

K. Jónasson

## Rhyolite volcanism in the Krafla central volcano, north-east Iceland

Received January 14, 1994 / Accepted June 8, 1994

**Abstract** At the Krafla central volcano in north-east Iceland, two main phases of rhyolite volcanism are identified. The earlier phase (last interglacial) is related to the formation of a caldera, whereas the second phase (last glacial) is related to the emplacement of a ring dike. Subsequently, only minor amounts of rhyolite have been erupted. The volcanic products of Krafla are volumetrically bimodal. Geochemically, there is a series of basaltic to basalto-andesitic rocks and a cluster of rhyolitic rocks. Rocks of intermediate to silicic composition (icelandites and dacites) show clear signs of mixing. The rhyolites are Fe-rich (tholeiitic), and aphyric to slightly porphyritic (plagioclase, augite, pigeonite, fayalitic olivine and magnetite). They are minimum melts on the quartz–plagioclase cotectic plane in the granite system (Qz–Or–Ab–An). The rhyolites at Krafla were produced by near-solidus, rather than near-liquidus fractionation. They are interpreted as silicic minimum melts of hydrothermally altered crust, mainly of basaltic composition. They were primarily generated on the peripheries of an active basaltic magma chamber or intrusive domain, where sufficient volumes of crust were subjected to temperatures favorable for rhyolite genesis (850–950°C). The silicic melts were extracted crystal-free from their source in response to crustal deformation.

**Key words** Rhyolite · petrology · minimum melts near-solidus fractionation · mixing · Krafla · Iceland

### Introduction

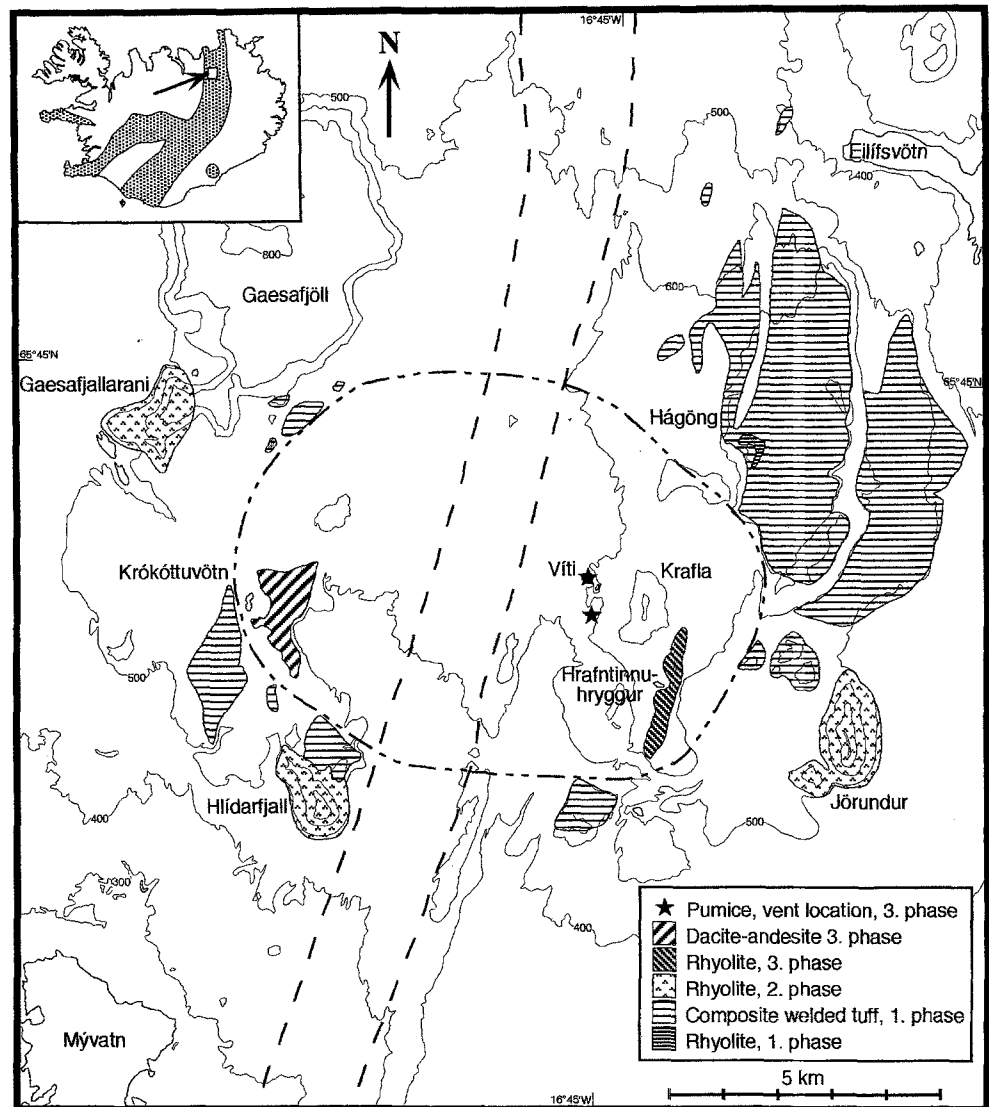
Understanding the production of silicic material in oceanic crust is essential in explaining the origin of continental crust. Significant amounts of silicic material

have been produced at numerous central volcanoes in Iceland, providing excellent opportunities to study processes that may have played a major part in generating ancient continental material.

Silicic rocks in Iceland range from subalkaline tholeiitic rhyolites and dacites to peralkaline silicic rocks (Jakobsson 1972; Óskarsson et al. 1982). Calc-alkaline silicic rocks have also been found (Jónasson et al. 1992). Theories about the origin of the silicic rocks in Iceland can be divided in two main groups. One explains the origin of the silicic rocks by fractional crystallization from basaltic magmas (e.g. Carmichael 1964; Macdonald et al. 1990), sometimes coupled with assimilation or contamination (e.g. Macdonald et al. 1987; Nicholson et al. 1991). The other explains the origin of the silicic rocks by partial melting of crustal basaltic rocks (e.g. O’Nions and Grönvold 1973; Sigvaldason 1974; Sigmarsson et al. 1991), often hydrothermally altered (e.g. Óskarsson et al. 1982; Condomines et al. 1983; Hemond et al. 1988). Partial melting of amphibolite facies rocks in the lower crust has been proposed to explain the origin of the calc-alkaline silicic rocks (Jónasson et al. 1992). Finally, modification of the silicic magmas by mixing (e.g. Sigurdsson and Sparks 1981; McGarvie et al. 1990) or by remelting (e.g. Sigurdsson 1977; Marsh et al. 1991) has often been invoked.

The present study attempts to shed some light on these processes by studying in detail the silicic components of volcanic products at a single extensively studied volcano: the Krafla central volcano in northern Iceland. The Krafla volcanic system is located in the northern rift zone of Iceland. The system is composed of a central volcano (Fig. 1), roughly 15 km in diameter, and a 100 km long fissure swarm that bisects the volcano. During the last interglacial ( $\approx 100\,000$  BP) a caldera was formed within the central volcano (Saemundsson 1991). The caldera is now filled to the rims with hyaloclastite rocks and lavas. The silicic rocks of the Krafla central volcano are tholeiitic, mainly rhyolites (Table 1). The volcano has been studied extensively during the last two decades because of geothermal exploration

**Fig. 1** Simplified geological map of the Krafla central volcano showing the main tectonic features and the occurrence of silicic volcanic rocks. The inset map shows the neo-volcanic zones (shaded) of Iceland and the location of the Krafla central volcano (arrow). The broken lines outline the caldera and the presently active fissure swarm. The geology is based on the maps of Saemundsson (1991) with some minor modifications. The silicic volcanic rocks are concentrated at and around the caldera rim and away from the central part of the fissure swarm



and development (e.g. Ármannsson et al. 1987) and because of a rifting event with nine basaltic eruptions between 1975 and 1984 (e.g. Björnsson 1985).

### Occurrence of silicic rocks

At the Krafla central volcano two main phases of rhyolitic volcanism can be observed, each producing about  $1 \text{ km}^3$  of rhyolite. The first rhyolite phase (around 100000 BP) produced a small ( $<0.05 \text{ km}^3$ ) rhyolitic dome in Hágöng (Fig. 1) and the silicic component of a composite welded tuff which has been related to the formation of the caldera at Krafla (Björnsson et al. 1977, 1979; Calderone et al. 1990; Saemundsson 1991). The tuff is of mixed intermediate composition and has a total volume of about  $2.5 \text{ km}^3$ , containing about  $1 \text{ km}^3$  of the rhyolitic component (Calderone et al. 1990). This rhyolite phase began with a dome extrusion of pitchstone and fine-grained rhyolite. Later, a Plinian eruption commenced, producing rhyolitic ash mixed

with basaltic scoria. This eruption continued with Strombolian activity, producing incompletely mixed rhyolitic and basaltic pyroclasts. The main source of the eruption was near the eastern caldera rim (Calderone et al. 1990). Abundant lithic fragments are found in the tuff layer, including gabbroic and granophyric xenoliths, basalt and other volcanoclastic fragments.

