

NORDIC VOLCANOLOGICAL INSTITUTE **8201**  
UNIVERSITY OF ICELAND

NORRÆNA ELDFJALLASTÖÐIN

**SPINEL MINERALS IN TRANSITIONAL  
AND ALKALI BASALTIC GLASSES  
FROM ICELAND**

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**Nordic Volcanological Institute  
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Abstract

Crystallization of spinel minerals in transitional and alkali basalts from Iceland can be related to FeO, MgO, TiO<sub>2</sub> and Cr content of the coexisting melt. Chromian spinel occurs in glasses with TiO<sub>2</sub> and FeO/MgO below 2.8 and 2, while titanomagnetite occurs above 4 and 2.7, respectively (all in wt% oxides). Chromian spinel only occurs in basalts with a chrome content above 200 ppm. It is suggested that non-equilibrium crystallization of chromian spinel occurs, together with crystallization of olivine, from a liquid with a olivine liquidus temperature from above 1200°C down to approximately 1150°C. A discontinuity in spinel crystallization follows until below 1100°C, where titanomagnetite starts to crystallize. Textural relations suggest a homogeneous nucleation for titanomagnetite, whereas chromian spinel nucleates heterogeneously, dependent on growth of olivine phenocrysts. It is suggested that local concentration gradients around growing olivine crystals may change melt structure and the solubility of chromian spinel in the melt, and may be the prime reason for the precipitation of chromian spinels subsequently trapped by growing olivine phenocrysts. Compositional variation in chromian spinel, attached to or included in homogeneous olivine phenocrysts, are also believed to stem from local depletion around nucleation sites and growing crystals.

## Introduction

Spinal minerals crystallize from basaltic liquids over a wide compositional range, and their stability end composition are controlled by temperature, oxygen activity, coexisting phases and composition of the melt. Chromian spinal crystallizes over a more restricted temperature and oxygen activity range than titanomagnetite (Hill & Roeder, 1974). Complete solid solution exists between chromian spinal and titanomagnetite, approximately above the NNO buffer curve, while a hiatus occurs between high temperature chromian spinal and lower temperature titanomagnetite (Hill & Roeder, 1974). The composition of chromian spinal may also be sensitive to pressure (Irvine, 1967; Sigurdsson & Schilling, 1976; Sigurdsson, 1977; Haggerty, 1979; Fisk & Bence, 1980).

A study of spinal minerals in basaltic rocks can therefore provide information on crystallization conditions in magmas. The present study is mainly concerned with compositional variation in coexisting spinal, olivine and liquid in transitional and alkali basalts from Iceland. Some physical and compositional parameters controlling the crystallization of titanomagnetite and chromian spinels are also evaluated. In order to minimize zoning and subsolidus re-equilibration effects, rapidly quenched glassy rocks were used. The present work forms part of an investigation of phase relations in alkali basaltic glasses from Iceland.

## Alkali basalts in Iceland

The regional petrology of the active volcanic zones of Iceland has been reviewed by Jakobsson (1972, 1979), Imsland (1978) and Seemundsson (1979). A model for rift zone petrogenesis has been proposed by Oskarsson et al. (1982). The location of the active volcanic zones is summarized in Fig. 1, based on the interpretation by Jakobsson (1972), Palmeson & Saemundsson (1974) and Imsland (1978), showing transitional and alkali basaltic areas in an off-rift position relative to the main tholeiitic rift zone.

Alkali basalts are found in the westward extending Sneefellsnes Volcanic Zone (Sn.V.Z.) and in the Vestmannaeyjar volcanic area, while tholeiitic and transitional basalts occur in the Sudurland Volcanic Zone (S.V.Z.) (Jakobsson, 1972, 1979; Imsland, 1978). The latter shows a sequence of compositions from alkali basalts to Fe-Ti basalts. The Fe-Ti basalts were referred to by Jakobsson (1979) as belonging to a transitional series between tholeiitic and alkalic basalts. Classification of Icelandic basalts is in general based on alkali-silica relations and normative composition (Jakobsson, 1979). All glasses considered in this study can be distinguished from tholeiites by their higher alkali content. However, only the Vestmannaeyjar basalts are consistently nepheline normative (Thy, 1982). For this reason the basalts are referred to as transitional and alkali basaltic or simply as alkalic (s.l.).

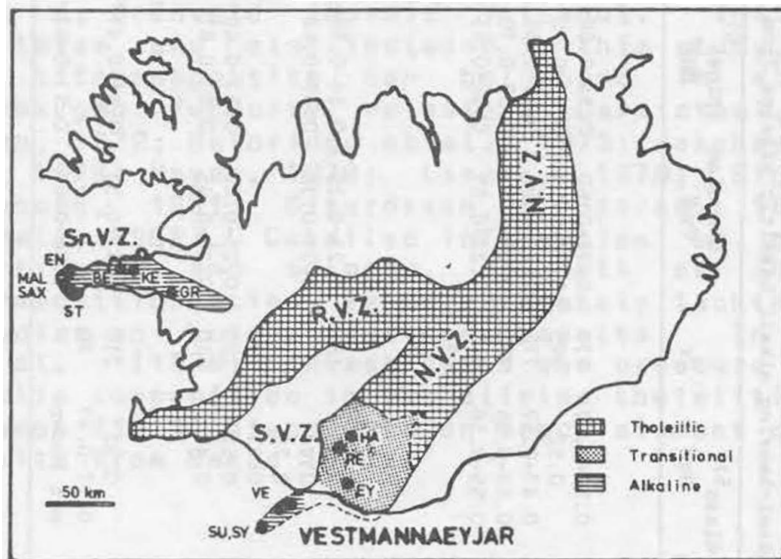


Figure 1. Sample localities and outline of the active volcanic zones of Iceland (Palmason & Saemundsson, 1974). Distribution of rock types after Jakobsson (1972) and Imslund (1978). Sn.V.Z. - Snæfellsnes Volcanic Zone; S.V.Z. - Sudurland Volcanic Zone; R.V.Z. - Reykjanes Volcanic Zone; N.V.Z. - Nordurland Volcanic Zone.

A petrographic and chemical study of phase relations in the glasses is given elsewhere (Thy, 1982) and only the main features will be reviewed here. The phenocryst assemblages comprise plagioclase, olivine, clinopyroxene and spinel minerals. The compositions of phenocrysts and glass show complex relations which can be related to polybaric origin. The glass composition can be related to pressure dependent phase relations in the experimental basalt system, and the silicate phenocryst assemblages also show a polybaric origin. There is a bimodality in olivine end plagioclase composition which is in part related to size distribution (micro- and macrophenocrysts). This bimodality is likely to be related to fractionation during phenocryst crystallization. Relatively high- $F_o$  olivine macrophenocrysts often contain inclusions of chromian spinels. Low- $F_o$  microphenocrysts more rarely contain chromian spinel, and are most likely to have crystallized from the enclosing glass.

TABLE 1. Summary of petrological and chemical characteristics of spinel-bearing alkali basaltic glasses from Iceland

Sample no.	Lithology <sup>1)</sup>	Anal. samples <sup>2)</sup>	Phenocryst <sup>3)</sup> phases				Mole% <sup>4)</sup>		Glass <sup>5)</sup>			Chromian Cr#	spinel <sup>6)</sup> Mg#	Total range <sup>7)</sup> in Fo
			An	Fo	wt. % Al <sub>2</sub> O <sub>3</sub>	Mg#	Cr							
Snaefellsnes Volcanic Zone														
KE	H	6 (80)	ol	pl	chr	71-76	81-87	14.4-15.3	0.52-0.57	570	0.32-0.60	0.54-0.66	84-89	
GR	H	2 (17)	ol	pl	chr	71-73	83	15.4-15.6	0.52	395	0.34-0.59	0.53-0.60	83-84	
BE	T	3 (30)	ol	pl	cpx chr	72-74	81-88	13.6-14.1	0.47-0.52	775	0.52-0.68	0.45-0.65	88-90	
EN	H	2 (23)	ol	pl	cpx chr	70-76	86-89	14.2-14.7	0.53-0.55	960	0.39-0.62	0.55-0.70	86-90	
ST	H	2 (14)	Ol	pl	cpx chr	75-82	81-87	15.1-15.4	0.55-0.58	410	0.28-0.55	0.58-0.65	86-88	
NAL	T	1 (6)	ol	pl	mt	63	66			33				
SAX	T	3 (7)	ol	pl	mt	60-63	66-69			64				
Vestmannaeyjar Volcanic Area														
VE 46	H	1 (6)	ol	pl	chr	69	80-84	15.9	0.46	185	0.23-0.29	0.54-0.58	80	
VE 76	H	1 (5)	ol	pl	chr	69	78	15.8	0.44		0.37-0.38	0.59-0.62	85	
VE 112	T	(14)	ol	pl	mt	55	60	14.3	0.28	5				
VE 156	T	(12)	ol	pl	mt	61	68	13.7	0.31	20				
SU 19	T	(5)	ol	pl	chr	75	80-86	15.6	0.52	220	0.25-0.41	0.53-0.60	80	
SY	T	(13)	ol	pl	chr	76	83	15.1	0.47	610	0.29-0.45	0.53-0.61	82-84	
Sudurland Volcanic Zone														
RE	H	1 (8)	ol	pl	cpx chr	75-79	80	14.3	0.52	210	0.47-0.53	0.41-0.49	81	
HA	H	2 (8)		pl	cpx mt	57-59		12.9-13.5	0.37-0.39	20				
EY	L	1 (9)	ol	pl	chr	59-73	78-89		not known		0.56-0.67	0.53-0.67	89-90	

1) H - hyaloclastite; T - tephra; L - lava.

2) Total numbers of samples analysed. The numbers in parenthesis are the total number of analyses used.

3) Phenocryst assemblages.

4) Total variation of average microphenocryst compositions.

5) Summary of microprobe analyses of glass. Mg# -Mg/(Mg+Fe<sup>2+</sup>). Cr (ppm) values are average of XRF analyses on bulk rock samples by S. Fregerslev (Univ. Aarhus, Denmark).

6) Summary of total range in analysed chromian spinels. Cr# -Cr/(Cr+Al); Mg# -Mg/(Mg+Fe<sup>2+</sup>). Representative analyses is given in Table 3, and average of titanomagnetite analyses in Table 2.

7) Total range in Fo mole % of olivine enclosing chromian spinel.

