Contrib Mineral Petrol (2007) 153:593–605
DOI 10.1007/s00410-006-0165-5

ORIGINAL PAPER

Crustal thermal state and origin of silicic magma in Iceland: the case of Torfajökull, Ljósufjöll and Snæfellsjökull volcanoes

E. Martin · O. Sigmarsson

Received: 4 July 2006 / Accepted: 15 November 2006 / Published online: 9 January 2007

Abstract Pleistocene and Holocene peralkaline rhyolites from Torfajökull (South Iceland Volcanic Zone) and Ljósufjöll central volcanoes and trachytes from Snæfellsjökull (Snæfellsnes Volcanic Zone) allow the assessment of the mechanism for silicic magma genesis as a function of geographical location and crustal geothermal gradient. The low δ^{18}O (2.4‰) and low Sr concentration (12.2 ppm) measured in Torfajökull rhyolites are best explained by partial melting of hydrated metabasaltic crust followed by major fractionation of feldspar. In contrast, very high ^{87}Sr/^{86}Sr (0.70473) and low Ba (8.7 ppm) and Sr (1.2 ppm) concentrations measured in Ljósufjöll silicic lavas are best explained by fractional crystallisation and subsequent ^{87}Rb decay. Snæfellsjökull trachytes are also generated by fractional crystallisation, with less than 10% crustal assimilation, as inferred from their δ^{18}O. The fact that silicic magmas within, or close to, the rift zone are principally generated by crustal melting whereas those from off-rift zones are better explained by fractional crystallisation clearly illustrates the controlling influence of the thermal state of the crust on silicic magma genesis in Iceland.

Communicated by J. Hoefs.

E. Martin (✉) · O. Sigmarsson
Laboratoire Magmas et Volcans, OPGC,
Université Blaise Pascal, CNRS, 5 rue Kessler,
63038 Clermont-Ferrand, France
e-mail: E.Martin@opgc.univ-bpclermont.fr

O. Sigmarsson
Institute of Earth Sciences, University of Iceland,
101 Reykjavik, Iceland

Keywords Silicic magma · Fractional crystallisation · Partial melting · Assimilation · Geothermal gradient · Iceland

Introduction

The proportion of silicic rocks in Iceland is exceptionally high for an oceanic island, representing about 5–10% of the exposed lavas (e.g. Walker 1966; Thordarson and Larsen 2006). Together with the geochemical arguments of O- and Th isotope ratios, which are found to be lower in the silicic magmas than in contemporaneous basalts for a given volcano, this strongly suggests a crustal melting origin for the silicic rocks (e.g. Nicholson et al. 1991; Sigmarsson et al. 1991). However, generation of silicic magma by fractional crystallisation is still favoured for some Icelandic volcanoes (e.g. Carmichael 1964; Macdonald et al. 1990; Furman et al. 1992; Prestvik et al. 2001). The question remaining to be answered is whether or not these two contrasting mechanisms of silicic magma formation are mutually exclusive and, if not, what determines which of the two mechanisms is dominant.

The model of crustal partial melting developed for Krafla, Askja and Hekla volcanoes (e.g. Sigmarsson et al. 1991; Jónasson 1994), which are located in central Iceland, may not be realistic at the periphery of the island where the crustal thermal gradient is lower. Variable geothermal gradients have been measured in boreholes and clearly reflect a stronger influence of the Iceland mantle plume inland, as well as higher gradients in the presence of the rift system, while there are colder regions closer to the coast, and away from the spreading axis (Flovenz and

Saemundsson 1993). Genesis of silicic magma by partial melting of the altered basaltic crust requires an elevated geothermal gradient so that incoming basaltic melts can raise the crustal temperature above its solidus. This condition is most likely to be met in central Iceland, whereas around the coast away from the rift zones the geothermal gradient is relatively low and the crustal temperature is likely to cool down incoming basaltic mantle melts. The lower geothermal gradient away from the rift zones is principally due to significantly lower volumes of incoming basalts. In this context, it would not be surprising to find different mechanisms for the generation of silicic magmas in regions with contrasting thermal gradients. Such an effect of the geothermal gradient on silicic magma formation has recently been discussed by Christiansen and McCurry (2006).

This study focuses on Pleistocene and Holocene felsic magmas from Torfajökull and Ljósufjöll and Snæfellsjökull volcanoes located respectively at the centre and the periphery of Iceland. The difference in geographical location of these volcanoes allows the potential influence of thermally contrasted settings on the mode of silicic magma formation in Iceland to be tested.

Geological setting

Iceland is located at a point of interaction between the Mid-Atlantic ridge and the Iceland mantle plume (Fig. 1). The centre of the mantle plume is deduced from seismic and gravity anomalies characterized by the location of high ^{3}He/^{4}He volcanics (e.g. Eysteinsson and Gunnarsson 1995; Wolfe et al. 1997; Breddam et al. 2000). The influence of the mantle plume on the geothermal gradient of the crust is most likely induced by higher basalt productivity at the centre of Iceland compared to the coastal areas. Jakobsson (1972) estimated that the volume of basalts erupted during the Holocene in central Iceland was two to three times greater than in the coastal areas. This concurs with significantly higher geothermal gradients inland, as expressed by the shallower depth to the inferred 1,200°C isotherm (Fig. 1; Flovenz and Saemundsson 1993). To the extreme east and north-west the maximum depth of the inferred 1,200°C isotherm is observed in the oldest regions of Iceland, but this isotherm is not symmetrical around the rift axis due to change in the ridge position in the past. For example, the anomalously shallow depth of this isotherm to the north of the Snæfellsnes Volcanic Zone (SNVZ) represents a relatively high geothermal gradient that coincides with the site of a former rift axis, before it relocated to the Reykjanes Rift Zone (RRZ; Fig. 1).

Torfajökull volcano, located in the propagating rift zone in South Central Iceland (T in Fig. 1), is the largest silicic complex in Iceland, covering an area of more than 400 km^2 (Gunnarsson et al. 1998). A large part of this complex was formed during the Pleistocene and several subglacial and silicic table mountains, such as Laufafell, are thought to represent syn-caldera eruptions on a ring fracture (e.g. McGarvie 1984). Its post-glacial activity is characterized by the formation of obsidian lava flows following an initial explosive phase and tephra production. One such eruption occurred approximately 8,000 years ago when the ca. 80 m-thick tephra pile and the Hrafntinnusker obsidian hill formed, while another one took place during the ninth century when the obsidian lava flow was erupted (Hrafntinnuhraun). Silicic magmas at Torfajökull are characterized by their peralkaline composition (e.g. Macdonald et al. 1990; Gunnarsson et al. 1998 and references therein).