In the second phase, which occurred during the last glacial period (Saemundsson 1991), three subglacial rhyolite ridges (Jörundur, Hlíðarfjall and Gaesafjallarani; Fig 1) formed outside the caldera rim. They form steep short ridges subparallel to the caldera rim. The ridges (subglacial equivalents of domes) rise about 300 m above their surroundings (800 m asl) with a total volume of  $0.7 \text{ km}^3$ . They are built up of lobes (Grönvold 1972; Saemundsson 1972) of lithic rhyolite with an outer layer of obsidian and a flow-banded and flow-folded, vesicular, glassy rhyolite in between. The lobes are enclosed in hyaloclastite that consists of fragments of the same material as found in the lobes (type 2 hyaloclastite: Furnes et al. 1980). Overall similarity in

**Table 1** Major and trace element composition of representative samples from the Krafla central volcano

	Basaltic rocks Krafla central volcano							Rhyolites							Mixed eruptives Króttuvötn							Granophytic rocks Krafla central volcano							Experi- mental melt
								1. phase							2. phase							3. phase							
	1	2	3	4	5	6	7	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26		
SiO <sub>2</sub>	49.70	50.75	50.53	51.14	50.84	53.85	72.22	72.58	74.46	74.70	74.08	74.25	74.13	75.28	74.38	73.24	73.23	50.01	52.70	62.25	66.85	73.70	74.27	71.84	72.44	73.09			
TiO <sub>2</sub>	0.84	1.80	1.79	1.95	2.21	1.87	0.46	0.61	0.37	0.39	0.32	0.30	0.36	0.35	0.29	0.42	0.48	1.16	1.18	0.93	0.83	0.43	0.48	0.69	0.54	0.22			
Al <sub>2</sub> O <sub>3</sub>	15.76	14.00	13.92	13.23	13.49	12.96	12.12	12.34	12.09	12.04	12.18	12.09	11.78	12.36	11.98	13.25	12.57	16.91	15.00	13.88	13.24	12.35	12.74	12.15	12.99	12.91			
FeO*	9.42	12.24	14.11	15.00	14.99	15.02	4.62	4.64	3.72	3.94	3.61	3.53	3.42	2.84	3.59	4.01	3.75	10.25	9.83	7.37	6.11	3.30	3.08	4.75	4.00	3.77			
MnO	0.18	0.22	0.25	0.26	0.27	0.25	0.14	0.10	0.09	0.09	0.10	0.09	0.09	0.09	0.09	0.10	0.11	0.09	0.20	0.18	0.17	0.07	0.08	0.11	0.11	0.05			
MgO	8.75	6.99	5.85	5.21	5.04	5.01	0.25	0.60	0.15	0.18	0.11	0.09	0.14	0.12	0.10	0.71	0.36	6.51	6.12	3.08	1.68	0.43	0.21	1.31	0.78	0.56			
CaO	13.20	11.35	10.49	10.09	9.70	7.38	2.10	2.74	1.95	2.07	1.93	1.90	1.92	1.94	1.80	2.70	2.68	12.93	11.61	6.87	4.85	2.10	1.92	3.04	3.20	2.32			
Na <sub>2</sub> O	1.76	2.00	2.35	2.34	2.53	2.39	5.07	3.64	4.20	3.69	4.63	4.70	4.96	3.92	4.60	2.61	3.54	1.60	2.48	3.43	3.86	4.25	3.94	4.01	3.84	4.32			
K <sub>2</sub> O	0.17	0.28	0.36	0.36	0.46	0.62	2.72	2.51	2.86	2.62	2.80	2.84	2.92	2.83	2.96	2.70	3.05	0.19	0.54	1.59	1.98	2.97	2.92	1.66	1.71	2.39			
P <sub>2</sub> O <sub>5</sub>	0.10	0.24	0.22	0.27	0.33	0.48	0.12	0.10	0.10	0.11	0.08	0.05	0.12	0.11	0.05	0.08	0.11	0.13	0.20	0.28	0.28	0.26	0.22	0.30	0.23	0.35			
mg #	65.3	53.6	45.6	41.3	40.5	40.3	10.0	20.9	7.5	8.4	6.0	4.9	7.8	7.8	5.2	26.5	16.1	56.2	55.7	45.8	35.8	20.8	12.1	35.8	28.2	23.2			
Rb	2	4	5	4	7	8	60	49	59	62	65	65	62	54	59	48	63	2	9	29	40	58	62	35	42				
Sr	108	155	139	141	153	182	125	108	102	112	124	112	119	137	112	110	107	139	131	129	126	146	147	142	201				
Ba	36	67	79	92	113	117	606	488	572	567	637	610	478	457	648	640	564	40	126	345	440	360	340	294	427				
Y	22	30	37	43	49	76	119	105	131	132	144	127	124	130	125	138	104	29	43	79	100	96	113	94	138				
Zr	54	94	104	114	157	156	601	437	549	557	481	407	490	434	512	515	470	74	187	443	578	321	293	258	411				
Ni	136	86	63	38	27	69	25	2	14	26	3	3	22	24	9	18	15	69	59	36	19	1	30	37	30				
Co	46	63	71	63	77	63	2	1	5	7	1	17	17	1	6	7	6	33	20	11	10	1	1	17	1				
Cr	305	209	55	20	19	20	1	6	6	8	7	9	1	3	6	17	15	230	186	83	38	1	25	46	13				
V	273	374	444	502	492	567	65	97	52	90	110	127	155	207	44	85	75	293	266	143	80	275	291	268	336				
Sc	44	47	50	52	49	33	7	12	8	9	7	4	6	6	6	9	8	47	43	25	19	7	10	9	7				
Cu	120	124	139	163	173	166	18	44	22	32	36	27	34	34	13	30	28	143	126	71	38	17	49	73	53				
Zn	71	108	124	140	146	158	124	114	108	121	148	129	125	162	142	139	82	84	94	114	120	97	88	74	103				

The samples were analysed at the Nordic Volcanological Institute by N. Óskarsson, using inductively coupled plasma atomic emission spectrometry, except for Rb, which was analysed by atomic absorption spectrometry. The analyses are presented as 100% volatile-free compositions. Oxide concentrations are shown in wt.%; total iron as FeO\*; Mg-number = 100 \* Mg/(Mg + Fe<sup>2+</sup>) with iron normalized to Fe<sub>2</sub>O<sub>3</sub>/FeO = 0.15; trace element concentrations are in ppm. Sample key (numbers in parentheses give original sample numbers): 1, olivine tholeiite from Gjástykkisbunga (KK 31); 2, scoria from eruption in 1984 (KK 152); 3, tholeiite from Eldá, Hverfjallseidar (KK 99); 4, tholeiite, south of Hvannstód (KR 10); 5, tholeiite from ridge west of Hrafninnuhryggur (KR 36); 6, basaltic andesite from ridge west of Hrafninnuhryggur (KR 40); 7, pitchstone from Hággöng (KK 7); 8, pumice from bottom layer of composite welded tuff, south-west of Elflisvötn (A1); 9, rhyolite from Gaesafjallarani (KR 14); 10, glassy rhyolite from Gaesafjallarani (KK 3); 11, rhyolite from Hlíðarfjall (KR 21); 12, obsidian from Hlíðarfjall (KR 24); 13, obsidian from Hrafninnuhryggur (KR 29); 14, obsidian from Hrafninnuhryggur (KR 42); 15, obsidian from Hrafninnuhryggur (A-THO); 16, pumice from Hveragl (6865); 17, pumice from Víti, AD 1724 (KK 56); 18, hyaloclastite beneath lava at Króttuvötn (KR 9); 19, porphyritic basaltic andesite from small lava flow in southern part of Króttuvötn lavas (KR 4); 20, andesite from Króttuvötn lavas (KR 11); 21, dacite from Króttuvötn lavas (KR 6); 22, granophyre xenolith from Víti, AD 1724 (3703); 23, granophyre xenolith from Víti, AD 1724 (7242); 24, granophyre from borehole KG 25, near center of caldera, 2080 m (KG 25-2080); 25, granophyre from borehole KJ 16, south-east part of caldera, 1074 m (KJ 16-1074); and 26, experimental melt composition, 10% melt of Kilauca tholeiitic basalt, 913°C, GCH buffer, 1 kb (87A; Spulber and Rutherford 1983)

**Table 2** Composition of representative mafic phases in rhyolites from the Krafla central volcano

	Rhyolite, 1. phase		Rhyolite, 2. phase				Rhyolite 3. phase (Víti pumice, 1724)						
	1 Augite	2 Augite	3 Augite	4 Olivine	5 Augite	6 Olivine	7 Augite	8 Augite	9 Pigeonite	10 Augite	11 Pigeonite	12 Pigeonite	13 Olivine
SiO <sub>2</sub>	52.46	50.58	49.20	30.12	48.84	29.77	50.88	50.91	50.15	50.92	50.82	50.15	30.66
TiO <sub>2</sub>	0.39	0.37	0.30		0.34		0.33	0.33	0.18	0.32	0.23	0.21	
Al <sub>2</sub> O <sub>3</sub>	1.58	1.08	0.82		0.82		0.94	1.09	0.43	0.85	0.50	0.42	
FeO	12.75	18.92	25.44	62.43	26.74	65.55	17.31	19.03	32.06	21.48	33.42	34.53	62.16
MnO	0.48	0.64	0.77	1.57	0.68	1.62	0.52	0.57	0.97	0.63	1.03	1.04	1.40
MgO	13.64	8.97	4.99	4.96	3.02	2.35	11.01	10.20	12.01	8.75	9.40	9.70	5.18
CaO	18.59	18.95	17.54	0.00	18.89	0.00	18.59	17.86	4.00	17.13	5.14	3.87	0.00
Na <sub>2</sub> O	0.33	0.22	0.21		0.19		0.19	0.20	0.04	0.19	0.10	0.05	
Total	100.23	99.73	99.28	99.08	99.51	99.29	99.78	100.18	99.84	100.27	100.62	99.98	99.40
Mg-no.	64.8	45.0	25.3	12.1	16.4	5.9	53.9	48.7	39.3	41.3	32.7	32.7	12.7
Wo	38.8	40.6	39.0		42.5		39.5	38.0	8.6	36.8	11.4	8.6	
En	39.6	26.7	15.4		9.4		32.6	30.2	35.9	26.1	29.0	29.9	
Fs	20.8	31.6	44.2		46.9		27.0	30.9	53.8	36.0	57.8	59.7	

The analyses were carried out at the Nordic Volcanological Institute on an ARL SEMQ electron microprobe. Oxide concentrations are shown in wt.%. Mg-number =  $100 * \text{Mg} / (\text{Mg} + \text{Fe}^{2+} + \text{Mn})$ , with iron(III) calculated to give stoichiometric results. Wo = Ca, En = Mg, Fs =  $\text{Fe}^{2+} + \text{Mn}$ . Key to analyses [numbers in parentheses give sample numbers (Table 1)]: 1, augite phenocryst, rhyolitic pumice from composite welded tuff layer (8); 2, augite phenocryst, pitchstone from Hágöng (7); 3, augite phenocryst, glassy rhyolite from Gaesafjallaráni (10); 4, fayalitic olivine phe-

nocryst, glassy rhyolite from Gaesafjallaráni (10); 5, augite phenocryst, obsidian from Hlíðarfjall (12); 6, fayalitic olivine phenocryst, obsidian from Hlíðarfjall (12); 7, augite phenocryst, pumice from Víti (17); 8, augite phenocryst, pumice from Víti (17); 9, pigeonite phenocryst, pumice from Víti (17); 10, augite phenocryst, pumice from Víti (17); 11, pigeonite phenocryst, pumice from Víti (17); 12, pigeonite phenocryst, pumice from Víti (17); and 13, fayalitic olivine phenocryst, pumice from Víti (17)

shape, composition and age indicates that the ridges (domes) represent a distinct short-lived phase in the history of Krafla, possibly related to the emplacement of a ring dike.

