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Fil.lic. avhandling i geologi mineralogi
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ABSTRACT

The gabbroic nodules from Miðfell, Sw Iceland consist of 53% plagioclase (An 88 mol%), 11% clinopyroxene (Fs 5, En 50, Wo 45 mol%), 7% olivine (Fo 87 mol%) and 29% interstitial ground mass. The nodule bearing host rock is a picritic pillow basalt with olivine, plagioclase, clinopyroxene and spinel as phenocryst phases. The glassy pillow rims have a MgO content of 9.7 w% and FeOt content of 9.2 w%. The composition of the glass is the most primitive found in Iceland so far. The minerals and the confining glass are apparently close to equilibrium. The liquidus temperatures calculated on the basis of the equilibrium compositions are about 1250 oC. There is no essential difference in the chemical composition of the nodule minerals and the host rock phenocrysts. The extremely narrow compositional variation in the minerals, lack of zoning and the extensively overlapping ranges are explained by crystallization in open magma systems from different batches of primitive magma from a common source with similar composition. The nodules could have originated from olivine gabbro dikes formed at pressures where crystallization proceeds along olivine and spinel, spinel and clinopyroxene, clinopyroxene and plagioclase equilibrium stages. Replenishment with new batches of primitive magma from the depth occasionally reversed the crystallization order in the nodules. The host used same channels when ascending to a slightly higher niveau where it reached olivine-plagioclase equilibrium stage.

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1. INTRODUCTION

The present study is on the mineralogy and chemistry of mantle derived gabbroic nodules and its host rock from Miðfell, SW Iceland. The nodules are found in a very primitive picritic basalt and they may therefore provide constraints on the evolution of the magmas within the oceanic upper mantle that more evolved nodules do not.

The hypothesis of sea-floor spreading and plate tectonics assumes the creation of new oceanic crust at mid-ocean ridges. In recent years attention has focused on to the nature of the magmas produced at the ridges and their ultimate origin. The location of Iceland across Mid-Atlantic Ridge axis favors the study of the processes taking place at crests of oceanic ridges. When Bowen (1928) suggested that basalts represent primary, unmodified magmas derived from the mantle, there were few arguments against this. With the theory of ocean-floor spreading, in the sixties, the question has been seriously discussed and reconsidered. The discussion is now very much focused on the characteristics of the mid-ocean ridge basalts (MORB) and to what extent they reflect chemical properties of the underlying mantle. The fundamental question is the nature of the mantle derived primary magmas and the processes involved in their formation and subsequent evolution.

Models on basalt petrogenesis at the ocean ridges can be divided into two opposing categories: The proponents of the one view point argue that the composition of the most primitive MORB:s is close to the primary magmas. These magmas are generated by partial melting in the mantle beneath mid-ocean ridges at the pressures from 8-10 kb. These magmas have been only slightly modified after their segregation from the source region (Green and Ringwood 1967; Kushiro, 1973; Frey et al., 1974; Blanchard et al., 1976; Bryan 1977; Langmuir et al., 1977; Bougault et al., 1979; Presnall et al., 1979; Rhodes et al., 1979; Fuji and Bougault, 1983; Takahashi and Kushiro, 1983). The opposing view maintains that the primary magmas are picritic, and have undergone extensive fractionation of mainly olivine to yield the most primitive MORB:s. As to the depth of generation of picritic magmas the opinions diverge and the suggested pressure region ranges from 5 to 30 kb (O'Hara, 1968, 1968a; Yoder, 1976; Elthon, 1979; Green et al., 1979; Jacques and Green, 1980; Stolper, 1980; Elthon and Scarfe, 1980, 1984). Presnall et al. (op.cit.) maintain that the existence of picritic liquids are of no importance in basalt petrogenesis. It has also been argued that even though picritic liquids might prevail at depth, they will not reach the surface because of their high density (Sparks et al., 1980; Stolper and Walker, 1980; Smewing, 1981).

The study of nodules could enhance our knowledge on the processes operating in depth. If the relation of the nodules to the host can be established this possibly will, depending on their origin, elucidate crustal magma processes, early fractionation history of the host as well as the physical and chemical properties of the crust and upper mantle. Many outstanding studies have been made of ultramafic mantle derived nodules in continental basalts and kimberlites, the compilation of which is found e.g. in Sobolev (1977), Boyd and Meyer (1979) and Hawkesworth and Norry (1983). These studies have helped in identifying possible primitive liquids, in constructing chemical and mineralogical models for mantle, in estimating pressure ranges for equilibriation of nodule assembladges, in interpreting deformation textures and in estimating velocity of eruption and volatile content of the nodule bearing host. Nodules, that undoubtly represent refractory mantle fragments, are most common in kimberlites and alkali basalts. At least one occurrence of spinel-lherzolite nodules is also reported in a tholeiitic basalt in Tasmania (Sutherland 1974).

In oceanic environment the reported occurrence of ultramafic xenoliths is restricted to alkaline basalts (Daly, 1965; White, 1966; Bennel, 1969; Jackson and Wright, 1970; Imsland, 1984). On the other hand mafic nodules and megacrysts of various origin are geographically widespread in oceanic tholeiites (Muir and Tilley, 1964, 1966; Nicholls et al., 1966; Aumento, 1968; Melson et al., 1968; Miyashiro et al., 1969; Melson and Thompson, 1971; Frey et al., 1974; Bougault and Hekinian, 1974; Donaldson and Brown, 1977; Rhodes et al., 1979). The scarcity of ultramafic mantle fragments in oceanic tholeiitic suites is explained by the rare occurrence of the primitive mantle derived basalts. In part it also may reflect the different nature of the upper mantle beneath the oceanic and continental crusts, the former having plagioclase peridotitic and the latter peridotitic composition.

Gabbroic nodules in Iceland have been reported previously by several authors (þ´orarinsson, 1953; Tryggvasson, 1963; Kristmannsd´ottir, 1971; Jakobsson et al., 1973; J´onsson, 1963, 1978; Jakobsson 1966, 1979; Sæmundsson, 1967; Larsen, 1979; Pedersen and Hald, 1982; Ragnarsd´ottir et al., 1982; Höj, 1984). Some nodules are interpreted as true xenoliths while others are considered to be cognate with the confining magma. A crustal origin is generally suggested.

Nodules from picritic rocks in Iceland have not been studied previously. Comparison shows unequivocally that the nodules in Miðfell are not of the same type as the above mentioned gabbro nodules. In the following the primitive nature and possibly deep origin of the Miðfell magma is discussed. Microprobe analyses of major elements in the nodule minerals and in the coexisting glass and mineral phases in the picritic host rock are reported. As the nodules are contained in a very primitive basalt they may provide means to evaluate the early fractionation of a mantle derived magma provided that a genetic relationship between the magma and the nodules can be established. Thus the primary aim of the work is to examine whether the nodules and the host rock are comagmatic or if they represent different magmas from a common source or if they are derived from totally different magmas.

2. GEOLOGICAL SETTING

A zone of active rifting and recent volcanism, the Neovolcanic Zone (NVZ), transverses Iceland from north in a N-S direction to central Iceland and then changes direction to SW fig. In central Iceland the zone is divided into two branches, the eastern and western volcanic zones. The eastern zone continues from the northern part and runs via Myrdalsjökull to Vestmannaeyjar and Surtsey. The western zone extends through þingvellir and the Reykjanes peninsula to the Reykjanes Ridge and connects the NVZ to Mid-Atlantic Ridge. The NVZ is flanked by Quartenary formations and Tertiary plateau basalts forming a roughly symmetrical pattern. The age of the rocks increases with increasing distance from the NVZ so that oldest rocks, about 16 m.y., are found in NW and in E Iceland (Sæmundsson 1978).

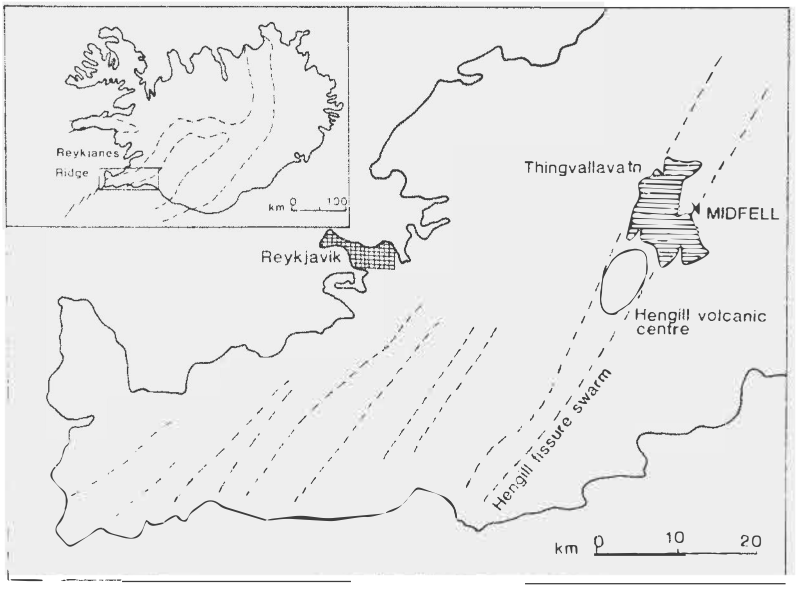


Figure 1. Index map showing the Neovolcanic Zone in Iceland, the fissure swarms on the Reykjanes peninsula and the sample location Miðfell at the lake þingvallavatn.

Miðfell is situated on the western branch of NVZ north of Reykjanes Peninsula. It is a part of the Hengill volcanic system. The Hengill central volcano is the easternmost of the five fissure systems of the Reykjanes peninsula arranged in NE-SW direction and en echelon along an E-W directed plate boundary (Sæmundsson 1978, Jakobsson et al. 1978).

Miðdfell is a hyaloclastite ridge formed in subglacial fissure eruptions. It is situated on the eastern shore of the lake Þingvallavatn (fig.1). The hill is about 300 meters high and extends 3 kilometers in SW-NE direction, the same general trend as observed for the fissures and topography in the Hengill area (Sæmundsson 1967). The ridge consists of two units of a slightly different age. The northeastern part is a plagioclase porphyritic hyaloclastite with only sparse olivine phenocrysts with or without inclusions of chromian spinel, and with microphenocrysts of plagioclase, olivine and spinel. As inferred from aerial photographs and field relations the southwestern part is older. It is a picritic pillow lava with abundant olivine phenocrysts, and is always spinel bearing. Plagioclase and clinopyroxene are also present in most samples.

The nodules are found in the picritic basalt in two different localities about 1 km apart on either side of the ridge. The size of the nodules ranges from few cm up to 10 cm. They are generally round, friable and appear corroded with interstitial ground mass derived from the host picritic basalt.

3. PETROGRAPHY

The petrography of the nodules and the host rock is first described and compared in this chapter. The chemical composition is discussed in the following two chapters.

3.1. Nodules

The gabbroic nodules are pieces of medium to coarse grained holocrystalline rocks with a hypidiomorphic inequigranular texture. The mineral assemblage consists of plagioclase, clinopyroxene and olivine. Spinel is found only as rare inclusions in olivine and in pyroxene and within the intervening host rock material. The minerals contain inclusions of glass in some cases. The nodule minerals show signs of resorption by a rounded outline. The averaged modal analysis of the nodules gives the mineral proportions as follows: 53 % plagioclase, 11 % clinopyroxene, 7 % olivine and 29 % interstitial ground mass.

In thin section the nodules seem to have disintegrated to a varying degree or they are in a state of reaction with the basalt. Some have compact cores while in others the intervening material has separated the nodule minerals from each other (fig. 2., 3.).

3.1.1. Plagioclase

Plagioclase is the dominating nodule mineral, up to 70 % of the nodule constituents. One of the sampled nodules is made almost entirely of plagioclase with only very minor olivine. The size of the individual plagioclase grains varies from 1 to 8 mm. The grains often occur in glomerocrysts. Both glomerocrysts and individual grains usually show well rounded surfaces with marginal resorption, in particular those close to the margins of the nodules. Oscillatory zoning is often very pronounced in the crystals. Liquid inclusions, sometimes zonally arranged, are an ubiquitous phenomenon in the plagioclases. The extent of resorption varies. In some samples plagioclase is extremely corroded, has rugged outlines and interstitial material penetrating into the grains. In others the crystals have fresh appearance, only signs of resorption being then rounded margins.

Figure 2. Photograph on the gabbro nodules in the picritic pillow basalt, as exposed at the sampling site. Miðfell, SW Iceland.

Figure 3 a). Microscope photograph on a gabbro nodule consisting of plagioclase, clinopyroxene and small olivine grains. Pronounced oscillatory zoning in plagioclase.

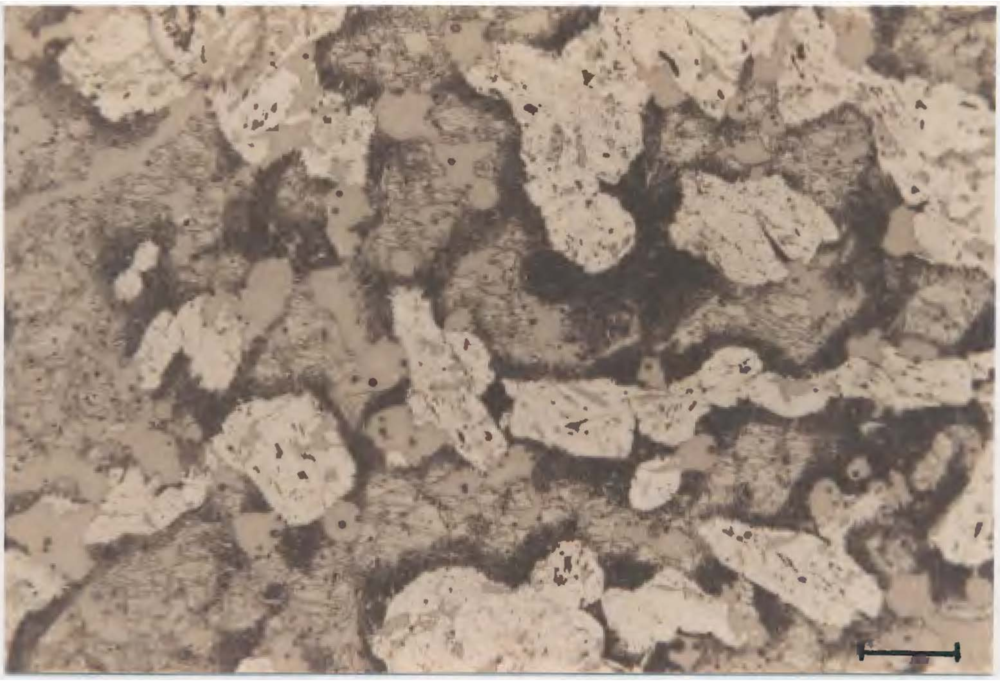


Figure 3 b). Disintegrating nodule consisting of plagioclase and clinopyroxene. Nicols not crossed.

3.1.2. Clinopyroxene

The clinopyroxene is a green, weakly pleochroic chromian diopside ranging in size up to 1 cm. It is always anhedral, sometimes almost spherical. It contains inclusions of olivine, plagioclase, glass and probably also of spinel. Twinning is common. The pyroxene is the most corroded nodule mineral and it shows the greatest degree of disintegration having interstitial host rock material intruded along the margins and the cleavage planes, sometimes to such an amount that only relics of the clinopyroxene are recognizable.

In some samples the clinopyroxene shows enough optical continuity, in spite of the extensive corrosion, that the plagioclase crystals can clearly be seen to be poikilitically enclosed by the large rounded clinopyroxene grains.

3.1.3. Olivine

The nodule olivines are smaller in size (from 0.5 to 4 mm, in average about 1.5 mm) than the plagioclase and clinopyroxene. The crystal habit is poorly developed as the grains are squeezed in between the main crystalline phases. The olivines contain occasional small inclusions of glass and spinel(?).

From the general appearance of the minerals in the thin sections it seems that the three phases, plagioclase, clinopyroxene and olivine, crystallized more or less simultaneously. As the pyroxene contains inclusions of olivine, at least some of the olivine must have crystallized prior to the clinopyroxene. The smallish subhedral olivine grains in between the larger pyroxenes and plagioclases are on the other hand products of a later crystallization. The sparse spinel inclusions and olivine included in clinopyroxene constitute the first crystalline phases.