Snæfellsjökull and Ljósufjöll volcanoes, in the Snæfellsnes Volcanic Zone (SNVZ; Fig. 1), are located

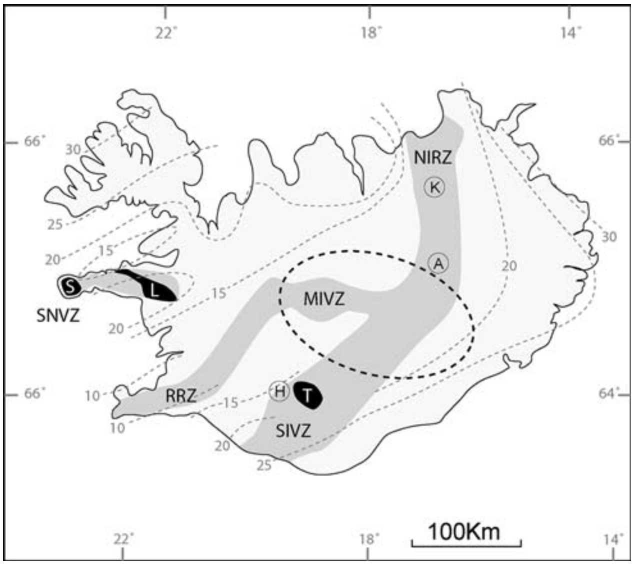

Fig. 1 Map of Iceland showing the location of Snæfellsjökull (*S*), Ljósufjöll (*L*) and Torfajökull (*T*) volcanoes. The stippled ellipse corresponds to the location of the gravity and seismic velocity anomalies, which are interpreted as representing the mantle plume centre (Eysteinsson and Gunnarsson 1995; Wolfe et al. 1997). *Dashed grey lines* correspond to inferred isobaths (km) of the 1,200°C isotherm (Flovenz and Saemundsson 1993) and illustrate the decreasing thermal gradient away from the rift zones. A noteworthy exception is observed north of the Snæfellsnes volcanic zone (*SNVZ*). *MIVZ* Mid-Iceland volcanic zone, *SIVZ* South-Iceland volcanic zone, *NIRZ* North-Iceland rift-zone, *RRZ* Reykjanes rift-zone, *K* Krafla, *A* Askja and *H* Hekla volcanoes

Table 1 Major- and trace element compositions together with $^{143}Nd/^{144}Nd$, $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ isotope compositions of samples from Torfajökull, Ljósufjöll and Snæfellsjökull volcanoes

		SiO_2	TiO_2	Al_2O_3	$Fe_2O_3^t$	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H_2O–	H_2O+	Total
Torfajökull	HRN-21	70.2	0.28	14.5	2.80	0.08	0.27	0.98	5.63	4.52	0.10	0.17	1.02	100.5
	Lauf	72.1	0.34	13.3	2.95	0.11	0.06	0.72	5.83	3.66	0.03	0.10	0.37	99.6
	Hsk8	72.6	0.19	12.3	3.36	0.07	0.07	0.40	5.63	4.44	0.02	0.16	0.37	99.6
	Hsk8	*73.3*	*0.20*	*12.4*	*3.39*	*0.07*	*0.07*	*0.40*	*5.68*	*4.48*	*0.02*			*100.0*
	Hsk7	71.4	0.20	11.9	3.41	0.08	0.09	0.44	5.08	4.29	0.03	0.34	3.71	101.0
	Hsk7	*73.6*	*0.20*	*12.3*	*3.52*	*0.08*	*0.09*	*0.45*	*5.25*	*4.43*	*0.03*			*100.0*
	Hsk4	70.9	0.20	12.1	3.49	0.07	0.08	0.42	4.66	4.51	0.03	0.37	3.62	100.4
	Hsk4	*73.5*	*0.21*	*12.5*	*3.61*	*0.08*	*0.09*	*0.43*	*4.83*	*4.67*	*0.03*			*100.0*
	Hsk3											0.94	4.42	
	Hsk1	68.5	0.22	12.9	3.75	0.07	0.23	0.47	2.83	4.31	0.03	0.88	5.22	99.4
	Hsk1	*73.4*	*0.23*	*13.9*	*4.01*	*0.08*	*0.25*	*0.51*	*3.04*	*4.62*	*0.03*			*100.0*
Ljósufjöll	L-1	70.0	0.25	13.6	4.20	0.16	0.02	0.45	5.72	4.86	0.05	0.08	0.19	99.6
	L-4	71.6	0.22	13.1	4.00	0.14	0.03	0.49	5.90	4.74	0.02	0.07	–0.05	100.2
	L-6	71.5	0.21	12.8	3.88	0.13	0.03	0.43	5.96	4.60	0.02	0.24	0.19	100.0
	L-5	71.3	0.21	12.7	3.87	0.13	0.03	0.43	5.86	4.58	0.02	0.06	0.24	99.4
	L-2	54.9	1.64	16.3	9.04	0.21	3.89	7.41	4.18	2.66	0.34	0.10	–0.35	100.3
	L-3	53.0	1.72	16.2	9.33	0.20	4.83	8.77	3.61	2.20	0.35	0.13	–0.14	100.2
	7640 $^{(1),(2)}$	47.5	1.96	13.4	11.4	0.16	11.2	11.9	2.27	0.67	0.30			100.8
	7637 $^{(1),(2)}$	47.2	2.49	14.9	12.4	0.18	6.81	11.8	2.95	0.76	0.52			100.0
Snæfellsjökull	SN-1	65.6	0.31	14.9	3.95	0.15	0.16	1.31	5.59	4.40	0.06	0.38	2.72	99.5
	SN-2	64.9	0.38	15.1	4.51	0.17	0.27	1.61	5.59	4.20	0.08	0.28	2.08	99.1
	SN89-23$^{(1),(2)}$	45.8	2.60	13.8	12.6	0.17	10.5	11.6	2.25	0.74	0.42			100.5
	SN89-27$^{(1),(2)}$	46.3	4.04	14.8	15.4	0.22	5.17	10.2	3.35	1.11	0.85			101.4

		Rb	Sr	Y	Zr	Nb	Ba	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy
Torfajökull	HRN-21	105.8	53.5	60.2	551.0	95.7	421.4	87.1	174.4	18.4	64.8	12.6	1.61	10.1	1.78	11.0
	Lauf	80.9	76.3	107.3	989.2	133.4	573.2	106.1	228.8	25.8	100.9	20.8	4.95	17.0	3.07	19.8
	Hsk8	119.8	12.2	116.9	1067.0	140.3	172.4	120.6	246.8	28.1	102.0	21.5	1.88	18.5	3.33	20.6
	Hsk7	110.4	14.1	111.4	1026.3	133.5	203.2	113.9	232.0	26.3	96.2	20.2	1.74	17.4	3.12	19.5
	Hsk4	114.0	13.8	115.9	1093.3	140.2	163.8	118.1	242.4	27.7	100.8	21.4	1.89	18.6	3.33	20.7
	Hsk3															
	Hsk1	102.2	14.5	112.2	1189.3	135.6	139.5	118.4	247.5	27.6	100.8	21.3	1.89	18.4	3.24	20.2
Ljósufjöll	L-1	161.72	1.79	91.3	1350.1	251.1	17.7	353.4	552.8	68.4	219.1	34.3	1.18	20.9	3.10	17.1
	L-4	178.85	1.17	86.7	1320.1	253.6	8.72	170.4	355.8	37.0	127.7	22.9	0.77	15.9	2.73	17.2
	L-6	184.17	1.31	111.3	1524.2	281.7	9.30	180.4	329.5	40.0	140.1	27.1	0.95	19.6	3.64	22.5
	L-5	184.94	1.60	141.9	1446.4	274.8	9.52	184.8	340.4	39.5	138.8	26.9	0.95	19.7	3.63	22.6
	L-2	46.28	345.35	36.9	357.2	74.9	691.5	82.9	161.4	17.8	63.9	10.7	2.74	7.93	1.17	6.85
	L-3	41.28	356.80	31.5	281.1	61.5	595.8	59.2	118.0	13.3	49.1	8.72	2.55	6.86	1.01	5.97
	7640 $^{(1),(2)}$	15.6	368.0	–	129.0	24.5	233.0	22.2	47.8	6.22	24.5	5.14	1.72	4.97	0.72	3.95
	7637 $^{(1),(2)}$	14.5	398.0	–	147.0	29.4	282.0	24.8	54.6	7.26	29.4	6.33	2.21	6.32	0.92	5.14
Snæfellsjökull	SN-1	98.7	83.3	61.0	646.9	117.6	1214.6	87.5	177.5	19.4	69.3	13.0	2.51	11.2	1.75	10.9
	SN-2	91.3	118.5	58.6	666.8	113.9	1216.8	85.0	172.6	19.0	68.1	12.7	2.75	10.9	1.71	10.7
	SN89-23$^{(1),(2)}$	15.8	453.0	–	152.0	38.1	309.0	25.1	54.2	7.13	28.0	5.79	2.01	5.39	0.79	4.25
	SN89-27$^{(1),(2)}$	23.4	546.0	–	242.0	61.6	534.0	42.0	91.4	12.0	47.3	9.55	3.33	8.84	1.28	6.83

