indicating a local concentration of larger intrusive masses at shallow levels in the crust. The coincidence of central volcanoes having a great bulk of shallow level intrusions, with the sites of shallow depth to layer 3 strongly suggests an intrusive nature for layer 3. With the exception of Hengill, the Stardalur central volcano is the youngest of these sites. That the site of the shallowest depth to layer 3 recorded so far in Iceland is in Stardalur may be explained by the easier accommodation of large intrusions in soft hyaloclastites than in the hard and brittle lava piles of the Tertiary provinces in Iceland (Friðleifsson 1973a). At the end of the life span of the Stardalur central volcano the level at which layer 3 is now recorded may have been at a depth of about 1.5km. Hengill is still within the active volcanic zone, and the high temperature area there indicates intrusive activity. The thick piles of hyaloclastites underneath Hengill may in the future host voluminous intrusive bodies.

e. Nature of basaltic intrusives in Iceland

The distribution of basaltic dykes in the Icelandic crust is fairly well understood. Walker (1960) demonstrated the progressive increase of dyke density with depth in the lava pile in eastern Iceland, and showed (Walker 1963) that narrow dyke swarms are commonly associated with the central volcanoes. Lenticular lava units produced by these swarms were recognised by Gibson (1966). Gibson and Piper (1972) suggested that the increase in dyke intensity with depth was non-linear, and that at a relatively shallow depth (a few km), perhaps determined by lithostatic pressure conditions, the intensity of dyke injection increases rapidly to produce 100% dyke intensity. This can, however, not be verified by measurements as exposures are too shallow.

The distribution of larger intrusive bodies in Iceland and their relations with the host rocks have gained less attention. Intrusions more than 20m thick are relatively rare, and are mainly found associated with the central volcanoes. A survey of much of the published literature and several unpublished theses has led the author to the conclusion that, apart from the availability of magma and the ratio of total magma pressure/ lithostatic pressure, the lithology of the host rock is the most important factor in determining the size and shape of the basaltic intrusive bodies. The majority of large basic intrusions in Iceland are intruded into soft and "structureless" host rocks. Examples of most of the main intrusion localities which have been studied in Iceland are listed to support this conclusion;

the areal extension of the intrusions is in most cases measured by the author from maps, and the authors cited are therefore not responsible for these approximate values.

a) The host rock is: "hot" and still partly liquid acid intrusive material*.

Examples: Vesturhorn (granophyre and gabbro), about 20km^2, (Roobol 1969, 1972); Austurhorn (granophyre and gabbro), about 11km^2, (Blake 1966).

b) The host rock is: tuffaceous hyaloclastites, vent and caldera agglomerates, sedimentary rocks.

Examples: Viðborðsfjall multiple gabbro and dolerite intrusion, about 10km^2, Valagil gabbro, about 2km^2, Geitafell gabbro, about 1.5km^2, Hauksheiði gabbro and dolerite intrusions, - all these intrusions are in Hornafjörður (Annels 1967); Kolgrafarmúli gabbro, about 6km^2, in Setberg centre I (Sigurðsson 1966, 1970a), which is intruded at the contact of the Setberg caldera with the surrounding (older) basalt lavas; Stóra-Laxá multiple dolerite intrusion, about 10km^2, in Hreppar (Friðleifsson 1970); Kjalarnes multiple dolerite intrusion, about 2km^2, Leiðhamrar-Þverfell-Esjuberg dolerite intrusions, about 4km^2, Stardalshnúkur dolerite, 2.3km^2, in Esja.

c) The host rock is: hydrothermally propylitized lavas in the vicinity of the cores of central volcanoes.

Example: Hríshóll gabbro, about 1km^2, in the Króksfjörður central volcano (Hald et al 1971). Intensely altered lavas are, in fact,

* The acid intrusions are apparently mainly emplaced by stoping and/or updoming of the country rock.

crumbly and "structureless" like agglomerates and tuffaceous hyaloclastites.

Noteable apparent exceptions to this generalization are: Hrappsey-Purkey dolerite sill, which forms islands in about 10km^2 area in Breiðafjörður, and intrudes basalt lavas (Kristmannsdóttir 1971); Hólar-Skessusæti, about 1.2km^2, and Borgarvirki, about 0.7km^2, eucrite intrusions in the Víðidalur-Vatnsdalur area (Annells 1968), which intrude basalt lavas. The eucrite intrusions are both within the boundaries of the cone sheet swarms of the Víðidalsfjall intrusive centre; the Borgarvirki eucrite is just inside the margin of the early set of cone sheets, and the Hólar-Skessusæti eucrite just outside the margin of the (inner)late set of cone sheets. Annells (1968) considers both intrusions to have been emplaced at the start of the early cone sheet phase; Lýsuskarð gabbro, about 2.5km^2, and Þorgeirsfell gabbro, about 3km^2, in Setberg Centre II, which are stock like and have an arcuate form, suggestive of a ring fracture around Setberg Centre II (Sigurðsson 1970a). The intrusions, which apparently cut lavas, coincide with the densest part of the cone sheet swarm associated with the centre, and may have taken advantage of a structural weakness associated with circular fractures in the central volcano. The Lýsuskarð gabbro is, in fact, enveloped by a granophyric sheet, which forms a 200 to 400m wide screen between the gabbro and the basaltic country rock in the south, west and north; the eastern margins of the gabbro are not exposed. The contacts between the gabbro and granophyre are often gradational, but chilling of granophyre against gabbro found at one locality indicates that the granophyre is a later intrusion (Sigurðsson 1970a). Contacts of the granophyre with the older basalt lavas are well exposed, but contacts of the gabbro and the country rock are poor. Sigurðsson's (1970a) descriptions suggest to the present author that the Lýsuskarð gabbro-granophyre intrusion may possibly be yet another example of gabbro intruding granophyre, i.e. that the granophyre intrusion is older, but, in places, rheomorphic.

Minor intrusions are abundant in the immediate vicinity of volcanic centres (Walker 1964). Cone sheet swarms have been found in the majority of the central volcanoes investigated to date in Iceland. Outside the altered core regions of central volcanoes, minor intrusions, other than dykes and thin regular sheets, appear to be preferentially accommodated in the softer country rocks such as tuff horizons, rhyolite flows and olivine tholeiite compound lavas. Examples of this can be found in descriptions of e.g. the Reyðarfjörður area (Walker 1959) and the Fáskrúðsfjörður area (Gibson et al 1966).