Subsequently, only small amounts of silicic material have been erupted in a few events, all occurring within the caldera. For practical reasons, they will be grouped together as the third phase of rhyolite volcanism. Hrafninnuhryggur, a small ridge (<0.05 km<sup>3</sup>) of rhyolite composition (Fig. 1), formed just inside the south-eastern caldera rim during late glacial time. The rhyolite is aphyric with abundant obsidian. The ridge is parallel to the main tectonic trend. Towards the end of the last glacial period, a mixed eruption occurred near Krókóttuvötn (Fig. 1), just inside the western caldera-rim, producing a series of hyaloclastites and lava flows composed of plagioclase porphyritic basalts, basaltic andesites, andesites and dacites. The basalts and basaltic andesites contain up to 15% plagioclase megacrysts (up to 5 mm). Such megacrysts are also found in the andesites and dacites. Two post-glacial eruptions of rhyolitic pumice occurred at around 9000 BP (Hveragil tephra; Saemundsson 1991), and in 1724 (Víti, Fig. 1). The 1724 eruption, which marked the beginning of the 'Mývatn fires' rifting event (Thoroddsen 1907–15; Saemundsson 1991), was phreatic, producing mostly thermally altered material, with minor juvenile rhyolitic pumice, basaltic scoria and granophyric xenoliths.

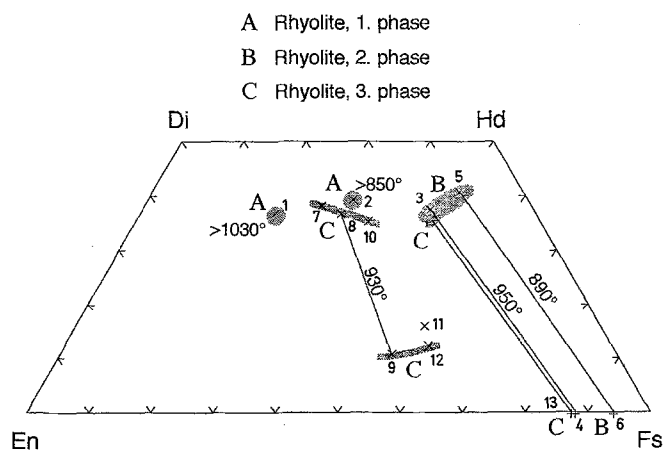
In addition to volcanic silicic rocks, granophyric rocks have been found in drill cuttings from geothermal wells in the central and south-eastern parts of the Kra-

fla caldera (Fig. 1) (Ármansson et al. 1987). They are all highly altered by hydrothermal activity. Granophyric rocks are also found as xenoliths in some of the rhyolite formations.

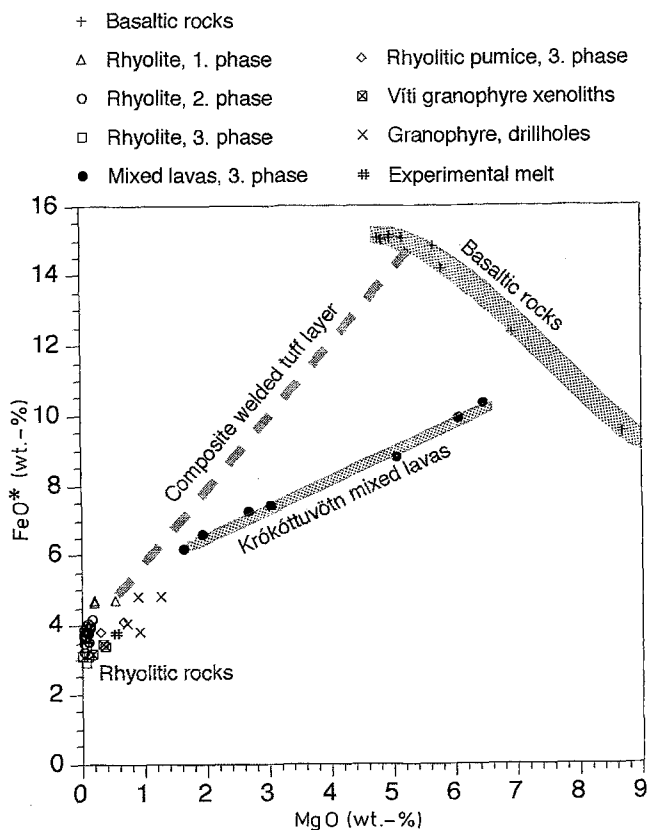
Heidarspordur, a rift segment to the south of the Krafla central volcano, produced basalt, basaltic andesite, icelandite and dacite in a series of eruptions at the onset of the Holocene (9000–10000 BP). These are, however, not directly related to the Krafla central volcano. It has been suggested that they are a part of a separate volcanic system (Saemundsson 1974, 1978, 1991).

### Petrography and mineral chemistry

The Krafla rhyolites are aphyric to slightly porphyritic. They are fine-grained or glassy (obsidian), occasionally vesicular or pumiceous. The phenocrysts are plagioclase, augite, pigeonite, fayalitic olivine and magnetite; microlites of apatite and zircon are also found. Sub- to euhedral crystals with compositions broadly compatible with the host liquid are regarded as equilibrium phases, whereas crystals showing signs of resorption, or having compositions clearly incompatible with the host liquid, are regarded as xenocrysts. The temperatures mentioned in the following are based on pyroxene thermometry (Lindsley 1983; Fig. 7 in Davidson and Lindsley 1989). Augite alone gives minimum temperatures, whereas two pyroxenes or augite–olivine gives equilibrium temperatures with an estimated accuracy of  $\pm 50^\circ$ .



**Fig. 2** Compositions of the pyroxene and olivine phenocrysts from the Krafla rhyolites. Pyroxene quadrilateral projection of Lindsley (1983: 486). Small numbers refer to Table 2. Numbers 1, 2, 3, 5, 7, 8 and 10 are augites; numbers 9, 11 and 12 are pigeonites; and numbers 4, 6 and 13 are fayalitic olivines. Shaded areas indicate observed variation. Thin lines represent apparent equilibrium tie-lines, with calculated temperatures



**Fig. 3** Total Fe as FeO versus MgO in rocks from the Krafla central volcano. Concentrations calculated as 100% volatile-free. Basaltic rocks range from olivine tholeiites to Fe-rich tholeiites and basaltic andesites. Clear signs of mixing or hybridism are seen in all rocks of andesitic to dacitic composition. Lines for composite welded tuff layer and basaltic rocks are based on analyses from Calderone et al. (1990) and Nicholson (1990). In addition to the composite welded tuff layer, some mixed tephra layers and hyaloclastites plot along the same line

The rhyolites from the first phase contain <0.5 vol.% of plagioclase ( $An_{42-45}$  and  $An_{60-67}$ ) and green augite (Table 2 and Fig. 2). The andesine plagioclase is chemically close to equilibrium with the rhyolite liquid, whereas the labradoritic plagioclase is clearly xenocrystic. There are also two populations of augite (Fig. 2), ferroaugites ( $Wo_{41}En_{27}Fs_{32}$ ), indicating a minimum temperature of  $850 \pm 50^\circ C$ , and xenocrystic augites ( $Wo_{39}En_{40}Fs_{21}$ ) which indicate minimum temperatures of  $1030 \pm 50^\circ C$ . The xenocrysts represent phenocrysts from the coerupted basaltic magma.

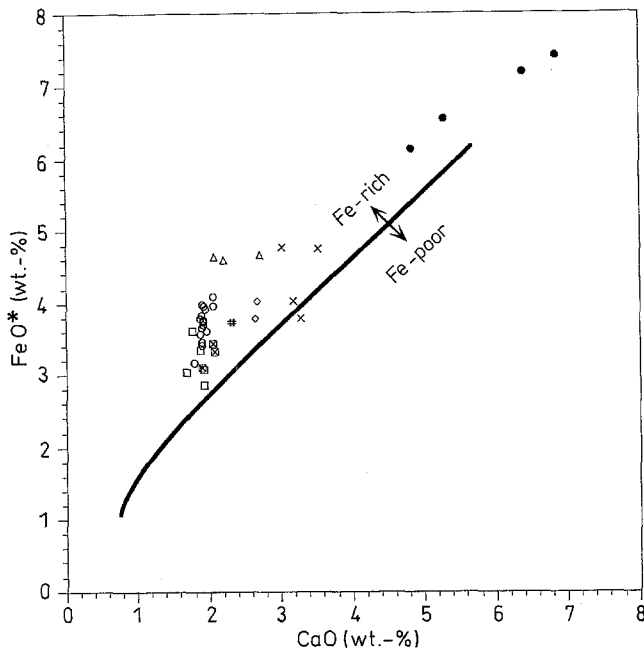
The rhyolites of the second phase (Jörundur, Hlíðarfjall and Gaesafjallaráni) contain up to 7 vol.% phenocrysts, mainly plagioclase ( $An_{34-43}$ ), with trace amounts of green ferroaugite (Table 2), fayalitic olivine ( $Fo_{5-13}$ , Table 2) and magnetite ( $Usp_{58-65}$ ); microlites of apatite and zircon are also found. The green ferroaugites and the fayalitic olivines were in equilibrium with the rhyolite liquids and indicate equilibrium temperatures of  $850 \pm 50^\circ C$  to  $950 \pm 50^\circ C$  (Fig. 2). The compositional variation from grain to grain and within grains is generally insignificant. In some instances, however, the cores of plagioclases display slight reverse zoning, followed by normal zoning at the rims (e.g.  $An_{38}-An_{42}-An_{40}$ ).