Table 1 continued

		Ho	Er	Tm	Yb	Lu	Hf	Ta	Th	U	Th/U	$^{87}Sr/^{86}Sr$	$^{143}Nd/^{144}Nd$	$\delta^{18}O$
Torfajökull	HRN-21	2.24	6.21	0.91	5.83	0.86	15.4	8.11	17.6	5.09	3.46	0.70345		
	Lauf	4.19	11.0	1.53	10.3	1.38	25.9	11.5	13.9	4.22	3.29	0.70332	0.51298	2.44
	Hsk8	4.19	11.4	1.60	10.9	1.60	28.6	12.2	20.0	5.96	3.36	0.70338	0.51298	3.87
	Hsk7	3.94	10.7	1.51	10.3	1.50	26.8	11.1	19.7	5.77	3.41	0.70355	0.51297	
	Hsk4	4.22	11.4	1.60	11.0	1.61	29.3	12.6	20.3	5.73	3.54	0.70353	0.51297	
	Hsk3								21.4	5.68	3.77	0.70341	0.51298	
	Hsk1	4.07	11.0	1.53	10.5	1.54	31.3	12.8	21.9	5.17	4.24	0.70386	0.51297	
Ljósufjöll	L-1	3.16	8.52	1.28	9.38	1.49	29.1	17.6	23.7	5.39	4.40	0.70473	0.51291	6.30
	L-4	3.66	10.3	1.57	11.3	1.81	31.8	18.7	24.9	7.06	3.53	0.70435	0.51293	6.15
	L-6	4.63	12.6	1.84	12.8	2.00	34.5	20.6	26.3	6.84	3.85	0.70463	0.51292	5.84
	L-5	4.75	13.0	1.88	12.8	1.99	32.4	19.4	24.9	6.56	3.80	0.70442	0.51292	
	L-2	1.36	3.62	0.51	3.52	0.59	7.69	5.45	6.16	1.59	3.87	0.70338	0.51292	5.67
	L-3	1.19	3.18	0.46	3.06	0.48	6.34	5.20	5.43	1.52	3.57	0.70347	0.51290	5.84
	7640 (1),(2)	0.78	2.05	0.30	1.87	0.28	3.03		1.96	0.586	3.34	0.70334	0.51296	5.39
	7637 (1),(2)	1.03	2.69	0.40	2.46	0.38	3.50		1.97	0.590	3.34	0.70330	0.51295	5.20
Snæfellsjökull	SN-1	2.30	6.37	0.95	6.79	1.09	15.3	8.64	12.3	3.50	3.51	0.70338	0.51297	5.10
	SN-2	2.24	6.27	0.95	6.55	1.04	15.5	8.29	11.5	3.29	3.50	0.70337	0.51296	4.92
	SN89-23(1),(2)	0.83	2.16	0.31	1.99	0.30	3.41	2.37	1.85	0.537	3.45	0.70336		5.40
	SN89-27(1),(2)	1.35	3.51	0.50	3.20	0.49	5.37	3.82	2.69	0.779	3.45	0.70336	0.51297	5.30

In *italics* are major element concentrations for Hrafntinnusker samples recalculated to 100%. *Underlined values* are for results obtained by isotopic dilution mass spectrometry. *HRN* Hrafntinnuhraun, *Lauf* Laufafell, *Hsk* Hrafntinnusker. (1) Sigmarsson et al. (1992b) and (2) Carpentier (2003) in western Iceland. Snæfellsjökull is a stratovolcano that has produced a complete alkaline magma series, with three Plinian eruptions releasing trachytic tephra layers of Postglacial age (Jóhannesson et al. 1982). To the east, the Ljósufjöll complex is principally composed of silicic extrusives with a few lavas of intermediate composition, all of which were formed in subglacial eruptions during the Pleistocene. Their age is estimated to be in the range of ~100 to ~800 ka (Flude et al. 2004; Gudmundsdóttir and Sigmarsson 2006) and all known silicic formations known have peralkaline compositions. In contrast to Torfajökull, Snæfellsjökull, and even more so Ljósufjöll, are constructed on a crust with a significantly lower geothermal gradient (Fig. 1).

Sample description

The Torfajökull samples come from three eruptive units: the subglacial table mountain Laufafell (Lauf, Table 1), the Hrafntinnuhraun eruption (HRN-21) and the early Holocene Hrafntinnusker (Hsk). The Lauf and Hsk8 samples consist of slightly porphyritic obsidians (3–5% phenocrysts: 80% plagioclase and 20% clinopyroxene) whereas all other samples are aphyric pumices. Hrafntinnusker samples come from a single 80 m-thick tephra section formed during a single eruptive phase shortly after, or contemporaneous with, deglaciation in this region (Martin et al., in preparation). Sample Hsk1 represents the first tephra emplaced during this eruption, whereas Hsk3 and Hsk4 sample the middle of the profile and Hsk7 the top. An obsidian lava flow (Hsk8) covers the tephra pile, forming a summit cap on the Hrafntinnusker mount, and represents the last stage of the eruption.

Samples from Snæfellsjökull volcano are pumices greater than 1 cm in diameter, collected from two tephra layers approximately 10 km NE of the volcano. The pumices (SN1, SN2) contain a few plagioclase phenocrysts as well as clinopyroxene (less than 5%) and, more rarely, a few phenocrysts of amphibole. In contrast, all samples from Ljósufjöll are compact lava samples from subglacial extrusive units; domes and lava flows. Their mineralogy is characterized by 1–2% feldspar phenocrysts that frequently form glomerocrysts showing signs of resorption, with trace amounts of Fe–Ti oxides, clinopyroxene, amphibole, zircon, monazite and quartz (Flude, personal communication). All samples look fresh under the microscope, with no visible alteration products.

Analytical techniques

Major element compositions of whole rock samples were measured on ULTIMA C ICP-AES, using purified lithium tetraborate fusion of rock powder. The ICP-AES analytical conditions are given in Cantagrel and Pin (1994) and international rock standards (BHVO, BR, GH and DRN) were used for instrument calibration. Trace element compositions were obtained on a PQ2 + ICP-MS instrument, on the same solution as the major element analyses. The ICP-MS analytical conditions will be given elsewhere. Internal standards (BR and BHVO) were used for instrument calibration. The Th and U concentrations were measured by isotopic dilution method on a CAMECA TSN 206 mass spectrometer using a mixed ^{235}U–^{230}Th spike (Condomines et al. 1982). Analytical error is estimated at 0.5% (2σ) for Th and U concentrations. The Sr and Nd isotope ratios were measured on a Thermo-Finnigan TRITON TI mass spectrometer and normalized to ^{86}Sr/^{88}Sr = 0.1194 and ^{146}Nd/^{144}Nd = 0.7219. During this study, the ^{87}Sr/^{86}Sr of NBS-987 was equal to 0.710258 ± 6 (2σ; $n = 8$) and ^{143}Nd/^{144}Nd of the AMES standard equal to 0.511960 ± 4 (2σ; $n = 9$). External error is about 2.10^{-5} (2σ) for both Sr and Nd isotopic ratios. For oxygen isotope measurements, 5–10 mg of whole rock powder was reacted overnight with BrF_5 at 680°C in Ni reaction vessels (Clayton and Mayeda 1963). After conversion to CO_2 gas, the O isotope ratios were analyzed on a VG SIRA 10 dual inlet instrument at "Géosciences Rennes" and normalized to NBS-28 = 9.6‰ (the mean value of the NBS 28 found during the experiments was 9.36‰). Total analytical uncertainties on δ^{18}O are estimated at 0.15‰ based on duplicate analysis.