3.2. Host Rock

The host rock of the gabbroic nodules is a picritic pillow basalt formed subglacially. The macro crystalline mineral phases are olivine, plagioclase, chromian spinel and clinopyroxene. The ground mass is glassy through cryptocrystalline. It has an intersertal to microgranular texture with small plagioclase laths, pyroxene, olivine and opaque minerals and glass. The amount of macrocrysts show a great variation ranging from about 10 to 50 %. This variation is largely due to the gravitational settling within the pillows after the eruption (fig. 4.). Similar occurrences have been described earlier from the Mælifell picrite in the Hengill area (Sæmundsson 1967) and from the Stapafell olivine tholeiite from Reykjanes (Matthews et al. 1964).

The average phenocryst mode of the picritic host rock estimated by point counting gives 13% olivine, 0.5% plagioclase, 1.3% spinel, 0.1% clinopyroxene and 84.6% ground mass for the samples from the glassy pillow rims and 25% olivine, 5% plagioclase, 5% clinopyroxene, 1% spinel and

64% crystallized ground mass for the samples taken from the inner parts of the pillows.

Figure 4. Photograph on the nodule bearing host pillow basalt showing the inhomogeneous distribution of the olivine phenocrysts within the pillows.

3.2.1. Olivine

Olivine is the dominating phenocryst phase in the picrite samples. The size of the phenocrysts varies between 0.5-6 mm, being in average 2-3 mm. In some cases the olivine shows a tendency towards a glomerophyric growth either with other olivines or with plagioclase and/or pyroxene. The grain size is usually somewhat smaller in the samples from the glassy pillow rims. The crystal form is mostly subhedral with rounded outline, but all gradations from euhedral to completely anhedral crystals are seen. Often the olivines show signs of mechanical stress. This is manifested by the occurrence of kinkbanded and microruptured grains as well as by a very pronounced undulating extinction in some crystals. Strongly rounded outlines again suggest resorption phenomena. Skeletal features implying a rapid crystallization are sometimes observed. Euhedral chromian spinel is commonly found as inclusion in the larger olivine grains and as microphenocrysts marginal to the olivines. The olivines also contain occasional glass inclusions. The morphology of the olivines is revealed in the photo series in fig. 5.

3.2.2. Plagioclase

Plagioclase phenocrysts are found in two size fractions. The larger megacrysts reach up to 5 mm, are often found as glomerocrysts of several crystals, sometimes together with olivine and/or pyroxene. Separate large grains are also encountered. The general appearance of these macrophenocrysts is similar to that of the nodule plagioclase (fig. 6.). These, too, are subhedral with rounded outlines, rich in inclusions, sometimes zonally arranged and with a frequent marginal oscillatory zoning. The smaller crystals (less than 0.5 mm) are almost euhedral and usually contain much fewer or no inclusions. Occasionally these are found as inclusions in pyroxene.

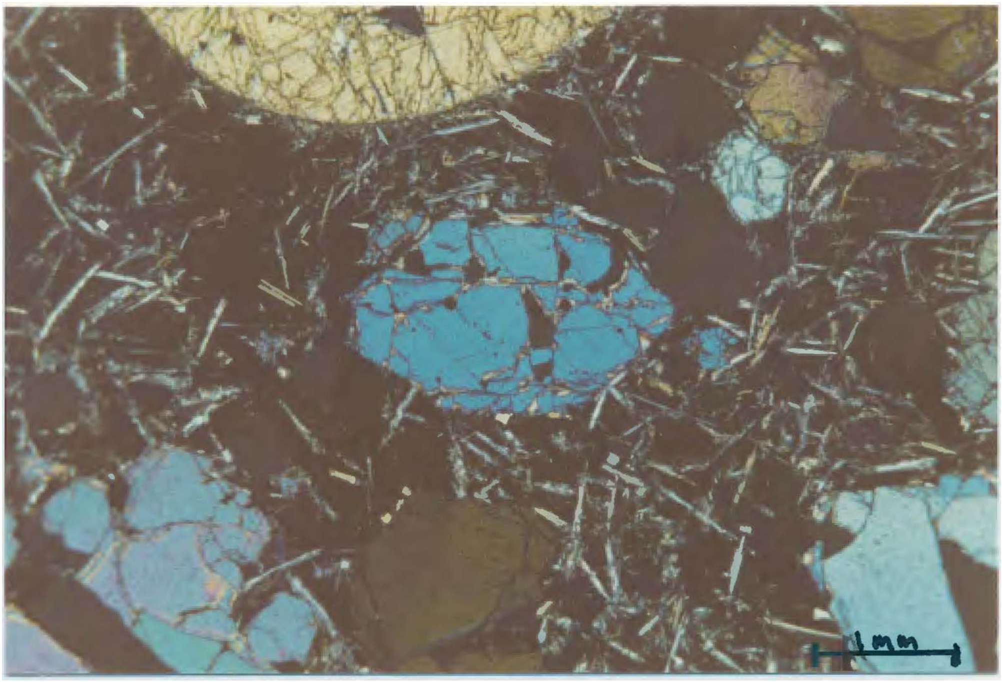


Figure 5 a). Euhedral olivine phenocryst with inclusions of chromian spinel.

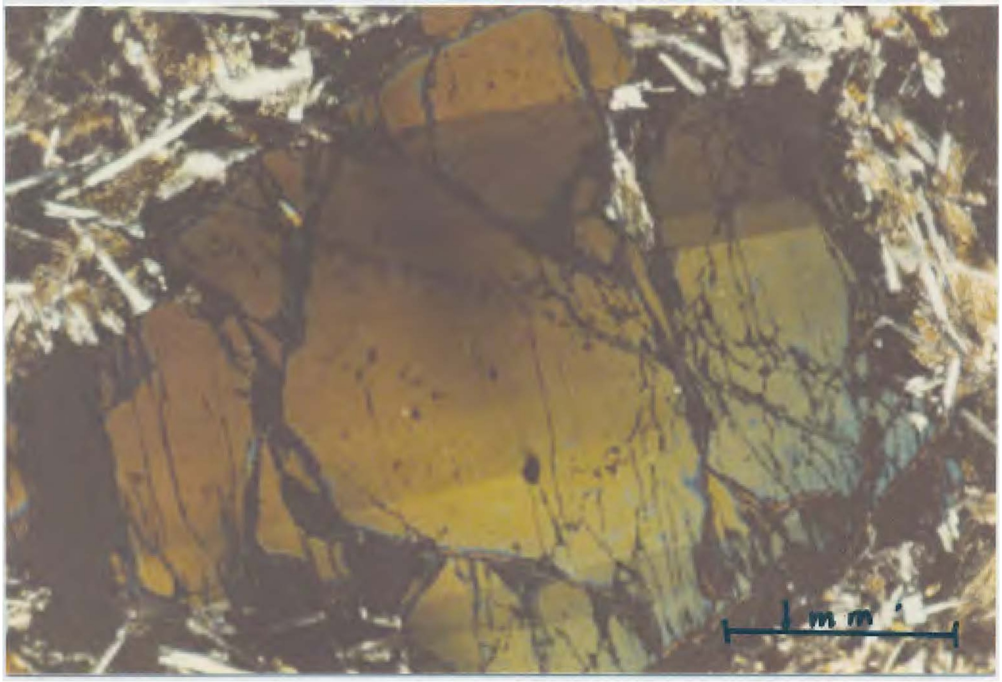


Figure 5 b). Kinkbanded anhedral olivine in the picritic host.

Figure 5 c). Anhedral, resorbed olivine in the host rock.

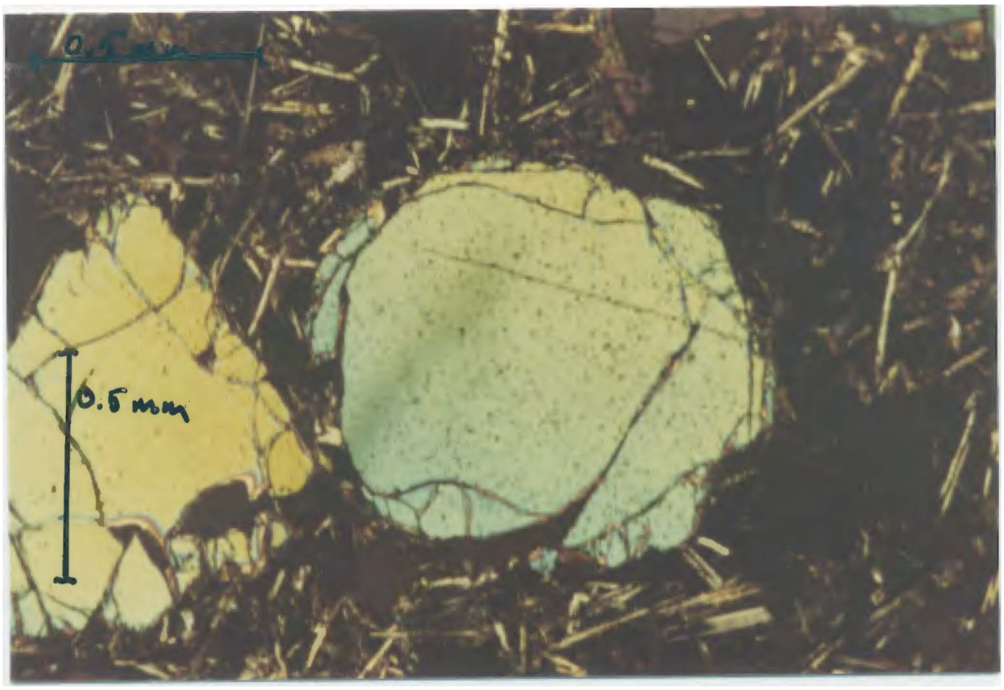
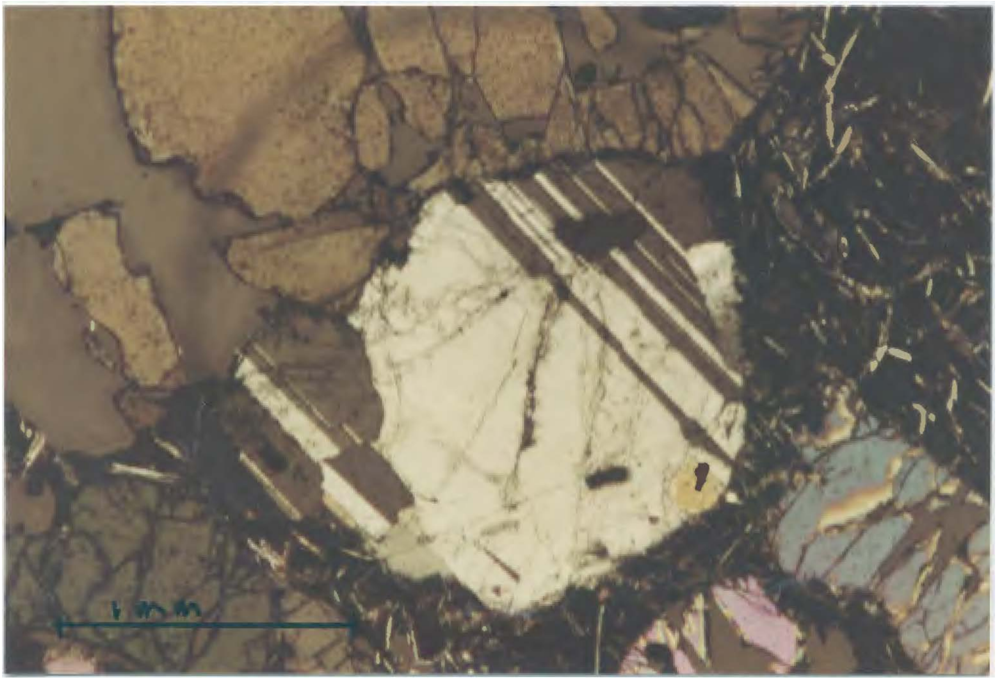


Figure 6. Glomerocrystic plagioclase megacryst in the host rock.



3.2.3. Clinopyroxene

Some clinopyroxene is present in most of the thin sections of the picritic basalt, but in the glassy rims it is a very minor constituent. In general appearance the pyroxene is similar to that in the nodules (fig. 6.) having rounded outlines and showing varying degree of disintegration. Glomerophyric clusters with other clinopyroxene and/or olivine and/or plagioclase are often found. The size of the grains varies from 2 to 7 mm. Twinning is often observed. The inclusions contained within the pyroxenes are olivine, spinel, plagioclase and glass.

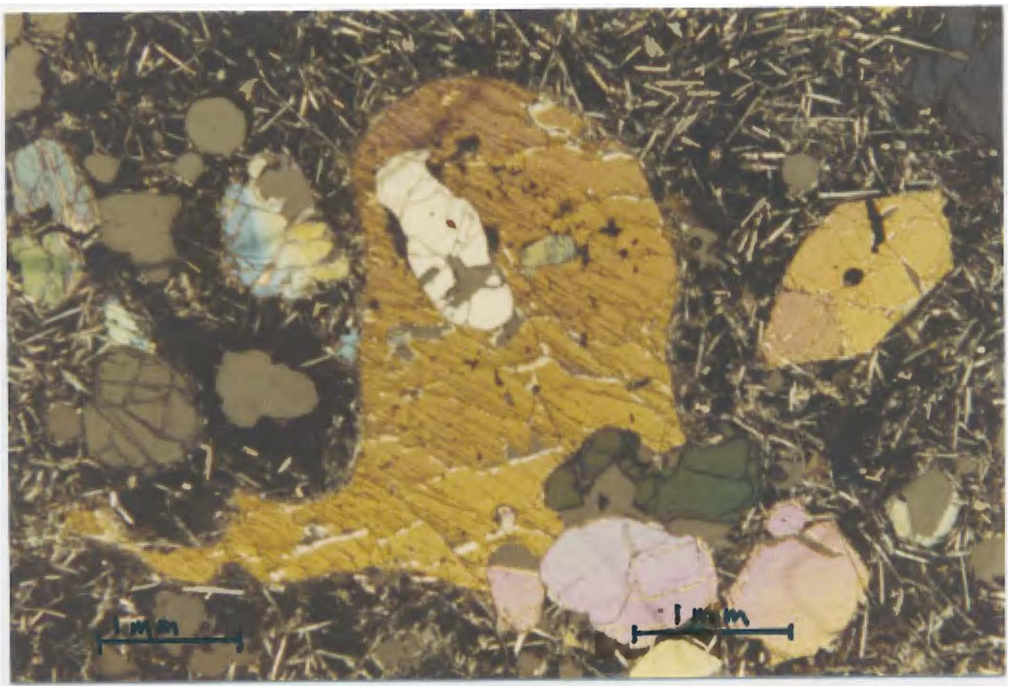


Figure 7. Clinopyroxene megacryst in the host rock containing inclusions of olivine and chromian spinel. The bigger olivine inclusion also has chromian spinel inclusions. The euhedral olivine to the right contains liquid inclusions.

3.2.4. Spinel

Spinel is found in the picrite basalt as phenocrysts and phenocryst clusters mainly marginal to olivine or as inclusions either in olivine or, more seldom in pyroxene (fig. 5., 7.). No spinel inclusions were found within the plagioclases.

The spinels are in most cases euhedral, but sometimes a slight rounding is observed. The size varies from less than 0.01 to 0.6 mm. In average it is about 0.2 mm. No difference in the sizes of the different types of spinels was observed. The color is deep brown.

4. MAJOR ELEMENT CHEMISTRY

Averages of the microprobe analyses of the different mineral phases in the nodules and in the host rock are given in tables 1, 2, 3 and 4 and of glass in table 5.

4.1. Plagioclase

Both core and rim of the nodule plagioclase were analyzed. The composition ranges from An83.1 to An92.5, most of the grains falling between the limits of An86-90 (fig. 8.). About one third of the analyzed grains showed a slight reversed compositional zoning the difference between core and rim being 0.6-4.1 mol%. In the rest no zoning was observed.

In the picritic host basalt An content of the plagioclase ranges from 86.0 to 90.9 mol%, this range thus being similar to that of most of the nodule plagioclase (fig. 8.). Despite considerable overlap the anorthite content in nodule plagioclase is generally slightly lower than in host rock plagioclase. This slight but significant difference will be discussed later. Zoning was observed in eight of the analyzed grains. In most cases it is reversed (6 grains) the difference between core and rim varying from 1.5 to 3.7 mol%. Normal zoning was insignificant (less than 1 mol%) or not observed.