## Previous work on spinel minerals in Icelandic lavas

A few chromian spinels from Icelandic lavas have been analysed by Carmichael (1967a), Steinthorsson (1972), Haggerty (1976), Fisk (1978), Jakobsson et al. (1978), Mäkipää (1978) and Maaløe & Jakobsson (1979). Most of these analyses [c. 10] are, where bulk rock composition is known, of spinels in picritic and tholeiitic basalts. Unpublished analyses by K. Grönvold (Nordic Volcanol. Inst.) from tholeiitic lavas are also included in this study. Several analyses of titanomagnetite can be found in studies on temperature-oxygen fugacity relations (Carmichael, 1967a,b; Steinthorsson, 1972; Baldrige et al., 1973; Jakobsson et al. 1973; Arney, 1978; Meyer, 1978; Larsen, 1979; Sigvaldason, 1979; Jørgensen, 1981; Sigurdsson & Sparks, 1978, 1981; Pedersen & Hald, 1982). Detailed information on coexisting liquid, silicates and spinels, as well as zoning and subsolidus reequilibration, is unfortunately lacking in most of these studies on holocrystalline basalts. In addition Osborn et al. (1979) investigated the pressure effect on titanomagnetite composition in two olivine tholeiitic lavas, and Sigvaldason (1974) gives data on trace element content in titanomagnetite from Hekla lavas.

## Petrography

A total of 40 basaltic and 2 hawaiitic-mugearitic glasses were collected from the transitional and alkali basaltic areas in the active volcanic zones of Iceland (Fig. 1 and appendix). Of these, 20 samples contained chromian spinel and only 8 titanomagnetite phenocrysts (Table 1). A holocrystalline picrite from Eyjafjallajökull volcano which contains chromian spinel was also investigated (EY, Table 1). The glass content of the samples range from 30-90%. Chromian spinel occurs in alkali basaltic glasses, whereas titanomagnetite occurs in Fe-Ti basaltic or hawaiitic-mugearitic glasses.

The spinels occur as euhedral octahedra included in silicates and glass. The chromian spinels are typically below 20  $\mu\text{m}$  in diameter, but may rarely reach 200  $\mu\text{m}$  as isolated crystals in glass. Clusters or isolated chromian spinel grains are mostly included in, or closely attached to, olivine phenocrysts, and very rarely occur enclosed by plagioclase or clinopyroxene. Grain-size of titanomagnetite frequently reaches 40  $\mu\text{m}$ , and occasionally 500  $\mu\text{m}$ . The titanomagnetite is included in all phenocryst types, but is not found in samples containing chromian spinel. Typical textural relations between the spinels, silicates and glass are shown in Fig. 2. Textural evidence indicates that the spinels crystallized relatively early, and that chromian spinel precipitation was dependent on olivine crystallization. Ilmenite only occurs as a groundmass phase in alkali lavas (e.g. Steinthorsson, 1972), and was not found as a phenocryst in the samples studied.

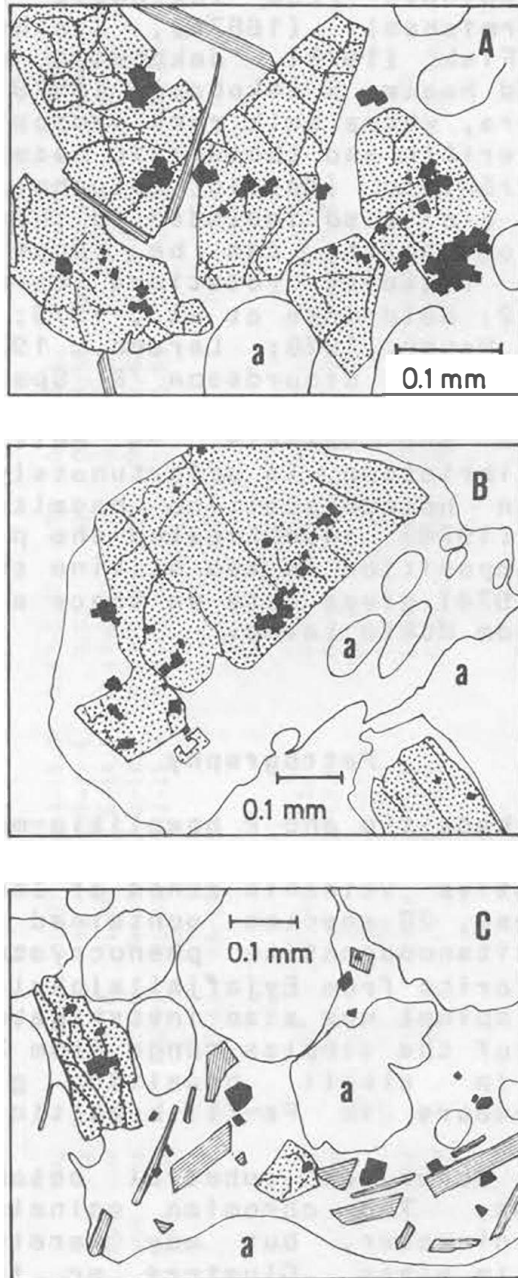


Figure 2. Typical textural relations of spinels and silicates in vesiculated alkalic glasses. 'a' - vesicle.

A. Olivine and plagioclase phenocrysts in an alkali glass from Snaefellsnes (ST). Chromian spinel octahedra included in or attached to olivine phenocrysts or more rarely in glass.

B. Olivine phenocryst with chromian spinel in an alkali basaltic glass fragment from Snaefellsnes (KE).

C. Olivine, plagioclase and titanomagnetite phenocrysts in a highly vesicular mugearitic glass from Vestmannaeyjar (VE 112).



The silicate phenocryst assemblages are plagioclase in all samples, olivine in most, and/or clinopyroxene in a few. Only minor zoning in phenocrysts can be observed, except in clinopyroxene from Sudurland, which occasionally have a low-Ti, low-Al core. Compositional gradients in silicate glass have not been detected. Details and interpretations of silicate and glass relations will be given elsewhere (Thy, 1982) and only a summary is presented in Table 1. Regional differences in phenocryst assemblages are shown in Fig. 3. Clinopyroxene occurs in the samples from western Snaefellsnes, but is absent in eastern Snaefellsnes and in the Vestmannaeyjar area. Clinopyroxene appears in most glasses from Sudurland, and olivine is a rare phenocryst in the Fe-Ti glasses.

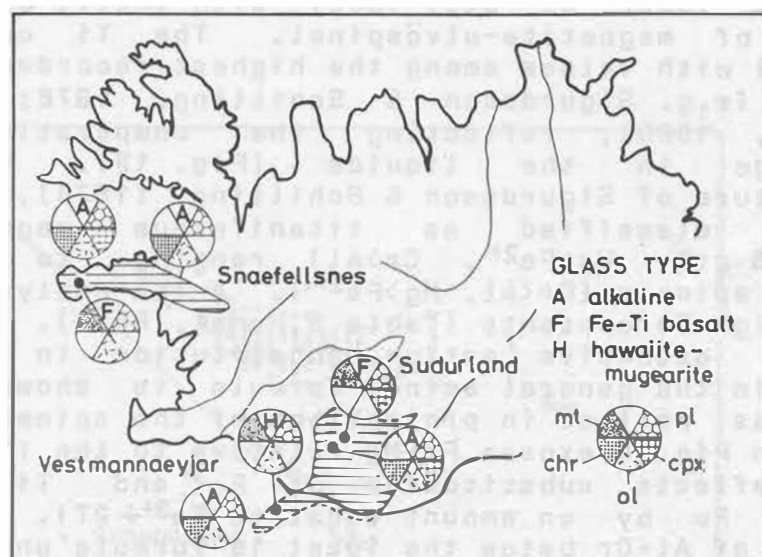


Figure 3. Summary of the main variation in phenocryst assemblages and glass types in the transitional and alkalic zones of Iceland. Data from Table 1. The hawaiite-mugearite glasses are from the 1973 Heimaey eruption (Jakobsson et al., 1973).

## Chemistry of the spinels

All analyses reported in this study were performed with a ARL-SEMQ microprobe, using natural minerals as standards. A total of 210 chromian spinel and 50 titanomagnetite analyses, together with simultaneously analysed coexisting olivine, are used in this study. The spinel analyses were recalculated on the basis of 32 oxygens in order to obtain structural formulae, and total Fe was distributed between  $\text{Fe}^{3+}$  and  $\text{Fe}^{2+}$  in proportions required for spinel stoichiometry (Finger, 1972). A full listing of chromian spinel analyses can be obtained from the author. Throughout the paper Mg# stands for  $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ , Cr# for  $\text{Cr}/(\text{Cr}+\text{Al})$  and  $\text{Fe}^{3+}\#$  for  $\text{Fe}^{3+}/(\text{Cr}+\text{Al}+\text{Fe}^{3+})$ . If the valence state of iron is not stated, Fe refers to total iron calculated as  $\text{Fe}^{2+}+\text{Fe}^{3+}$ .

Titanomagnetites (Table 2) are comparable in composition to other analyses from basalts, showing a ulvöspinel component between 35-60 mole %, and relatively high Mg# values. The  $\text{FeCr}_2\text{O}_3$  in solid solution is generally low (Table 2). The chemistry of the titanomagnetite will not be considered further in this section.

The chromian spinels (Table 3) belong to the chromite solid solution series: chromite-magnesiochromite-spinel-hercynite (Deer et al., 1962), with small, but significant amounts of magnetite-ulvöspinel. The Ti content varies (Table 3) with values among the highest recorded from abyssal basalts (e.g. Sigurdsson & Schilling, 1976; O'Donnell & Presnall, 1980), reflecting the comparatively larger  $\text{TiO}_2$  range in the liquids (Fig. 13). Following the nomenclature of Sigurdsson & Schilling (1976), the spinels can be classified as titaniferous magnesiochromites ( $\text{TiO}_2 > 0.5$  wt%,  $\text{Mg} > \text{Fe}^{2+}$ ,  $\text{Cr} > \text{Al}$ ) ranging to titaniferous chromian spinels ( $\text{Cr} < \text{Al}$ ,  $\text{Mg} > \text{Fe}^{2+}$ ). A few analyses show high-Ti and high-Fe contents (Table 3, anal. RE 1).