Results

Major and trace elements

Major and trace element compositions are listed in Table 1 together with the isotope ratios of Sr, Nd and O for Torfajökull, Snæfellsjökull and Ljósufjöll samples. All samples are silicic with SiO_2 ranging from 70.2 to 73.6 wt% (on an anhydrous basis) for Torfajökull rhyolites, 70 to 71.6 wt% for Ljósufjöll samples, and 67–68.1 wt% for the trachytes of Snæfellsjökull (Fig. 2). The Torfajökull samples are all peralkaline rhyolites (except Hsk1) with an agpaitic index (molecular [(Na_2O + K_2O)/Al_2O_3]) higher than 1 (from 1.02 to 1.15) and between 1.13 and 1.5 wt% of

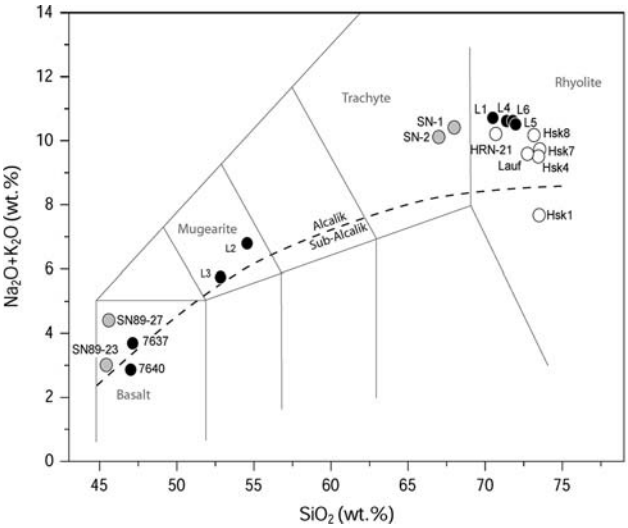

Fig. 2 Total alkali versus SiO_2 diagram with rock classification boundaries from Le Bas et al. (1986). The limit between alkalic and sub-alkalic fields (*dashed line*) is that of Miyashiro (1978). *Filled, grey and open circles* represent samples from Ljósufjöll Snæfellsjökull and Torfajökull, respectively

CIPW normative acmite. The same holds for the peralkaline rhyolites of Ljósufjöll, which have an agpaitic index of ~1.1, and 0.9–2% normative acmite. In contrast, the Snæfellsjökull trachytes have an agpaitic index of less than 1 (0.91 and 0.94).

The compositional similarity between Torfajökull and Ljósufjöll rhyolites is also observed in their trace element patterns. In a primitive mantle normalized multi-element diagram (Fig. 3) rhyolites from both volcanoes display important negative anomalies for Ba, Sr and, to a lesser extent, for Eu, whereas the Snæfellsjökull trachytes display a strong negative anomaly for Sr and a weak one for Eu. Incompatible trace elements, such as Th, have concentrations reaching record high values in the peralkaline rhyolites, up to 21.9 ppm and 26.3 ppm in Torfajökull and Ljósufjöll rhyolites, respectively, whereas silicic rocks elsewhere in Iceland rarely exceed 11 ppm Th (Sigmarsson et al. 1991). The trachytes of this study have 11.5–12.3 ppm.

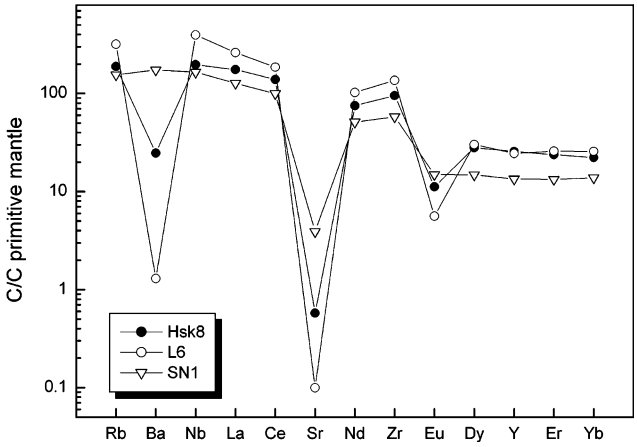

Fig. 3 Primitive mantle normalized (Sun and McDonough 1989) multi-element diagram for Torfajökull (Hsk8) and Ljósufjöll (L6) peralkaline rhyolites and Snæfellsjökull trachyte (SN1). These patterns illustrate the strong Ba, Eu and Sr anomalies in typical samples from the first two volcanoes

The very high water concentrations in some of the Hsk samples from Torfajökull (Table 1) are most probably due to water absorption of the volcanic glass during, or shortly after, their emplacement. The detailed process of glass hydration and consequences for trace and isotopic composition is outside the scope of this paper and will be discussed elsewhere. Nevertheless, the concentrations of Al_2O_3, $Fe_2O_3^t$, CaO, Zr (Fig. 4a), Hf and Th decrease from sample Hsk1 to Hsk7, that is from the base of the profile to the top. In contrast, Na_2O, Ba, Rb and U concentrations increase upwards through the profile. Other elements show insignificant variations throughout the profile. Since Th and U concentrations are negatively correlated, the Th/U strongly decreases from bottom (Th/U = 4.24 in Hsk1) to top (Th/U = 3.36 in Hsk8) of the Hrafntinnusker section (Fig. 4b). As a result, only those elements from the Torfajökull samples, that are not fluid-mobile will be discussed in this paper.

The lava samples from Ljósufjöll contain little water despite their subglacial origin (Table 1). When compared with basalts from the same volcanic system (7637 and 7640) and mugearites (L2, L3; Fig. 2), all the peralkaline rhyolites are depleted in TiO_2, $Fe_2O_3^t$, CaO, P_2O_5 and Sr but enriched in alkalis (Na_2O and K_2O) and in the most incompatible trace elements. Only Ba deviates from this general behaviour, since the highest Ba contents are observed in the mugearites (Table 1). The rhyolites have very similar compositions except L1, which has significantly higher La/Yb and Th/U (37 and 4.41 compared to 14–15 and 3.53–3.85, respectively; Fig. 4b, c).

O, Nd and Sr isotopes

The whole-rock δ^{18}O-values measured in Hsk8 and Lauf are 3.9 and 2.4‰, respectively. The latter value is significantly lower than those measured by Gunnarsson (1987) and Gunnarsson et al. (1998) in Laufafell (4.2‰). In stark contrast, the peralkaline rhyolites from Ljósufjöll have δ^{18}O-values ranging from 5.8 to 6.3‰, and the mugearites have values of 5.8‰. The trachytes of Snæfellsjökull have slightly lower δ^{18}O-values of 4.9–5.1‰.