TABLE 1. Average microprobe analyses on plagioclase minerals from Miðfell, SW Iceland.

no	SiO2	Al2O3	FeOt	CaO	Na2O	K2O	total	An%
nodule plg:								
1	46.9	32.7	0.46	16.6	1.65	0.02	98.3	85.0
2	46.7	33.4	0.49	17.2	1.54	0.01	99.3	86.1
3	47.3	32.9	0.51	16.7	1.67	0.02	99.1	84.7
4	46.4	34.0	0.46	17.5	1.42	0.02	99.8	87.2
host rock plg:								
5	46.2	34.2	0.49	17.7	1.36	0.02	100.0	87.8
6	44.3	33.6	0.47	17.6	1.47	0.02	99.5	86.9
7	47.6	33.7	0.53	17.6	1.49	0.02	101.0	86.7
8	47.5	33.4	0.38	16.7	1.36	0.02	99.4	87.2
9	47.0	33.2	0.49	17.5	1.52	0.03	99.7	86.4
12	46.4	34.0	0.46	17.5	1.42	0.02	99.8	87.2
11	47.5	33.5	0.40	17.6	1.38	0.04	100.4	87.6

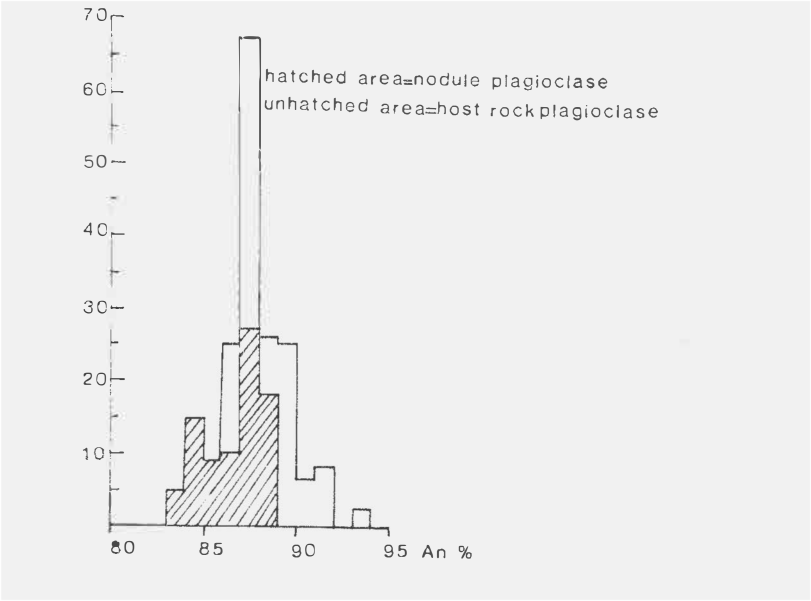


Figure 8. Histogram of the plagioclase composition given as percentage anorthite for all the analyzed samples from Miðfell.

4.2. Olivine

In nodule olivines the forsterite content ranges from 86.4 to 88.8 mol% (fig. 9.). The average for core and rim is similar, but within individual grains a very slight compositional zoning is commonly observed the cores having a Fo-content of 86.4-87.3 and the rims of 86.4-88.1 mol%. This zoning is both normal and reversed.

The host rock olivines also show very narrow compositional variation (fig. 9.). The great majority of the phenocrysts, both from the crystallized parts and the glassy ground mass, as well as the microphenocrysts from the quenched rims fall within the limits of 86-91mol% Fo. The range is thus similar for nodule and host rock olivines. Some host rock grains had a slightly higher Fo content than the nodule olivines. This extends the range in Fo content of the host rock olivines few mol% higher. The total range of the host rock olivines is from Fo81.8-90.7. Only one of all the analyzed grains had a markedly lower Fo content (81.8 to 85 mol%). In the olivines from the glassy pillow rims no difference between the composition of the core and rim was observed. A slight compositional zoning is, however, frequently observed in the olivines in crystallized samples. The zoning is more often reversed than normal. The difference between core and rim within the individual grains varies from 0.3 to 3.2 mol% in the reversely zoned and from 0.2 to 1.1 mol% in the normally zoned crystals. The strained and partly resorbed olivines are compositionally indistinguishable from the rest (fig. 9.). Fo content ranges from 85.5 to 90.1 mol%, and minor zoning, when detected, is reversed.

TABLE 2. Average microprobe analyses on olivines from Miðfell, SW Iceland.

no	SiO2	FeOt	MnO	MgO	CaO	NiO	total	Fo%
nodule ol:								
1	40.0	11.5	0.19	47.8	0.24	0.24	100.0	88.1
3	40.2	11.8	0.19	46.6	0.24	0.18	99.2	87.6
host rock ol:								
5	39.3	11.9	0.18	47.7	0.25	0.21	99.5	87.7
6	39.6	11.8	0.18	46.9	0.23	0.25	99.0	87.6
7	39.7	12.0	0.19	46.6	0.24	0.23	99.0	87.4
8	40.3	11.2	0.18	48.9	0.24	0.22	101.0	88.6
9	40.4	11.5	0.18	47.6	0.25	0.26	100.0	88.0
10	39.9	11.0	n.a.	47.3	n.a.	n.a.	98.4	88.5
11	40.3	10.7	n.a.	46.9	n.a.	n.a.	97.8	88.7
13	40.7	12.0	0.17	47.1	0.26	0.27	100.5	87.5
14	39.8	11.8	0.18	46.7	0.24	0.26	99.0	87.6

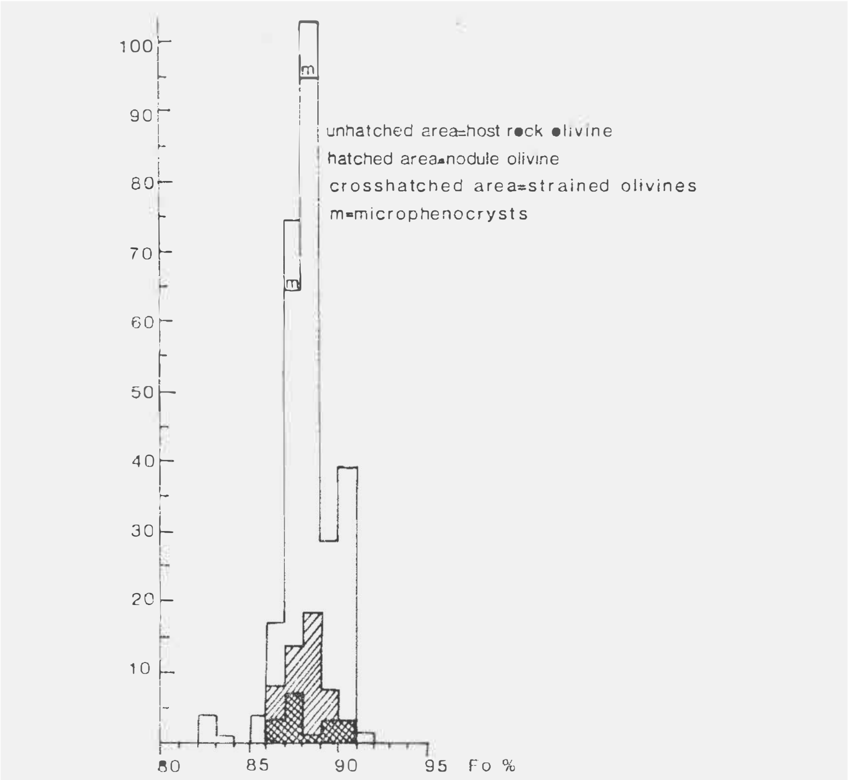


Figure 9. Histogram of the olivine composition given as percentage forsterite for all the analyzed samples from Miðfell.

4.3. Clinopyroxene

The analyses of nodule clinopyroxene show a rather narrow chemical variation (fig. 10. table 3.). Compositional zoning is negligible except for sample Mið-11, where in two cases the rims proved to be 1-2 mol% lower in Fs and higher in En than the cores. Chromium shows no consistent variation between the rims and the cores.

Only two of the analyzed host rock grains showed lower chromium content (0.25-0.39 w% Cr2O3) and at the same time the highest values for Fs (9-10 mol%). Also in the host rock the compositional zoning is generally nonexistent, insignificant or at least inconsistent, where observed.

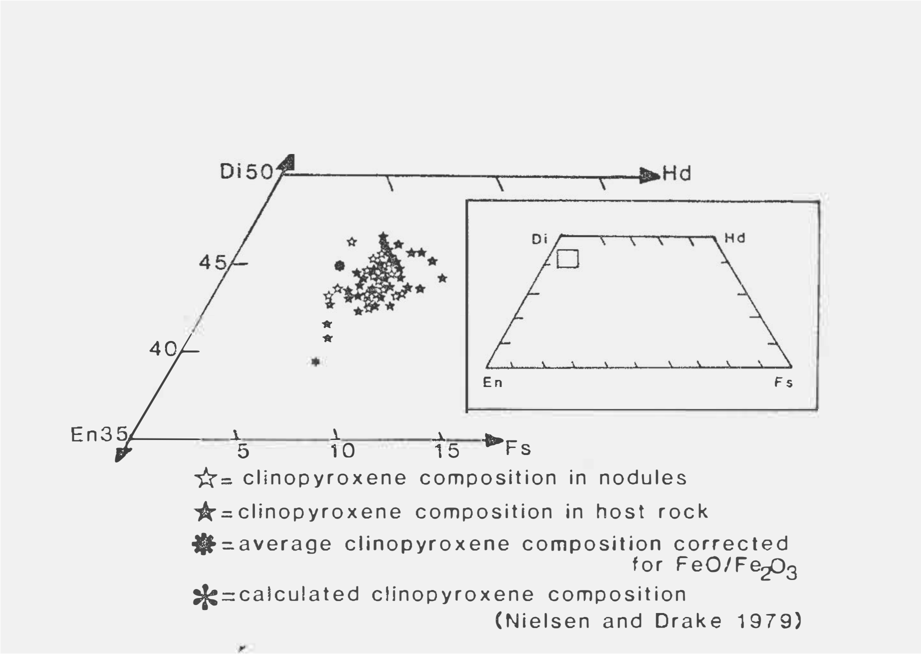


Figure 10. Summary of the clinopyroxene composition in the Miðfell samples plotted in the En-Fs-Hd-Di quadrilateral. As a reference the area occupied by the chromian diopsides from Miðfell within the whole quadrilateral is given.

The ranges in Fs%, En% and Wo% contents for the host rock clinopyroxene and the calculated, equilibrium composition are seen in fig. 10. In general the chromian diopsides show remarkably little variation as to the Fs, En, Wo and Cr_2O_3 content. Essentially they also are identical in the nodules and in the host rock megacrysts. Chromium content ranges from 0.49 to 1.23 w% for nodule and from 0.25 to 1.32 w% for host rock clinopyroxene. The majority of the grains have Cr_2O_3 content about 1 w%. Compositionally all the clinopyroxenens fall within the limited diopside field in the pyroxene quadrilateral (fig. 10.).

TABLE 3. Microprobe analyses on clinopyroxenes from Miðfell, SW Iceland.

no	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeOt	MnO	MgO	CaO	NaO	total	Fs%	En%	Wo%
nodule cpx:													
1	51.2	0.35	4.39	1.02	4.88	0.12	16.9	20.9	0.23	100.0	7.9	48.8	43.3
	51.3	0.27	4.63	0.94	4.25	0.09	16.6	21.1	0.20	99.4	7.0	48.6	44.4
3	51.0	0.30	3.92	0.58	4.69	0.13	16.4	21.1	0.19	98.3	7.7	48.0	44.3
4	51.1	0.24	3.97	0.80	4.23	0.09	16.6	21.2	0.18	98.4	6.9	48.5	44.6
host rock cpx:													
	51.1	0.29	4.96	1.17	4.23	0.10	16.3	21.4	0.20	99.8	7.0	47.9	45.1
6	50.6	0.26	4.76	1.06	4.30	0.10	16.7	20.9	0.20	98.9	7.1	48.9	44.0
	51.0	0.23	4.72	0.90	4.17	0.11	16.8	21.1	0.19	99.2	6.8	49.0	44.2
8	50.8	0.33	4.93	0.98	4.74	0.11	16.8	20.9	0.20	99.8	7.7	48.7	43.6
	50.8	0.28	4.84	1.04	4.46	0.10	16.5	20.9	0.20	99.1	7.4	48.5	44.1
10	50.6	0.27	5.32	1.17	4.07	0.10	16.4	21.1	0.23	99.3	6.7	48.5	44.8
11	51.7	0.27	4.70	1.20	4.07	0.10	16.7	21.2	0.21	100.2	6.7	48.8	44.5
16	51.1	0.32	4.62	0.90	4.47	0.11	16.4	20.3	0.21	98.4	7.5	49.0	43.5
17	50.3	0.26	5.10	0.94	4.17	0.11	16.8	20.6	0.21	98.5	6.9	49.5	43.6

4.4. Spinel

Spinel phenocrysts, phenocrysts marginal to olivine and spinel inclusions in olivines and in pyroxenes were analyzed (table 4.). Because of the small grain size it was only exceptionally possible to make a distinction between the core and the rim. Where this was done, the core proved in general, but not without exception, to be more chromian than the rim, whereas FeOt content was generally slightly higher in the rim.

TABLE 4. Microprobe analyses on spinels from Miðfell, SW Iceland.

no	TiO_2	Cr_2O_3	MnO	FeOt	MgO	Al_2O_3	total
inclusions in olivine	0.26	30.5	0.19	16.9	16.2	34.9	99.0
separate phenocrysts	0.29	26.7	0.19	18.4	17.2	36.5	99.3
5							
inclusions in olivine	0.26	35.3	0.17	16.9	15.5	30.4	98.5
inclusions in pyroxene	0.25	37.5	0.17	17.0	15.7	28.6	99.2
marginal phenocrysts	0.24	35.9	0.20	16.7	15.7	31.0	99.7
6							
inclusions in olivine	0.27	33.7	0.20	17.4	15.9	32.0	99.5
inclusions in pyroxene	0.40	33.6	0.22	19.1	15.2	30.4	98.9
separate phenocrysts	0.27	30.0	0.20	17.5	15.3	34.7	98.0
7							
inclusions in olivine	0.25	30.5	0.20	17.3	16.2	34.1	98.1
8							
inclusions in olivine	0.26	34.6	0.21	16.6	16.4	31.8	99.8
separate phenocrysts	0.25	35.4	0.20	17.0	16.0	31.0	99.9
9							
inclusions in olivine	0.26	34.0	0.19	17.3	15.7	31.7	99.2
marginal phenocrysts	0.25	34.9	0.21	17.1	15.6	30.2	98.3
10							
separate phenocrysts	0.25	29.0	0.18	16.6	16.4	35.6	98.0
11							
inclusions in olivine	0.26	31.3	0.19	16.8	16.6	33.8	99.0
inclusions in pyroxene	0.42	31.8	0.20	19.0	15.6	33.1	100.1
separate phenocrysts	0.22	27.8	0.18	16.5	16.4	38.2	99.4
17							
separate phenocrysts	0.28	33.2	0.20	16.7	16.3	32.3	99.0
14							
separate phenocrysts	0.24	31.5	0.17	17.4	16.3	32.9	98.5