The silicic rocks of the third phase are more variable. The Hrafninnuhryggur rhyolite is aphyric and varies from being very fine-grained to glassy (obsidian). Phenocrysts in the Krókóttuvötn dacites and andesites are all of basaltic affinity. The silicic end-member was probably aphyric at the time of mixing and eruption. In the AD 1724 Víti pumice, plagioclase phenocrysts ( $An_{34-48}$ ) are found together with augite (Table 2), pigeonite, fayalitic olivine ( $Fo_{13}$ ) and magnetite phenocrysts. Augite and pigeonite equilibrated at  $930 \pm 50^\circ C$  (Fig. 2). The augites and pigeonites form a trend with varying Mg-number. This trend is roughly parallel with the isotherms (Fig. 7 in Davidson and Lindsley 1989) and zonation can be in either direction.

The types and compositions of the ferromagnesian silicates of the Krafla rhyolites indicate oxygen fugacities on or below the FMQ oxygen buffer curve (Carmichael 1967; Warshaw and Smith 1988). This is also indicated by the bulk chemical composition.

## Geochemistry

Whole rock analyses of rocks from the Krafla central volcano span the whole range from olivine tholeiites to rhyolites. Volumetrically, however, the products are bimodal. Basaltic rocks predominate, ranging from olivine tholeiites to Fe-rich tholeiites and basaltic andesites. Rhyolites are also present in significant amounts (Fig. 3). Andesitic to dacitic rocks are scarce and they all show clear signs of mixing or hybridism between silicic and basaltic end-members. These rocks are mainly the mixed lavas at Krókóttuvötn and the composite welded tuff layer, as well as mixed tephra layers (e.g.



**Fig. 4** Total Fe as FeO versus CaO in silicic rocks from the Krafla central volcano. Concentrations calculated as 100% volatile-free. Symbols as in Fig. 3. The line separates higher Fe types from lower Fe types (Warshaw and Smith 1988). The Krafla rocks are of the higher Fe type

Víti AD 1724) and hyaloclastites. Nicholson et al. (1991) reported smooth, continuous variation trends in the chemistry of rocks from the Krafla area. Their variation diagrams included the mixed rocks from the Krafla central volcano and intermediate to silicic rocks from the Heidarspordur rift segment, south of Krafla, which are not directly related to the Krafla central volcano. Removing these samples from the plots leaves a bimodal distribution (Daly gap) with no rocks having  $\text{SiO}_2$  contents in the range 57–71 wt.-%.

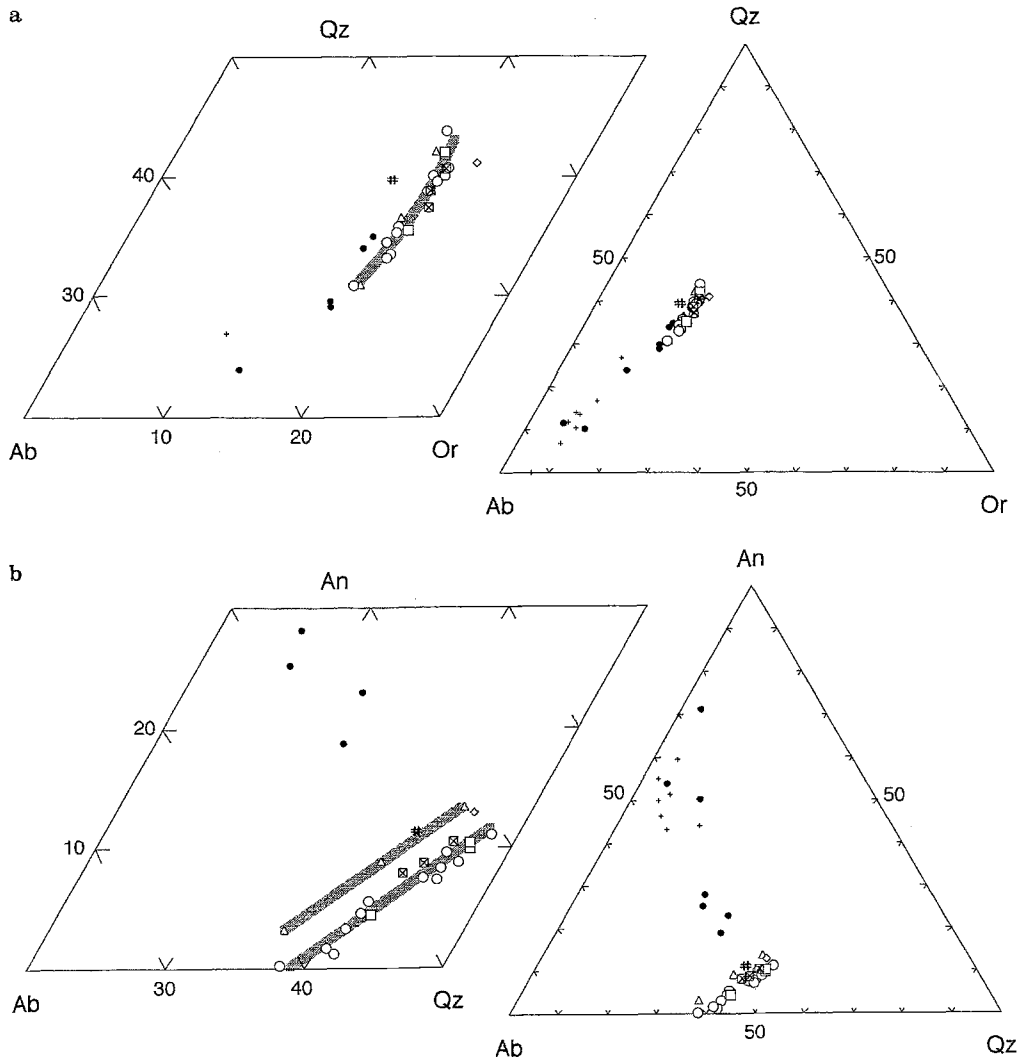
The rhyolites from Krafla have CaO contents around 2 wt.%, with FeO varying between 3 and 4.6 wt.% (Table 1 and Fig. 4). The pumice samples and the granophyres from drillholes have higher Ca contents, most likely due to contamination with basaltic material or hydrothermal calcite. The silicic rocks from Krafla are all of the higher Fe type (Warshaw and Smith 1988). Rocks of the higher Fe type usually precipitate orthopyroxene or fayalite, are either peralkaline or metaluminous to subaluminous, and have oxygen fugacities on or below the FMQ oxygen buffer curve (Warshaw and Smith 1988). Melting experiments have shown that if such rocks are generated by equilibrium crystallization or melting, then the water pressure was less than 1 kb. Higher water pressures lead to the production of rocks with a lower Fe content (e.g. Helz 1976; Spulber and Rutherford 1983; Thy et al. 1990). The compositions of the Krafla rhyolites compare well with the composition of a 10% melt (Table 1) of a Kilauea basalt at 913°C, 1 kb total pressure and 0.67 kb

water pressure, leaving a residue of plagioclase (49 wt.%), high Ca pyroxene (32 wt.%), low Ca pyroxene (12 wt.%) and Fe–Ti oxides (7 wt.%) (Spulber and Rutherford 1983). The main difference is that the experimental melt contains slightly less  $\text{TiO}_2$  and more  $\text{P}_2\text{O}_5$ . This will be addressed in the discussion.

The chemistry of the Krafla rhyolites is best described by projections in the granite system (Ab–An–Or–Qz– $\text{H}_2\text{O}$ ). These components represent around 90 wt.% of the compositions of the rhyolites. The Krafla rhyolites define a cotectic trend from the minimum point at An=0 up along the quartz–feldspar plane (Fig. 5). They are apparently close to being in equilibrium with plagioclase of similar composition to the phenocrysts they contain, and with quartz (e.g. Nekvasil 1988). Quartz phenocrysts have, however, not been observed in the rhyolites. The rhyolites are rather low in the Or component and far from being saturated with potassic feldspar. The rocks from the first phase of rhyolitic volcanism define a trend that is slightly displaced away from the Qz corner relative to the trend that the other rocks define (Fig. 5b). This indicates that the earliest rhyolites equilibrated at higher total pressures, or lower partial pressures of water ( $P_{\text{H}_2\text{O}}/P_{\text{Total}}$ ), than later rocks (Nekvasil 1988).

The elements that are not included in the granite system, primarily Fe, are not abundant in the Krafla rhyolites, and they do not correlate with the trend defined in the granite system. Trace and minor elements in the rhyolites have poor correlations internally and with major elements (Figs. 6 and 7). The rhyolites plot as a cluster rather than defining a clear chemical trend. Each unit has its own distinct signature with respect to some elements (Fig. 7), however, a feature also seen in other Icelandic central volcanoes, such as Torfajökull (Marsh et al. 1991). This could indicate separate, slightly different, source compositions, even though the genetic processes are the same.