The majority of the larger intrusions are in sheet-form (as opposed to laccolith form), which reflects the weight of the load, and have formed by multiple injection of magma, which may vary in composition. The grain size commonly decreases towards the edge of each sheet, but true chilled contacts between multiple sheets are rare, reflecting the shortness of the intervals between the individual injections. The degree of consolidation of the host rock controls the regularity of the shape of the intrusions; unconsolidated tuffs, for example, do not respond to regional stress patterns such as cone fracturing.

f. Nature of crustal layer 3

The cause of the higher seismic velocity in layer 3 in Iceland relative to the overlying rocks is not well understood. From a comparison of the depth to layer 3 with crustal temperatures, as inferred from borehole data, Pálmason (1971) suggested that the layer 2/layer 3 boundary might be metamorphic, layer 3 perhaps consisting of amphibolite facies metabasics, as was suggested by Cann (1968) for the oceanic layer. The highly altered cores of the volcanic centres (Walker 1964), indicating a rise in the isotherms in the vicinity of the centres, seemed compatible with the relatively shallow depths to layer 3 observed in the volcanic centres.

It is interesting in this context to consider the density variations likely to occur in tholeiitic lavas undergoing low grade metamorphism, and to compare this with the density difference between layers 2 and 3 estimated as 0.18 and 0.19 g/cm^3 in two localities in Iceland (Pálmason 1971). Table 7 (data from Deer, Howie and Zussman 1966) shows the specific gravity of the main primary and secondary minerals in the tholeiitic rocks. It is clear that the replacement of the primary minerals of a non-porous basalt (density say $3.00 g/cm^3$) by secondary minerals of the zeolite or greenschist facies, will in all cases reduce the density. In the case of a porous basalt, in filling of pore spaces in the rock by secondary minerals will not make it as dense as a non-porous basalt of the same composition. Epidote is found in the core regions of nearly all the central volcanoes and garnet is found in some, but

Table 7

Specific gravity* of primary and secondary minerals in thöleiitic rocks

	g/cm^3		g/cm^3
Quartz	2.65	Zeolites	2.05 - 2.40
Plagioclase	2.63 - 2.76	Analcite	2.24 - 2.29
Pyroxene	2.96 - 3.52	Calcite	2.72
Olivine	3.22 - 4.39	Chlorite	2.6 -- 3.3
Ilmenite	4.70 - 4.78	Serpentine	2.55
Magnetite	5.20	Prehnite	2.90 - 2.95
		Epidote	3.12 - 3.52
		Garnet	3.13 - 4.32
		Pyrite	4.95 - 5.03
		Hornblende	3.02 - 3.45

hornblende is rare and it is unlikely that a true amphibolite facies would be reached at the depths inferred from the seismic data. Christensen and Salisbury (1972) have shown how progressive alteration changes the density and hence seismic velocities of basaltic volcanics. Secondary alteration alone can hardly be a major factor in producing the density difference of about 0.2g/cm^3 between layers 2 and 3. The density difference can, however, be readily explained by a transition from a porous, but altered lava layer to a layer largely composed of non-porous intrusives. But how can the sharpness of the boundary between layers 2 and 3 be explained?

Gibson and Piper (1972) proposed that there was perhaps a preferred level in the crust to which the intruding magma tended

* From Deer, Howie and Zussman, 1966.

to rise, and this level might be governed by some form of hydrostatic equilibrium. But could not the alteration state of the crust also control this level?

The evidence of the larger intrusives being accommodated in the softer rock types (soft and "structureless" hyaloclastites as well as propylitized lavas) suggests that magma would tend to spread out laterally within the highly altered lavas at the base of layer 2 rather than penetrate the progressively harder lava pile. The "metamorphic boundary" would thus primarily function as a degree of alteration at which the lava pile looses its strength and accommodates large non-porous intrusives, and not as a density boundary in itself as proposed by Pálmason (1971).

This model would be compatible with the correspondence between observed depths to layer 3 and crustal temperatures as inferred from borehole data. The pressure field conditions would favour inclined sheets as the predominating intrusion form. The uplift evidenced by normal faults with downthrow towards the active volcanic zones as seen in most of Iceland may be caused by intrusions accommodated at depth at the layer 2/ layer 3 interface in the altered flanks of the volcanic zones.

Model calculations for crustal growth by dyke injection and surface lavas show that the lower crust in Iceland should consist almost entirely of intrusives (Pálmason 1973). The steady-state model used, however, requires the intrusive fraction to be elevated in the volcanic zone relative to the adjacent lithospheric plates, but no such elevation of layer 3 is indicated in the present volcanic zones in Iceland (Pálmason 1973). Considering the stress control of intrusive processes at extensional plate margins, Piper and Gibson (1972) found that if magma is generated across a narrow zone it will tend to be injected predominantly as vertical dykes. As the width of magma generation is increased a greater percentage of the intrusions into the crust will be deflected away from the spreading axis thus forming inclined sheets. This implies a much wider zone of intrusive activity than extrusive volcanic activity. This may provide an explanation of the discrepancy between Pálmason's model calculations (in which it is assumed that all dykes reach the surface in the active volcanic zones) and the observed depth to layer 3 under the volcanic zones.