The extensive cation substitution in the chromian spinels in the general spinel formula is shown in Figs. 4 and 5, as well as in projections of the spinel prism (Figs. 6-8). In Fig. 4 excess Fe+Mg relative to the ideal 8 formula units reflects substitution of  $\text{Fe}^{3+}$  and  $\text{Ti}^{4+}$  for  $\text{Al}+\text{Cr}$ , reducing Fe by an amount equal to  $\text{Fe}^{3+} + 2\text{Ti}$ . In Fig. 5 the position of  $\text{Al}+\text{Cr}$  below the ideal 16 formula units is roughly equivalent to the excess Fe+Mg (Fig. 4), indicating magnetite in solid solution. The amount of ulvöspinel in most analyses ( $\text{Ti} \sim 0.2$  formula units) is too low for this rough estimate. An exception is the chromian spinel in the Sudurland alkali glass (RE 1) with higher Fe and Ti contents and magnetite-ulvöspinel component (Figs. 4 and 5).

The total range in the Al-Cr substitution amounts to 7 formula units (Fig. 5), while the total range in  $\text{Fe}^{2+}$ -Mg substitution amounts to 2 formula units (Fig. 4). A dependency of the total amount of  $\text{Fe}^{3+}+\text{Ti}$  substitution (from 1.2 to 2.3 formula units) on the increase in Al or decrease in Cr can be seen from Fig. 5.

T A B L E 2. Average of microprobe analyses of titanomagnetite

Sample no.	VE 112	VE 156	HA 3	HA 6	MAL 3	SAX 1
SiO <sub>2</sub>			0.11	0.07	0.07	0.13
TiO <sub>2</sub>	20.33	21.87	18.99	16.91	22.38	14.41
Al <sub>2</sub> O <sub>3</sub>	5.01	4.95	3.16	4.15	4.41	7.30
Cr <sub>2</sub> O <sub>3</sub>		0.05		0.03	0.90	4.46
FeO (1)	43.47	44.00	42.07	39.21	44.17	36.85
Fe <sub>2</sub> O <sub>3</sub> (1)	25.98	23.49	31.14	33.44	22.00	31.34
MnO	0.60	0.58	0.45	0.35	0.47	0.38
MgO	4.32	4.94	4.56	4.99	5.16	5.80
Total	99.71	99.88	100.48	99.15	99.56	100.67
Ulv mole % (2)	54.7	58.6	48.0	41.4	60.8	35.5
Mt mole % (2)	45.3	41.4	52.0	58.6	39.2	64.5

(1) Calculated assuming stoichiometry in the spinel solid solution series.

(2) Calculated as proposed by Bowles (1977).

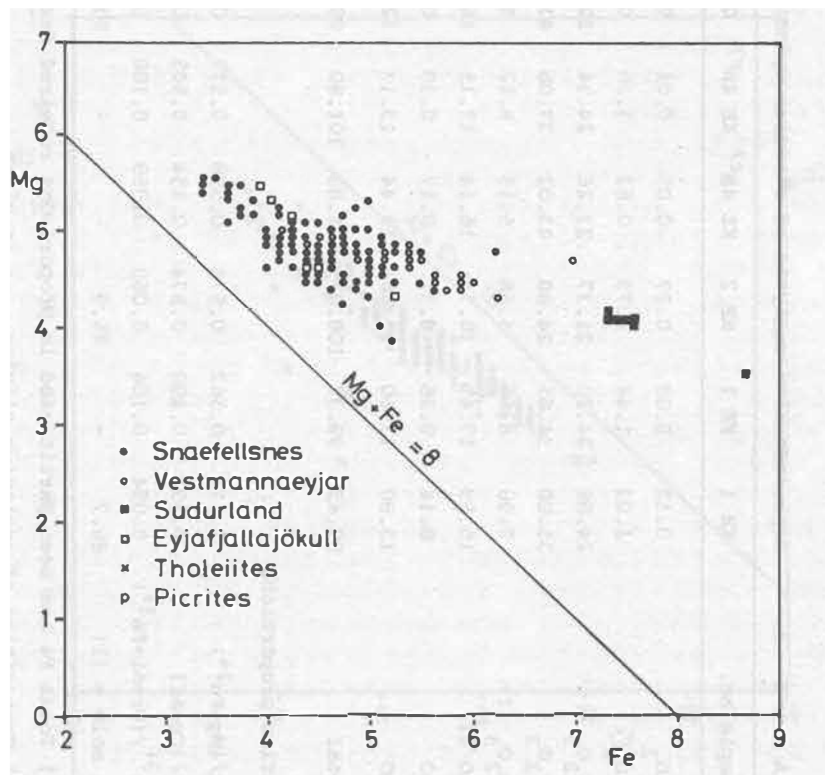


Figure 4. Substitution of Fe-Mg in chromian spinels. Iron is given as total Fe. Calculated to 32 oxygen and 24 cations. Analyses from tholeiites and picrites are not shown on this figure or on Fig. 5, but are given in subsequent figures.

T A B L E 3. Representative analyses of chromian spinel

Sample no.	KE 1	KE 1	KE 2	KE 4B <sup>c)</sup>	KE 4B <sup>r)</sup>	GR	BE 1	BE 3	EN 3	EN 5	ST 1	ST 2	ST 2	VE 46
SiO <sub>2</sub>	0.17	0.08	0.27	0.05	0.04	0.15	0.17	0.21	0.09	0.16	0.10	0.06	0.13	0.16
TiO <sub>2</sub>	1.03	1.48	0.79	0.62	1.26	0.54	0.55	0.70	0.82	0.69	0.99	1.67	0.84	1.26
Al <sub>2</sub> O <sub>3</sub>	24.88	24.20	21.37	23.26	24.34	22.86	20.75	16.23	19.25	25.42	26.51	35.10	27.99	38.06
Cr <sub>2</sub> O <sub>3</sub>	35.88	34.65	24.88	43.02	37.06	41.72	45.17	51.73	46.33	38.75	35.36	20.03	31.96	17.41
Fe <sub>2</sub> O <sub>3</sub> (1)	7.96	8.65	5.68	5.13	8.52	5.21	4.47	3.49	4.86	6.89	7.56	11.96	8.78	9.75
FeO (1)	15.59	17.45	16.71	16.14	17.15	16.40	15.31	14.36	13.54	12.55	15.02	14.80	15.53	18.31
MnO	0.14	0.26	0.31	0.17	0.30	0.20	0.36	0.28	0.50	0.68	0.21	0.17	0.19	0.14
MgO	13.80	12.60	12.86	13.44	13.13	12.85	13.30	13.79	14.19	15.55	14.35	15.57	13.96	13.26
Total	99.45	99.37	100.87	101.83	101.80	99.93	100.08	100.79	99.58	100.69	100.10	99.36	99.38	98.35

Cation proportions

Mg/(Mg+Fe <sup>2+</sup> )	0.612	0.563	0.578	0.598	0.577	0.583	0.608	0.631	0.651	0.688	0.621	0.652	0.616	0.564
Cr/(Cr+Al)	0.492	0.490	0.574	0.554	0.505	0.550	0.594	0.681	0.618	0.506	0.472	0.277	0.434	0.235
Fe <sup>3+</sup> /(Cr+Al+Fe <sup>3+</sup> )	0.094	0.104	0.068	0.059	0.100	0.062	0.053	0.042	0.058	0.079	0.088	0.136	0.102	0.111
Fo mole % (2)	86.7		85.9			83.3		90.4	89.3	86.1	87.0		87.6	79.7

(1) Total Fe has been partitioned in proportions required for spinel stoichiometry (Finger, 1972).

(2) Composition of enclosing olivine. in glass.

c) Core.

r) Rim.

T A B L E 3 (continued)

	VE 76	SU 19	SY	RE 1	RE 1	EY
SiO <sub>2</sub>	0.16	0.16		0.23	0.08	0.09
TiO <sub>2</sub>	0.29	1.19		3.71	3.30	1.16
Al <sub>2</sub> O <sub>3</sub>	30.47	38.02	26.64	21.13	20.43	18.38
Cr <sub>2</sub> O <sub>3</sub>	27.33	18.77	29.68	27.70	26.63	45.22
Fe <sub>2</sub> O <sub>3</sub> (1)	10.17	9.76	13.94	14.17	15.75	6.25
FeO (1)	15.97	17.79	16.43	21.98	20.73	13.16
MnO	0.24	0.20	0.29	0.17	0.30	0.22
MgO	13.27	13.79	12.54	10.90	10.75	14.62
Total	97.90	99.68	99.52	99.99	97.97	99.10
Cation proportions						
Mg/(Mg+Fe <sup>2+</sup> )	0.597	0.580	0.576	0.469	0.480	0.665
Cr/(Cr+Al)	0.376	0.249	0.428	0.468	0.467	0.623
Fe <sup>3+</sup> /(Cr+Al+Fe <sup>3+</sup> )	0.118	0.110	0.161	0.186	0.208	0.076
Fo mole % (2)	84.5	80.2				89.6

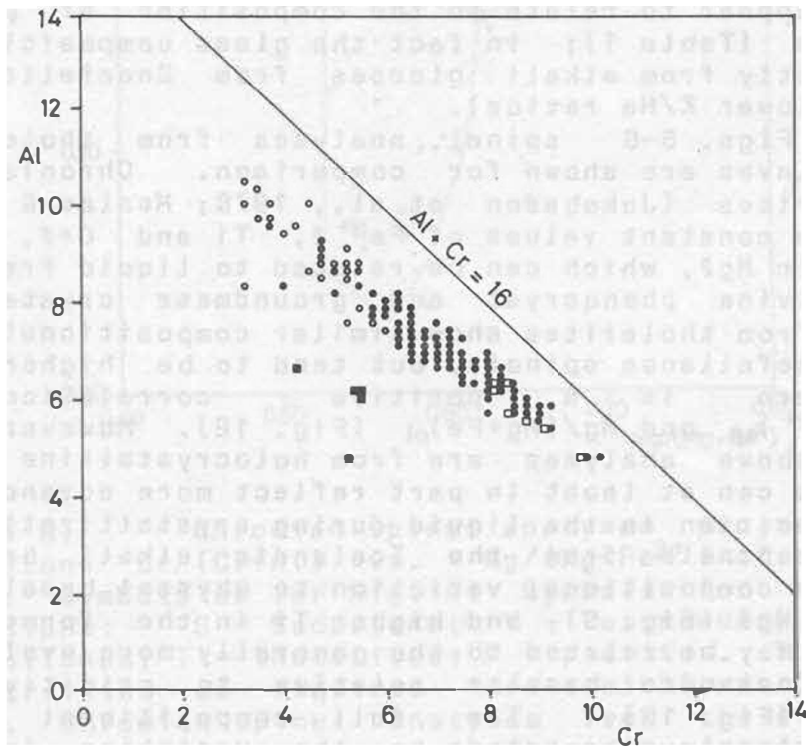


Figure 5. Substitution of Al-Cr in chromian spinels. Calculated to 32 oxygens and 24 cations. See Fig. 4 for legend.