When trace element concentrations in the Krafla rhyolites are compared with those of a representative basalt from Krafla (e.g. No. 4, Tables 1 and 3), the rhyolites are seen to be enriched in Rb, Th and U by a factor of 8–12 (Table 3). Assuming that these elements are incompatible in all phases present in the Krafla rocks, the rhyolites could be produced by 90% crystallization from a basaltic magma or 10% melting of basaltic crust. Other elements show less enrichment. The elements Zr, Nb, Hf and the light rare earth elements (REEs) (La–Nd) are enriched by a factor of about five compared with the basalt. The middle to heavy REEs (Sm–Lu), together with Y, are enriched by a factor of about three, with the exception of Eu, which is only slightly enriched. Sr is slightly depleted in the rhyolites (Table 3). The REE patterns (Fig. 8) can be described in terms of the ratios La/Sm and Sm/Lu. The La/Sm ratio in the rhyolites is almost twice as high as in the tholeiite basalt (Table 3), whereas the Sm/Lu ratio is about the same. The transition metals Ni, Co, V, Cr and Sc are depleted in the rhyolites, the concentrations being



**Fig. 5a, b** Composition of rocks from the Krafla central volcano in the 'granite' system. Symbols as in Fig. 3. Values are calculated normative wt.% of the minerals quartz, albite, orthoclase and anorthite. The plots are in full size on the right, whereas on the left the area of the silicic rocks has been enlarged to show details. **a** Qz, Ab and Or. **b** An, Ab and Qz. The Krafla rhyolites form cotectic trends on the quartz-feldspar plane. The two separate trends depicted in **b** are not separable in **a**. The rhyolites are not saturated with-K-feldspar

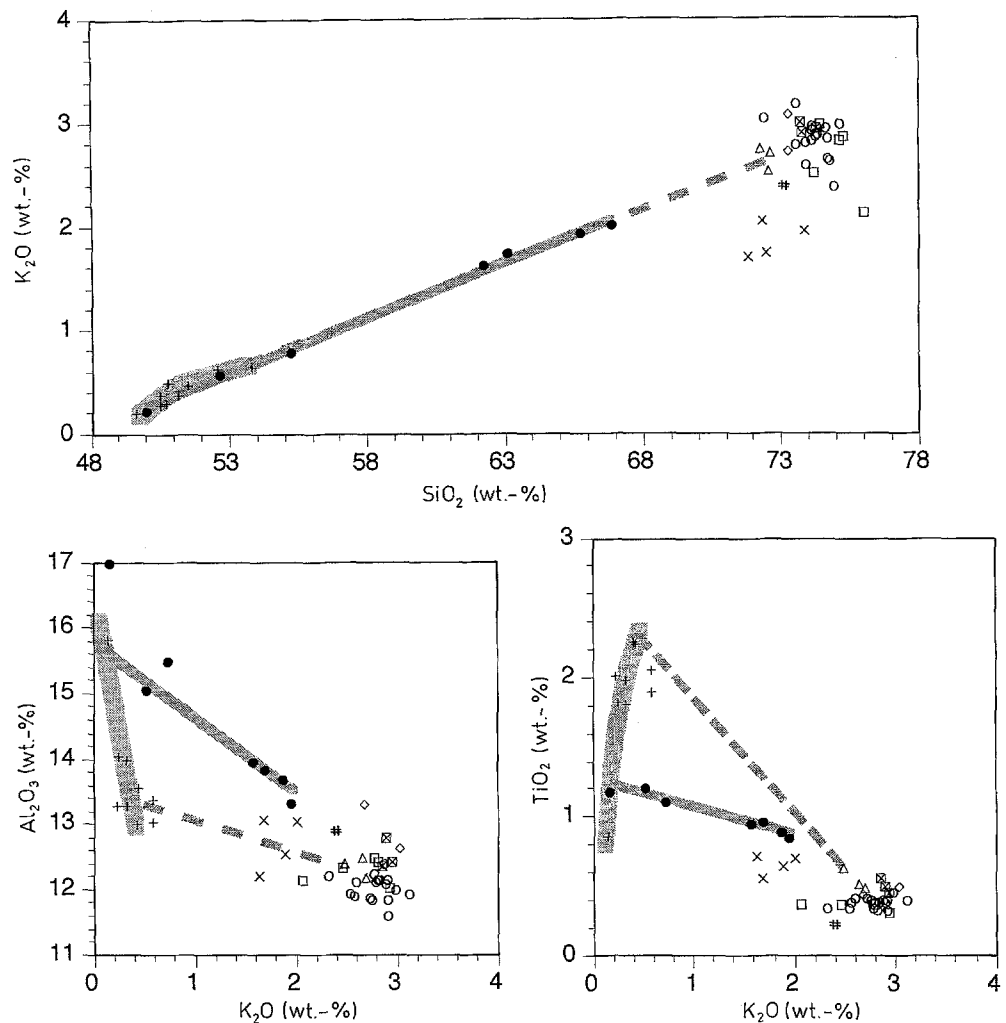
in general about 10 times lower than the concentrations in a representative basalt (Table 1).

Measured partition coefficients for silicic systems vary by an order of magnitude due to variations in melt structure (Arth 1976; Mahood and Hildreth 1983; Nash and Crecraft 1985). Relative to one another, however, the partition coefficients for different trace elements display a consistent pattern dominated by crystal-chemical controls (Michael 1988; Sisson 1991). It is therefore appropriate to compare the pattern produced in the rhyolites with the general partitioning patterns of minerals present in the Krafla system.

The depletion of Sr and the distinct negative Eu anomaly in the rhyolites are a strong indication of plag-

ioctase control. The overall pattern of the REEs (light REE enrichment and flat, slightly concave pattern for the middle to heavy REEs) indicates that clinopyroxene was an important phase in the petrogenesis of these rocks. This is supported by the depletions of Cr, Ni, Co and Sc. Other phases do not seem to have affected the trace element pattern of the rhyolites significantly. Partition coefficients for the high field strength elements (HFSEs) (e.g. REEs, Y, Zr, Nb and Hf) in orthopyroxene and olivine are smaller than for clinopyroxene and the overall patterns are similar. Their influence is therefore difficult to discern. The depletion of V indicates magnetite control. Zircon fractionates strongly between the middle and heavy REEs, the partition coefficient for Lu being almost two orders of magnitude greater than for Sm (Mahood and Hildreth 1983). The concentrations of Zr and Hf are relatively high in the rhyolites and the Sm/Lu ratios are similar in the rhyolites and the tholeiite basalt. Zircon has thus evidently not had much influence on the trace element pattern of the Krafla rhyolites. The REEs and Y partition strongly into apatite, whereas Zr is incompatible (Pearce and Norry 1979). As the enrichment of Zr in

**Fig. 6** Composition of rocks from the Krafla central volcano in terms of  $K_2O$ ,  $SiO_2$ ,  $Al_2O_3$  and  $TiO_2$ . Concentrations calculated as 100% volatile-free. Symbols and lines as in Fig. 3. The Krafla rhyolites plot as a cluster, rather than defining a clear trend. This relationship holds for all elements except for those that are included the 'granite' system



the rhyolites is similar to that for the REEs, it does not seem likely that apatite played a significant part in their genesis, even though  $P_2O_5$  contents are lower in the rhyolites than in the basalts. The apparent  $P_2O_5$  anomaly will be addressed in the discussion.

Least-squares modeling of fractional crystallization (method of Wright and Doherty 1970), based on representative mineral compositions of plagioclase, augite, olivine, orthopyroxene, magnetite, ilmenite and apatite, indicated that about 80% crystallization from a basaltic parent ( $MgO \approx 6$  wt.%) could produce a magma similar to the Krafla rhyolites (Nicholson et al. 1991). Excess enrichment of the elements K, Rb, Th and U (indicating 90% crystallization) were explained by assimilation of a hydrothermally altered wallrock contaminant (Nicholson et al. 1991).

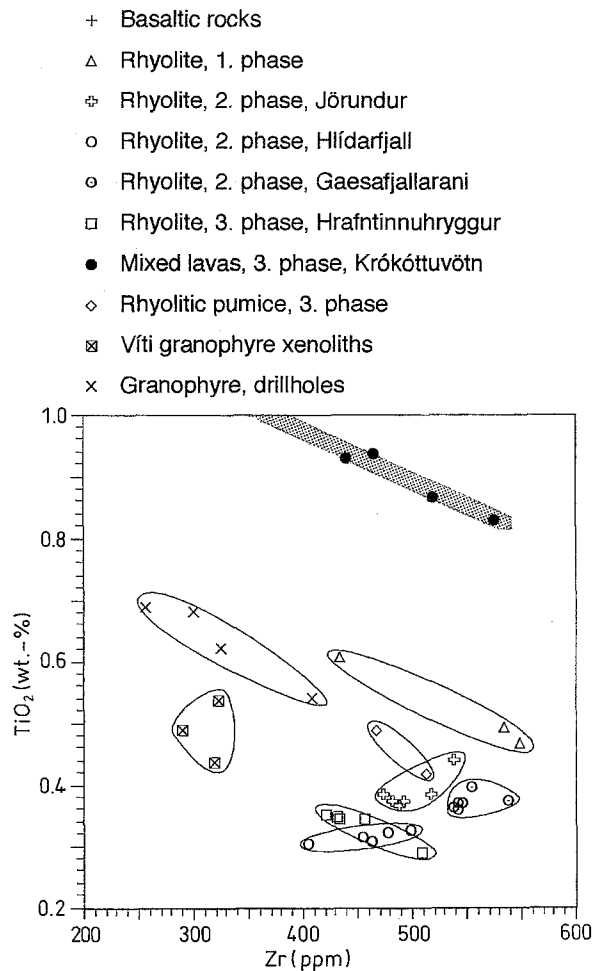
The isotope ratios of Sr (0.70317–0.70326) and Nd (0.51295–0.51311) do not vary with composition in volcanic rocks from Krafla (Nicholson et al. 1991; Sigmarsson et al. 1991; Hemond et al. 1993). In contrast,  $\delta^{18}O$  values are significantly lower in the silicic rocks (1.0–2.4) than in the basaltic rocks (3.0–4.5), and the same holds for ( $^{230}Th/^{232}Th$ ) activity ratios (0.955–1.003 and 1.048–1.145) in post-glacial samples (Nicholson et al.

1991). The low  $\delta^{18}O$  values indicate a hydrothermal component in the rhyolites, whereas the Th isotope activity ratios indicate a component of relatively old crust, at least 50–140 ka (Nicholson et al. 1991; Sigmarsson et al. 1991).