It is of interest to note that in Pálmason's (1971) map (Fig.35) of the depth to layer 3 in Iceland, a large area in southwest Iceland has depths of less than 3km. This area coincides fairly closely with the outcrops of Quaternary to recent volcanics in southwest Iceland and can be explained by large intrusives being accommodated in the subglacial hyaloclastite units. The depth to layer 3 in the Reykjanes peninsula is about 2.6km. A direct comparison cannot be made between the peninsula and the Reykjanes ridge as velocity values typical of layer 3 have not been detected in the axial zone of the ridge (Talwani et al 1971); in the flank zones of the ridge, however, layer 3 is at about 2km depth. From the evidence on the nature of basaltic intrusives in Iceland and assuming a constant lithostatic/magma pressure ratio the depth to layer 3 in the crust could be expected to increase southwards along the Reykjanes ridge as with increasing water depth the tuff/pillow lava ratio decreases and a higher degree of alteration would be needed to make the pillow lavas "hospitable" to large intrusions. Data is, however, too scanty to test this. Layer 2 is recorded all over Iceland except on the Reykjanes peninsula, where it appears to be missing (or is too thin to be detected (Pálmason 1971)). This may be explained if the bulk of the volcanics are hyaloclastites, as has been indicated by drilling (Tómasson and Kristmannsdóttir 1972); alteration of hyaloclastites would not produce densities and seismic velocities corresponding to an altered lava layer. Intrusives will be easily accommodated in the hyaloclastite piles. The high level of the intrusions is probably reflected in the four high temperature areas in the peninsula, which are the only present high temperature areas in Iceland not connected with central volcanism (Pálmason and Sæmundsson 1973).

An explanation of the conspicuous, progressive increase in the depth to layer 3 in central southern Iceland may perhaps be found in that at the time of the trough formation through eastern Iceland preceding the shift of the volcanic zone in northern Iceland (Sæmundsson 1973), layer 3 had already formed, partly by intrusions into the highly altered base of the lava pile. As this pile had drifted away from the then active western volcanic zone and was now in an area of a relatively low thermal gradient the process of alteration of the lavas at the base of layer 2 did not match the rate of subsidence and hence the palaeoboundary of layer 2/ layer 3 sank as the thickness of the overlying layers increased, layer 1 increasing faster in thickness than layer 2. The scanty available seismic data on the relative thicknesses of layers 1 and 2 appears to support this model.

When volcanism commenced in the trough the magma either penetrated the whole way to the surface and was erupted, or was accommodated well below the layer 2/ layer 3 palaeoboundary, as the "old" intrusives had already filled the "readily available space" in the altered lava layer. It is interesting to note that in this area of exceptional depth to layer 3, some of the largest volumes of postglacial lavas in Iceland have been erupted, such as the Lakagígar eruption with a volume of about 12.5km^3 (Thorarinsson 1969) and the Eldgjá eruption(s) producing between 10 and 20km^3. The large volume produced by the Eldgjá fissure is particularly surprising in view of the relatively small quantities of transitional alkali basalts produced during postglacial times (Jakobsson 1972), and lends support to the suggestion that the bulk of the lava produced by partial melting in one batch below the crust was brought to the surface, and that a substantial percentage of the total volume was not intruded at the base of layer 2, as may be the case in areas with a "normal" layer 2/ layer 3 boundary. The remarkable chemical uniformity of the Lakagígar lava (with plagioclase, olivine and clinopyroxene phenocrysts) throughout the 1783 eruption (Grönvold 1972, Bell 1973) suggests an undelayed passage of the magma from the zone of partial melting to the surface.

The northern part of the eastern volcanic zone is characterised by several elongated fissure swarm units, each unit accompanied by normal faults, and both eruptive and open fissures passing through a central volcano (Sæmundsson 1973). This situation, according to the proposed hypothesis, is believed to be typical of areas overlying a "normal" layer 2/ layer 3 boundary, where large intrusives are emplaced in sheet-form at the base of layer 2 and fissures at the surface, representing crustal extension, are not always filled. In the south, however, such open fissures are rare, supporting the suggestion that here large intrusives are not emplaced at the base of layer 2 (now at greater depth), the magma either being confined to the base of layer 3, or brought to the surface filling all available tensional fissures. Crustal drift is probably slower in southern than in northern Iceland (drift is concurrently occurring in the Reykjanes-Langjökull volcanic zone), but the absence of open fissures may also indicate a linear increase in dyke dilation with depth (every dyke reaches the surface) and an equilibrium between the drift rate and the volume of intrusives emplaced in the crust.

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Appendix I   SAMPLE LOCALITIES AND SHORT DESCRIPTIONS

1. Tholeiite lava by sheep fold south of Ártún (C6, 50m, unit 3).* Fine-grained tholeiite with accessory amounts of plagioclase phenocrysts and rare, small phenocrysts of pyroxene.

2. Tholeiite lava in Lokufjall (D7, 220m, unit 3). Virtually aphyric tholeiite with rare plagioclase and very rare pyroxene phenocrysts.

3. Tholeiite sill, about 2m thick, east of Ártún (C7, 50m). Virtually aphyric, fine-grained tholeiite, but with very rare microphenocrysts of plagioclase, pyroxene and olivine.

4. Tholeiite sill, just east of main road, north of Ártún (C7, 50m). Very fine-grained aphyric tholeiite.

5. Dolerite dyke 4m thick in Lokufjall (C8, 100m). Plagioclase-pyroxene-phyric tholeiite dolerite.

6. Dolerite sill (Fig.14) in Melafjall (D8, 220m). Plagioclase-pyroxene-phyric tholeiite dolerite.

7. Tholeiite sill, 30m thick, east of Ártún (C7, 200m). Plagioclase-phyric tholeiite with about 9% (modally) of euhedral plagioclase phenocrysts. Accessory amounts of pyroxene microphenocrysts.

8. Plagioclase-phyric tholeiite lava, about 20m thick, south of Tindstaðir (E8, 220m, unit 5). Euhedral plagioclase phenocrysts (10%) and microphenocrysts of pyroxene (4%) are set in a fairly coarse-grained tholeiite groundmass.

9. Tholeiite pillow breccia on Kjalarnes peninsula (B5, 20m, unit 9 or 10). Very fine-grained aphyric tholeiite.

10. Tholeiite pillow breccia in fault zone north of Brautarholt (B6, 15m, unit 9 or 10). Very fine-grained aphyric tholeiite.

* The samples are listed in stratigraphical-chronological order, and the geological map coordinates and altitude of sample localities are given as well as the unit number in the Esja stratigraphic column (Table 1).

11. Prominent, coarse-grained tholeiite dolerite sheet by the sea on the northern side of Borgarvík (B5, 2m).

12. Icelandite dyke, 3m thick, in Lokufjall by Tíðaskarð (C8, 150m). Rare phenocrysts of plagioclase and very rare phenocrysts of olivine and pyroxene in a fine-grained matrix.