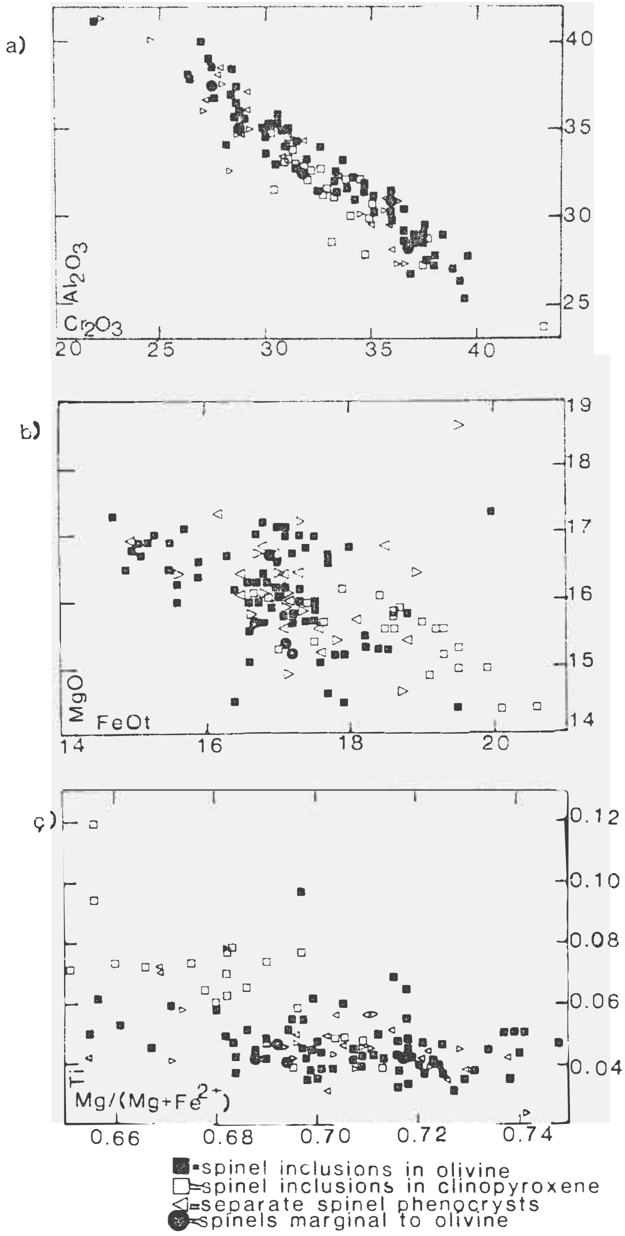
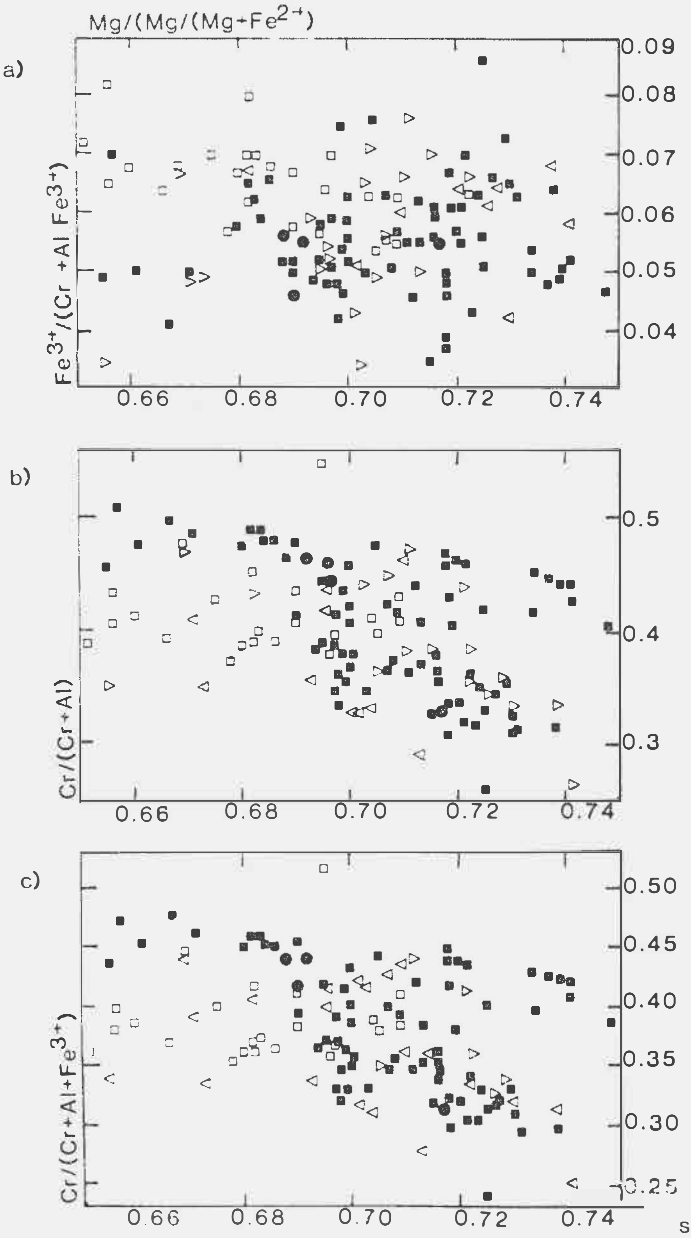


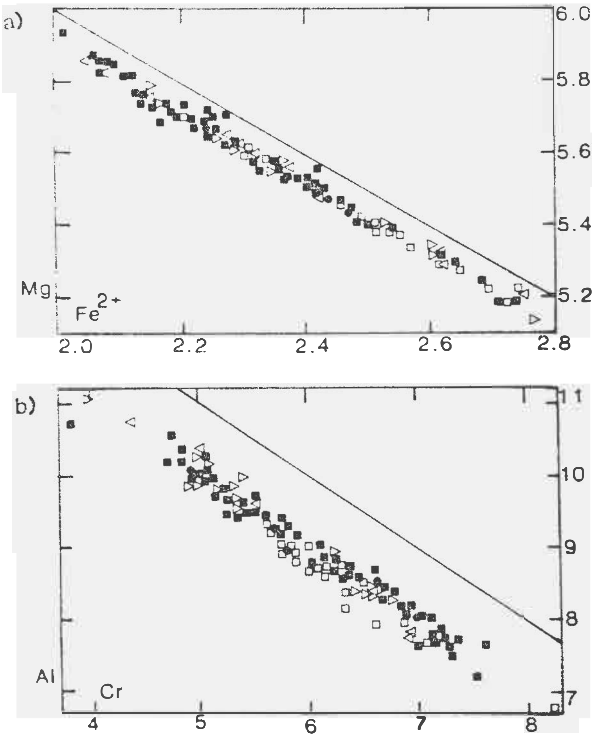
Figure 11. Microprobe analyses the chromian spinels from Miðfell. In fig. 11 c) the spinel analyses are calculated to 32 oxygens an 24 cations, total iron is redistributed in proportions required for the spinel stochiometry (Finger, 1972).

The chemical composition of the spinels is highly variable in contrast to other mineral phases. This is especially true for the Al_2O_3/Cr_2O_3 ratio which shows a large variation (fig. 11 a.). The MgO/FeOt ratio is more constant and the variation is mainly due to the FeO content (fig. 11 b.). The chromium content varies from 21.8 to 39.5 weight% in the inclusions within olivines, from 22.1 to 36.5 weight% in the separate (micro)phenocrysts, from 27.5 to 35.9 weight% in the marginal phenocrysts and from 30.0 to 43.2 weight% in the inclusions within the pyroxenes.



symbols as in fig. 11.

Figure 12. Microprobe analyses on the chromian spinels from Miðfell. The analyses are calculated to 32 oxygens and 24, total iron redistributed between Fe2+ and Fe3+ in stochiometric proportions (Finger, 1972).



symbols as in fig. 11. and 12.

Figure 13. Chromian spinels from Miðfell. The microprobe analyses recalculated as in fig. 11 c) and 12. The ideal sums for normal spinel stochiometry Mg+Fe2+=8 and Al+Cr=16 are shown by the lines in fig. 13 a) and b).

Spinel inclusions in the pyroxenes are slightly richer in Fe and Ti than the other spinels (fig. 11 b., 11 c.). The range in chromium content also extends higher. The same general trend is obvious when the ratios between the trivalent cations is plotted against the ratios between the divalent cations (fig. 12.). The difference is small but consistent throughout when all the analyses are considered. There is, however, extensive overlap when all the grains of different spinel types are considered. In spite of this the clinopyroxenes that contain spinel inclusions are identical in chemical composition with those without spinels.

In fig. 13. the substitution of Mg and Fe2+ and of Al and Cr are shown. The Miðfell spinels are very close to the ideal sums, 8 respectively 16, for the normal spinel structure. Even though they deviate slightly from these sums, the slope is exactly the same and continuous, a fact which indicates a strict stochiometric control.

4.5. Glass chemistry

The glassy pillow rims are porous and friable, and black in hand specimen. In thin section the glass is pale brown, translucent, isotropic sideromelane. Opaque, tachylytic glass is sometimes found in the inner parts of the rims and within the pillows. The analyzed glasses contain phenocrysts and microphenocrysts of olivine, spinel and frequently of some plagioclase whereas clinopyroxene is only rarely found. The amount of phenocrysts in the analyzed samples is in average about 16%.

TABLE 5. Average microprobe analyses on the glass phase from Miðfell, SW Iceland.

no	SiO_2	TiO_2	Al_2O_3	FeOt	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	tot
13	48.4	0.76	15.7	9.21	0.16	9.74	15.4	1.54	0.00	0.04	101.0
14	48.5	0.76	15.5	9.17	0.18	9.74	15.5	1.54	0.01	0.04	100.9
14	48.7	0.76	15.5	9.21	0.17	9.58	15.3	1.55	0.01	0.03	100.8
15	48.5	0.80	15.4	9.12	0.17	9.62	15.2	1.57	0.00	0.04	100.4

The glass analyses are listed in table 5. In figures 14 and 15 they are compared with the glasses from other areas. The glass is shown to represent the most basic and the most primitive found in Iceland. The low FeOt/MgO ratio of 0.94-0.95 and FeOt/(FeOt+MgO) of 0.48-0.49 and the very low K2O content of less than 0.01 weight% and the low contents of TiO2 (0.76-0.80 weight%) and P2O5 (0.03-0.04 weight%) indicate the primitive nature of glass. The exceptionally high content of CaO and Al2O3, both exceeding 15 weight% and the ratio being close to unity, is worth pointing out as a distinguishing property of the Miðfell magma.

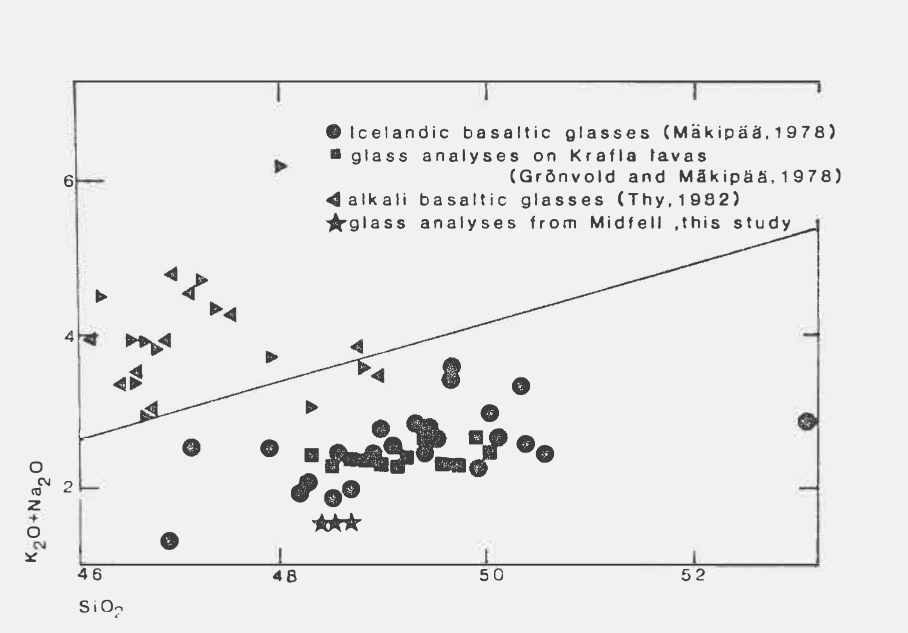
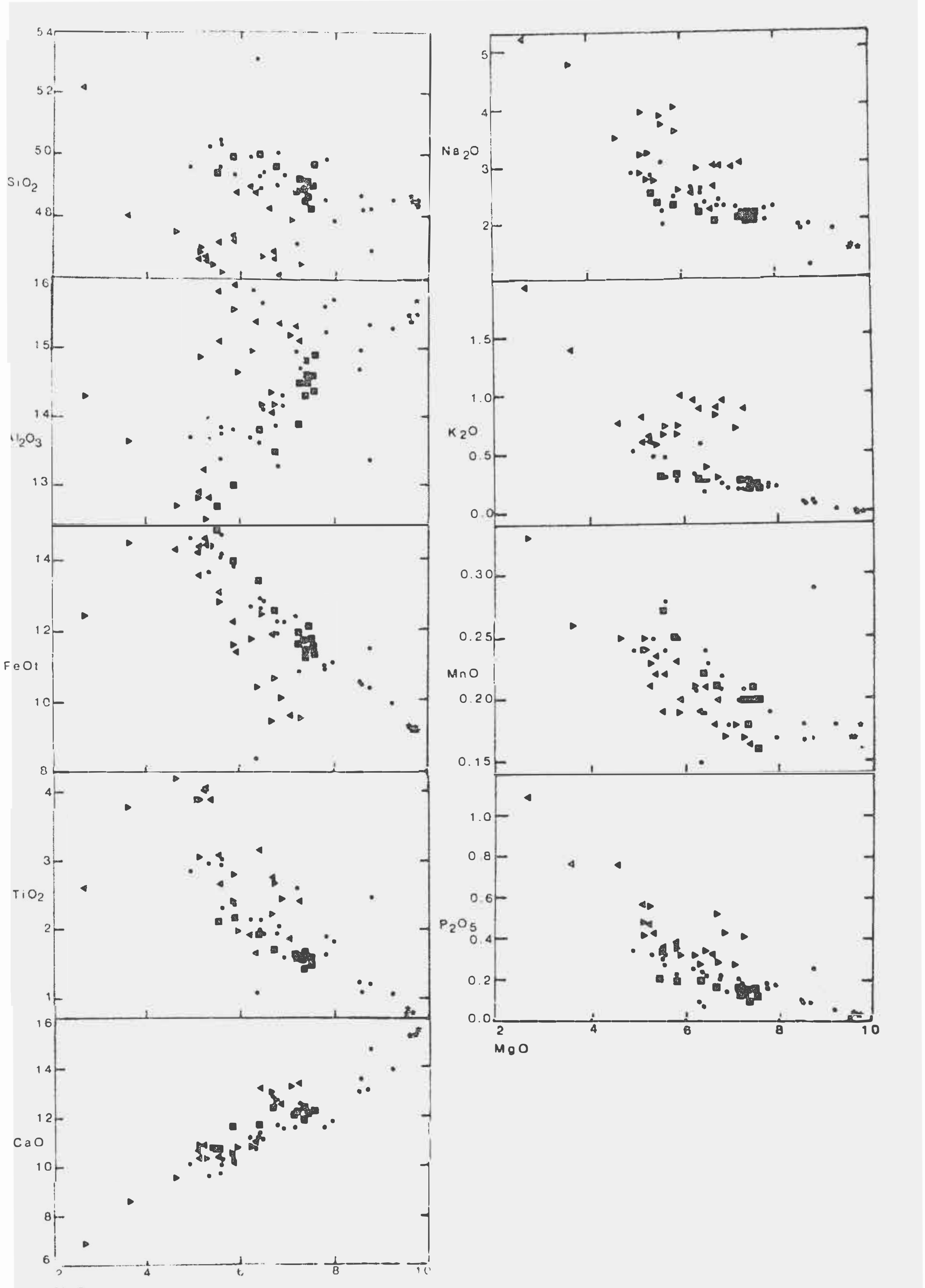


Figure 14. Total alkalies versus silica. The Hawaiian dividing line from MacDonald and Katsura (1964).



symbols as in fig. 14.

Figure 15. MgO variation diagrams. The homogeneous and primitive composition of the Miðfell glass is compared with available data on other Icelandic basaltic glasses.

On MgO variation diagrams the Miðfell glasses follow the general trends, characteristic for the tholeiitic basalts in Iceland (fig. 15.), i.e. Al_2O_3 and CaO increase with increasing MgO, while FeO_t P_2O_5, K_2O, Na_2O and TiO_2 decrease with MgO. This is also clearly seen when the total alkalies are plotted against the SiO_2 content (fig. 14.). For comparison and to get an idea on the general compositional trends in Icelandic basaltic glasses the following analyses have been included in the diagrams: basaltic glasses(Makipaa 1978a), glass analyses on the Krafla lavas (Grönvold and Makipaa 1978) and analyses on the alkali basaltic glasses (Thy 1982). The major element analyses of the Miðfell glass show no significant variation suggesting that the magma was homogeneous on eruption. No correlation between the amount of phenocrysts and the glass composition was observed.

The results show that olivine, clinopyroxene and plagioclase both in the nodules and in the host rock are remarkably homogeneous. Compositional zoning is almost nonexistent or it is very minor and inconsistent where observed. There is no significant difference in the very primitive chemical composition of nodule and host rock minerals. On the basis of similar appearance and chemical composition and, recalling that the nodules in some cases can be seen in a state of disintegration, the host's glomerocrysts as well as clinopyroxene, plagioclase and some of olivine megacrysts are most likely xenocrysts originating from the more fully disintegrated nodules. This is confirmed when the Miðdfell samples are compared with other picritic rocks: in picrites olivine and spinel are obligatory phenocryst phases, plagioclase is rare and clinopyroxene is not generally found (Blanchard et al., 1976; Sigurdsson and Shilling, 1976; Jakobsson et al., 1978; Jakobsson, 1983).