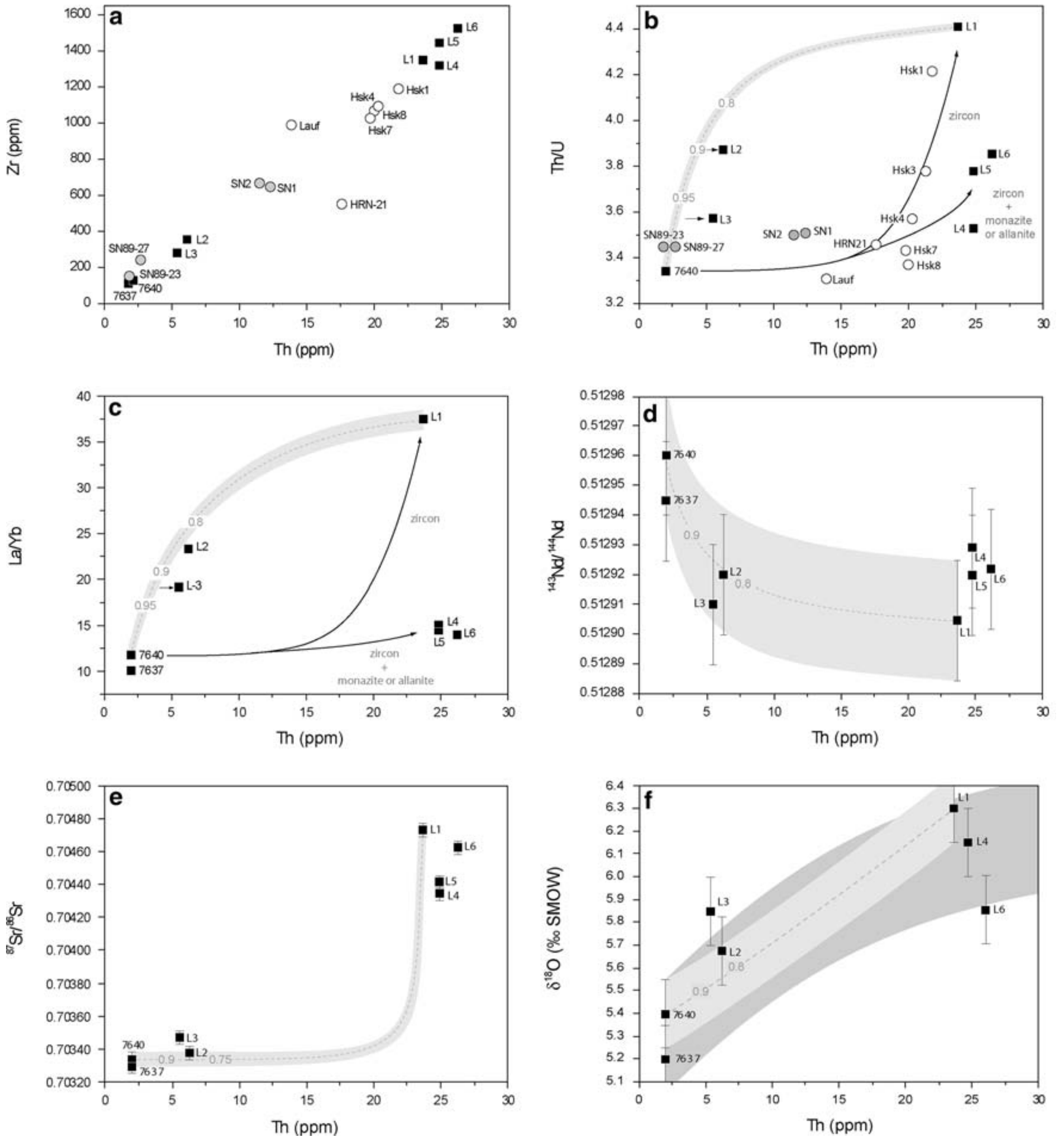

Fig. 4 The Th concentration versus **a** Zr concentration, **b** Th/U, **c** La/Yb, **d** ^{143}Nd/^{144}Nd, **e** ^{87}Sr/^{86}Sr and **f** δ^{18}O. *Open and grey circles* represent samples from Snæfellsjökull and Torfajökull, respectively and *black squares* the Ljósufjöll ones. **a** The observed linear correlation precludes important fractionation of zircon during magma differentiation with the exception for sample HRN21. High zirconium in the Lauf sample suggests an accumulation of zircon. **b–f** *Solid arrows* or *dark grey* areas show the computed fractional crystallisation evolutions and *light grey shaded* area represent the calculated mixing curve between basalt 7640 and rhyolite L1. The curve width represents the 2σ analytical uncertainty

^{143}Nd/^{144}Nd does not show any significant variation in the samples of this study (Torfajökull: 0.51296–0.51298, Ljósufjöll: 0.51290–0.51296, Snæfellsjökull: 0.51296–0.51297; Table 1). The results for Torfajökull fall within the range observed for a larger sample set by Stecher et al. (1999), whereas those from Ljósufjöll, together with those for the Öræfajökull rhyolites, SE Iceland (Prestvik et al. 2001), display the lowest Nd isotope ratios measured in Icelandic rocks.

^{87}Sr/^{86}Sr ratios show large variations from 0.70332 at Laufafell, through 0.70337 at Snæfellsjökull to 0.70386 in the first tephra (Hsk1) of the Hrafntinnusker

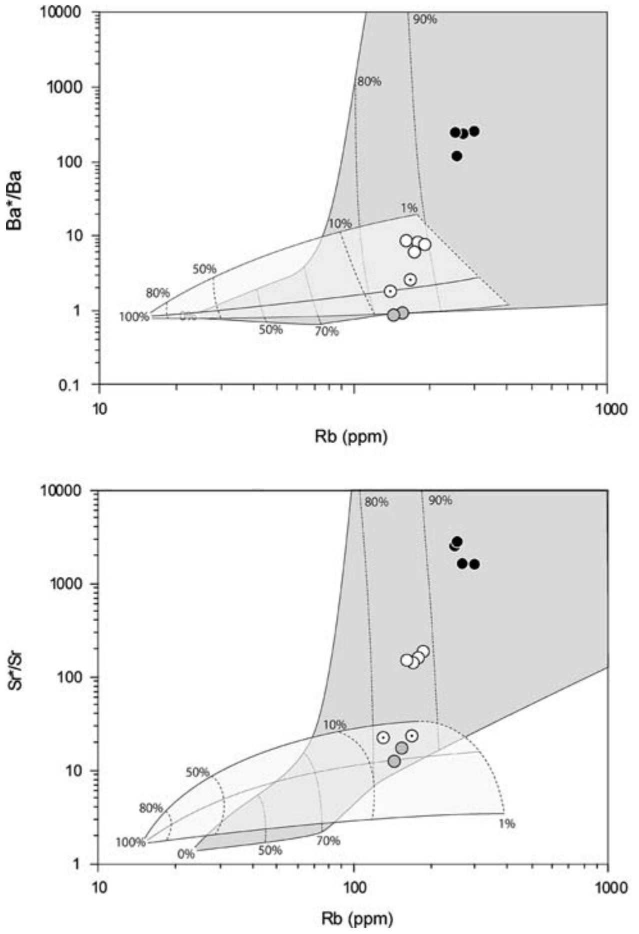

Fig. 5 Diagrams of Ba*/Ba and Sr*/Sr versus Rb concentration showing different compositional fields for fractional crystallisation and crustal partial melting. *Light grey area* is the domain for evolved melts generated by partial melting of the Iceland basaltic crust, whereas the *dark grey area* represents the domain formed by fractional crystallisation from a basalt of Ljósufjöll. The value for the average crust is taken from drill cuttings at 800 m depth in the geothermal field of Krafla, N-Iceland (Ingvarsson and Sigmarsson, in preparation). The Ba* is defined as $10^{(\mathrm{Log[Rb]+Log[Nb]})/2}$ and Sr* as $10^{(\mathrm{Log[Ce]+\ Log\ [Nd]})/2}$ and all concentrations are normalized to primitive mantle values (Sun and McDonough 1989). Symbols as in Fig. 2 except *open circles* with a central point for Lauf- and HRN- samples (see text for further discussion)

eruption (Torfajökull) and up to 0.70473 at Ljósufjöll. This last value is by far the highest $^{87}Sr/^{86}Sr$ ever measured in Icelandic rocks.