13. Tholeiite dyke, 10m thick, east of sheep fold south of Ártún (C6, 50m). Small phenocrysts of plagioclase (9%) and pyroxene (5%) occur both in glomeroporphyritic clusters and separately.

14. Tholeiite dyke, 2m thick, east of Ártún (C7, 260m). Fine-grained and virtually aphyric tholeiite with very rare plagioclase and pyroxene microphenocrysts.

15. Plagioclase-phyric tholeiite plug in hyaloclastite on western side of Kerlingagil (E8, 450m, unit 10). Strongly zoned plagioclase phenocrysts form about 15% (modally) of the rock, but microphenocrysts of pyroxene are less than 1%.

16. Plagioclase-pyroxene-phyric tholeiite plug in hyaloclastite in western Hrútadalur (F8, 210m, unit 10). Plagioclase phenocrysts constitute 15%, pyroxene phenocrysts 10%, but microphenocrysts of olivine make up less than 1% of the rock.

17. Tholeiite hyaloclastite, solid mass in plug area in Dýjadalshnúkur (E8, 530m, unit 10). Fine-grained tholeiite with accessory amounts of microphenocrysts of plagioclase (3%), pyroxene (2%) and very rare microphenocrysts of olivine.

18. Tholeiite plug in hyaloclastite in southwest Tindstaðahnúkur (E7, 550m, unit 12). Fine-grained tholeiite with accessory amounts of microphenocrysts of plagioclase (7%), pyroxene (2%) and rare microphenocrysts (mostly acicular) of olivine.

19. Tholeiite plug in hyaloclastite, on the ridge south of Blikdalur (D6, 560m, unit 12). Fine-grained tholeiite with accessory amounts of microphenocrysts of plagioclase and pyroxene and rare microphenocrysts of olivine.

20. Tholeiite pillow lava in river east of Flekkudalur farm (I8, 80m, unit 12). Rare phenocrysts of plagioclase, but

accessory amounts of microphenocrysts of plagioclase (8%) and pyroxene (6%); the latter are commonly strained.

21. Plagioclase-phyric tholeiite lava, 35m thick, forming a prominent ledge in the northern side of Sneiðar (H9, 180m, unit 13). Large (up to 4mm long) euhedral phenocrysts of plagioclase make up about 15% of the rock; smaller pyroxene phenocrysts (7%) are commonly strained.

22. Olivine tholeiite lava, 35m thick, succeeding lava 21 in Sneiðar (H9, 220m, unit 13). The olivine is mostly in phenocrysts, and there are accessory amounts of plagioclase phenocrysts.

23. Olivine tholeiite lava forming ledge of waterfall in Flekku-dalur (I8, 150m, unit 13). This is probably the same lava as sample 22.

24. Small tholeiite plug north of Kerhólakambur (E6, 650m, unit 13). Fine-grained tholeiite, rare euhedral phenocrysts of plagioclase.

25. Plagioclase-phyric tholeiite plug west of Kerhólakambur (D5, 660m, unit 13). Euhedral phenocrysts of plagioclase make up 10% of the rock, but microphenocrysts of pyroxene are rare.

26. Plagioclase-phyric tholeiite lava at the top of cliffs southwest of Kerhólakambur (E5, 700m, unit 13). Plagioclase phenocrysts and microphenocrysts make up about 20% of the rock.

27. Very coarse-grained plagioclase-phyric tholeiite dolerite sill in Leiðhamrar (E3, 20m). Plagioclase phenocrysts are up to 5mm in diameter, and pyroxene phenocrysts are up to 2mm in diameter. The groundmass is coarse-grained and consists of plagioclase and pyroxene. Ore is mainly in the form of skeletal ilmenite. The rock is fairly altered.

28. Coarse-grained tholeiite dolerite sheet in the main gulley northeast of Esjuberg farm (E4, 100m). The texture of the dolerite is ophitic with rare large phenocrysts of plagioclase.

29. Olivine tholeiite dolerite; from the top of a sheet east of Kvensöðlar (D5, about 250m). The modal composition is: plagioclase 46%, pyroxene 40%, olivine 7%, ore 7%.

30. Olivine tholeiite dolerite; from the same sheet as sample 29, but some 10m below top of sheet. The modal composition is: plagioclase 45%, pyroxene 38%, olivine 12%, ore 5%.

31. Coarse-grained olivine tholeiite dolerite sheet in Þverfell, north of gravel quarry (E4, about 100m).

32. Chilled, glassy margin of a thin tholeiite dyke in Gljúfurdalur (E5, 570m). Rare microphenocrysts of plagioclase and minute ore specks in the devitrified glass.

33. Olivine tholeiite lava, 34m thick, northeastern Þórnýjartindur (G8, about 500m, unit 16). Fine-grained lava with accessory amounts of plagioclase phenocrysts but rare phenocrysts of pyroxene. Some of the olivine is in the form of tabular phenocrysts, but most of it is in small microphenocrysts.

34. Olivine tholeiite lava which "ends" in the middle of the slope of western Skálatindur (H7, about 400m, unit 16). Fine-grained lava with most of the olivine as microphenocrysts.

35. Olivine tholeiite lava in Þverárkotsháls (I4, 180m, unit 16). Accessory amounts of plagioclase phenocrysts.

36. Olivine tholeiite lava, at the base of flow units on the western side of Eilífsdalur (G7, about 550m, unit 19).

37. Olivine tholeiite lava in northern Þórnýjartindur (G7, about 600m, unit 19).

38. Olivine tholeiite lava in Nónbunga (H8, about 520m, unit 19).

39. Tholeiite lava from north of Helgafell (G2, 60m) which is south of the area mapped. This very fine-grained lava is tentatively assigned to unit 19. Rare phenocrysts of plagioclase and pyroxene.

40. Tholeiite lava in a small valley in unit 18 hyaloclastite in Þverárdalur (I5, 480m, unit 19). Very fine-grained and virtually aphyric with very rare plagioclase and pyroxene microphenocrysts. Similar to the basaltic andesites.

41. Tholeiite lava at 126m depth in drillhole H-1 in Stardalur (L4, unit 19). Fine-grained tholeiite with rare plagioclase

and pyroxene phenocrysts. The sample is fairly altered with zeolites and quartz in cracks.