The exception from this uniformity are the spinels with highly variable composition. This feature remains unexplained, but not unique as it is reported also in other sample suites (Frey et al., 1974; Sigurdsson and Shilling, 1976; Pallister and Hopson, 1981). The slight difference observed between spinel inclusions in clinopyroxene and the other types of spinels, is already pointed out, and a possible explanation will be given later.

The glass is primitive. It is the most magnesian glass found in Iceland, and it has a low content of alkalies and incompatible elements indicating low degree of crystal fractionation. The primitive composition of the minerals and of the glass suggests a possible relationship between the two.

5. TRACE ELEMENT ANALYSES

Trace elements, six transition metals and Sr, were analyzed in order to (a) further evaluate the state of evolution in magma and nodules, (b) to see if there is a difference in trace element composition of nodule and host rock minerals, (c) to further test host-nodule relations and their origin and (d) to ultimately use trace element composition as an indication of primary liquid composition.

Some of the trace elements have very high mineral-liquid distribution coefficients (D values), and they are preferentially concentrated in specific minerals. The trace element abundances in the liquid are therefore strongly affected by the fractionation of these particular phases. Relative changes in trace element concentrations during fractionation are much larger than in case of major elements. Trace elements thus provide a sensitive test for fractional crystallization models. It also has been claimed that Ni content in mantle derived olivine is uniform (Sato 1977) and that the oxidation state of Ni is insensitive over a wide range of oxygen fugacities of basaltic magmas (Seward 1971). This makes Ni especially suitable when fractionation history is to be deciphered.

The elements Ni, Cr,and Co, preferentially included in the mafic minerals, and the less compatible elements Mn, Ti and Zn were analyzed with microprobe in olivine, clinopyroxene, spinels and glass. In addition Sr in nodule plagioclase and ground mass and in host's glass was analyzed with XRF. Trace elements were considered to be homogeneously distributed within the liquid. This may be an oversimplification as analyses of the glass phase showed apparent inhomogeneities within and between samples. Table 6 gives the average trace element content for glass and the ranges for the minerals as well as the ranges in the calculated distribution coefficients.

With the exception of Co the average content of trace elements in glass is comparable with the most primitive glasses (MgO > 9 w%) analyzed from oceanic tholeiites (Bougault and Hekinian, 1974; Frey et al., 1974; table 6.). The composition of the mineral phases show a rather wide range in trace element contents. This range is not bimodal excluding the possibility of two chemically distinct sources for the nodules and the host. Neither do the megacrysts represent refractory mantle xenocrysts, because in that case a uniform composition would be expected. The continuous ranges in the trace element content instead indicate a cumulate origin. It is also suggested on the basis of the trace element variation that some of the olivines are not in equilibrium with the confining liquid.

No major difference in the trace element content of nodule and host rock minerals can be seen, and in all cases the ranges overlap. The analytical data show, however, some differences, the significance of which is discussed later:

The calculated distribution coefficients are

-for nodule olivine $D_{cr} < 1$, for host rock olivine $1 < D_{cr} > 1$

-for nodule olivine D_{ni} is usually lower than for host olivine

The incompatible elements Ti, Zn and Mn showed no difference in behaviour when nodule and phenocryst minerals were compared.

All the elements showed some degree of zoning in some cases when cores and rims of the mineral grains were compared. There was no regularity in the zoning, and its significance is doubtful.

TABLE 6. Average trace element content in the glass phase, ranges in trace element contents in the minerals and distributions coefficients.

	glass	nod.ol.	host ol.	nod cpx.	host cpx	spinel
Ni	150	1110-1891	1044-3097	81-455	96-389	812-1202
1*	200					
2*	165					
Cr	381	108-257	131-644	2236-8826	3598-8403	
1*	470					
2*	505					
Co	164	167-485	209-466	90-512	168-449	542-1203
1*	45					
2*	42					
Ti	4045	8-258	9-272	1217-2891	1302-2779	2200-2597ip 1506-1570io
1*	4600					
2*	4860					
Mn	1260	1100-1721	806-1996	549-1063	595-1003	1400-1764
1*	1317					
2*	1324					
Zn	91	12-85	11-82	23-75		111-562
2*	52					
Dni		7.4-12.4	7.0-20.6	0.5-3.0	0.6-2.4	6.0-8.0
Dcr		0.28-0.67	0.34-1.69	5.87-23.2	9.44-22.1	
Dco		1.01-2.96	1.27-2.84	0.55-3.12	1.02-2.74	6.12
Dti				0.30-0.71	0.32-0.69	0.3-0.5
Dmn		0.91-1.43	0.64-1.58	0.44-0.84	0.47-0.80	1.11-1.40
Dzn		0.13-0.93	0.12-0.90		0.25-0.82	1.22-6.18

	glass(host)	ground mass(nod)	plg	Dsr
Sr	105-110	105-110	250	2.35
1*	100			

io=inclusion in oli, ip=inclusion in cpx
1*=basaltic glass from Atlantic Ocean (Frey et al., 1974)
2*=picritic glass from Atlantic Ocean (Bougault and Hekinian, 1974)

6. EQUILIBRIUM RELATIONS AND TEMPERATURE ESTIMATES

The compositional similarity of the nodule and host rock minerals suggests that they could be derived from a same or very similar type of magma. Whether the crystallization of the observed minerals has taken place in equilibrium with the confining liquid can be evaluated by comparing the coexisting mineral-mineral and mineral-liquid pairs with experimental results. If the analyzed compositions correspond to the equilibrium compositions, these pairs can then be used as geothermometers to infer the crystallization temperatures. As the crystallization takes place over a temperature interval, the highest crystallization temperatures may approach the liquidus temperature.

The equilibrium relations have been evaluated by using coexisting olivine-liquid, clinopyroxene-liquid, clinopyroxene-olivine and plagioclase-liquid pairs by comparing them with experimentally determined relations. The equilibrium composition for clinopyroxene was calculated on the basis of the liquid composition according to the method of Nielsen and Drake (1979). As the microprobe does not differentiate between Fe2+ and Fe3+ these components in glass, ground mass and clinopyroxene were chemically analyzed. FeO in glass was 9.03 w%, in ground mass 4.33 w% and in clinopyroxene 2.89 w% The corrected microprobe analyses are the basis for equilibrium, temperature and oxygen fugacity calculations.

6.1. Olivine-liquid

The MgO and FeO contents in olivine and in melt according to the method deviced by Roeder and Emslie (1970) gives equilibrium constants (KD values) of 0.3052-0.3057 for host rock olivine and of 0.3052-0.3058 for nodule olivine. These values are in good agreement with the results of Roeder and Emslie (op.cit.) and so indicate equilibrium conditions. KD is calculated for the exchange reaction:

$$(MgO)ol + (FeO)l \longleftrightarrow (MgO)l + (FeO)ol$$

parentheses denote molar fractions

As the composition of olivine is thought to be dependent on the liquid composition only and since the liquidus temperature is also dependent on composition (Roeder and Emslie, op. cit.) the temperature of equilibration can be calculated using either the distribution of MgO or FeO between the liquid and the olivine. The nodule and phenocryst olivine compositions give estimated average temperatures of 1214 oC for MgO and and 1243 oC for FeO (table 7.).

Oxygen fugacity was calculated according to the formula of Roeder and Emslie (op.cit.). In olivine all iron was assumed to be ferrous. The values so obtained show some scatter, the negative logarithms of fO2 for the host ranges from 8.2507-12.0152 and for the nodules from 8.7118-12.5901. The average value of about 10 would suggest that both the nodule and host have crystallized at fugacities below the QFM buffer.

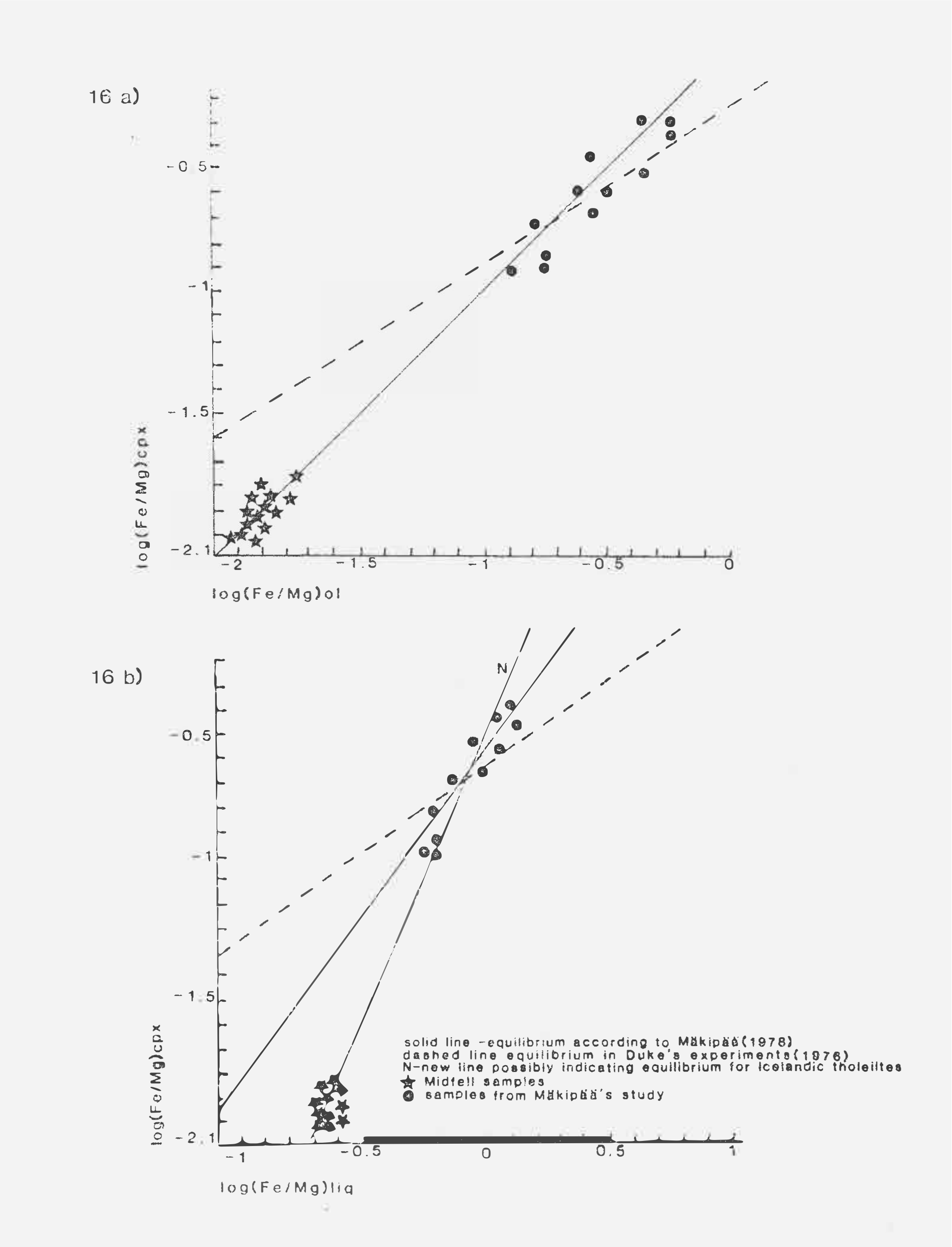


Figure 16. Logarithmic values of Fe/Mg in clinopyroxene plotted against those in olivine (16 a).. The Miðfell samples coincide with continuation of the equilibrium line of the Icelandic basalts. Fe/Mg in clinopyroxene versus Fe/Mg in liquid (16 b) shows poor correlation with earlier experiments, but the new line probably indicates equilibrium for the whole range of Icelandic tholeiites.

6.2. Olivine-clinopyroxene

The mole ratios of FeOt/MgO in coexisting clinopyroxenes and olivines when plotted against each other, give KD values close to unity (fig. 16 a.). The Miðfell samples coincide with the continuation of the line indicating equilibrium in Icelandic tholeiitic basalts (Makipaa 1978). The temperatures were calculated using Dni values of coexisting olivines and clinopyroxenes according to Hakli and Wright (1967), and a reasonable agreement was attained (table 7.).

6.3. Clinopyroxene-liquid

When the FeOt/MgO ratios of the coexisting clinopyroxene is plotted against that in the liquid the Miðfell samples show no correlation with the experimental results of Duke (1976) and they also fall off the line of equilibration extrapolated from the results of Makipaa (op.cit.; fig.16 b.). If all the data from Makipaa (op.cit.) and Miðfell are compiled together, an apparently reasonable trend results (fig. 16 b.). The new relation so obtained might represent equilibrium conditions for the whole range of the Icelandic tholeiitic basalts.

The clinopyroxene equilibrium composition can be estimated using the iterative method of Nielsen and Drake (1979) starting with a known glass composition. For the Miðfell glass the calculated equilibrium clinopyroxene composition is about En 53.5%, Fs 6.8% and Wo 40%, not markedly different from the analyzed composition (fig. 10.). The clinopyroxene therefore seems to be close to equilibrium with the confining glass. The temperature calculated with this method is 1210 oC.

Temperature can also be estimated on basis of the known equilibrium composition of melt and pyroxene (Nielsen and Drake 1979). Reciprocal of absolute temperature is correlated with the equilibrium constant (K) with the generalized linear regression equation; $\ln K=A/T+B$, A and B are constants and T is absolute temperature. K has been calculated for the reaction:

$$(CaO)l+(MgO)l+(FeO)l+(3SiO2) \longleftrightarrow (CaSiO3)px+(MgSiO3)px+(FeSiO3)px.$$

Parentheses denote activities, which are for the melt components:

$aSiO2=XSiO2/(XNaOo.5+XKO0.5+XSiO2)$

$aMgO=XMgO/((XMgO+XFeO+XCaO+(XAlO1.5-XNaO0.5-XKO0.5)$

$aFeo=XFeO/$ -"-

$aCaO=XCaO/$ -"-

and for pyroxene:

$aMgSiO3=XMg/XMg+XFe+XCa+2XNa+XCr+XTi+1/2(XAl+XNa))$

$aCaSiO3=XCa/$ -"-

$aFeSiO3=XFe/$ -"- (X=atomic fractions)

Nielsen and Drake (op.cit.) give different values for the constants A and B depending on the assumed mixing model for the melt components. In clinopyroxene the mixing at only single macroscopic 'M' site is considered here. When simple oxides in the melt are assumed to mix ideally the equation of linear regression to correlate K and T is:

$\ln K = 13596/T - 4.47$

This mixing model was first deviced by Roeder and Emslie (1970). Nielsen and Drake (op.cit.) showed that it is adequate only for low alkali basalts.

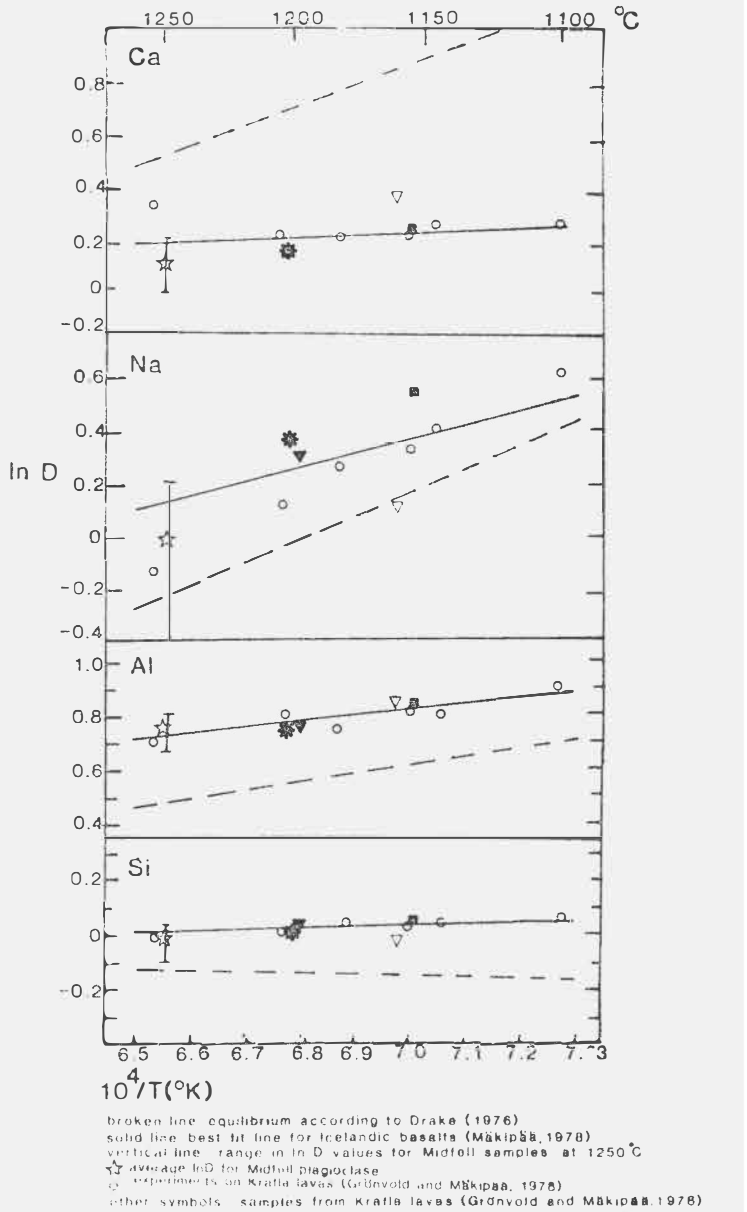
When the same macroscopic 'M' site mixing model for clinopyroxene components as assumed above and a mixing model modified from Bottinga and Weill (1972) for the melt is used, the equation is written:

$\ln K = 14600/T - 9.92$

Bottinga and Weill considered the melt to consist of network forming (NF) and network modifying (NM) components. When K, Na, Ba, Sr, and Ca, in this order, combine with the tetrahedrally coordinated Al in the melt, the components $KAlO_2$, $NaAlO_2$, $BaAl_2O_4$, $SrAl_2O_4$ and $CaAl_2O_4$ are formed (Bottinga and Weill, op.cit.; Drake, 1976). These, together with SiO_2, make up the network forming components. MgO, FeO and the CaO excess after having combined with Al, are the most important network modifiers. In the modified Bottinga and Weill model (Nielsen and Drake, op.cit.) all the Ca is assumed to act as network modifier instead of being distributed between $CaAl_2O_4$ (NF) and CaO (NM). The clinopyroxene temperatures are seen in table 5. The modified Bottinga-Weill mixing model gives good correlation with the FeO temperatures for olivine (table 7.).