The analysed samples were collected from three geographical and tectonic areas (Fig. 1). There are small but significant differences in lava composition from these areas (Jakobsson, 1972, 1979; Imslund, 1978) but only minor differences are revealed in the spinel chemistry. Compared with the Snaefellsnes spinels, those from Vestmannaeyjar have higher Al/Cr and Fe/Mg ratios (Figs. 4 and 5), consistent with differences in glass and olivine composition (Table 1, Fig. 12).

Further details of the overall compositional variation and the regional differences are given in Figs. 6-8. The main variation is in the Cr# ratio (Fig. 6), followed by some variation in Mg#, but minor variation in Ti (Fig. 7) and in Fe<sup>3+</sup># (Fig. 8). Some differences are revealed between the Vestmannaeyjar and the Snaefellsnes spinels, notably in that the Vestmannaeyjar spinels have higher Fe<sup>3+</sup>#, lower Cr# and somewhat lower Mg#. This is consistent with the 'more evolved nature' of the Vestmannaeyjar basalts as revealed by glass and olivine compositions (Table 1). The Snaefellsnes spinels span most of the compositional range observed (except for the high-Fe and Ti spinel from the Sudurland alkali glass). The Sudurland basalts rarely contain chromian spinels (Jakobsson, 1979) and only one sample was found containing a few chromian spinels (RE 1) which depart markedly from all other analyses, with high-Fe<sup>3+</sup>#, high-Ti, low-Mg# and intermediate Cr# (Figs. 6-8). This composition does not appear to relate to the composition of phenocrysts and glass (Table 1); in fact the glass composition departs only slightly from alkali glasses from Snaefellsnes (e.g. slightly lower K/Na ratios).

In Figs. 6-8 spinel analyses from tholeiitic and picritic lavas are shown for comparison. Chromian spinels from picrites (Jakobsson et al., 1978; Maaløe & Jakobsson, 1979) show constant values of Fe<sup>3+</sup>#, Ti and Cr#, but some scatter in Mg#, which can be related to liquid fractionation during olivine phenocryst and groundmass crystallization. Spinel from tholeiites show similar compositional variation to the Snaefellsnes spinels, but tend to be higher in Mg#, and there is a positive correlation between Mg/[(Mg+Fe<sup>2+</sup>)<sub>sp</sub>] and Mg/[(Mg+Fe)<sub>ol</sub>] (Fig. 12). However, as most of the above analyses are from holocrystalline lavas, the variations can at least in part reflect more advanced stages of fractionation in the liquid during crystallization.

The spinels from the Icelandic alkali basalts have comparable compositional variation to abyssal basalts, except for lower Mg# (Fig. 9), and higher Ti in the former. These features may be related to the generally more evolved nature of these Icelandic basalts relative to primitive abyssal basalts (Fig. 13). The full compositional variation, however, closely corresponds to the variation in chromian spinels from the calc-alkalic island of Grenada, Lesser Antilles (Arculus, 1978).

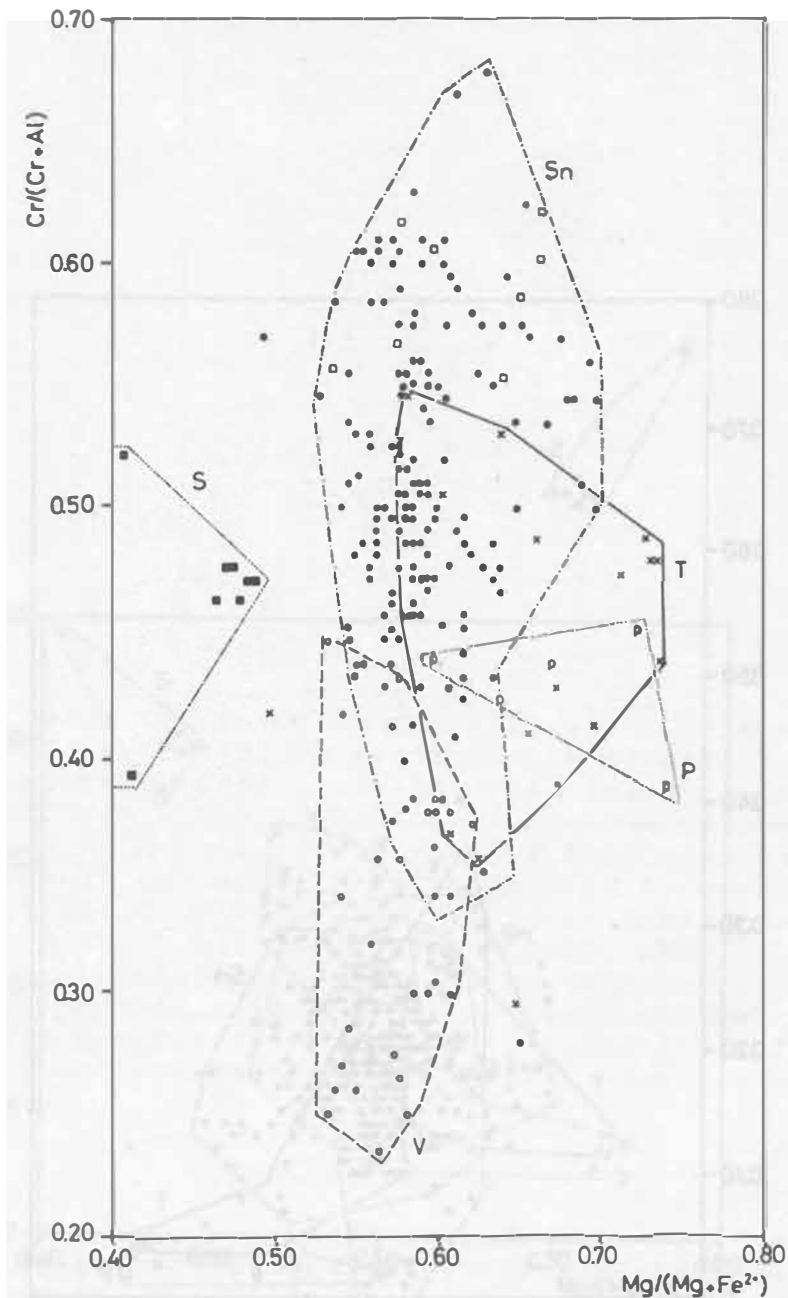


Figure 6. Chromian spinal analyses projected onto the plane  $\text{Cr}/(\text{Cr}+\text{Al})$  vs.  $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$  of the spinal prism. Symbols as for Fig. 4. Symbols for the total variations: S - Sudurland; V - Vestmannaeyjar; Sn - Snaefellsnes; T - tholeiites; P - picrites. Total iron redistributed as required for stoichiometry (Finger, 1972). Chromian spinal analyses from tholeiites and picrites are taken from various sources as given in text, together with unpublished analyses by K. Grönvold (Nordic Volcanol. Inst.).

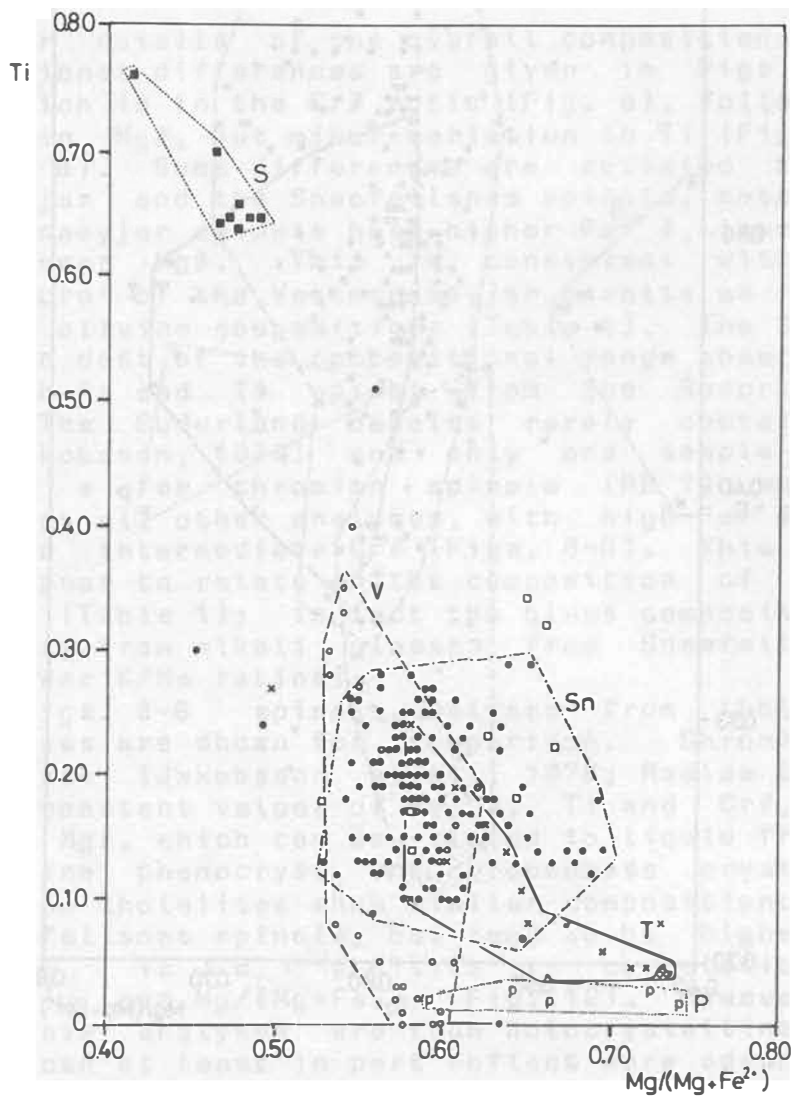


Figure 7. Ti vs.  $Mg/(Mg+Fe^{2+})$  for chromian spinels. Calculated to 32 oxygens and 24 cations. Symbols as given in Figs. 4 and 6.