42. Tholeiite lava at 67m depth in drillhole H-1 in Stardalur (L4, unit 19). Fine-grained tholeiite. The sample is very altered with zeolites and quartz in cracks and vesicles.

43. Icelandite lava at the head of the Stardalur valley (L4, 270m, unit 19). Very fine-grained lava with rare phenocrysts of plagioclase and olivine in a glassy matrix.

44. Basaltic andesite hyaloclastite; lithic mass in hyaloclastite on the north bank of Leirvogsá river 200m west of Tröllafoss (J4, 140m, unit 21). Very fine-grained with rare plagioclase phenocrysts.

45. Basaltic andesite hyaloclastite; lithic mass in hyaloclastite about 900m east of Tröllafoss (K4, 160m, unit 21). Fine-grained with rare phenocrysts of plagioclase.

46. Basaltic andesite plug in Múli (L4, 330m, unit 21). Very fine-grained with rare phenocrysts of plagioclase. The bimodal magnetite grain size is very clear in this sample. The sample is slightly altered.

47. Basaltic andesite plug west of road to Svínaskarð (K5, 400m, unit 21). Very fine-grained with rare phenocrysts of plagioclase. The sample is slightly altered.

48. Basaltic andesite plug, Bláhnúkur (J5, 580m, unit 21). Very fine-grained with rare phenocrysts of plagioclase and pyroxene. The bimodal magnetite grain size is very clear in this sample.

49. Basaltic andesite plug, Bláhnúkur (J5, 580m, unit 21). Very fine-grained with rare phenocrysts of plagioclase and pyroxene.

50. Plagioclase-phyric tholeiite dolerite dyke in Múli (L4, 250m). Euhedral plagioclase phenocrysts make up about 15% of the rock. The sample is fairly altered.

51. Tholeiite dolerite sheet in burn 900m southeast of Skeggjastaðir (J3, 160m). The sample is very altered.

52. Tholeiite dolerite, from a "solid" part of a coarse-grained dolerite sheet in the Leirvogsá gorge southwest of Tröllafoss (J4, 140m). The sample is fairly altered.

53. Tholeiite dolerite, from a more "sugary" part of the same sheet as rock 52. The texture of the coarse-grained dolerite is ophitic. The sample is fairly altered.

54. Tholeiite dolerite sill, from the fan-shaped columns north of the dam in Leirvogsá southwest of Tröllafoss (J4, 130m). Coarse-grained dolerite with ophitic texture. The sample is fairly altered.

55. Tholeiite dolerite sheet west of track crossing Skarðsá river north of Haukafjöll (J4, 140m). Very coarse-grained "sugary" dolerite. The sample is fairly altered.

56. Tholeiite dyke a little further down river from sample 55. The tholeiite contains accessory amounts of plagioclase and pyroxene phenocrysts. The sample is fairly fresh, and the dyke could be considerably younger than the Stardalur sheets, perhaps from the Gilsá normal polarity event.

57. Tholeiite dolerite sill of Gráhnúkur (J5, 280m). Coarse-grained dolerite. The sample is fairly altered.

58. Olivine tholeiite dolerite; from a column in the southern slope of Stardalshnúkur laccolith (K4, 340m). Very coarse-grained dolerite with olivine crystals up to 2mm in diameter and poikilitic augite crystals up to 5mm in diameter. The olivine phenocrysts are fresh, but glassy pockets in the matrix are altered and zeolites fill cracks in the rock.

59. Tholeiite pillow breccia west of Eyjadalur (J8, 480m, base of unit 22). The sample is a devitrified glass, with microphenocrysts of plagioclase and minute ore specks. The sample is fairly altered with zeolites in vesicles.

60. Tholeiite lava in northern Þórnýjartindur (G7, 670m, unit 22). Fine-grained, aphyric tholeiite.

61. Rhyolite from composite sheet in northeast Múli in Svínadalur (L7, 260m, unit 23). Glassy rhyolite, altered and devitrified; zeolites, calcite and quartz in vesicles.

62. Rhyolite from the same sheet as rock 61, but at a lower level further north. This sample is much less altered and is of virtually aphyric rhyolite. Vesicles are partly filled with quartz.

63. Icelandite, from the same composite sheet as rocks 61 and 62 in Múli, sample collected just above the track at the northern end of the sheet. Rare plagioclase and very rare pyroxene phenocrysts. The sample is altered with interstitial devitrified glass and zeolites in vesicles.

64. Basaltic andesite, lithic mass in hyaloclastite in gorge climbing west, south of Trana (K6, 400m, unit 23). Fine-grained rock, but with accessory amounts of plagioclase (5%) and pyroxene (2%) phenocrysts.

65. Basaltic andesite, lithic mass in gulley running west at the mouth of Trönudalur (K7, about 400m, unit 23). Fine-grained with rare phenocrysts of plagioclase.

66. Basaltic andesite plug in western Trönudalur (K7, 450m, unit 23). Fine-grained with rare phenocrysts of plagioclase.

67. Basaltic andesite pillow lava in western Eyjadalur (J7, 540m, unit 23). Fine-grained with rare phenocrysts of plagioclase.

68. Tholeiite lava in the eastern slope of Trana (K6, about 500m, unit 24). Euhedral plagioclase phenocrysts make up 8% of the rock.

69. Plagioclase-phyric tholeiite in southern Kistufell (H4, 650m, unit 24). Plagioclase phenocrysts make up 12% of the rock. Rare pyroxene phenocrysts.

70. Tholeiite lava, 16m thick, in cliff in northeastern Gunnlaugsskarð (G6, about 700m, unit 24). Accessory amounts of olivine microphenocrysts; on borderline of tholeiite and olivine tholeiite. Rare phenocrysts of plagioclase.

71. Tholeiite lava on the top of Esja, below cairn north of Kistufell (G5, 900m, unit 24). Very fine-grained lava with

accessory amounts of very small microphenocrysts of olivine; on borderline of tholeiite and olivine tholeiite. Rare phenocrysts of plagioclase.

72. Olivine tholeiite plug in southwest Skálafell (K5, 490m, unit 25). Very fine-grained rock with rare phenocrysts of plagioclase and olivine.

73. Olivine tholeiite plug in eastern Dalhólar, Grímmannsfell (J2, 180m, unit 25). Very fine-grained rock with rare phenocrysts of plagioclase.