6.4 Plagioclase-liquid

Equilibrium relations can be estimated using the distribution of Ca, Na, Al and Si in plagioclase and melt. The temperature must be estimated using some other method independent of plagioclase composition (Drake 1976). Here the temperature estimated with olivine and clinopyroxene geothermometers were used.



(redrawn after Grönvold and Mäkipää, 1978)

Figure 17. Distribution coefficients D for plagioclase-glass in the Miðfell samples compared with those determined by Drake (1976), Makipaa (1978) and Grönvold and Makipaa (1978).

The distribution of the major elements is compared with the experimental data of Drake (1975, 1976), Makipaa (1978) and Grönvold and Makipaa (1978; fig 17.). There is a close agreement with the values determined for Icelandic basalts indicating that the plagioclase composition is in equilibrium with the confining melt. Because of the negligible slope of the regression lines the major element distribution can not be used as a geothermometer, with the possible exception of Dna. The disagreement of the experimental data sets plotted clearly demonstrates that the bulk composition has a marked effect on distribution of the major elements.

The temperature estimates using major element composition base on the method of Kudo + Weill (1970) and it's different modifications (Mathez, 1973; Drake, 1976; Grönvold, 1984). The generalized form for the equation is:

$\ln\lambda/\mathcal{E} + A\emptyset'/T = BT - C$

A, B, C = constants

T = absolute temperature

$\emptyset' = xCa + xAl - xNa - xSi$

$\lambda = xNa\ xSi/xCa\ xAl$

$\mathcal{E} = XAb/XAn$

x - atomic fraction in liquid

X - mole fraction in plagioclase

All the temperatures except for those calculated for a partial H_2O pressure of 5 kb are considerably higher than the temperatures arrived at by using clinopyroxene and olivine geothermometers. Only the temperatures calculated on basis of dry experiments are compiled in table 7. An agreement was achieved only by using equation from Grðnvold (1984). That equation is based on experiments on Icelandic tholeiites and demonstrates an obvious compositional dependence of temperature estimates.

Sr distribution between plagioclase and melt can give temperature estimates when used unmodified. The equations of Drake (1972) and Drake and Weill (1975) give temperatures of about 1220 oC, but that of Sun et al. (1974) gives temperature of only about 1140 oC . Temperatures can be calculated also using the molar fraction distribution coefficients according to Sun et al. (op.cit.), but these also give rather low values, about 1110 to 1125 oC. This discrepancy in Sr temperatures was recognized also by Sun et al. (1974), and they regard it to be mainly due to the effect of the bulk composition on the Sr distribution.

Equilibrium constant Ksr was calculated for reaction:

SrAl204(l)+2Si02(l)<--->Al2Si208(plg)

Using this K value and assuming simple mixing of complex melt components temperatures based on equation of Drake and Weill (1975) range from 1270 to 1310 oC. They are so in reasonable agreement, even though on high side,and comparable with the temperatures calculated with other geothermometers. The compound equilibrium constant KSr/Ca for the exchange reaction:

(CaAl2Si208)plg+(SrAl204)l<--->(SrAl2Si208)plg+(CaAl204)l gives far too low temperatures, from about 1000 to 1150 oC. Also the compound distribution coefficient Dsr/ca gives similarly low temperatures. The temperatures calculated for Dca and Kca values defined in the same manner as the corresponding values for Sr (Drake and Weill 1975) are all close to and above 1400 oC and so obviously unrealistic. The results from the temperature estimates are summarized in table 7.

The use of Sr distribution between plagioclase and liquid to estimate temperature results therefore in great discrepancies between the (compound) equilibrium constants, the molar distribution coefficients and the straight forward distribution coefficient. This shows that the compositional effects can not be disregarded. Involving many compositional parameters does not seem to improve the results and the discrepancy is more pronounced for the major elements than for Sr. The simple weight ratio Sr distribution coefficient can be used as a geothermometer when the caution is taken not to extrapolate the temperature dependence of the Sr partitioning to the rocks of very different composition.

TABLE 7: Temperature estimates with different geothermometers

	Geothermometer	Temperature	Reference
1	olivine (Mg)	1210-1220	Roeder and Emslie, 1970
2	olivine (Fe)	1243-1264	-"-
3*	clinopyroxene	1210	Nielsen and Drake, 1979
4*	clinopyroxene	1248-1261	"-"
5*	clinopyroxene	1291-1301	"-"
6*	clinopyroxene	1270-1280	"-"
7	plagioclase	1380-1345	Kudo and Weil, 1970
8	plagioclase	1329-1369	Mathez, 1973
9	plagioclase	1342-1379	Drake, 1976
10	plagioclase	1243-1291	Grönvold, 1984
11	plagioclase(Dsr)	1220	Drake, 1972
12	plagioclase(Ksr)	1271-130	Drake and Weill, 1975
13	plagioclase(Dsr)	1138	Sun et al., 1974
14*	plagioclase(Ksr)	1106-1120	"-"
15	plagioclase(Ksr/ca)	989-1153	Drake and Weill,1975
16	plagioclase(Dca)	1377-1505	"-"
17	plagioclase(Kca)	1382-1439	"-"
18	olivine/clinopyroxene	1230-1289	Hakli and Wright, 1967

3	Calculated from the liquid composition
4	Modified Bottinga-Weill mixing model of melt components
5	Simple oxide mixing model of melt components
6	Same as 5, but activities for clinopyroxene components are approximated with their molar fractions
14	K denotes molar fraction distribution coefficient

See text for further explanations.

6.5. Summary

The inconsistency of the calculated plagioclase temperatures reflects the role of other factors (P, PH2O, fO2, composition) which probably have an effect on the crystallization of plagioclase. These have not been adequately counted for in the equations of Kudo and Weill (1970), Mathez (1973) and Drake (1976). It seems that the major element composition of plagioclase alone should not be used as a geothermometer, but only in conjunction with independent temperature estimates as an affirmative parameter.

The plagioclase geothermometer tends to give somewhat higher temperatures than the rest, a fact which has been noted also by others. On the other hand the Dsr temperature coincides with the lowest olivine temperatures and is close to those clinopyroxene temperatures where the liquid composition was used in calculations. When clinopyroxene analyses are corrected for FeO, the calculated temperatures are higher. Even though the clinopyroxene composition generally speaking is very homogeneous, there is some variation (of 1-2 w%) in the FeOt content. In the composition sensitive temperature calculations this is bound to cause some scatter.

The temperatures obtained with different geothermometers seem to show an internal agreement accepting the difference of few tens of degrees in the calculated temperatures. This shows that the crystallization of the different mineral phases has taken place within a limited temperature range. The liquidus temperatures obtained are close to 1250 oC.

Higher crystallization temperatures than 1250 oC may be inferred by comparing Dmn values for olivines from this work with those experimentally determined by Makipaa (1980). Within temperature interval 1125-1250 oC Makipaa reported Dmn values consistently higher than 1. As Dmn decreases with increasing temperature the negative Dmn values in some of the Miðfell samples might indicate crystallization at temperatures above this upper limit.

The fundamental question is whether the phenocrysts are in equilibrium with the melt. Comparison with experimental data unequivocally shows that the composition of the minerals corresponds to equilibrium composition. The homogeneous composition of the minerals and the negligible compositional zoning indicate this, too. But a composition corresponding to equilibrium composition does not necessarily imply that the phenocrysts crystallized from this particular magma, as they could have crystallized also from another, but very similar type of magma.

One of the purposes of the trace element work was to see, if the trace element composition can be used as a test for equilibrium by comparing distribution coefficients (D) and equilibrium constants (K) to those experimentally determined. If equilibrium between the liquid and the mineral phase concerned or between two coexisting mineral phases can be ascertained, trace element composition can be used as a reliable geothermometer, since large amount of experimental data is available. A good review of the literature on the subject is given by Irving (1978). In the case of Miðfell this proved, however, impossible or at least dubious because of the large scatter in trace element contents. It is obvious that some of the olivines are not in equilibrium with confining liquid. With such wide ranges there are always some minerals with D values that agree with the experimental values. To attain reliable and comparable equilibrium D:s experimental work on with these particular rocks would be necessary. Trace elements can, however, be used to infer fractionation history from the olivines.

7. DISCUSSION

When compared with the available data on gabbroic nodules from Iceland the Miðfell nodules show some very distinctive features. The mineral assemblage is limited to only clinopyroxene, plagioclase, olivine and probably spinel. The nodule minerals show very limited variation in major element mineral composition. They are more primitive with much higher content of both An and Fo in plagioclase and in olivine respectively (fig. 8., 9.). They also contain Cr-diopside with with a restricted composition instead of the compositionally variable augite usually found (fig. 10.). These nodules therefore appear to be of a more primitive nature than the other nodules previously studied from tholeiitic and olivine tholeiitic lava suites. The mineral assemblage in the Miðfell nodules does not correspond to a low pressure crystallization (Bender et al., 1978; Presnall et al., 1979; Fisk et al., 1980; Olafsson, pers. comm.). Furthermore the Miðfell clinopyrexene is a chromian diopside, which has been reported only as a high pressure mineral phase (Arculus, 1975; Donaldsson and Brown, 1977; Eggler et al., 1979; Wass, 1979; Imsland, 1984).

The main difference between the mineralogy of the host basalt and the nodules is the common occurrence of chromian spinel in the host rock and it's almost total absence in the nodules. The general appearance of clinopyroxene and plagioclase and also of some olivine megacrysts in the lava suggests that they were not crystallizing during the eruption but disintegrating or reacting with the liquid.

The close similarity of these to the nodule minerals suggests genetic relationship or at least that they are derived from a similar magma. On basis of the appearance and identical chemical composition the glomerocrysts and large clinopyroxene and plagioclase megacrysts as well as some olivine megacrysts are taken to be xenocrysts which represent pieces of more fully disintegrated nodules.

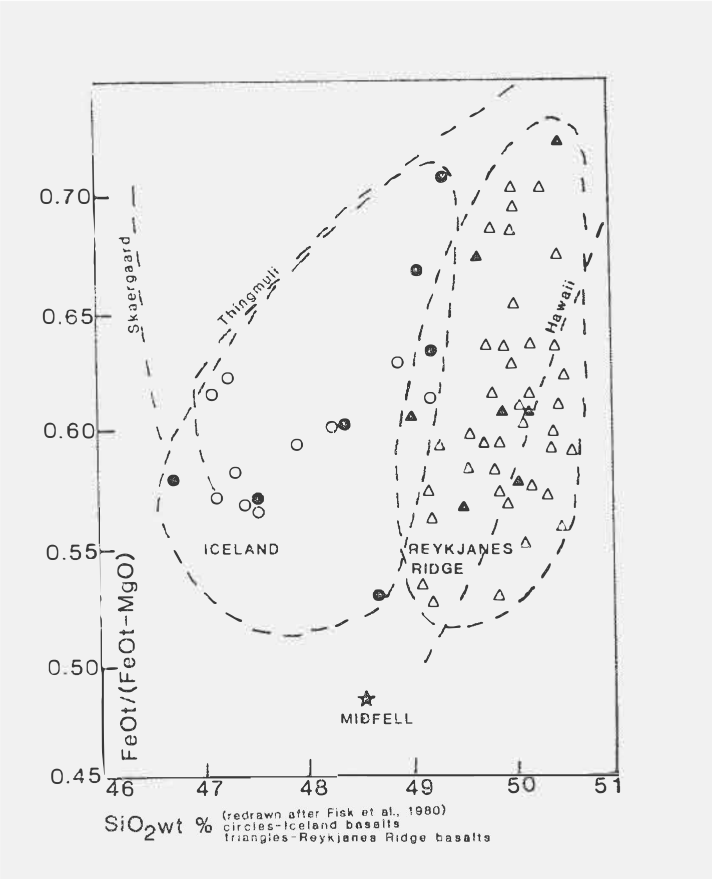
The composition of spinel is highly variable. There is a slight, but consistent difference in spinels included in clinopyroxene and other type of spinels (fig. 11., 12., 13.). This difference can be due to a slightly different evolutionary stage between the two groups of spinels.

The temperature estimates based on calculations with equilibrium compositions give an approximate minimum temperature for the formation of the nodules of about 1250 oC. The geothermal gradient in Reykjanes area is about 100 oC/km for the crust and 1 oC/km for the mantle (Beblo and Björnsson, 1980). Estimated thickness of the crust is about 8-9 km (Palmasson and Sæmundsson, 1974). This indicates that the nodules have been formed at the approximate minimum depth of crust mantle boundary or in the upper mantle.

The glass analyses (table 5, fig. 14. and 15.) show that the liquid that carried the nodules to the surface is most magnesian glass found in Iceland. It belongs to those rare of oceanic glasses analyzed that have MgO content greater than 9 w%. Low degree of fractionation is indicated by the extremely low content of incompatible elements. The trace element content of the glass is similar to that found in the least fractionated ocean-floor basaltic glasses. Ni content in the glass is about half of the estimated average for primary oceanic magmas (Sato, 1977). These features can, of course, partly reflect the original source composition and degree of partial melting.

The high Al_2O_3 and CaO content in the Miðdfell glass are naturally explained by an early enrichment when only olivine along with spinel was fractionated. Comparison with the mineral phases in other picrites confirms this. The ascending host magma is thus a picritic liquid which only has reached the stage of fractionating olivine and spinel, and continues to do so when it comes in contact with olivine gabbro yielding the nodules. It represents primitive melt from the mantle as has been suggested to be the case for Icelandic picrites and rift zone rocks in general (Jakobsson et al., 1978; Maalöe and Jakobsson, 1980; Oskarsson et al., 1982; Harðard´ottir, 1984). The primitive liquid giving rise to the Miðfell picrite is accordingly also of mantle derivation and is "primary", at least on a local scale. Comparison with the Icelandic tholeiites suggests even, that the Miðfell samples represent the mantle derived parent, from which the whole range of Icelandic tholeiites have been derived and which has not been sampled before (fig. 15., 18.).

The nodules represent olivine-clinopyroxene-plagioclase-(spinel) equilibrium stage. They are not cumulates from the host. But, as judged from the trace element contents, nodule and phenocrystic olivine have crystallized from very similar type of magma. comagmatic, xenocrystic origin seems thus most plausible for the nodules, and they represent crystallization products within the stability field of plagioclase in the upper mantle.