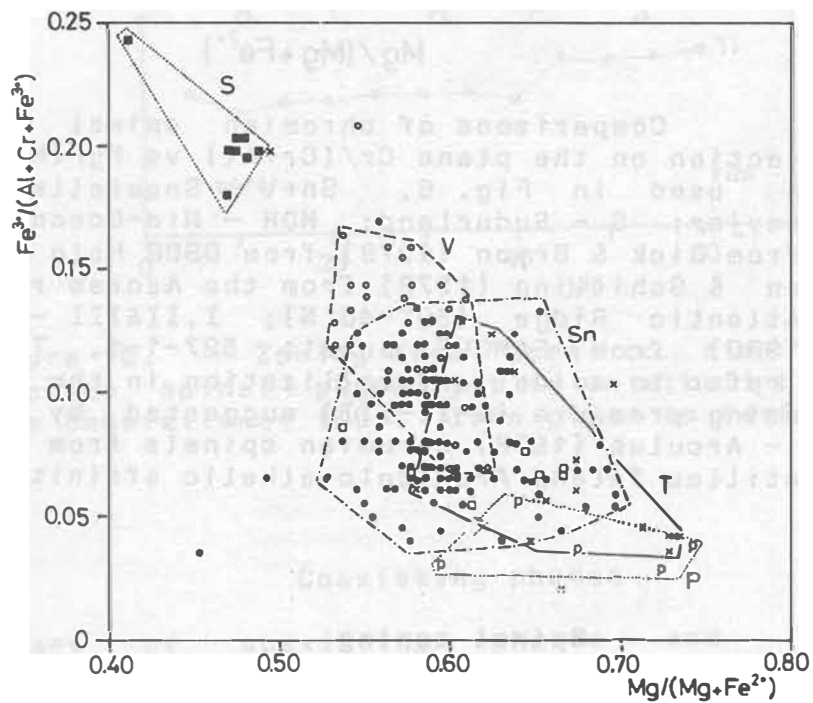


Figure 8. Chromian spinels projected onto the plane  $\text{Fe}^{3+}/(\text{Cr}+\text{Al}+\text{Fe}^{3+})$  vs.  $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$  in the spinel prism. Symbols as for Figs. 4 and 6.

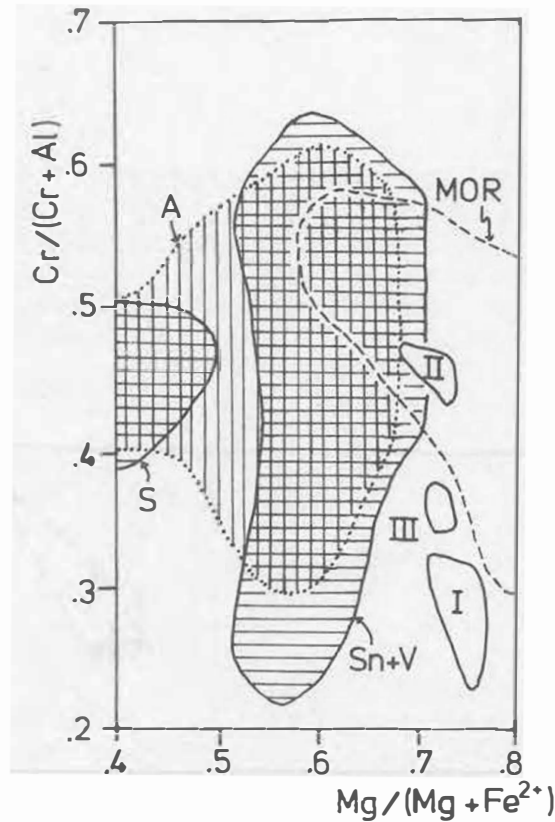


Figure 9. Comparisons of chromian spinel analyses in a projection on the plane  $\text{Cr}/(\text{Cr}+\text{Al})$  vs  $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ , as also used in Fig. 6. Sn+V - Snaefellsnes and Vestmannaeyjar; S - Sudurland; MOR - Mid-Ocean Ridge spinels from Dick & Bryan (1978) from DSDP Hole 3968 and Sigurdsson & Schilling (1976) from the Azores region of the Mid-Atlantic Ridge ( $30^{\circ}$ - $40^{\circ}$ N); I, II & III - Fisk & Bence (1980) from FAMOUS basalt 527-1-1. The Roman numerals refer to spinel crystallization in the sequence of decreasing pressure (I-II-III) suggested by Fisk & Bence; A - Arculus (1978) chromian spinels from Grenada, Lesser Antilles Island Arc (calc-alkalic affinity).

#### Spinel zoning

Zoning has not been detected in the titanomagnetite phenocrysts. The small size of most of the chromian spinel crystals prohibited detailed examination of zoning, but slight marginal zoning ( $10$ - $20 \mu\text{m}$ ) occurs in a few large isolated ( $100$ - $200 \mu\text{m}$ ) chromian spinels enclosed in glass. An example is shown in Fig. 10 (and Table 3, anal. KE 48), where a fall in Cr# and rise in  $\text{Fe}^{3+}$  # and Ti towards the rim is apparent (i.e. towards titanomagnetite). Zoning, as well as systematic arrangements in the compositions of chromian spinels included in olivine phenocrysts, was looked for, but not found.

Zoning similar to the present results (Fig. 10) have been reported by, amongst others, Ridley et al. (1974) from DSDP lavas, Sigurdsson & Schilling (1976) and O'Donnell & Presnall (1980) from glassy basalts from the MID-Atlantic Ridge, and Sigurdsson (1977) from LEG 37 basalts.

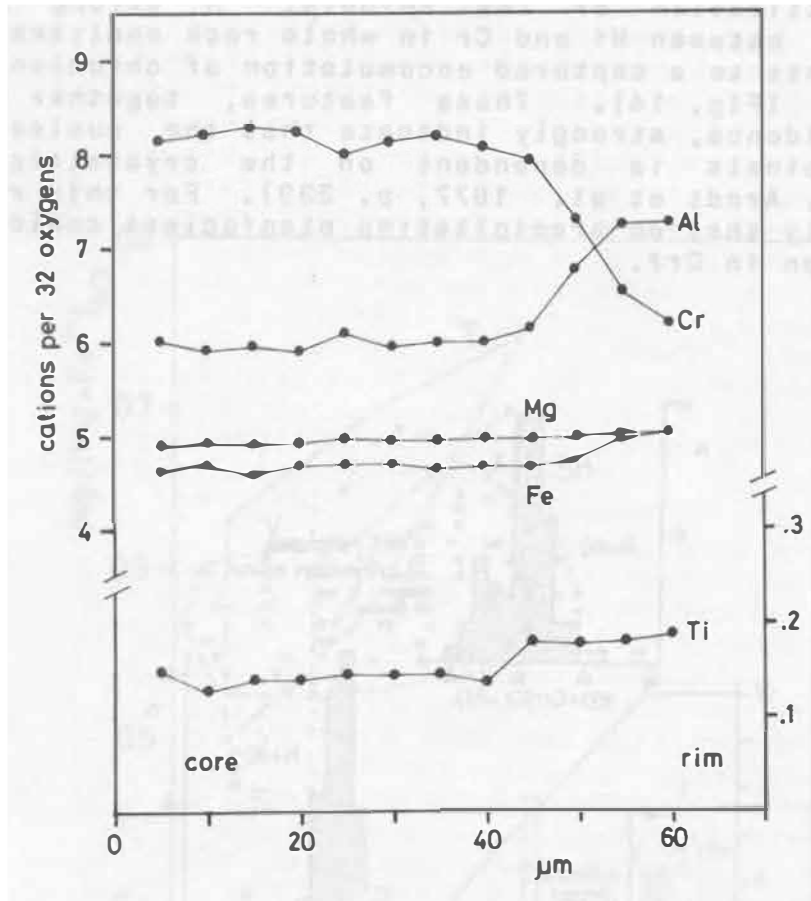


Figure 10. Zoning from core to rim in a 100  $\mu\text{m}$  chromian spinel phenocryst in an alkali basaltic glass from Snaefellsnes (KE). Iron given as  $\text{Fe}=\text{Fe}^{2+}+\text{Fe}^{3+}$ .

#### Coexisting phases

A summary of coexisting phases and compositional characteristics of olivine, plagioclase and glass is given in Table 1. The host olivine of chromian spinel is generally high in forsterite (Fo) compared to the total range in olivine composition (Fig. 11 and Table 1). These high-Fo olivine phenocrysts (macrophenocrysts) are believed to have crystallized from a liquid of a less 'evolved' nature (i.e. higher Mg# and Cr contents). This suggests that chromian spinel and high-Fo olivine phenocrysts crystallized relatively early during evolution of the liquid. The low-Fo olivines (microphenocrysts) probably crystallized from the chrome depleted enclosing glass and only contain a few chromian spinels. No clear relationship is observed between the disappearance of spinels and the appearance of clinopyroxene (Table 1). A peritectic reaction relationship between chromian spinel and clinopyroxene has been suggested by Irvine (1967) and Hill & Roeder (1974), causing a discontinuity in spinel crystallization, which is well

documentated in layered intrusions (e.g. Irvine, 1967; Irvine & Smith, 1969). It can be seen from Table 1 that the crystallization of clinopyroxene does not appear to influence the crystallization of the spinels. A strong positive correlation between Ni and Cr in whole rock analyses (unpub. anal.) points to a captured accumulation of chromian spinel in olivine (Fig. 14). These features, together with the textural evidence, strongly indicate that the nucleation of chromian spinels is dependent on the crystallization of olivine (cf. Arndt et al. 1977, p. 339). For this reason it is not likely that co-precipitating plagioclase could explain the variation in Cr#.