74. Olivine tholeiite columnar intrusion west of Helgufoss, Grímmannsfell (J2, 150m, unit 25). Very fine-grained rock with accessory amounts of plagioclase phenocrysts and rare olivine phenocrysts.

75. Rhyolite from a composite dyke in a subsidiary gulley of Leirvogsá, west of Tröllafoss (J4, 130m). Aphyric rhyolite, fairly altered.

76. Rhyolite intrusion in Grafardalur (I4, 220m). Felsitic rhyolite with rare microphenocrysts of pyroxene and olivine.

77. Rhyolite, black pitchstone column in the southern slope of the easternmost peak of Móskarðshnúkar (K5, about 750m, unit 26). Accessory amounts of plagioclase phenocrysts in glassy pitchstone. The sample is very altered.

78. Rhyolite, from columnar-jointed grey pitchstone mass in Grímmannsfell (I1, 200m, unit 26). Aphyric glassy pitchstone. The sample is fairly altered.

79. Olivine tholeiite pillow lava in gulley in southern Mosfell (H2, 110m, Post-erosional).

80. Olivine tholeiite lava by bridge at Leirvogsvatn (L3, 200m, Post-erosional).

Appendix II ANALYTICAL METHODS

The rock samples selected for chemical analysis were cut to approximately 1cm^3 pieces in a hydraulic rock-splitter and selected fresh pieces crushed in a jaw-crusher and subsequently in a swing mill using an agate container for 58 samples but a tungstencarbide container for 22 samples. A test was carried out to check on possible oxidation during crushing on one sample (55) using the procedure described by Fitton and Gill (1970), but no significant oxidation was found to have taken place. This is in agreement with the results of Grönvold (1972) who applied the same test to less oxidised rocks crushed in the same manner.

The finely-ground rock powder was pressed at 5-10 tons/in^2 in a hydraulic ram into pellets with boric acid powder forming the back and rims of each pellet. Two pellets were made from each powdered sample.

The chemical analyses were carried out using Philips 1540 and Philips 1410 spectrometers with a 2kw generator. Al, Si, P and K were analysed using a Cr-tube and PET-crystal. Ca, Ti, Mn and Fe were analysed using a W-tube and LiF 200-crystal. Mg was analysed using a Cr-tube and RAP-crystal. Na was analysed using a Cr-tube and gypsum-crystal on the Philips 1540, and a Cr-tube and RAP-crystal on the Philips 1410.

In order to ensure reproducability and to monitor electronic drift affecting the count rate, "internal standard" pellets were analysed at frequent intervals along with the unknown samples, the frequency depending on the stability of the count rate, but never at longer intervals than once between every ten samples. Ratios were subsequently calculated for unknown sample/internal standard and these ratios were used to obtain analyses from calibration curves made from standards (analysed by "wet-chemical" methods) which were treated in the same way as the unknown samples. For elements where good reproducability could not be obtained between the Philips 1540 and 1410 spectrometers, all the samples were analysed on the newly installed 1410 spectrometer. Each sample was counted four times, twice on each of the duplicate pellets.

A combination of 7 U.S.G.S. standards and 5 "wet-chemically" analysed samples from the Esja collection were used as standards. Two sets of standards were used; one for rocks with SiO_2 wt% higher than 57% and another for those of lower silica content. A test made to see if a combination of both sets would give better values for 55% to 65% SiO_2 showed such procedure to be unneccessary. For the acid rocks, the U.S.G.S. standards G-2, AGV-1 and BCR-1 and Esja standards 62, 63 and 17 were used. For the basaltic rocks, the U.S.G.S. standards PCC-1, W-1, BCR-1 and AGV-1 and the Esja standards 74, 17 and 49 were used. The recommended values of Fleischer (1969) and Flanagan (1969) were used for the U.S.G.S. standards.

The five Esja standards were analysed by "wet chemical" methods (Table 8). The analyses were carried out with a flame photometer (Na_2O, K_2O), colorimetrically (Al_2O_3, total iron,

TiO_2, P_2O_5), and by atomic absorption (MgO, CaO and MnO) methods following the Oxford geochemical laboratory manual, and with much assistance from Mr. N. Killingback. Due to the author's inexperience the classical gravimetric method of analysing SiO_2 gave too low values, and the SiO_2 values of the Esja standards reported here were obtained by XRF methods using a variety of high quality standards.

A computer program kindly provided by Dr. J.G. Holland of Durham University was used to make least square fit calibration curves and inter-element mass absorption corrections. The method has been described by Holland and Brindle (1966). One drawback of the method is that it recalculates the analyses to 100% on a water-free basis with total iron as Fe_2O_3. If the rock has a water-free total of less than e.g. 98% and the conversion of

Table 8

Chemical analyses of rock standards from Esja

Sample	74	17	49	63	62
SiO_2*	48.13	47.87	50.83	58.68	71.66
Al_2O_3	15.36	13.79	13.38	13.19	13.00
FeO	8.30	11.37	13.14	6.70	1.34
Fe_2O_3	2.50	3.11	2.66	3.82	3.27
MgO	8.61	5.71	3.71	2.60	0.63
CaO	12.38	11.32	7.97	6.56	1.80
Na_2O	2.04	2.59	3.10	3.07	4.21
K_2O	0.15	0.27	0.68	0.99	2.42
TiO_2	1.52	2.72	3.54	2.08	0.51
MnO	0.19	0.25	0.28	0.22	0.11
P_2O_5	0.12	0.34	0.50	0.65	0.10
Total	99.30	99.34	99.79	98.56	99.05

* SiO_2 analysed by XRF

FeO into Fe_2O_3 does not make the water-free total reach 100% then an increment in all the analysed element percentages results. In fresh rocks with a water-free total (with total iron as Fe_2O_3) of more than 100%, all the element percentages would be brought down. To recalculate the analyses again with an independently determined FeO% value would not correct this latter discrepancy as the true FeO% value would have to be subtracted from a lowered total iron (as Fe_2O_3) value, and hence the recalculated analyses would give a false FeO/Fe_2O_3 ratio. In view of this it was decided to report the analyses with total iron as Fe_2O_3, the FeO% values determined independently using the metavanadate method being listed in a separate column in the table of analyses (Table 2).