Discussion

Petrogenesis

Deciphering which of the two principal mechanisms of silicic magma production, crustal melting or crystal fractionation, is dominant is important to be related to the thermal state of the crust. A cold crust would need a large input of heat from basalts to reach its solidus, whereas hotter crust is more easily melted. At the same time, cold crust cools down basaltic intrusions more than hotter crust, thus favoring the formation of evolved magma. The arguments for or against each mechanism of silicic magma production will be evaluated for each volcanic system studied.

Torfajökull

The most recent studies of the petrogenesis of the large Torfajökull silicic complex are those of Macdonald et al. (1990) and Gunnarsson et al. (1998). The former reached the conclusion that fractional crystallisation explains the compositions of the silicic lavas whereas the latter prefer partial melting of a silicic intrusion produced by fractional crystallisation. The relatively low $\delta^{18}O$ of the silicic rocks are interpreted as resulting from partial melting of hydrothermally altered crust, which may contain plagiogranites (Gunnarsson et al. 1998; Sigurdsson 1977) or be in the amphibolite facies (Oskarsson et al. 1982). Our Hrafntinnusker obsidian has similar $\delta^{18}O$ to those of Gunnarsson et al. (1998) whereas the low value for Laufafell (2.4‰) is clearly lower than those published for the basalts of Torfajökull (4.5–5‰). A lower $\delta^{18}O$-value in silicic compared to mafic magma is similar to that which has been observed for Krafla and Askja volcanoes (Nicholson et al. 1991; Sigmarsson et al. 1991). Application of the crustal melting model based on results from the neighbouring volcano, Hekla (Sigmarsson et al. 1992a), is therefore appropriate to the Torfajökull case. In this model, the hydrothermally altered basalts in the amphibolite facies experience amphibole dehydration melting to produce a dacitic melt. This melt separates from its source and fractionates towards a rhyolitic composition upon cooling during ascent through the crust.

The magmatic evolution of the Torfajökull volcanic system is faithfully recorded by the concentrations of the highly incompatible element Th. The Th concentration varies by over an order of magnitude, from 2.65 ppm in basaltic inclusions in the Hrafntinnuhraun obsidian lava (Sigmarsson unpublished results), to 21.9 ppm in the first emitted silicic pumice of the Hrafntinnusker eruption (Hsk1). The ca. 80 m-thick tephra sequence of this eruption has Th concentrations that decrease only slightly down to 19.7 ppm in the most recent tephra at the top of the sequence. Thorium shows a very good correlation with Zr (Fig. 4a), which suggests that no or very limited zircon fractionation has

Table 2 Partition coefficients used in trace element modelling (from Brooks et al. 1981; Fujimaki et al. 1984; Mahood and Hildreth 1983; Nash and Crecraft 1985; Paster et al. 1974)

	Olivine	Clinopyroxene	Plagioclase	Orthoclase	Ilmenite	Zircon	Alanite
La	0.0067	0.056–0.19	0.302–0.38	0.08	0.098–1.223	4.18–16.9	820–2,362
Ce	0.006	0.15–0.5	**0.005**–0.221–0.27–**0.62**	**0.017**–0.037–**0.095**	0.11–1.64	4.31–16.75	635–2,063
Nd	0.0059	0.31–1.11	**0.008**–0.149–0.21–**0.29**	**0.009**–0.035–**0.093**	0.14–2.267	4.29–13.3	463–1,400
Sm	0.007	0.5–1.67	0.102–0.13	0.025	0.15–2.833	4.94–14.4	205–756
Eu	0.0074	0.51–1.56	0.376–1.4	2.6	0.1–1.013	3.31–16	81–122
Gd	0.01	0.61–2	0.067–0.097	0.025	0.14–2.3	6.59–12	130
Dy	0.013	0.68–1.93	0.05–0.064	0.055	0.145–3.267	47.4–101.5	50–123
Er	0.0256	0.65–1.8	0.045–0.055	0.006	0.15–2.2	99.8–135	25–60
Yb	0.0491	0.62–1.8	0.041–0.049	0.03	0.17–1.467	128–527	8.9–24.5
Rb	0.0098	0.013–0.032	**0.016**–0.04–0.07–**0.46**	**0.11–1.75**			
Ba	0.0099	0.026–0.131	**0.03**–0.16–2.5–**19.5**	**1–44**	0.00034		
Nb	0.01	0.005–0.8	**0.008**–0.025–0.035–**0.27**	**0.01–0.15**	0.8–6.58		
Sr	0.014	0.06–0.516	**1.3**–1.8–4.7–**33**	**1.6–22**	0.1–0.5		
Zr	0.012	0.1–0.6	0.01–0.135	0.03	0.28–1.38		
Y	0.01	0.9–4	0.06–0.13	0.03	0.0045–0.3		
Th		0.36–2.89	0.01–0.048	0.023	0.00055–0.463	62–91	168–420
U		0.12–0.33	0.01–0.09	0.048	0.0082–0.517	298–383	6–14

Values in *bold* characters correspond to the highest and the lowest partition coefficients as compiled in the *Earth Science reference data and models* (http://www.earthref.org) and were used for the models in Fig. 5

taken place, in agreement with the high zirconium solubility in peralkaline magmas (Watson 1979). The slightly decreasing Th concentration of the crystal-free pumice during the Hrafntinnusker eruption imply both a small and finely stratified liquid magma layer that was inverted upon eruption, together with a very efficient method of liquid-crystal separation. This situation is comparable to that proposed for high-silica rhyolites elsewhere, for which a gas-filter pressing mechanism is advocated for liquid-crystal separation (e.g. Sisson and Bacon 1999; Bachmann and Bergantz 2004).

In a primitive mantle normalized multi-element diagram (Fig. 3) Hrafntinnusker peralkaline rhyolites show important negative anomalies for Ba, Sr and to a lesser extent, for Eu. These anomalies can be shown to be inconsistent with crustal partial melting alone, whereas feldspar fractionation can readily account for the observed Ba, Sr and Eu depletions (Fig. 5). In Fig. 5 two compositional fields are shown for an evolved magma derived from: (1) fractional crystallisation of a transitional basalt (sample 7637) and (2) partial melting of a hydrothermally altered basalt the latter being considered as a representative average of the Icelandic basaltic crust. In addition, in order to access the model boundaries, the highest and the lowest partition coefficients of Sr and Ba (Table 2) have been considered for plagioclase and orthoclase. Figure 5 clearly demonstrates that in Hrafntinnusker rhyolites the Ba anomaly can be produced by both mechanisms, whereas the Sr anomaly reflects either high degrees (85–90%) of fractional crystallisation or partial crustal melting with subsequent crystal fractionation. Hence, fractional crystallisation is an important process during the formation of the silicic magma at Torfajökull. However, the O-isotope ratios clearly show that, prior to the overprint of the final crystal fractionation stage, the magmas were originated from crustal partial melting. Thus, a two-stage process can be envisaged: (1) partial melting of hydrated metabasaltic crust generating a silicic magma having $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ more or less comparable to those of contemporaneous and spatially related basalts, but with lower $\delta^{18}O$-values; (2) subsequent fractional crystallisation of plagioclase and/or orthoclase accounting for the strong Sr depletion (as well as Ba and Eu) observed in Hrafntinnusker rhyolites. The principal difference with the case of Hekla (Sigmarsson et al. 1992a) is that, during the second stage, the role of zircon fractionation is significantly reduced because of the peralkaline nature of the silicic magma in Torfajökull.