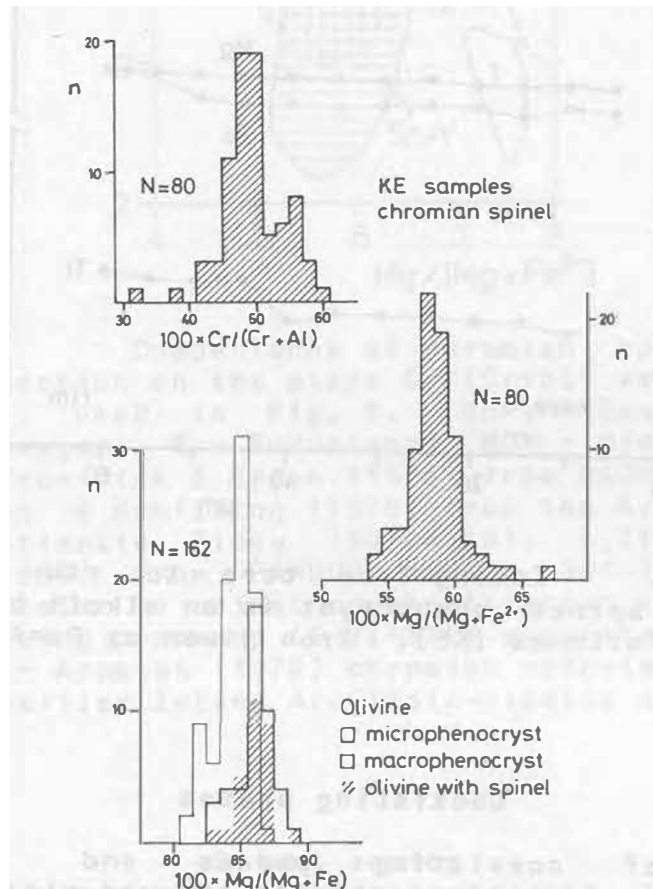


Figure 11. Summary histograms of compositional variation in coexisting chromian spinels and olivine in the KE samples (6 samples). Lower left histogram shows the compositional variation in microphenocrysts and macrophenocrysts, as identified on size (0.30 mm). The macrophenocrysts shows a tendency to extend towards higher Fo than the microphenocrysts. Also shown are the compositional variation of olivines enclosing chromian spinels (analysed simultaneously with the spinels). It can be seen that the spinel containing olivines belong to the macrophenocrysts (e.g. higher Fo). The two other histograms show the total compositional variation in Mg# and Cr# for the chromian spinels. Spinel enclosed within homogeneous olivine can show the full variation in Mg# and Cr#. See text for further discussion.

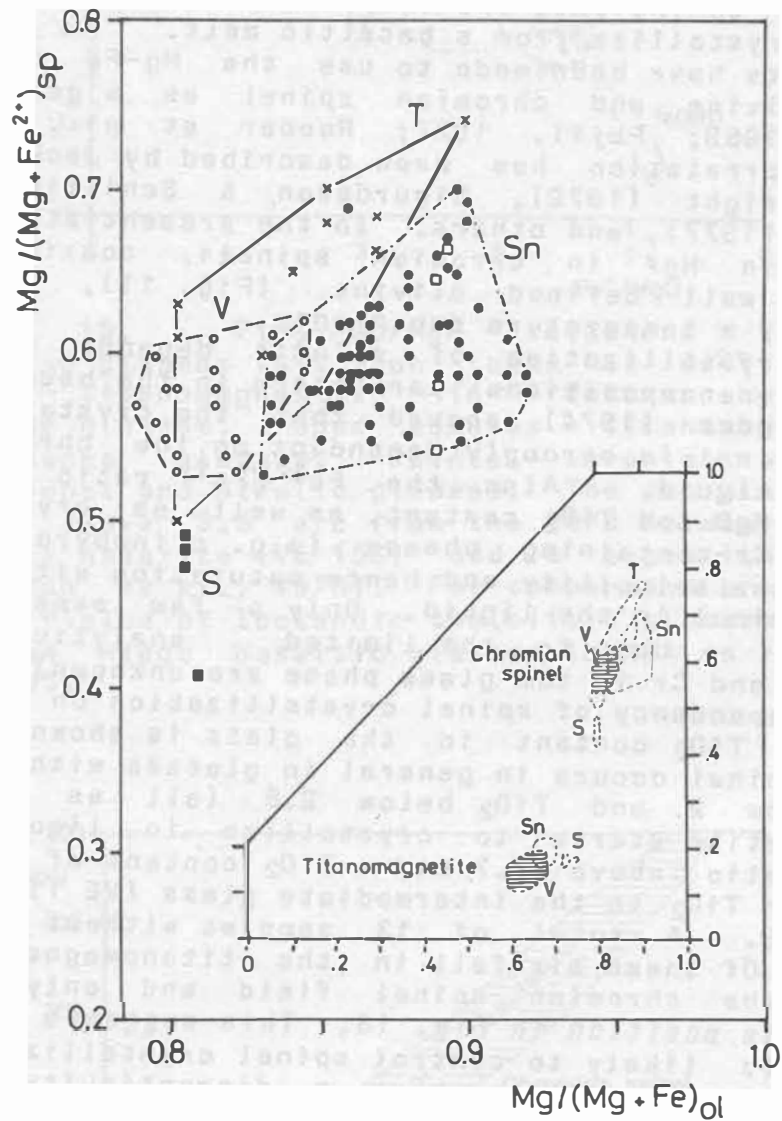


Figure 12. Mg-Fe distribution between coexisting spinels and olivine.  $Fe^{2+}$  has been calculated based on spinel stoichiometry, and all iron assumed as  $Fe^{2+}$  for olivine. The Sudurland glasses with titanomagnetite do not contain olivine (Table 1), and a composition has been inferred from a corresponding glass composition. See text for sources of chromian spinels from Icelandic tholeiites, and Figs. 4 and 6 for symbols.

The relation between  $Mg/(Mg+Fe^{2+})_{sp}$  and  $Mg/(Mg+Fe)_{ol}$  is shown in Fig. 12. Differences appear between the alkalic areas. Chromian spinels from Snaefellsnes have a broad variation and trend towards higher Mg# ratios than those from the other areas. The Vestmannaeyjar chromian spinels have somewhat lower Mg# ratios, while those from Sudurland have the lowest Mg# ratios observed in this study. The field of chromian spinels in tholeiitic basalts defines an elongate field, suggesting a systematic distribution of Mg-Fe between the two phases which is not observed for the alkalic glasses. High values for both ratios are found in glasses containing titanomagnetite in this study (Fig. 12, inset). These values are likely to indicate the composition when titanomagnetite starts to crystallize from a basaltic melt.

Attempts have been made to use the Mg-Fe distribution between olivine and chromian spinel as a geothermometer (Jackson, 1969; Fujii, 1977; Roeder et al., 1979). A positive correlation has been described by Jackson (1969), Evans & Wright (1972), Sigurdsson & Schilling (1976), Sigurdsson (1977), and others. In the present study the wide variation in Mg# in chromian spinels, coexisting with a relatively well defined olivine (Fig. 11), cannot be explained by a temperature dependency.

The crystallization of spinels depends among other parameters on compositional parameters in the basaltic melt. Hill & Roeder (1974) showed that the crystallization of chromian spinel is strongly dependent on the chrome content of the liquid. Also the  $Fe^{3+}/Fe^{2+}$  ratio (or oxygen fugacity), MgO and  $TiO_2$  content, as well as crystallization of other Cr-containing phases (e.g. clinopyroxene), will influence the solubility and hence saturation with respect to chromian spinel in the liquid. Only a few aspects can be evaluated, due to the limited analytical results ( $Fe^{3+}/Fe^{2+}$  and Cr in the glass phase are unknown).

The dependency of spinel crystallization on the  $FeO/MgO$  ratio and  $TiO_2$  content in the glass is shown in Fig. 13. Chromian spinel occurs in general in glasses with a  $FeO/MgO$  ratio below 2 and  $TiO_2$  below 2.8 (all as wt% oxides). Titanomagnetite starts to crystallize in liquids with a  $FeO/MgO$  ratio above 2.7 and a  $TiO_2$  content of 4. A marked decrease in  $TiO_2$  in the intermediate glass (VE 112) is seen in Fig. 13. A total of 13 samples without spinels were examined. Of these six fall in the titanomagnetite field, four in the chromian spinel field and only three in an intermediate position in Fig. 13. This suggests that other factors are likely to control spinel crystallization. Fig. 13 nevertheless suggests that a discontinuity occurs in spinel crystallization in the sequence of basaltic glass compositions in this study. This discontinuity is unrelated to clinopyroxene crystallization. The compositional parameter most likely to control chromian spinel crystallization is the chrome content of the melt. The four samples without spinels which lie in the chromian spinel field in Fig. 12, all have whole rock chrome contents below approximately 200 ppm (Table 1, Fig. 14). This value is similar to the maximum solubility found by Hill & Roeder (1974) at  $1200^\circ C$  and  $f_{O_2} = 10^{-8}$ .

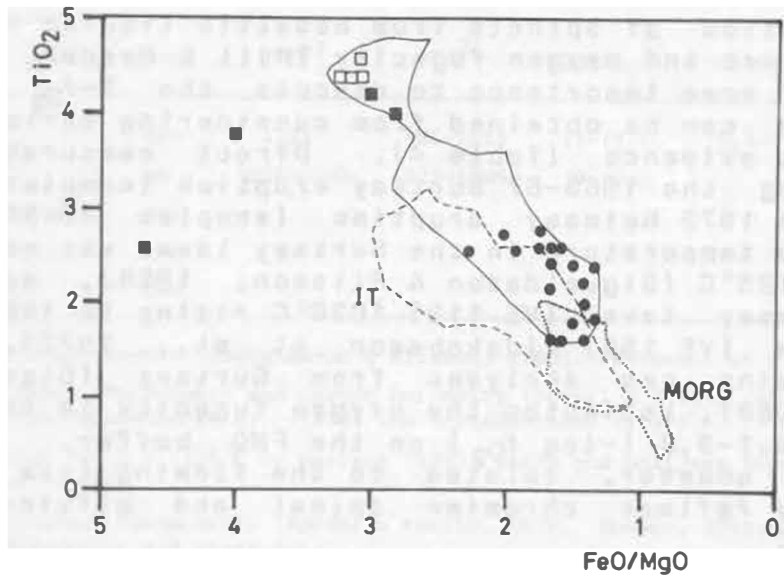


Figure 13.  $TiO_2$ - $FeO/MgO$  relations in spinel containing glasses (all iron taken as  $Fe^{2+}$ ). Filled squares - titanomagnetite in Vestmannaeyjer and Sudurland glasses. Open squares - titanomagnetite in Snaefellsnes glasses. Points - chromian spinels in transitional end alkalic glasses. The two glasses with  $FeO/MgO$  above 3.5 are from the 1973 Heimeey eruption, and are a hawaiiite (VE 156) and a mugeerite (VE 112) (Jakobsson et al., 1973). For comparison are shown the general fields of Icelandic tholeiitic glasses (IT) and Mid-Ocean Ridge basaltic glasses (MORG) as represented by FAMOUS.