In order to make a direct comparison possible between the chemical data presented in this work and the data produced by Grönvold (1972), four Kerlingarfjöll samples analysed by "wet chemical" methods and used as XRF-standards by Grönvold were analysed again by XRF-methods along with the Esja rocks. The "wet chemical" analyses and the two sets of XRF analyses of these samples are shown in Table 9. Grönvold used eight "wet chemically" analysed standards to make XRF calibration curves by least square fitting and did not make mass absorption corrections.

Table 9

Comparison of Kerlingarfjöll (Grönvold 1972) and Esja analyses.

Sample		SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5
K 77	Kerl. wet	47.9	15.2	12.2	8.4	13.4	2.1	0.09	1.0	0.19	0.09
"	" XRF	49.3	15.2	11.9	8.7	13.2	2.2	0.07	1.1	0.19	0.08
"	Esja XRF	49.1	16.8	11.2	7.1	12.2	2.4	0.06	0.9	0.18	0.08
K 245	Kerl. wet	48.7	13.7	15.0	5.4	9.7	2.9	0.82	2.4	0.24	0.33
"	" XRF	48.3	13.6	15.1	5.6	9.6	2.9	0.84	2.6	0.23	0.40
"	Esja XRF	49.4	14.5	14.2	5.2	9.8	3.2	0.73	2.4	0.24	0.37
K 44	Kerl. wet	54.3	13.5	11.6	3.1	6.7	3.7	1.3	2.5	0.26	0.94
"	" XRF	52.7	13.3	11.7	3.3	6.5	3.9	1.3	2.5	0.25	0.95
"	Esja XRF	54.4	13.6	13.2	3.1	7.3	3.6	1.3	2.7	0.27	0.77
K 169	Kerl. wet	68.4	13.1	6.6	0.09	2.5	4.9	2.7	0.45	0.20	0.07
"	" XRF	69.5	13.5	6.2	0.18	2.4	5.0	2.7	0.43	0.23	0.06
"	Esja XRF	69.4	12.7	6.4	0.00	2.4	5.8	2.6	0.47	0.20	0.08