## Discussion

The Krafla rhyolites are chemically very similar to experimentally produced melts of tholeiitic basalts, and O-isotopes indicate a hydrothermal component. Therefore, the following model is proposed and tested to explain the generation of the rhyolites of the Krafla central volcano. The rhyolites were produced by near-solidus, rather than near-liquidus fractionation (i.e. by removing a small amount of interstitial melt from a rigid crystalline framework, rather than by removal of crystals from an essentially liquid magma). They are silicic minimum melts of hydrothermally altered, mainly basaltic crust, equilibrated at  $P_{H_2O} < 1$  kb, temperatures of 850–950°C and oxygen fugacities on or below the FMQ oxygen buffer. They are primarily generated on the peripheries of an active basaltic magma chamber or intru-





**Fig. 7**  $\text{TiO}_2$  versus Zr in silicic rocks from the Krafla area. Concentrations calculated as 100% volatile-free. Each silicic unit has its own distinct signature, indicating separate, slightly different sources. The concentration of  $\text{TiO}_2$  (and/or Zr) in the silicic end-member of the mixed Krókóttuvötn lavas is much higher than in the other silicic rocks

sive domain, where sufficient volumes of crust are subjected to temperatures favorable for rhyolite genesis (850–950°C). The silicic melts are extracted crystal-free from their source in response to crustal deformation. Alternative models include fractional crystallization coupled with crustal assimilation (Nicholson et al. 1991). The present data will now be discussed in relation to these models.

The low  $\delta^{18}\text{O}$  values measured in the Krafla rhyolites indicate a component of hydrothermally altered crust. Therefore, the effect of hydrothermal alteration on the bulk chemistry of basaltic rocks must be evaluated. The circulating hydrothermal fluid at Krafla is of meteoric origin. It is dilute and the composition of alteration minerals reflects the parental composition (Sveinbjörnsdóttir 1992). The absence of hydroxychloride and other soluble complexing ions common to seawater indicates a lesser tendency for mobilization of metal ions. In the Reydarfjörður drill core, eastern Ice-

land, redistribution of  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  as a result of alteration is evident (Flower et al. 1982). Nevertheless, the bulk chemistry of the rocks is still basaltic. The trace elements Rb, Ba, Sr and the transition metals, are possibly mobile during alteration, whereas the HFSEs (REEs, Y, Zr, Nb, Hf and Th) are considered to be stable (e.g. Wood et al. 1976; Gibson et al. 1982). Hydrothermally altered basaltic rocks found in Krafla drill-holes have  $\delta^{18}\text{O}$  values as low as  $-10$  (Hattori and Muehlenbachs 1982). The values found in the rhyolites are much higher (1–2.4). Therefore, if the rhyolites originated by partial melting, the protolith has not been severely altered. Experiments have shown that the effect of slight changes in bulk composition on the composition of a small volume partial melt is minor compared with the effect of  $P_{\text{H}_2\text{O}}$  (Helz 1976).

The paucity of intermediate rocks representing true liquids, and the fact that the rhyolites plot as a cluster rather than defining a chemical trend, cast some doubt on the use of fractional crystallization models for explaining the Krafla data. Furthermore, although it is possible to model the evolution of the Krafla rocks with a least-squares fitting method (Nicholson et al. 1991), some of the basic assumptions are questionable. An icelandite sample used in the calculations (KK35) comes from the Heidarspordur rift segment to the south of Krafla, and is not directly related to the Krafla central volcano. A sample of evolved basalt used in the calculation (KK08) comes from a composite welded tuff, and does not represent a true liquid composition. To produce the icelandite from the evolved basalt, 30% fractional crystallization is required of a phase assemblage containing 30% magnetite (Nicholson et al. 1991), which seems rather improbable.

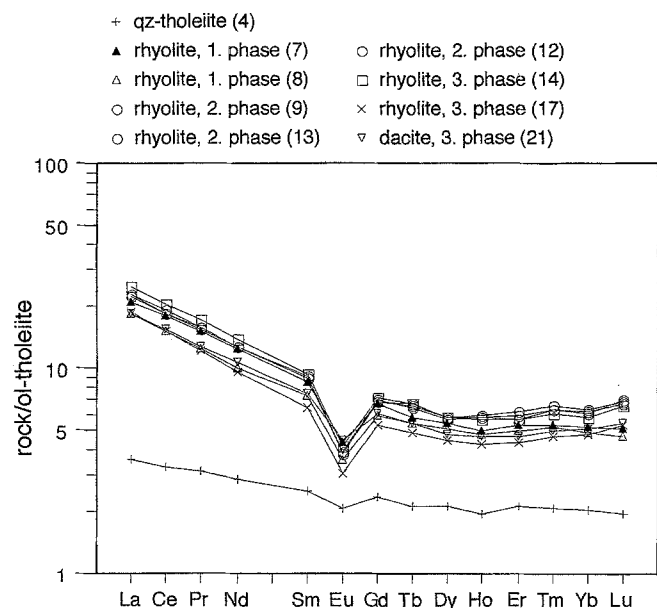
Melting experiments (e.g. Spulber and Rutherford 1983; Thy et al. 1990) have shown that equilibrium melting or equilibrium crystallization of a tholeiitic basalt at  $P_{\text{H}_2\text{O}} < 1$  kb, temperatures of approximately 900°C and oxygen fugacities on or below the FMQ oxygen buffer, can yield a silicic melt similar in composition to the Krafla rhyolites. This melt coexists with plagioclase and high Ca pyroxene, together with small amounts of low Ca pyroxene or olivine, and Fe–Ti oxides (Spulber and Rutherford 1983). The REE patterns of the Krafla rhyolites (Fig. 8) are consistent with such a scenario. Pyroxene thermometry indicates equilibrium temperatures of approximately 900°C, again broadly consistent with the results of the melting experiments. The Krafla rhyolites contain more  $\text{TiO}_2$  and less  $\text{P}_2\text{O}_5$  than the experimental melt (Table 1). The Krafla rhyolites crystallized fayalite and magnetite, but no ilmenite, and are therefore assumed to have had oxygen fugacities close to the FMQ buffer. The experiments were conducted at lower oxygen fugacities (GCH buffer) and had ilmenite as a residual phase (Spulber and Rutherford 1983), thus depleting the coexisting melt in  $\text{TiO}_2$ . The experimental melt contains 0.3 wt.%  $\text{P}_2\text{O}_5$ , three times as much as the Krafla rhyolites and much more than can be dissolved in the

**Table 3** Trace element concentrations in samples from the Krafla central volcano

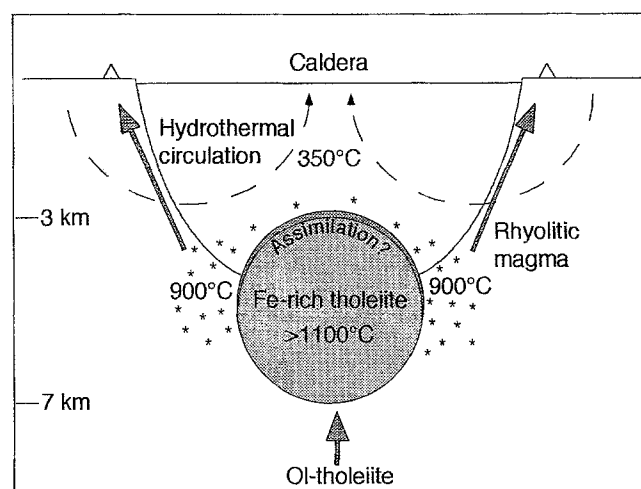
	Basaltic rocks		Rhyolites							Dacite	EF
	1	4	7	8	9	12	13	14	17	21	
Rb	1.78	5.47	65.2	55.1	65.9	74.6	70.2	66.3	47.5	70.7	11.6
Sr	105	137	112	96.1	88.7	92.2	88.4	97.4	133	89.6	0.7
Ba	20.1	69.8	495	411	501	512	480	524	379	483	6.8
Th	0.17	0.69	6.35	5.94	7.33	7.77	7.16	7.36	5.18	7.12	9.7
U	0.06	0.24	1.92	1.75	2.11	2.19	2.09	2.21	1.59	2.16	8.3
Y	16.1	32.5	82.1	77.3	95.1	101	98.1	98.2	75.4	71.5	2.8
Zr	32.7	95.9	524	387	497	449	512	475	510	421	5.0
Nb	2.38	8.26	55.1	37.3	47.3	46.8	47.4	52.4	40.7	33.4	5.7
Hf	1.01	2.73	13.9	11.3	14.4	13.5	14.0	13.3	12.7	11.2	4.9
La	2.29	8.22	48.27	41.69	52.15	52.47	50.83	56.69	43.30	41.88	6.0
Ce	6.033	20.14	107.9	91.85	111.6	114.7	111.1	124.7	91.74	92.17	5.3
Pr	0.877	2.76	13.13	10.94	13.95	13.87	13.48	15.21	10.72	11.15	4.7
Nd	4.545	13.24	56.65	45.92	57.98	57.49	58.12	63.42	43.84	48.38	4.1
Sm	1.54	3.92	13.17	11.27	13.78	14.03	13.97	14.56	9.95	11.63	3.3
Eu	0.65	1.36	2.84	2.32	2.53	2.58	2.50	2.71	2.02	2.89	1.8
Gd	1.866	4.45	12.72	11.06	13.09	12.79	13.00	13.37	9.89	11.37	2.8
Tb	0.373	0.79	2.17	2.02	2.39	2.55	2.48	2.52	1.84	2.01	2.9
Dy	2.60	5.51	14.12	13.26	14.85	15.05	15.19	15.04	11.70	12.46	2.6
Ho	0.63	1.23	3.13	3.00	3.64	3.77	3.67	3.58	2.70	2.97	2.7
Er	1.579	3.37	8.34	7.90	9.07	9.83	9.32	9.06	6.91	7.37	2.6
Tm	0.242	0.51	1.30	1.27	1.54	1.61	1.54	1.47	1.13	1.21	2.8
Yb	1.626	3.30	8.39	8.02	9.91	10.35	10.01	9.47	7.74	8.21	2.8
Lu	0.231	0.45	1.17	1.09	1.62	1.61	1.55	1.54	1.24	1.27	3.1