Ljósufjöll

Rhyolite genesis

The peralkaline rhyolites of Ljósufjöll have several characteristics in common with those of Torfajökull. In a primitive mantle normalized multi-element diagram (Fig. 3) they display very strong depletions in Ba, Sr and, to a lesser extent, in Eu, due to feldspar fractionation. In contrast with Torfajökull volcano, the

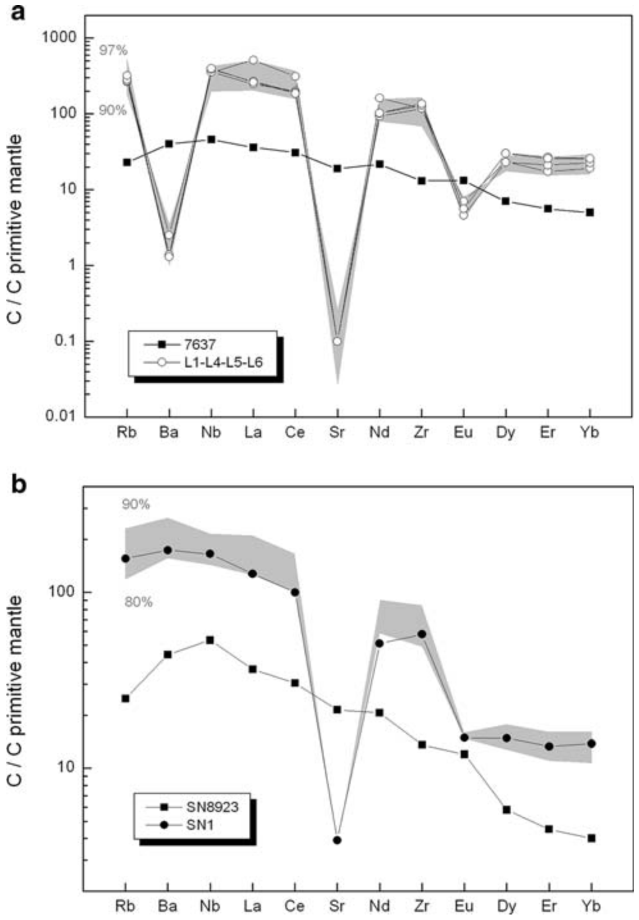

Fig. 6 Primitive mantle normalized (Sun and McDonough 1989) multi-element diagram comparing **a** samples L1–L4–L5–L6 with the calculated trace elements composition (*grey area*) of liquid obtained by 90–97% fractional crystallisation from the 7637 basalt. **b** Trachyte SN1 compared with calculated trace element compositions for 80–90% fractional crystallisation of SN89–23 basalt from the vicinity of Snæfellsjökull volcano. The good fit between calculated and measured concentrations suggests that fractional crystallisation fully accounts for the compositional variability of the silicic rocks of Ljósufjöll and Snæfellsjökull

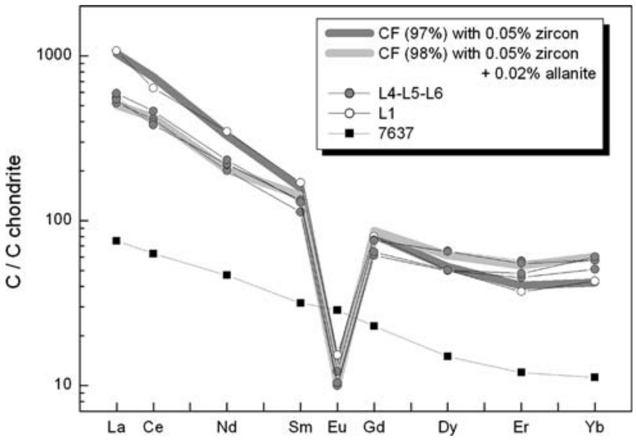

Fig. 7 Chondrite normalized REE patterns comparing rhyolites L1–L4–L5–L6 with the calculated trace element compositions of liquid obtained by 97–98% fractional crystallisation including small proportions of zircon and allanite

δ^{18}O is higher in the silicic lavas (5.8–6.3‰, Table 1) than in associated basalts from the Ljósufjöll volcanic system (5.2–5.4‰). In this case, there is no reason to invoke crustal partial melting since fractional crystallisation can fully explain the observed geochemical variations.

The genesis of rhyolites by fractional crystallisation of a basaltic magma can be tested and modelled in two ways. The mechanism is first verified using mass balance calculations (Stormer and Nicholls 1978). Based only on major elements, this algorithm calculates both the modal and chemical compositions of the crystallizing assemblage, which fractionated from the magma, as well as the degree of crystallisation. The mass balance calculation relates the genesis of a rhyolite (L4) to a basaltic parental magma (7637) with a fractionating mineral assemblage of plagioclase (50%) + clinopyroxene (28%) + Fe–Ti oxides (15%) + olivine (4%) + orthoclase (3%), yielding a low sum of squared residuals ($\Sigma r^2 = 0.06$). The necessary degree of fractional crystallisation is very high, 94%. When the mineralogical composition of the fractionating mineral assemblage is reintroduced into trace element modelling, the outcome fits perfectly with the analytical data of the rhyolites for degrees of fractionation ranging from 90 to 97% (Fig. 6a). However, slight differences appear when the REE are considered. For example, sample L1 is LREE-richer and HREE-poorer than other rhyolites (L4, L5 and L6) and has higher Th/U (Figs. 4b, c and 7). Fractionation of accessory mineral phases must therefore be taken into consideration. Only a very small amount of zircon fractionation (0.05%) is needed in order to account for the observed REE, without causing much decrease in Zr (Fig. 7), and it also accounts for the observed Th/U in sample L1, if $D_{\mathrm{Th}}^{\mathrm{zircon/melt}}$ and $D_{\mathrm{Th}}^{\mathrm{zircon/melt}}$ are as high as 76 and 340, respectively (Table 2). However, the REE patterns, specifically the LREE, of the other rhyolites (samples L4, L5 and L6) require the fractionation of a different accessory mineral such as monazite or allanite. The high U and Th partition coefficients for allanite ($D_{\mathrm{Th}}^{\mathrm{allanite/melt}}$ and $D_{\mathrm{Th}}^{\mathrm{allanite/melt}}$ are ~283 and ~12, respectively; Table 2) reduce the Th/U ratio in the melt. Figure 7 illustrates the effect of 0.02% allanite fractionation on REE patterns when added to the same fractionating mineral assemblage as for L1 (including 0.05% of zircon). This modified model fits the REE compositions of samples L4, L5 and L6. It also predicts a Th/U of ~3.7 (Th~25.4 and U~6.8) which is consistent with the Th and U content of these rhyolites (24.9–26.3 and 6.56–7.06, respectively). Monazite is similar to allanite, in that it has a Th/U typically between 5 and 50 (e.g. Deer et al. 1963; Förster 1998). When monazite is considered instead of allanite, a smaller proportion (around 0.002%) is required to provide the same effect. Such low fractionation rates are coherent with the high monazite solubility in peralkaline magmas (Montel 1986).