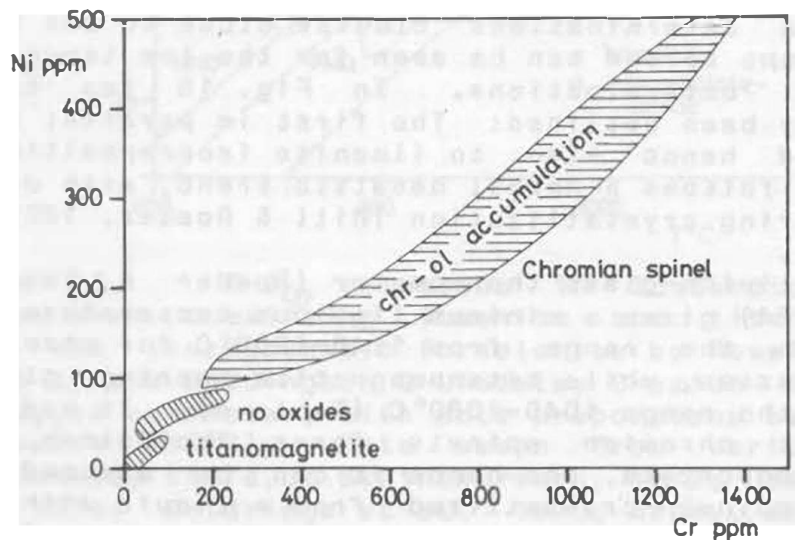


Figure 14. Ni-Cr relations in transitional and alkali glasses from Iceland (unpub. bulk rock anal.). Fields for crystallization of titanomagnetite, no oxides and chromian spinels are shown. The strong positive correlation is interpreted as reflecting a captured accumulation of chromian spinel by olivine (see text).

## Estimates of temperature and oxygen fugacities

Crystallization of spinels from basaltic liquids is affected by temperature and oxygen fugacity (Hill & Roeder, 1974). It is hence of some importance to discuss the  $T$ - $f_{O_2}$  relations. Information can be obtained from considering various sources of indirect evidence (Table 4). Direct measurements were made during the 1963-67 Surtsey eruption (samples SY and SU 19) and the 1973 Heimaey eruption (samples VE 112 and VE 156). The temperature in the Surtsey lavas was estimated to be about 1125°C (Sigvaldason & Elisson, 1968), and in the early Heimaey lavas (VE 112) 1030°C rising to 1080°C in the later ones (VE 156) (Jakobsson et al., 1973). Gerlach (1980), using gas analyses from Surtsey (Sigvaldason & Elisson, 1968), estimated the oxygen fugacity in the lavas in the range 9.1-9.8 ( $-\log f_{O_2}$ ) on the FMQ buffer. All these estimates, however, relates to the flowing lava and do not necessarily reflect chromian spinel and olivine liquidus conditions.

Using the Buddington & Lindsley 2-oxide thermometer and oxygen barometer (Powell & Powell, 1977), end analyses from Jakobsson et al. (1973) and Steinthorsson (1972), temperatures of 1025-1085°C and fugacity values of 9-10.2 for Heimaey, and temperatures of 985-1015°C and oxygen fugacities of 11-11.6 for Surtsey are obtained (Table 4, Fig. 15). These are, however, groundmass values (i.e. ilmenite does not crystallize together with titanomagnetite). The general  $T$ - $f_{O_2}$  field for Icelandic rocks is summarized in Fig. 15, showing most high temperature determinations on or between the NNO and FMQ buffer curves. The Heimaey determinations lie close to the NNO buffer, while the values for Surtsey lie just below the FMQ buffer curves. It is hence reasonable to assume a high temperature position at or in the vicinity of the FMQ end below the NNO buffer curves for basaltic rocks, as proposed by Gerlach (1980) from analyses of the Surtsey gases. Even though the high temperature determinations cluster close to the FMQ and NNO buffers, some spread can be seen for the low temperature and subsolidus determinations. In Fig. 15 two trends have tentatively been outlined: The first is parallel to the NNO buffer and hence also to ilmenite isocompositional lines. The second follows a normal basaltic trend, with decreasing  $T$  and  $f_{O_2}$  during crystallization (Hill & Roeder, 1975; Morse et al., 1980).

The olivine-glass thermometer (Roeder & Emslie, 1970; Roeder, 1974) gives a minimum liquidus temperature for low- $F_o$  olivine in the range from 1140-1200°C for chromian spinel bearing glasses, while titanomagnetite bearing glasses give values in the range 1040-1090°C (Table 4). It has been shown above that chromian spinels forms inclusions in high- $F_o$  olivine phenocrysts, and hence it can be assumed that the chromian spinel crystallized from a liquid with a slightly higher olivine-liquid temperature than that for low- $F_o$  olivine crystallization.



TABLE 4. Summary of estimates of temperature and oxygen fugacity on alkali basalts from Iceland

	Spinel	T°C <sup>1)</sup>	T°C <sup>2)</sup>	-logf <sub>O<sub>2</sub></sub> <sup>2)</sup>	-logf <sub>O<sub>2</sub></sub> <sup>3)</sup>	ol-glass T°C <sup>4)</sup>
Surtsey	chr	1125	985-1015	11-11.6	9.1-9.8	1150-1175
Heimaey <sup>5)</sup>	mt	1030-1080	1025-1085	9-10.2		1040-1090
Vestmannaeyjar	chr					1140-1180
Snaefellsnes	chr					1140-1190
Sudurland	chr					1200-1205

- 1) Direct measurements (Sigvaldason & Elisson, 1968; Jakobsson et al., 1973).
- 2) Fe-Ti oxide thermometry and oxygen barometry (Powell & Powell, 1977) on groundmass data from Jakobsson et al. (1973) and Steinthorsson (1972).
- 3) Estimates for T = 1125°C by Gerlach (1980) using gas analyses from Sigvaldason & Elisson (1968).
- 4) Olivine-glass thermometry (Roeder & Emslie, 1970; Roeder, 1974) using unpublished microphenocryst and glass data.
- 5) The low values are from the early lavas (VE 112), and the high from the late lavas, during the 1973 eruption (Jakobsson et al., 1973).

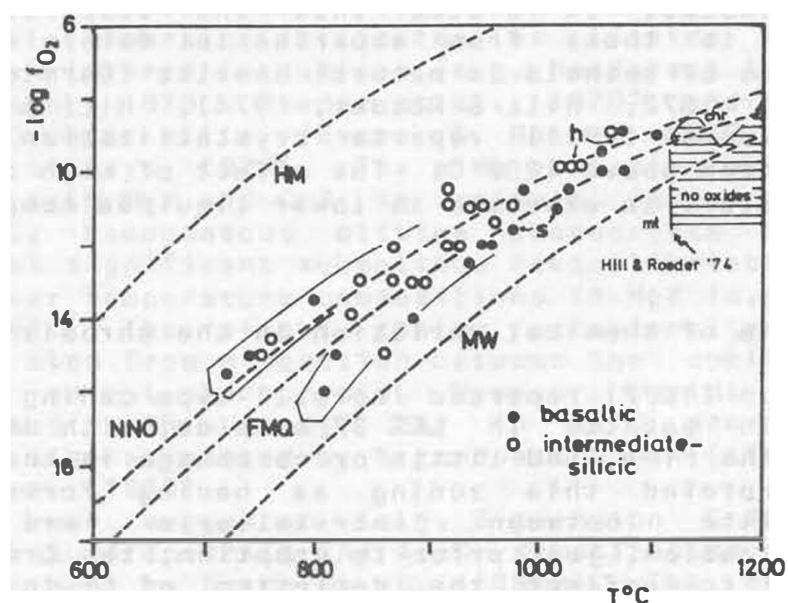


Figure 15. T-f<sub>O<sub>2</sub></sub> relations in Icelandic lavas as deduced from coexisting ilmenite and titanomagnetite using the thermodynamic formulation by Powell & Powell (1977) of the Buddington & Lindsley 2-oxide thermometer and oxygen barometer, with mole proportions as suggested by Bowles (1977). Data taken from various sources (Carmichael, 1967a,b; Steinthorsson, 1972; Jakobsson et al. 1973; Baldrige et al. 1973; Arney, 1978; Meyer, 1978; Larsen, 1979; Jørgensen, 1981; Pedersen & Hald, 1982). The fields for titanomagnetite, chromian spinel and no oxide crystallization from Hill & Roeder (1974). See text for further discussion. s - Surtsey (Steinthorsson, 1972); h - Heimeey (Jakobsson et al., 1973).

When the above estimates are compared to the 1 atm experimental results of Hill & Roeder (1974), a general model for spinel crystallization can be proposed. The fields of chromian spinel and titanomagnetite crystallization are shown in Fig. 15, taken from Hill & Roeder (1974, fig. 9). It is particularly significant to note that just below the NNO buffer there is no solid solution between the two spinels, and an oxide-free field occurs.

Crystallization of chromian spinels can be estimated to occur in liquids with olivine liquidus temperature from above 1200°C, and continue down to 1150°C. Crystallization of titanomagnetite starts just below 1100°C, followed by groundmass titanomagnetite and ilmenite at still lower temperatures. It is therefore interesting to note that a break in spinel crystallization exists from 1100°C to 1150°C. It is very likely that this break in T-f<sub>O<sub>2</sub></sub> relations correspond to the compositional break shown by TiO<sub>2</sub> and FeO/MgO in Fig. 12. The above temperatures are based on olivine-liquid relations, and the issue is further complicated by the possibility of a non-equilibrium heterogeneous nucleation of the chromian spinel, by inclusion in high-Fo olivine and not in equilibrium low-Fo olivine, and by the early crystallization of titanomagnetite relative to olivine. For these reasons the temperatures only represent best estimates. It is clear that the temperatures can be compared to those from experimental determinations on the appearance of spinels in natural basalts (Carmichael, 1967a; Thompson, 1972; Hill & Roeder, 1974). Hill & Roeder (1974) and Sigurdsson (1981) reports crystallization of chromian spinel from above 1250°C. The effect of high alkali content can, however, be expected to lower liquidus temperatures.