The samples were analysed at the centre de Recherches Pétrographiques et Géochimiques, France, by J. Govindaraju, using inductively coupled plasma mass spectrometry. The concentrations are in ppm. Sample numbers are the same as in Table 1. The enrich-

ment factor (EF) is the average concentration in the rhyolites divided by the concentration in a representative tholeiite basalt from Krafla (sample 4)



**Fig. 8** Rare earth element patterns of rocks from the Krafla central volcano. They are normalized to the composition of an olivine tholeiite from Krafla (1 in Table 1). The chondrite-normalized pattern of this sample is flat with values of about eight times chondrite values. It was selected as a representative composition for primitive mantle-derived magmas in the area. The overall pattern of the rhyolites indicates that Ca-rich pyroxene was a controlling phase in their genesis, whereas the distinct Eu anomaly indicates plagioclase control. The patterns of the rhyolites from the second phase are virtually identical to each other. Slight variations are seen between the other rhyolites. The numbers in the legend refer to sample numbers in Table 1



**Fig. 9** Schematic representation of the thermal environment of the active magma chamber or intrusive domain at a period of rhyolite genesis. Stars indicate the crustal domains where generation of rhyolitic magma is feasible. The thermal gradient away from the magma chamber is much steeper upwards than to the sides, due to the cooling effect of the hydrothermal system. The rhyolitic melts are generated in the temperature interval 850–950°C. At lower temperatures there is hydrothermal alteration; at higher temperatures the rocks lose their brittleness and bulk assimilation of crust into basaltic magma can occur. The temperature interval favorable for rhyolite genesis is much wider to the sides of the magma chamber than above it, resulting in the formation of a 'halo' of rhyolites around the volcano

melt according to apatite solubility studies (Harrison and Watson 1984). This probably results from the short time-scale of the experiments and the slow diffusivity of  $P_2O_5$ .

The trace element pattern of the Krafla rhyolites, compared with a representative tholeiite basalt, indicates that clinopyroxene and plagioclase were the main controlling phases in the formation of the rhyolites. This is consistent with the above model, though not conclusive. A similar pattern could possibly be produced by near-liquidus fractionation from basaltic magma. The main difference between the two models is that during near-solidus fractionation, the minerals only equilibrate with a silicic liquid, whereas during near-liquidus fractionation, a large part of the minerals removed equilibrate with basaltic to intermediate liquids. Partition coefficients for HFSEs are generally much higher in silicic liquids than in basaltic liquids. Near-liquidus fractionation would therefore produce a greater enrichment in these elements, and less fractionation between different elements, compared with near-solidus fractionation. Owing to the large variation in reported partition coefficients for silicic systems, this will not be modeled quantitatively.

Experimentally determined solubilities of zircon (Watson and Harrison 1983) and apatite (Harrison and Watson 1984) indicate that the Krafla rhyolites are very close to saturation in both phases (at about 500 ppm Zr and 0.1 wt.%  $P_2O_5$ , respectively). It is therefore not surprising to find them as small inclusions in phenocrysts in the rhyolites. When the Zr and  $P_2O_5$  contents of the rhyolites are compared with the concentrations in a likely source (No. 4 in Tables 1 and 3) it can be seen that although Zr is enriched in the rhyolites by an amount comparable with that for Nb and Ce,  $P_2O_5$  is depleted. It is unlikely that apatite could have been present in a basaltic source. The most likely reason for the depletion in  $P_2O_5$  is that as the partial melt was saturated with  $P_2O_5$ , most of the  $P_2O_5$  was left where it was originally. Owing to the lack of data on the partitioning of P into minerals other than apatite, this remains a problem.

Quartz phenocrysts have not been observed in the rhyolites, although projections in the granite system indicate that the rhyolites were in equilibrium with plagioclase and quartz (Fig. 5). If the rhyolites originated by a small degree of partial melting, they would be expected to lie on, or close to, the plagioclase-quartz cotectic plane even though there was no quartz in the residue. If the rhyolites were extracted crystal-free from their source and then migrated upwards, the decreasing pressure would have shifted the plagioclase-quartz cotectic plane towards the Qz corner. Thus only plagioclase would crystallize during ascent in addition to ferromagnesian phases. The amount of plagioclase phenocrysts in the rhyolites varies from trace amounts to about 7 vol.%, without affecting their whole rock composition. Some of them, mainly the rhyolites of the second phase, might thus have been stored in a reservoir

somewhere between the source and the surface, where *in situ* plagioclase crystallization (i.e. crystallization without fractionation) would occur, before being erupted. If the rhyolites originated by near-liquidus fractionation in a deeper magma chamber, they would be expected to contain quartz phenocrysts. They could be resorbed during ascent, but traces should still be found in the rhyolites. Furthermore, near-liquidus fractionation would be expected to produce a significant degree of normal zoning in the plagioclase phenocrysts. The slight reverse zoning observed speaks strongly against equilibration at slightly cooling conditions, but rather indicates a prograde process.

Silicic liquids have very high viscosities (Shaw 1972), unless they are very rich in volatiles. There is no indication of a very high volatile content in the Krafla rhyolites. This imposes great difficulties in using most differentiation models to explain the generation of the rhyolites (e.g. McKenzie 1985), especially models of near-liquidus fractionation (fractionation of crystals from a magma close to its liquidus temperature). It is possible, however, to use models of near-solidus fractionation (segregation of a silicic minimum melt from a near-solidus crystal mush).

When the melt fraction of a rock is below a critical melt fraction of approximately 0.25, the rock as a whole behaves as brittle material (Arzi 1978; van der Molen and Paterson 1979). This allows the relative movement of small fractions of silicic melt in response to deformation. Cracks can form or reopen in the presence of the high fluid pressure of the melt, and propagate as melt-filled 'hydrofractures' (Paquet and François 1980), thus producing silicic veins and segregations. It is interesting to note that this process requires the melt fraction to be below 0.25. When the melt fraction becomes higher the protolith disintegrates and becomes a magma. Although 0–25% melting of a basaltic rock produces rhyolitic and dacitic liquids, a higher degree of melting inhibits the separation of intermediate melt and matrix. Basaltic magma (mush) is formed instead, which could be assimilated by mantle-derived magmas to produce crustally contaminated tholeiites (e.g. Óskarsson et al. 1982).

The hydrothermal signature (low  $\delta^{18}O$ ) found in the rhyolites favors partial melting of hydrothermally altered crust, though remelting of hydrothermally altered silicic veins (e.g. Marsh et al. 1991) cannot be excluded. The low Th isotope activity ratios (in post-glacial rhyolites), however, suggest that at least 50 000 to 140 000 years old crust (Sigmarsson et al. 1991) was involved, and therefore support an origin of the rhyolites by partial melting of older crust. The trace and minor element signatures of the rhyolites indicate that they were formed from separate, slightly different sources, though the genetic process may have been the same. Granite phase relations indicate that the earliest rhyolites (from the first phase) equilibrated either at greater depths than the later rhyolites or at lower partial pressures of water.

The distribution of rhyolitic volcanic rocks at Krafla (at or around the caldera rim, away from the fissure swarm; Fig. 1) can be explained by the long-standing thermal environment of the active basaltic magma chamber or intrusive domain at the time of formation (Fig. 9). The thermal gradient away from the magma chamber is much steeper upwards than to the sides, due to the cooling effect of the hydrothermal system. The rhyolitic melts are generated in the temperature interval 850–950°C. At lower temperature there is hydrothermal alteration; at higher temperatures the rocks lose their brittleness and bulk assimilation of crust into basaltic magma can occur. The temperature interval favoring rhyolite formation is much wider to the sides of the magma chamber than above it, resulting in the formation of a 'halo' of rhyolites around the volcano. Alternatively this distribution could be the result of ring dike or cone sheet formation, providing pathways to the surface for the rhyolite magmas, although this fails to explain why rhyolites have not been erupted in significant amounts in the fissure swarm that cuts through the volcano.

## Conclusions

The rhyolitic rocks of Krafla are generated by near-soliquid fractionation. The source is most likely hydrothermally altered crust, mainly of basaltic composition, but remelting of hydrothermally altered granophyric segregations cannot be excluded. The generation of rhyolitic melts occurs primarily to the sides of the magma chamber, where large volumes of the crust are subjected to the thermal conditions suitable for rhyolite generation (850–950°C). Above the magma chamber the temperature decreases sharply because of the cooling effect of the hydrothermal system. At the wall and roof of the magma chamber wholesale melting and assimilation of hydrothermally altered crust can occur, resulting in the formation of crustally contaminated tholeiites and basaltic andesites. The rhyolitic melts are extracted crystal-free from their source in response to deformation. They migrate upwards, crystallizing mainly plagioclase in response to decreasing pressure. They may collect in reservoirs, erupting when encountered by ascending basaltic magma, or erupt directly, during the deformation episode that extracted them.

**Acknowledgements** The author thanks the staff of the Nordic Volcanological Institute for help during this research project, especially Karl Grönvold, Niels Óskarsson, Gudmundur E. Sigvaldason, Ágúst Gudmundsson and Henrik Schiellerup for fruitful discussions. Kristján Saemundsson is thanked for providing field maps of the area, and Asgrímur Gudmundsson for drill cuttings of granophyric rocks. Sveinn Jakobsson and Sigurdur Steinthórsón are thanked for reviewing early versions of the manuscript. The final version of the manuscript benefitted from helpful reviews by Tanya Furman and Eric Christiansen.