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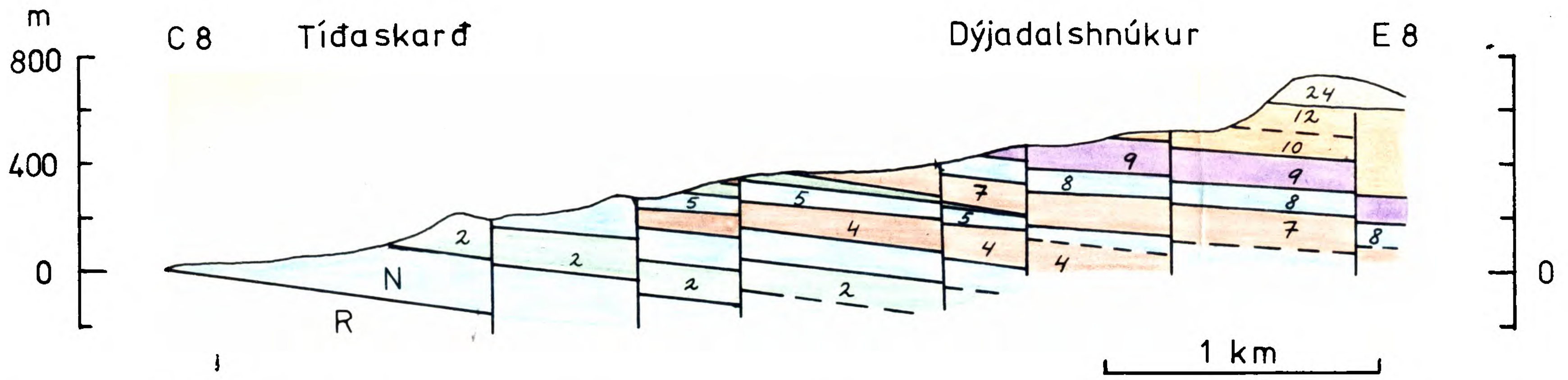
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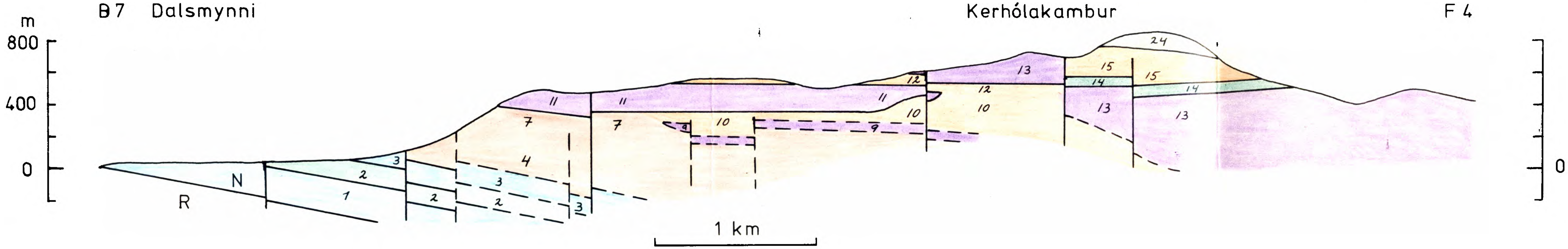
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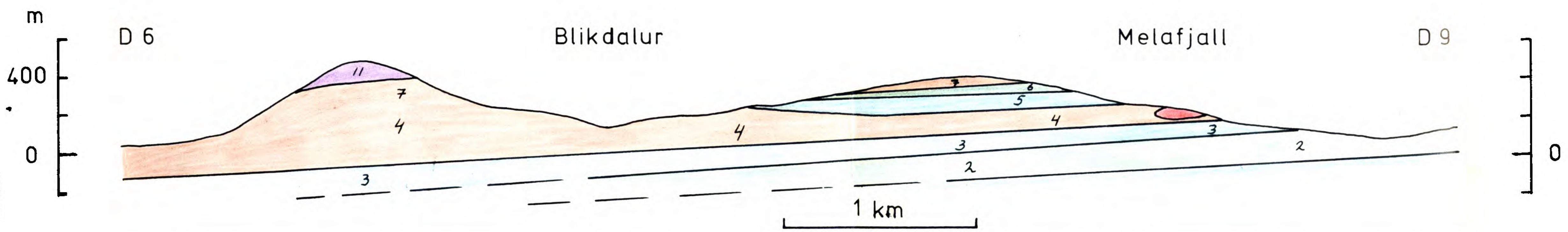
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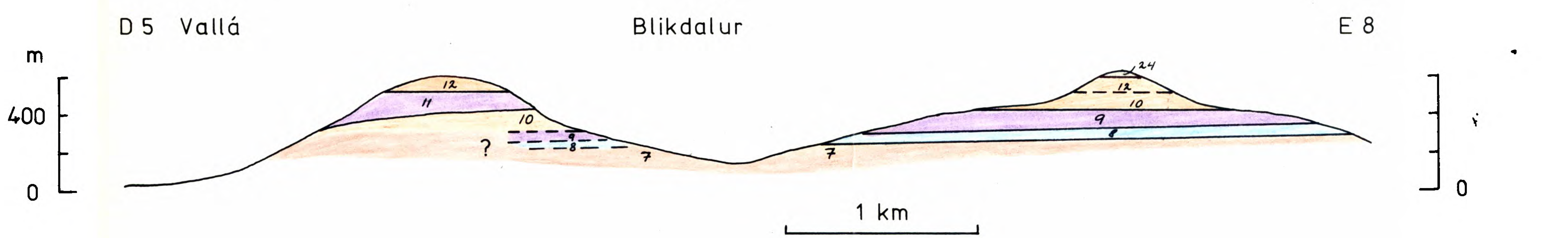
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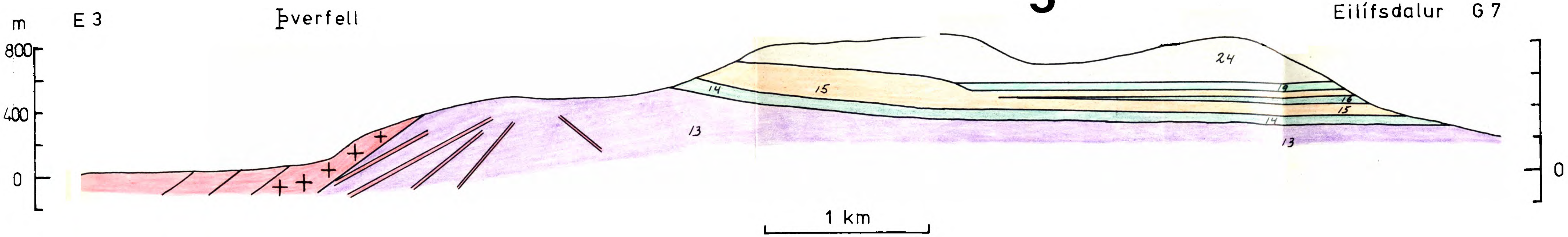
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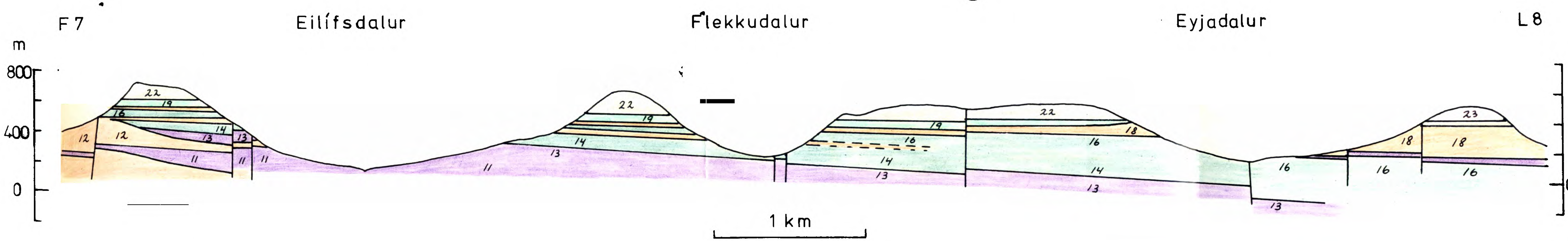
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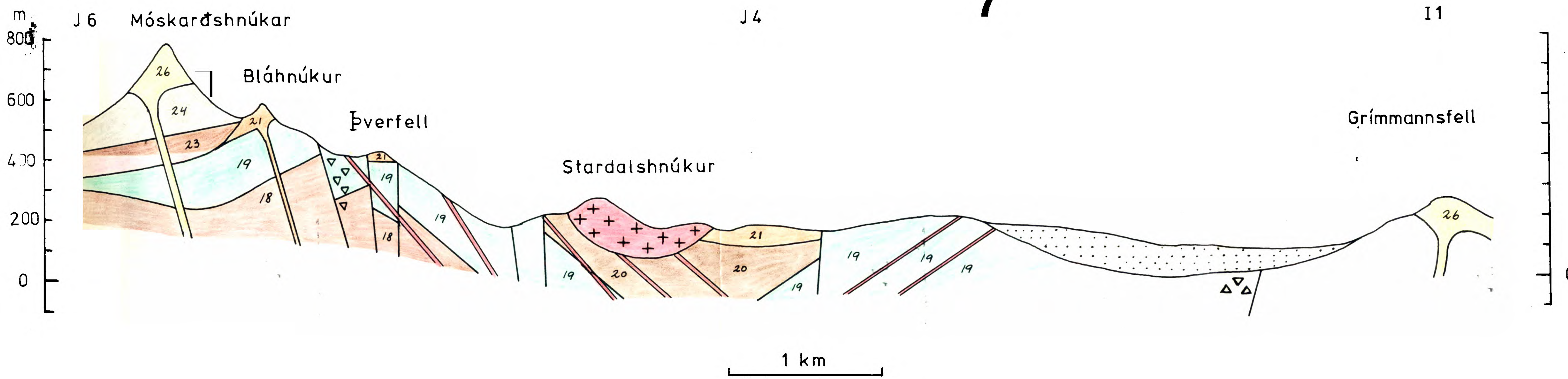
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