#### Origin of chemical variation in the chromian spinels

Sigurdsson (1977) reported 'normal' type zoning in chromian spinel in basalts in LEG 37 samples, with decrease in Cr# towards the rims, and little or no change in the Mg# ratios. He interpreted this zoning as having formed at a stage intermediate between intratelluric and groundmass crystallization, just prior to eruption; the Cr# decrease was believed to reflect the depletion of Cr in the melt with subsequent substitution of Al in octahedral sites in the spinel structure. The constant Mg#, Ti and Fe<sup>3+</sup># was taken to indicate limited Mg fractionation in the melt, and that no variation in temperature and oxygen fugacity was involved. A second type of zoning described by Sigurdsson showed decrease in Mg# and increase in Fe<sup>3+</sup>#, interpreted to reflect Fe/Mg fractionation in the melt together with post-eruptive increase in Ti. In a study of spinels in glassy Mid-Atlantic Ridge basalts, Sigurdsson & Schilling (1976) reported a third type of zoning with increase in Cr<sub>2</sub>O<sub>3</sub> and decrease in Al<sub>2</sub>O<sub>3</sub> towards the rims (i.e. 'reverse' zoning of Sigurdsson (1977)). Dick (1976), Dick & Bryan (1978) and Fisk & Bence (1980) found three generations of spinel crystallization in Mid-Ocean Ridge basalts: low Cr/Al; high Cr/Al; intermediate Cr/Al phase, related to high pressure, low pressure and groundmass crystallization, respectively.

In summary, two important types of zoning in chromian spinels seem to have been observed in abyssal basalts: a high-Al to high-Cr inverse zoning, and a normal zoning with decreasing Cr/Al and zoning towards titanomagnetite. The former type has been related to variations in pressure during crystallization [from high to low pressure], and the latter type to groundmass crystallization at elevated oxygen fugacity, decreasing temperature and changing composition of the melt. The fact that only rare examples of the second type of zoning have been found in this study is likely to reflect quenching of the liquids.

The implications of the interpretation by Sigurdsson (1977) for the zoning observed in this study (Fig. 10) are that it reflects decreasing Cr in the melt, together with rising oxygen fugacity (i.e. increasing magnetite-ulvöspinel component in the spinel structure), occurring just prior to eruption and quenching. An overgrowth of titanomagnetite on chromian spinels has been described in lava groundmass by Haggerty (1976), from Vestmannaeyjar by Steinthorsson (1972) and from Thingmuli by Carmichael (1967a). With rare exceptions (Table 2, anal. SAX 1), titanomagnetite contains less than 1 wt% Cr<sub>2</sub>O<sub>3</sub>, and mostly below the analytical limit of detection (Table 2 and analyses in the publications listed in text to Fig. 15).

Solidus and subsolidus reactions or reequilibration between spinel minerals, silicates or liquids is well documented from layered intrusions (e.g. Henderson & Suddaby, 1971; Henderson, 1975; Roeder et al., 1979), and in basalt lava lakes, flows or sills (Evans & Moore, 1968; Ridley, 1977; Hamlyn & Keays, 1979). The wide variation observed in this study of both Mg# and Cr# in spinels included within compositionally homogeneous olivine phenocrysts in glassy rocks precludes significant subsolidus reequilibration with olivine to lower temperature compositions in Mg# (e.g. Roeder et al., 1979). The slight zoning occasionally observed (Fig. 10) may stem from a reaction between the cooling melt and chromian spinel phenocrysts. However, rapidly quenched basalts, as used in this study, may in part prevent such processes, as also suggested by Sigurdsson & Schilling (1976) and Fisk & Bence (1980).

It has been suggested (e.g. Sigurdsson & Schilling, 1976) that pressure controls the Al-Cr substitution, based on a single experiment by Green et al. (1972). The effect of high pressure can be expected to increase octahedral sites in the melt and consequently decrease chrome in the spinel structure (Burns, 1975; Mysen & Kushiro, 1979). Unfortunately no clear experimental support is available for basaltic rocks (Dickey & Yoder, 1972; Fujii, 1977; Huebner et al., 1976; Osborn et al., 1979). For peridotite compositions a pressure effect on chromian spinel have been shown by Jaques & Green (1980). It is furthermore not very likely that a complex polybaric history is recorded in a non-systematic way in the chromian spinels included within individual olivine phenocrysts. Regional differences between the Vestmannaeyjar and the Snaefellsnes spinels (e.g. in Al-Cr) could be explained as a pressure effect, but could just as well be explained as reflecting compositional differences in the melt (low-Cr content, Table 1).

In order to explain the extensive variation in Cr# it is

reasonable to suggest that local depletion in chrome in the melt may impose a compositional effect on closely spaced nucleation and crystallization sites (Fig. 2). This is due to the very low chrome content, the solubility of chromian spinel in the melt, and the high preference for octahedral sites shown by chrome (Burns, 1970). Such an effect may explain the variation in Cr# from grain to grain, included in homogeneous olivine phenocrysts (Fig. 11). This does not, however, explain the variation in other cations (Mg, Fe<sup>2+</sup>, Fe<sup>3+</sup> and Ti). Small variations or fluctuations in temperature and oxygen fugacity can give rise to some variation in spinel chemistry (Hill & Roeder, 1974, figs. 6 and 7), but even though the amount is unknown, it is not considered likely to explain the total range in e.g. Mg# or Ti.

It can be suggested, mainly from textural relations, that the nucleation of chromian spinel depends on the growth of olivine since chromian spinels are almost always included in, or closely attached to, olivine phenocrysts (Fig. 2). An interpretation of the textural relations may be based on the experimental work by Irvine (1975).

Irvine suggested that chromite-rich layers in stratiform basic intrusions precipitated as a result of sudden blending of a basic magma with a silicic liquid. Increase in silica and alkalis would increase polymerization in the liquid and result in lower frequency of octahedral sites in the melt. A significantly lower solubility for chromian spinel in the liquid would result, and chromian spinel may precipitate suddenly. Irvine based his model on experiments in the system K<sub>2</sub>O-MgO-Cr<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>, and argued that these results should be applied to basaltic magmas. It will tentatively be suggested that crystallization of olivine in the melt can have the same qualitative effect on the local composition, structure and distribution of octahedral sites in the melt. A zone immediately surrounding growing olivine phenocrysts may be locally saturated with respect to chromian spinel. Concentration gradients in the melt adjacent to olivine in a DSDP basaltic glass were examined by Donaldson (1975) who found evidence for Si, Na, Ca, Al enrichment, and significant decrease in FeO/MgO close to the growing crystal-melt interface of a 100 μm crystal in a 20 μm wide zone. Concentration gradients in glass adjacent to olivine phenocrysts have been looked for in this study but not found. This is probably because the growth of olivine macrophenocryst and chromian spinel was terminated prior to eruption and quenching. This is evidenced by a non-equilibrium composition of the olivine macrophenocrysts (Thy, 1982).

The precipitation of chromian spinels in Icelandic alkali glasses cannot be related to details in variation in physical and compositional parameters of the average melt, but may be related to local compositional relations in the melt adjacent to growing crystals. The chemistry may not reflect the average composition of the enclosing glass, but is more likely to reflect local compositional gradients in the melt, e.g. in Ti, Fe<sup>3+</sup>, Cr/(Cr+Al) and Mg/(Mg+Fe<sup>2+</sup>).

The Ti and Fe-rich chromian spinels from the Sudurlend glass impose some additional problems. It is suggested that their variation can be explained by low initial chrome in the

melt, or depletion by previous clinopyroxene crystallization. A higher oxygen fugacity would be a possibility (Hill & Roeder 1974, fig. 6), but this is not reflected in the appearance of titanomagnetite, as compared to the Snaefellsnes glasses (Fig. 13). The first possibility is therefore favoured, and is consistent with the appearance of clinopyroxene in most of the Sudurland glasses. The rare occurrence of chromian spinels in the Sudurland transitional series suggests that this phase never or rarely reached saturation in these rocks.

#### Acknowledgements

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APPENDIX

List of sample localities

Snaefellsnes

KE 1	Kerlingarfjall	hyaloclastite
KE 2	Kerlingarfjall	hyaloclastite
KE 3A	Kerlingarfjall	hyaloclastite (1)
KE 3B	Kerlingarfjall	hyaloclastite (2)
KE 4A	Kerlingarfjall	hyaloclastite (1)
KE 4B	Kerlingarfjall	hyaloclastite (2)
GR 1	Grettisbaeli	hyaloclastite
GR 2	Grettisbaeli	hyaloclastite
BE 1	Kothraunskula	tephra
BE 2	Kothraunskula	tephra
BE 3	Kothraunskula	tephra
ST 1	Stapafell, Stapi	hyaloclastite
ST 2	Stapafell, Stapi	hyaloclastite (3)
EN 3	Enni, Olafsvik	hyaloclastite
EN 5	Enni, Olafsvik	hyaloclastite
MAL 1	Purkholar, Malarrif	tephra
SAX 1	Saxholar	tephra
SAX 2	Saxholar	tephra
SAX 3	Saxholar	tephra

Vestmannaeyjar

VE 46	Brandur	hyaloclastite
VE 76	Saefell, Heimaey	hyaloclastite (3)
SY 1	Syrtingur	tephra (4)
SU 19	NE Surtsey	tephra (5)
VE 112	Eldfell, Heimaey	tephra (6)
VE 156	Eldfell, Heimaey	tephra (7)

Sudurland

RE 1	Reidskard	hyaloclastite
HA 3	Hagafell, S of Laufafell	hyaloclastite (3)
HA 6	Hagafell, S of Laufafell	hyaloclastite
EY	Krossardalur, Eyjafjallajökull	lava

- (1) Relatively unconsolidated and coarse grained.
- (2) Consolidated mineral-rich, fine grained layer.
- (3) In part palagonitized.
- (4) Tephra fall June 6, 1965.
- (5) Tephra fall Nov. 1963 - Jan. 1964.
- (6) Tephra fall Jan. 23, 1973.
- (7) Tephra fall May 2-3, 1973.

All samples from Vestmannaeyjar collected by S.P. Jakobsson,  
Museum of Natural History, Reykjavik.

EY collected by K.A. Jørgensen, Univ. Copenhagen (sample 22087 W).