## References

- Ármannsson H, Gudmundsson Á, Steingrímsson BS (1987) Exploration and development of the Krafla geothermal area. *Jökull* 37:13–30
- Arth JG (1976) Behavior of trace elements during magmatic processes – a summary of theoretical models and their applications. *J Res US Geol Surv* 4:41–47
- Arzi AA (1978) Critical phenomena in the rheology of partially melted rocks. *Tectonophysics* 44:173–184
- Björnsson A (1985) Dynamics of crustal rifting in NE Iceland. *J Geophys Res* 90:10151–10162
- Björnsson A, Saemundsson K, Einarsson P, Tryggvason E, Grönvold K (1977) Current rifting episode in north Iceland. *Nature* 266:318–323
- Björnsson A, Johnsen G, Sigurdsson S, Thorbergsson G (1979) Rifting of the plate boundary in North Iceland 1975–1978. *J Geophys Res* 84:3029–3038
- Calderone GM, Grönvold K, Óskarsson N (1990) The welded air-fall tuff layer at Krafla, northern Iceland: a composite eruption triggered by injection of basaltic magma. *J Volcanol Geotherm Res* 44:303–314
- Carmichael ISE (1964) The petrology of Thingmúli, a Tertiary volcano in eastern Iceland. *J Petrol* 5:435–460
- Carmichael ISE (1967) The iron–titanium oxides of salic volcanic rocks and their associated ferromagnesian silicates. *Contrib Mineral Petrol* 14:36–64
- Condomines M, Grönvold K, Hooker PJ, Muehlenbachs K, O'Nions RK, Óskarsson N, Oxburgh ER (1983) Helium, oxygen, strontium and neodymium isotopic relationships in Icelandic volcanics. *Earth Planet Sci Lett* 66:125–136
- Davidson PM, Lindsley DH (1989) Thermodynamic analysis of pyroxene–olivine–quartz equilibria in the system CaO–MgO–FeO–SiO<sub>2</sub>. *Am Mineral* 74:18–30
- Flower MFJ, Pritchard RG, Brem G, Cann JR, Delaney J, Emmerman R, Gibson IL, Oakley PJ, Robinson PT, Schmincke H-U (1982) Chemical stratigraphy, Iceland research drilling project Reydarfjörður, Eastern Iceland. *J Geophys Res* 87:6489–6510
- Furnes H, Fridleifsson IB, Atkins FB (1980) Subglacial volcanics – on the formation of acid hyaloclastites. *J Volcanol Geotherm Res* 8:95–110
- Gibson IL, Kirkpatrick RJ, Emmerman R, Schmincke H-U, Pritchard G, Oakley PJ, Thorpe RS, Marriner GF (1982) The trace element composition of the lavas and dikes from a 3-km vertical section through the lava pile of Eastern Iceland. *J Geophys Res* 87:6532–6546
- Grönvold K (1972) Structural and petrochemical studies in Kerlingarfjöll region, Central Iceland. PhD Thesis, Univ Oxford:1–237
- Harrison TM, Watson EB (1984) The behavior of apatite during crustal anatexis: equilibrium and kinetic considerations. *Geochim Cosmochim Acta* 48:1467–1477
- Hattori K, Muehlenbachs K (1982) Oxygen isotope ratios of the Icelandic crust. *J Geophys Res* 87:6559–6565
- Helz RT (1976) Phase relations of basalts in their melting range at  $P_{H_2O} = 5$  kb as a function of oxygen fugacity, part II. Melt compositions. *J Petrol* 17:139–193
- Hemond C, Condomines M, Fourcade S, Allègre CI, Óskarsson N, Javoy M (1988) Thorium, strontium and oxygen isotopic geochemistry in recent tholeiites from Iceland: crustal influence on mantle-derived magmas. *Earth Planet Sci Lett* 87:273–285
- Hemond C, Arndt NT, Lichtenstein U, Hofmann AW, Óskarsson N, Steinthorsson S (1993) The heterogeneous Iceland plume: Nd–Sr–O isotopes and trace element constraints. *J Geophys Res* 98:15833–15850
- Jakobsson SP (1972) Chemistry and distribution pattern of recent basaltic rocks in Iceland. *Lithos* 5:365–386
- Jónasson K, Holm PM, Pedersen AK (1992) Petrogenesis of silicic rocks from the Króksfjörður central volcano, NW Iceland. *J Petrol* 33:1345–1369

- Lindsley DH (1983) Pyroxene thermometry. *Am Mineral* 68:477–493
- Macdonald R, Sparks RSJ, Sigurdsson H, Matthey DP, McGarvie DW, Smith RL (1987) The 1875 eruption of Askja volcano, Iceland: combined fractional crystallization and selective contamination in the generation of rhyolitic magma. *Min Mag* 51:183–202
- Macdonald R, McGarvie DW, Pinkerton H, Smith RL, Palacz CA (1990) Petrogenetic evolution of the Torfajökull volcanic complex, Iceland. I. Relationship between the magma types. *J Petrol* 31:429–459
- Mahood G, Hildreth W (1983) Large partition coefficients for trace elements in high-silica rhyolites. *Geochim Cosmochim Acta* 47:11–30
- Marsh BD, Gunnarsson B, Congdon R, Carmody R (1991) Hawaiian basalt and Icelandic rhyolite: indicators of differentiation and partial melting. *Geol Rundsch* 80:481–510
- McGarvie DW, Macdonald R, Pinkerton H, Smith RL (1990) Petrogenetic evolution of the Torfajökull volcanic complex, Iceland II. The role of magma mixing. *J Petrol* 31:461–481
- McKenzie D (1985) The extraction of magma from the crust and mantle. *Earth Planet Sci Lett* 74:81–91
- Michael PJ (1988) Partition coefficients for rare earth elements in mafic minerals of high silica rhyolites: the importance of accessory mineral inclusions. *Geochim Cosmochim Acta* 52:275–282
- Nash WP, Crecraft HR (1985) Partition coefficients for trace elements in silicic magmas. *Geochim Cosmochim Acta* 49:2309–2322
- Nekvasil H (1988) Calculated effect of anorthite component on the crystallization paths of H<sub>2</sub>O-undersaturated haplogranitic melts. *Am Mineral* 73:966–981
- Nicholson H (1990) The magmatic evolution of Krafla, NE Iceland. PhD Thesis, Univ Edinburgh:1–237
- Nicholson H, Condomines M, Fitton JG, Fallick AE, Grönvold K, Rogers G (1991) Geochemical and isotopic evidence for crustal assimilation beneath Krafla, Iceland. *J Petrol* 32:1005–1020
- O'Nions RK, Grönvold K (1973) Petrogenetic relationships of acid and basic rocks in Iceland: Sr isotopes and rare earth elements in late postglacial volcanics. *Earth Planet Sci Lett* 19:397–409
- Óskarsson N, Sigvaldason GE, Steinhórnsson S (1982) A dynamic model of rift zone petrogenesis and the regional petrology of Iceland. *J Petrol* 23:28–74
- Paquet J, François P (1980) Experimental deformation of partially melted granitic rocks at 600–900°C and 250 Mpa confining pressure. *Tectonophysics* 68:131–146
- Pearce JA, Norry MJ (1979) Petrogenetic implications of Ti, Zr, Y, and Nb variations in volcanic rocks. *Contrib Mineral Petrol* 69:33–47
- Saemundsson K (1972) Notes on the geology of the Torfajökull area. *Náttúrufræðingurinn* 42:81–99 [in Icelandic]
- Saemundsson K (1974) Evolution of the axial rifting zone in Northern Iceland and the Tjörnes fracture zone. *Geol Soc Am Bull* 85:495–504
- Saemundsson K (1978) Fissure swarms and central volcanoes of the neovolcanic zones of Iceland. *Geol J Spec Issue* 10:415–432
- Saemundsson K (1991) Geology of the Krafla system. In: Gardarsson A, Einarsson Á (eds) *Náttúra Mývatns. Hid Íslenska Náttúrufræðifélag, Reykjavík*, pp 25–95 [in Icelandic]
- Shaw HR (1972) Viscosities of magmatic silicate liquids: an empirical method of prediction. *Am J Sci* 272:870–893
- Sigmarrson O, Hémond C, Condomines M, Fourcade S, Óskarsson N (1991) Origin of silicic magma in Iceland revealed by Th isotopes. *Geology* 19:621–624
- Sigurdsson H (1977) Generation of Icelandic rhyolites by melting of plagiogranites in the oceanic layer. *Nature* 269:25–28
- Sigurdsson H, Sparks RSJ (1981) Petrology of rhyolitic and mixed magma ejecta from the 1875 eruption of Askja, Iceland. *J Petrol* 22:41–84
- Sigvaldason GE (1974) The petrology of Hekla and origin of silicic rocks in Iceland. *Soc Sci Islandica* V:1–44
- Sisson TW (1991) Pyroxene-high silica rhyolite trace element partition coefficients measured by ion microprobe. *Geochim Cosmochim Acta* 55:1575–1585
- Spulber SD, Rutherford MJ (1983) The origin of rhyolite and plagiogranite in oceanic crust: an experimental study. *J Petrol* 24:1–25
- Sveinbjörnsdóttir AE (1992) Composition of geothermal minerals from saline and dilute fluids – Krafla and Reykjanes, Iceland. *Lithos* 27:301–315
- Thoroddsen Th (ed) (1907–15) Reports on the Mývatn fires 1724–1729. *Safn til sögu Íslands* 4:385–411 [in Icelandic]
- Thy P, Beard JS, Lofgren GE (1990) Experimental constraints on the origin of Icelandic rhyolites. *J Geol* 98:417–421
- van der Molen I, Paterson MS (1979) Experimental deformation of partially-melted granite. *Contrib Mineral Petrol* 70:299–318
- Warshaw CM, Smith RL (1988) Pyroxenes and fayalites in the Bandelier Tuff, New Mexico: temperatures and comparison with other rhyolites. *Am Mineral* 73:1025–1037
- Watson EB, Harrison TM (1983) Zircon saturation revisited: temperature and composition effects in a variety of crustal magma types. *Earth Planet Sci Lett* 64:295–304
- Wood DA, Gibson IL, Thompson RN (1976) Elemental mobility during zeolite facies metamorphism of the Tertiary basalts of Eastern Iceland. *Contrib Mineral Petrol* 55:241–254
- Wright TL, Doherty PC (1970) A linear programming and least squares computer method for solving petrographic mixing problems. *Geol Soc Am Bull* 81:1995–2008

Editorial responsibility: W. Hildreth