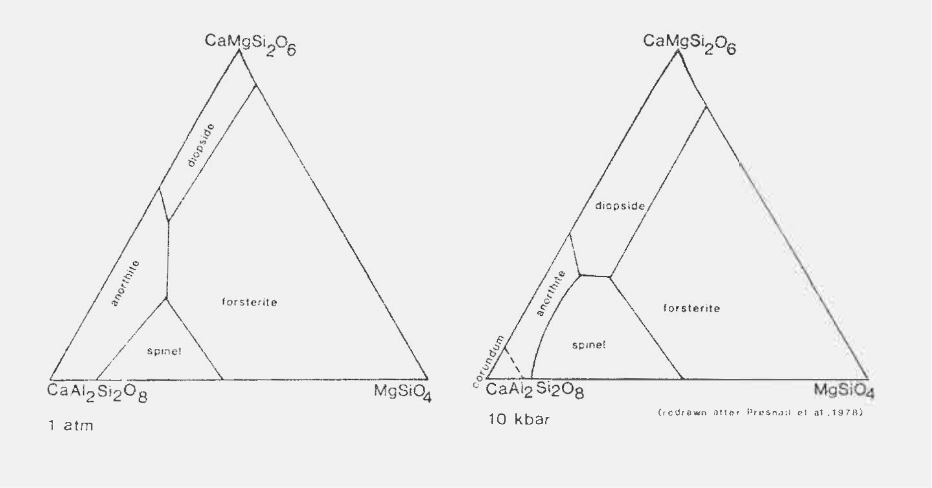
(redrawn after Fisk et al., 1980)
circles-Iceland basalts
triangles-Reykjanes Ridge basalts

Figure 18. FeOt/(FeOt+MgO) versus SiO2 for Icelandic and Reykjanes Ridge basalts and for the Miðfell samples. Fractionation trends of Skaergaard intrusion (Wager, 1960),Thingmuli (Carmichael, 1964) and Kilauea volcano (Tilley, 1960) are shown for comparison.

The upper pressure limit for the formation of the nodules is set by presence of plagioclase. According to Green and Hibberson (1970) and to Fisk (1978) this can not greatly exceed 7-8 kb. The partial H2O pressure, the effect of other volatiles and of the minor element contents within the minerals, which are unknown in this case, might have unpredictable consequences, though (Presnall et al., 1978; Biggar, 1983). Presnall et al. (1978) have demonstrated that olivine, plagioclase, clinopyroxene and spinel can coexist only within rather narrow pressure limits between 5-10 kb. The fine stochiometry exhibited by the spinels (fig. 13.) is also likely to indicate a physico-chemically strictly controlled environment. Within this pressure range the crystallization proceeds along different lines from those at atmospheric pressures: olivine and spinel, spinel and clinopyroxene and clinopyroxene and plagioclase instead of olivine and spinel, olivine and plagioclase and plagioclase and clinopyroxene (fig. 19.).

The highest Ni contents in Miðfell olivines are close to the highest values reported in the olivines from ocean-floor basalts (Muir and Tilley, 1964). The whole range in the Miðfell samples is similar to that of olivines from basal cumulus dunite, wehrlitic rocks interlayered with cumulus gabbro and in some of the olivines within the cumulus gabbro itself described from the Samail ophiolite complex. These olivines with high Ni contents represent the earliest crystallization products from primary magmas after separation from mantle peridotite (Sato, 1977; Pallister and Hopson, 1981). Because of the high Dni value for olivine/liquid Ni is rapidly concentrated in the first forming olivines. According to Sato (op.cit.) only 6-12 % fractionation of olivine depletes Ni to this point.

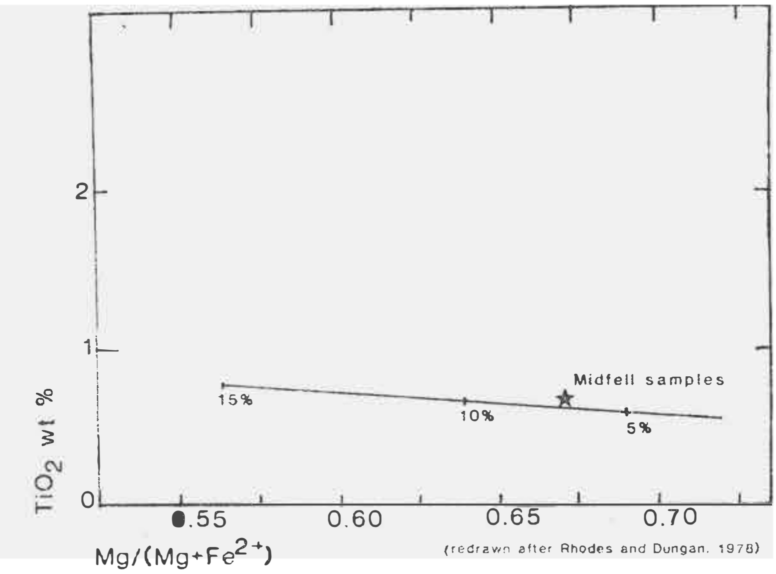
Fractionation not much exceeding 5% is also indicated when TiO2 is plotted against Mg/(Mg+Fe2+) (fig. 20).



(redrawn after Presnall et al.,1978)

Figure 19. Phase diagrams showing crystallization paths at 1 kb and 10 kb pressures. See text for further explanations.

Cr content in some of the Miðfell olivines is far greater than reported in olivines e.g. from the Samail ophiolite sequence (Pallister and Hopson, 1981). This may be partly due to contamination with minute undetected chromian spinel inclusions. Mainly it indicates, however, that clinopyroxene was not fractionating when the first olivines crystallized. The drop in the Cr content in olivines marks the beginning crystallization of chromian spinel and later that of chromian diopside.



(redrawn after Rhodes and Dungan, 1978)

Figure 20. Weight% TiO2 versus Mg/(Mg+Fe2+) in the glass indicating degree of olivine fractionation.

The ranges in Dni and Dcr values of the host rock olivines as well as the the ranges of anorthite and forsterite contents in some of the host's phenocrysts extend slightly higher than those in the nodule minerals (fig. 8., 9., table 6.). This small difference might be of importance as it indicates that some of the host minerals have crystallized from a magma with a slightly more primitive composition. This crystallization has also taken place in a pressure environment, where olivine-spinel-plagioclase cotecticum was reached (fig. 19.).

The nodule crystallization has proceeded in the main along clinopyroxene- plagioclase cotectic. However, olivine, plagioclase and spinel are found as inclusions in clinopyroxene, and olivine is a minor constituent of the nodule mineral assemblage. Accordingly the crystallization must have taken place very close to, if not exactly, at olivine-spinel-clinopyroxene-plagioclase invariant point.

The close similarity in the chemical composition of the nodule and the host rock minerals with extensively overlapping ranges can be explained only by crystallization in open magma systems within rather limited pressure interval. Extremely limited chemical variation and the negligible compositional zoning, often reversed when detected, of the minerals are a further evidence of this, as is the pronounced oscillatory zoning in plagioclases.

A plausible mechanism for the formation of the nodules is as follows: The primary liquid after separation from the mantle source, fractionates olivine and chromian spinel. This fractionation brings the melt composition via olivine-spinel-clinopyroxene equilibrium to the clinopyroxene-plagioclase cotectic. Some of the last crystallizing spinels are trapped as inclusions in clinopyroxene thus explaining the observed compositional trend when different types of spinels are compared. Repeated replenishment and mixing with primitive magma from the depth prevents the melt from evolving towards more fractionated compositions. Instead, influxes of new primitive magma drive melt composition back and forth across phase boundaries, occasionally even causing a reversed crystallization order. Crystal aggregates formed from the first batches i.e. the nodules, consist of plagioclase, clinopyroxene and olivine. They were plastered against the walls of the magma conduits, where they eventually form olivine gabbro dikes cutting the mantle peridotite. The signs of mechanical stress in some of host's megacrystic olivines imply that the rock from which these olivines were derived once was in a solid state. Most likely they belong to the nodule assemblage and are derived from the same olivine gabbro dikes. New batches of primitive magma ascend along the same channels towards somewhat higher levels and reach olivine-spinel-plagioclase equilibrium stage. During ascent earlier formed olivine gabbro is incorporated and partly assimilated causing the resorbed appearance of the nodules. During the Miðdfell eruption a more sudden surge of primitive magma was forceful and rapid enough to carry pieces of the dike rock as nodules to the surface.

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(foot note) THE ANALYTICAL PROCEDURE: The analyses were performed at the Nordic Volcanological Institute in Reykjavik with an automatic ARL-SEMQ microprobe. The sample currents used were 25 nA for glass and plagioclase, 75 nA for pyroxene and 90 for olivine and spinel. Natural minerals were used as standards. The results were recalculated according to the correction method by Bence and Albee (1968). Counting time was 10-20 s for the peak and 4 s for the background. For trace elements counting time from 40 to 100 s for the peak and 40 s for background were used. The sample currents were similar. Higher accelerating potential of 25 kilo volts did not improve precision and trace element analyses were also performed mainly with an accelerating potential of 15 kilo volts.

(foot note) TERMINOLOGY: To avoid too hasty genetic inferences I have used throughout the work the neutral term "nodule" to designate recognizable mineral aggregates, which in appearance differ from the main mineral assemblage of the host rock.

SVENSK SAMMANFATTNING: GABBRONODULER FRAN EN PIKRITISK PILLOW BASALT, MIÐFELL, SV ISLAND.

En subglacial pikritisk "pillow basalt" från Miðfell, SV Island, innehåller gabbronoduler, som ar av en annan typ än de tidigare från Island undersökta gabbronodulerna. Mineralsällskapet i Miðfell-nodulerna består av plagioklas, klinopyroxen (kromdiopsid) och olivin. Spinell förekommer som en sällsynt inklusion i klinopyroxen eller i olivin.

Den nodulbärande basalten har olivin, spinell, plagioklas och kromdiopsidisk klinopyroxen som makrokristallina mineralfaser. Glaset från de avkylda pillowkanterna är mycket primitivt. Det ar det mest magnesiumrika glaset som påträffats på Island och det innehåller mycket låga koncentrationer av alkalier och inkompatibla element, vilket indikerar en låg fraktioneringsgrad.

Den kemiska sammansättningen av olivin, plagioklas och klinopyroxen ar mycket begränsad och primitiv. Det finns inte någon betydande skillnad i sammansättning mellan nodul- och "host rock" -mineralerna. Vardbasaltens glomerokryster, klinopyroxen- och plagioklas megakryster samt några av dess olivinmegakryster tycks vara xenokryster härstammande från de mera fullständigt disintegrerade nodulerna. Detta grundas på det likartade utseendet, den identiska kemiska sammansättningen och att nodulerna i vissa fall kan ses ett tillstånd av disintegration. Den primitiva sammansättningen av glaset och mineralen hänvisar till att de är genetiskt relaterade.

I motsats till de övriga mineralfaserna är spinellernas sammansättning mycket varierande. Trots den vidsträckta överlappningen finns det en svag men konsekvent skillnad mellan spinellinklusionerna i klinopyroxen å ena sidan och spinellinklusionerna i olivin och skilda spinellfenokryster å den andra. Detta förklaras med olika evolutionära stadier av spinellerna.

I fråga om huvudelementen tycks mineralerna befinna sig i jämvikt med vätskan. Temperaturer kalkylerade med olika metoder, utgående från jämviktsammansättningen hos de samexisterande mineral-mineralparen och mineral-vätskeparen ger en liquidustemperatur på över 1250 oC. Betraktande den geotermiska gradienten och skorpans tjocklek i området, betyder detta att nodulerna har bildats i gransen mellan skorpan och manteln eller i den övre manteln.

Spårelementsammansättningen uppvisar stor variation. Variationen ar kontinuerlig och ej bimodal. Den är också väsentligen den samma för både nodul- och "host rock" -mineralen. Den kontinuerliga variationen indikerar att mineralerna har bildats genom en kumuluskristallisation och att i fråga om spårelementen en del kumuluskristaller inte ar i jämvikt med vätskan.

Mineralsällskapet i nodulerna representerar inte ett lågtryckssällskap. Den övre tryckgränsen för uppkomsten av nodulerna bestäms av plagioklasens narvaro till ca 5-10 kb, motsvarande ett djup av 15-30 km. I detta tryckområde följer kristallisationen en annan vag än vid atmosfäriskt tryck: olivin-spinell, spinell-klinopyroxen och klinopyroxen-plagioklas i stället för olivin-spinell spinell-plagioklas och plagioklas-klinopyroxen.

Den påfallande likheten i den kemiska sammansättningen hos nodul- och "host rock" -mineralen, den ytterst begränsade kemiska variationen och den obetydliga kemiska zoneringen förklaras genom en kristallisation i ett öppet magmasystem inom ett begränsat tryckområde.

En sannolik mekanism för uppkomsten av nodulerna ar: Den uppstigande primära vätskan fraktionerar, efter separering från mantelkällan, olivin och spinell tills vätskesammansätttningen når plagioklas-klinopyroxen -kotektikum. Förnyelse och blandning med nya primitiva magmapulser från djupet hindrar en vidare utveckling av vätskan. I stället håller sig vätskesammansättningen i närheten av invarianta punkten för klinopyroxen-plagioklas-olivin-spinell. Tillfälligtvis kan tillflöde av primitiv magma även förorsaka en omvänd kristallisationsordning. De först bildade mineralaggregaten bildar olivingabbrogångar i magmakanalerna i övre manteln. Senare pulser av primitiv magma använder samma kanaler vid uppstigning till en något högre niva, de når olivin-plagioklas-spinell -kotektikum. Härvid assimileras den tidigare bildade olivingabbron delvis, vilket förorsakar det resorberade utseendet hos nodulerna. Miðfell -eruptionen var tillräckligt kraftig och snabb för att kunna bära stycken av gångbergarten som noduler till ytan.

REFERENCES

Arculus, R.J., 1975: Melting behaviour of two basanites in the range 10-35 kbar and the effect of TiO2 on the olivine-diopside reactions at high pressures. Carnagie Inst. Wash. Year Book 74, pp 512-514.

Aumento, F., 1968: The Mid-Atlantic Ridge near 45 o N, II. Basalts from the area of Confederation Peak. Can. J. Earth Sci. 5,1.

Beblo, M. and Björnsson, A. 1980: A model of electrical resistivity beneath NE-Iceland, correlation with temperature. J.Geophys. 47:184-190.

Bence, A.E. and Albee, A.L., 1968: Empirical correction factors for the electron microanalysis of silicates and oxides. J. Geol. 76:382-403.

Bender, J.F., Hodges, F.N. and Bence, A.E., 1978: Petrogenesis of basalts from the project FAMOUS area: anexperimental study from 0 to 15 kbars. Earth Planet.Sci.Lett. 41:277-302.

Bennell, M.J., 1969: Ultrabasic and basic nodules from the Azores. Proc. Geol.Soc.Lond. 1658:252-253.

Biggar, G. M., 1983: Crystallization of plagioclase, augite and olivine in synthetic systems and in tholeiites. Min. Magaz. 47:161-176.

Blanchard, D.P., Rhodes, J.M., Duncan, M.A., Rodgers, K.V., Donaldson, C.H., Brannon, J.C., Jacobs, J.W., and Gibson, E.K. 1976: The chemistry and petrology of basalts from Leg 37 of the Deep Sea Drilling Project. J. Geophys. Res. 81:4231-4246.

Bottinga, Y. and Weill D.F., 1972: The viscosity of magmatic silicate liquids: a model for calculation. Amer.J.Sci. 272:438-475.

Bougault, H. and Hekinian, R., 1974: Rift valley in the Atlantic Ocean near 36o50'N: petrology and geochemistry of basaltic rocks. Earth Planet.Sci.Lett.24:249-261.

Bougault, H., Cambon, P., Corre, O., Joron, J,L. and Treuil, M., 1979: Evidence for variability of magmatic processes and upper mantle heterogeneity in the axial region of the Mid-Atlantic Ridge near 22oN and 36oN. Tectonophys. 55:11-34.

Boyd, F.R. and Meyer, H.O.A.,(ed.) 1979: The mantle sample: inclusions in kimberlites and other volcanics. Litho-Crafts Inc., Chelsea, Michigan, pp 423.

Bowen, N.L. (1928, 1. ed.)1956: The evolution of the igneous rocks. Dover Publications, Inc., New York, pp 332.

Bryan, W.B., 1979: Regional variation and petrogenesis of basalt glasses from the FAMOUS area, Mid-Atlantic Ridge. J. Petrol. 20:293-325.

Carmichael, I.S.E., 1964: The petrology of Thingmuli volcano, a Tertiary volcano in eastern Iceland. J.Petrol. 5:435-460.

Daly, R.A., 1965: The geology of Ascension Island. Proc.Am.Acad.Arts & Sci. 60:1-80.

Donaldson, C.H. and Brown, R.W., 1977: Refractory megacrysts and magnesium-rich melt inclusions within spinel in oceanic tholeiites: indicators of magma mixing and parental magma composition. Earth Planet. Sci. Lett. 37:81-89.

Drake, M.J., 1972: The distribution of major and trace elements between plagioclase feldspar and magmatic silicate liquid: an experimental study. Ph.D. dissertation, University of Oregon, 190 pp.

Drake, M.J., 1975: Plagioclase thermometry and barometry in igneous systems. In: Extended abstracts: International Conference on Geothermometry and Geobarometry, Pennsylvania State University, Pennsylvania.

Drake, M.J., 1976: Plagioclase-melt equilibria. Geochim. Cosmochim. Acta 40:457-465.

Drake, M.J. and Weill, D.F., 1975: The partition of Sr, Ba, Ca, Y, Eu2+, Eu3+ and other REE between plagioclase feldspar and magmatic silicate liquid: an experimental study. Geochim.Cosmochim.Acta 39:689-712.

Duke, J.M., 1976: Distribution of the period four transition elements among olivine, calcic clinopyroxene and mafic silicate liquid: experimental results. J.Petrol. 17: 499-521.

Dungan, M.A. and Rhodes, J.M., 1978: Residual glasses and melt inclusions in basalts from DSDP Legs 45 and 46: Evidence for magma mixing. Contrib.Mineral. Petrol. 67:417-431.

Eggler, D.H., McCallum, M.E., and Smith, C.B., 1979: Megacryst assemblages in kimberlite from northern Colorado and southern Wyoming: petrology, geothermometry-barometry, and areal distribution. In " The Mantle Sample: Inclusions in Kimberlites and other Volcanics (Boyd, F.R. and Meyer, H.O.A., ed.), LithoCrafts Inc., Chelsea, Michigan, pp213-226.

Elthon, D., 1979: High magnesia liquids as the parental magma for ocean floor basalts. Nature 278,5:514-518.

Elthon, D. and Scarfe, C.M., 1980: High-pressure face equilibria of a high-magnesia basalt: implications for the origin of mid-ocean ridge basalts. Carnagie Inst. Wash. Yearbook 79:277-281.

Elthon, D. and Scarfe, C.M., 1984: High-pressure phase equilibria of a high-magnesia basalt and the genesis of primary oceanic basalts. Am. Mineral.69:1-15.

Finger, L. W., 1972: The uncertainty in the calculated ferric iron content of a microprobe analysis. Carnagie Inst. Wash. Year Book 71:600-603.

Fisk, M.R., 1978: Melting relations and mineral chemistry of Iceland and Reykjanes Ridge basalts. University of Rhode Island, Ph.D. thesis.

Fisk, M.R., Shilling, J.-G. and Sigurdsson, H. 1980: An experimental investigation of Iceland and Reykjanes Ridge tholeiites: I. Phase relations. Contrib.Min.Petrol. 74:361-374.

Frey, F.A., Bryan, W.B. and Thompson, G. 1974: Atlantic ocean floor, geochemistry and petrology of basalts from Leg 2 and 3 of the Deep Sea Drilling Project. J.Geophys.Res.:79:5507-5527.

Fuji, T. and Bougault, H., 1983: Melting relations of a magnesian abyssal tholeiite and the origin of MORBs. Earth Planet.Sci.Lett.62:283-295.

Green, D.H. and Hibberson, W., 1970: The instability of plagioclase in peridotite at high pressure. Lithos 3:209-221.

Green, D.H., Hibberson, W.O. and Jaques, A.L., 1979:Petrogenesis of mid-ocean ridge basalts. In: "The Earth: Its Origin, Structure and Evolution", McElhinney, M.W.(ed.), Academic Press, London, pp 265-299.

Green, D.H. and Ringwood, A.E., 1967: The genesis of basaltic magmas. Contrib. Mineral.Petrol. 15:103-190.

Grönvold, K., 1984: Myvatn fires 1724-1729. Chemical composition of the lava. Nordic Volcanological Institute 8401. University of Iceland.

Grönvold, K. and Makipaa, H., 1978: Chemical composition of Krafla lavas 1975-1977. Nordic Volcanological Institute 78, 16. University of Iceland.

Hakli, T.A. and Wright, T.L., 1967:The fractionation of nickel between olivine and augite as a geothermometer. Geochim.Cosmocim. Acta 31:877-884.

Harðard´ottir, V., 1984: The petrology of the Hengill volcanic system, southern Iceland. McGill University, M.sc. thesis (in prep.)

Hawkesworth, C.J. and Norry, M.J.(ed.) 1983: Continental basalts and mantle xenoliths. Shiva Publishing Ltd., 272 p.

Höj, J., 1984: Gabbro-anorthosite nodules from Krisuvik fissure system and active zone of rifting, Reykjanes peninsula, SW. Iceland. Nordic Volcanological Institute 840 . University of Iceland.

Imsland, P., 1984: The petrology, mineralogy and evolution of the Jan Mayen magma system. V´sindaf´elag Islendica 43, 322 pp. Prentsmiðjan Oddi, Reykjavik.

Irving, A.J., 1978: A review of experimental studies of crystal/liquid trace element partitioning. Geochim.Cosmochim.Acta 42:743-770.

Jackson, E.D. and Wright, T.L., 1970: Xenoliths in the Honolulu volcanic series, Hawaii. J.Petrol. 11:405-430.

Jacques, A.L. and Green, D.H., 1980: Anhydrous melting of peridotite at 0-15 kb pressure and the genesis of tholeiitic basalts. Contrib.Mineral.Petrol. 73:287-310.

Jakobsson, S., 1966: The Grimsnes lavas, SW-Iceland. Acta Naturalia Islandica 2:6.

Jakobsson, S.P., 1979: Outline of the petrology of Iceland. Jökull 29:57-73.

Jakobsson, S.P., 1979: Petrology of recent basalts in the eastern volcanic zone, Iceland. Acta Naturalia Islandica 26.

Jakobsson, S.P., 1983: Islenskar bergtegundir I, Pikr´it (´ocean´it). N´att´urufræðingurinn 52:80-85.

Jakobsson, S.P., Pedersen, A.K., Rönsbo, J.G. & Larsen, L.M., 1973: Petrology of mugearite-hawaiite: Early extrusives in the 1973 Heimaey eruption, Iceland. Lithos 6:203-14.

Jakobsson, S.P., J´onsson, J. and Shido, F., 1978: Petrology of the Western Reykjanes Peninsula, Iceland. J. Petrol. 19:669-705.

J´onsson, J., 1963: Hnydlingar i islenzku bergi. N´att´urufræðingurinn 33,1:9-20.

J´onsson, J., 1978: Jarðfræðikort Reykjanesskaga. Report Orkustofnum OS 7HD 7831.

Kristmannsd´ottir, H., 1971: Anorthosite inclusions in tertiary dolerite from the island groups Hrappsey and Purkey, West Iceland. J. Geol. 79,741-748.

Kudo, A.M. and Weill, D.F., 1970: An igneous plagioclase thermometer. Contrib. Mineral.Petrol. 25:52-65.

Kushiro, I., 1973: Origin of some magmas in oceanic and circum-oceanic regions. Tectonophysics 17:211-222.

Langmuir, C.H., Bender, J.F., Bence, A.E. and Hanson, G.N., 1977: Petrogenesis of basalts from the FAMOUS area: Mid-Atlantic Ridge. Earth planet.Sci.Lett. 36:133-156.

Larsen, J.G., 1979: Glass-bearing gabbro inclusions in hyaloclastites from Tindfjallajökull, Iceland. Lithos 12, 289-302.

Maalöe, S. and Jakobsson, S.P., 1980: The PT phase relations of a primary oceanite from the Reykjanes peninsula, Iceland. Lithos 13:237-346.

MacDonald, G.A. and Katsura T., 1964: Chemical composition of Hawaiian lavas. J.Petrol. 5:82-133.

Makipaa, H.T., 1978: Mineral equilibria, geothermometers and geobarometers in some Icelandic hyaloclastites. Bull.Geol. Soc. Finland 50:113-134.

Makipaa, H., 1978a: Petrological relations in some Icelandic basaltic hyaloclastites. Bull. Geol. Soc. Finland 50:81-112.

Makipaa, H.T., 1980: Partitioning of chromium, manganese, cobalt and nickel between olivine and basaltic liquid: an experimental study. Bull.Geol.Soc. Finland 52,2:175-191.

Mathews, W.H., þorarinsson, S. and Church, N.B., 1964: Gravitative settling of olivine in pillows of an Icelandic basalt. Am. J. Sci. 262:1036-1040.

Mathez, E.A., 1973: Refinement of the Kudo-Weill plagioclase thermometer and its application to basaltic rocks. Contrib. Mineral. Petrol. 41:61-72.

Melson, W.G., Thompson, G. and Van Andel, T.H., 1968: Volcanism and metamorphism in the Mid-Atlantic Ridge, 22oN latitude. J.Geophys.Res. 73:5925-5941.

Melson, W.G. and Thompson, G., 1971: Petrology of a transform fault zone and adjacent ridge segments. Philos.Trans.Roy.Soc.Lond, A268:423-441.

Miyashiro, A., Shido, F. and Ewing, M., 1969: Diversity and origin of abyssal tholeiite from the Mid-Atlantic Ridge near 24o and 30o north latitude. Contrib.Mineral.Petrol. 23:38-

Muir, I.D. and Tilley, C.E., 1964: Basalts from the northern part of the rift zone of the Mid-Atlantic Ridge. J.Petrol. 5:409-434.

Muir, I.D. and Tilley, C.E., 1966: Basalts from the northern part of the Mid-Atlantic Ridge, II. The Atlantis collection near 30oN. J.Petrol. 7:193-201.

Nicholls, G.D., Nawalk, A.J. and Hays, E.E., 1964: The nature and composition of rock samples dredged from the Mid-Atlantic ridge between 22oN and 52oN. Mar.Geol. 1:333-

Nielsen,R.L. and Drake, M.J., 1979: Pyroxene-melt equilibria. Geochim. Cosmochim. Acta 43:1259-1272.

O'Hara, M.J., 1968: The bearing of phase equilibria studies in synthetic and natural systems on the origin and evolution of basic and ultrabasic rocks. Earth Sci.Rev. 4:69-133.

O'Hara, M.J., 1968a: Are ocean floor basalts primary magma? nature 220:683-686.

O'Hara M.J., Saunders, M.J. and Mercy, E.L.P., 1975: Garnet-peridotite, primary magma and eclogite: interpretation of upper mantle processes in kimberlite. Physics and Chemistry of the Earth, 9:571-604.

Oskarsson, N., Sigvaldason, G.E. and SteinÞ´orsson, S., 1982: A dynamic model of rift zone petrogenesis and the regional petrology of Iceland. J. Petrol. 22:28-74.

Pallister, J.S. and Hopson, C.A., 1981: Samail ophiolite suite: Field relations, phase variation, cryptic variation and layering, and a model of a spreading ridge magma chamber. J.Geophys.Res. 86:2593-2644.

Palmasson, G. and Sæmundsson, K. 1974: Iceland in relation to the Mid-Atlantic Ridge. In:"Annual Reviews of Earth planet.Sci. 2:25-50". (ed. F.A. Donath)

Pedersen, A.K. and Hald, N., 1982: A cummingtonite-porphyritic dacite with amphibole-rich xenoliths from the Tertiary central volcano at Krofjo´rdur, NW Iceland. Lithos 15:137-159.

Presnall,D.C., Dixon, S.A., Dixon, J.R., O'Donnell, T.H., Brenner, N.L., Schrock, R.L. and Dycus, D.W., 1978: Liquidus phase relations on the join diopside-forsterite-anorthite from 1 atm to 20 kbar: their bearing on the generation and crystallization of basaltic magma. Contrib. Mineral. Petrol. 66:203-220.

Presnall, D.C., Dixon, J.R., O'Donnell, T.H. and Dixon, S.A., 1979: Generation of mid-ocean ridge tholeiites. J.Petrol. 20:3-35.

Ragnarsd´ottir, K.V., Bishop, F.C. and Jakobsson, S.P., 1982: Crystallization processes in a shallow magma chamber beneath Seyðish´olar, Grimsnes, Iceland. (Abs.). In "Abstracts Generation of major basalt types". IAVCEI-IAGC Scientific Assembly. Reykjavik.

Rhodes, J.M., Dungan, M.A., Blanchard, D.P. and Long, P.E., 1979: Magma mixing at mid-ocean ridges: evidence from basalts drilled near 22oN on the Mid-Atlantic Ridge. Tectonophys. 55:35-62.

Roeder, P.L. and Emslie, R.F., 1970: Olivine-liquid equilibrium. Contrib. Mineral. Petrol. 29:275-289.

Sato, H., 1977: Nickel content of basaltic magmas: identification of primary magmas and a measure of the degree of olivine fractionation. Lithos 10:113-120.

Seward, F.M., 1971: The distribution of transition elements in the system CaMgSi2O6-Na2Si2O5-H2O at 1000 bars pressure. Chem.Geol. 7:73.95.

Sigurdsson, H. and Shilling, J-G., 1976: Spinels in Mid-Atlantic Ridge basalts: Chemistry and occurrence. Earth Planet.Sci.Lett. 29:7-20.

Smewing, J.D., 1981: Mixing characteristics and compositional differences in mantle-derived melts beneath spreading axes: Evidence from cyclically layered rocks in the ophiolite of North Oman. J.Geophys.Res. 86:2645-2659.

Sobolev, N.V., 1977: Deep-seated inclusions in kimberlites and the problem of the composition of the upper mantle. Edwards Brothers Inc., Ann Arbor, Michigan, p 279.

Sparks, R.S.J., Meyer, P. and Sigurdsson, H., 1979: Density variation amongst mid-ocean ridge basalts: Implications for magma mixing and the scaricity of primitive lavas. Earth Planet.Sci.Lett. 46:419-430.

Stolper, E., 1980: A phase diagram for mid-ocean ridge basalts: preliminary results and implications for petrogenesis. Contrib.Mineral.Petrol. 74:13-27.

Stolper, E. and Walker, D., 1980: Melt density and the average composition of basalt. Contrib.Mineral.Petrol. 74:7-12.

Sun, C-O., Williams, R.J. and Sun, S-S. 1974: Distribution coefficients of Eu and Sr for plagioclase-liquid and clinopyroxene-liquid equilibria in oceanic ridge basalt: an experimental study. Geochim.Cosmochim.Acta 38:1415-1433.

Sutherland, F.L., 1974: High-pressure inclusions in tholeiitic basalt and the range of lherzolite-bearing magmas in the Tasmanian volcanic province. Earth Planet.Sci.Lett. 24:317-324.

Sæmundsson, K., 1967: Vulkanismus und Tektonik des Hengill-Gebietes in Sudwest-Island. Acta Naturalia Islandica 2,7.

Sæmundsson, K., 1978: Fissure swarms and Central volcanoes of the neovolcanic zones of Iceland. In "Crustal evolution in northwestern Britain and adjacent regions. (Bowes, D.R. and Leake, B.E., ed.). pp 415-432.

Sæmundsson, K., 1979: Outline of the geology of Iceland. Jökull 29:7-28.

Takahashi, E. and Kushiro, I., 1983: Melting of a dry peridotite at high pressures and basalt genesis. Am.Mineral. 68:859-879.

Thompson, R.N. and Tilley, C.E., 1969: Melting and crystallization relations of Kilauean basalts in Hawaii. The lavas of the 1959-60 Kilauea eruption. Earth Planet. Sci. Lett.5:469-477.

Thy, P., 1982: Phase realations in transitional and alkali basaltic glasses from Iceland. Nordic Volcanological Institute 8202.

Tilley, C.E., 1960: Differentiation of Hawaiian basalts: some variants in lava suites of dated Kilauean eruptions. J.Petrol. 1:47-55.

Tryggvason, T., 1957: The gabbro bombs at lake Grænavatn. Bull.Geol.Inst. Uppsala, 30:273-320.

Yoder, H.S., 1976: Generation of basaltic magma. National Academy of Sciences, Washington D.C., 265 p.

þorarinsson, S., 1953: Grænavatn and Gestsstaðavatn. Geografisk Tidskrift 52:292-302.

Wager, L.R., 1960: the major element variation of the layered series of the Skaergaard intrusion and re-estimation of the average composition of the hidden layered series and successive residual mgmas. J.Petrol. 70:111-125.

Wass, S.Y., 1979: Multiple origins of clinopyroxenes in alkali basaltic rocks. Lithos 12:115-132.

White, R.W., 1966: Ultramafic inclusions in basaltic rocks from Hawaii. Contr. Min.Petrol. 12:245-314.