The results from the major and trace element modelling can be tested for the O isotope results on the same samples. The minimal δ^{18}O increase (≤1‰) from the basalts to the peralkaline rhyolites can be accounted for by the 94% fractional crystallisation, using the phase proportions (assumed to be constant during the differentiation process) from the major and trace element modelling and mineral-magma isotopic fractionation coefficients estimated from Harris and Smith (2000) and Eiler (2005) ($\Delta_{\text{plagioclase-melt}} = 0.2$, $\Delta_{\text{clinopyroxene-melt}} = -0.2$, $\Delta_{\text{orthoclase-melt}} = 0.4$, $\Delta_{\text{olivine-melts}} = -0.6$, $\Delta_{\text{Magnetite/Ilmenite-melts}} = -2$; Fig. 4f).

The high degree of fractional crystallisation at the origin of the rhyolites in Ljósufjöll results in very low Sr contents and the highest Rb/Sr ever analysed in Icelandic rocks (90–150). Considering the large variability in Rb/Sr between rhyolites and basalts and the K–Ar ages (~100 to ~800 ka) obtained on the Ljósufjöll central volcano by Flude et al. (2004), the time corrected ^{87}Sr/^{86}Sr of rhyolites and basalts are similar, as would be expected with the fractional crystallisation mechanism.

An interesting corollary of the Sr-isotope results from Ljósufjöll is that old silicic formations in Iceland, which were formed by fractional crystallisation should have elevated ^{87}Sr/^{86}Sr if their Rb/Sr has been increased by plagioclase fractionation. This could be a useful criterion when discussing the petrogenesis of Tertiary volcanoes.

Intermediate rocks genesis

The intermediate compositions of samples L2 and L3 are not readily explained by a simple fractional crystallisation scenario. This is especially true for their relatively high Th/U (3.57–3.87) compared to those of the basalts (Th/U = 3.34, Sigmarsson et al. 1992). However, magma mixing between previously formed rhyolites and incoming basalts of a similar composition to that presented in Table 1, appears to satisfy all geochemical parameters. Figure 4 shows the results of a mixing model between a basaltic magma (7640) and rhyolite L1. In order to account for the Th/U, La/Yb, ^{87}Sr/^{86}Sr, ^{143}Nd/^{144}Nd and δ^{18}O of the intermediate samples, mixing proportions of 80–95% basalt with 5–20% rhyolite are required. However, the composition of the rhyolite end-member must have been slightly less affected by feldspar fractionation than sample L1, in order to match the Ba, K, Na and Eu variations. Complex zonation and resorption edges of the feldspars are compatible with such magma mixing.

Snæfellsjökull trachytes

Both basic and acid magmas from Snæfellsjökull have identical Sr and Nd isotopes compositions (^{143}Nd/^{144}Nd ~0.51297 ± 1 and ^{87}Sr/^{86}Sr ~0.70337 ± 1; Table 1), which could indicate that they derived from a common source. As for the Ljósufjöll samples, the genesis of trachyte SN-1 by fractional crystallisation of alkali basalt was tested for both major and trace elements. The extracted mineral assemblage which best fits the data consists of Plagioclase (39%) + Clinopyroxene (37%) + Olivine (13%) + Fe-Ti oxides (11%). The Σr^2 is low (0.2) and the degree of fractional crystallisation is 87%. This mineralogical assemblage is without alkali feldspar and the extracted plagioclase is of a smaller amount than in Hrafntinnusker and Ljósufjöll samples. When the mineralogical composition of the crystallising assemblage is reintroduced into trace element modelling, it appears that the model fits analytical data for degrees of fractionation ranging from 80 to 90% (Fig. 6b). In the same way as at Ljósufjöll, very small amounts (0.01%) of fractionating zircon are required to adjust the Th and U concentrations. This results in the calculated Th and U concentrations (Th~12.3 ppm, U~3.5 ppm and Th/U~3.5) being consistent with SN-1 and SN-2 concentrations (Th = 12.3 and 11.5 ppm, U = 3.5 and 3.29 ppm, Th/U = 3.51 and 3.50, respectively). A closed system differentiation mechanism operating from basalts to trachytes appears thus to be the case for Snæfellsjökull volcano.

The δ^{18}O-values of the trachytes (4.9–5.1‰) are slightly lower than those of the Holocene basalts (5.3–5.4‰; Sigmarsson et al. 1992b and references therein) erupted around the Snæfellsjökull stratovolcano. As already discussed for Ljósufjöll magmas, large degrees of fractional crystallisation are expected to slightly increase the δ^{18}O in the derived magma when significant amounts of Fe–Ti oxides fractionation takes place. The O-isotopes of Snæfellsjökull lavas cannot, therefore, be explained by a closed-system fractional crystallisation process alone. However, slight crustal contamination from the metabasaltic crust of the ancient rift zone in Snæfellsnes is capable of explaining the δ^{18}O. In an assimilation-fractional crystallisation (AFC) process, assuming an average crustal δ^{18}O of 1–3‰ (Gautason and Muehlenbachs 1998) the mass ratio of crystal extract to assimilated crust would be in the range of 10–30 depending on the exact O-isotope fractionation factors between minerals and liquid. Such a high mass ratio of fractionating crystals over assimilated melt reflects a differentiation process occurring in a relatively "cold" environment. This illustrates the importance of the role of crystal fractionation compared to that of crustal partial melting. Accordingly, we propose that the compositions of the Snæfellsjökull trachytes are best explained by fractional crystallisation with minor crustal assimilation. Thus it appears that the genesis of the silicic magma at Snæfellsjökull is likely to lie between the contrasting mechanisms inferred for Ljósufjöll and Torfajökull, namely crustal cooling effect leading to fractional crystallization and crustal melting due to a hotter environment.

Geodynamic environment

Figure 1 depicts the depth to a inferred 1,200°C isotherm, which is estimated to be 20–25 km beneath Ljósufjöll and only 15–20 km beneath Snæfellsjökull. For Torfajökull, this isotherm can be assumed to be even shallower due to the influence of the propagating rift zone entering the silicic complex (e.g. Oskarsson et al. 1982). Partial melting of hydrated metabasaltic crust would only occur where the geothermal gradient is high and large degrees of a fractional crystallisation would be restricted to areas with low geothermal gradient.

Existing work on the genesis of silicic rocks supports the model presented in this paper. For instance, Sigmarsson et al. (1991, 1992a) proposed that the central volcanoes within or close to the rift zone resulted from partial melting of basaltic crust. Whereas, Prestvik et al. (2001) suggested that the silicic melts of the off-rift Öræfajökull central volcano (South-East Iceland) were generated by fractional crystallisation of basaltic magmas. Both these explanations are fully consistent with the geothermal gradient controlling which mechanism prevails.

Conclusion

The interaction between the mid-Atlantic ridge and a mantle plume beneath Iceland gives rise to high magma productivity, resulting in an elevated geothermal gradient. The lower amphibolitic crust is likely to cross its hydrous solidus at depth and will consequently undergo partial melting, generating silicic magmas within, or close to, the rift-zone. In contrast, at a distance from the rift zones or the plume centre, the geothermal gradient is significantly lower; there, the crust is unlikely to cross its hydrous solidus. Under the latter conditions, silicic magmas are generated by a high degree of fractional crystallisation of basaltic magmas in an off-rift context, as in the Ljósufjöll and Öræfajökull areas. The intermediate geothermal gradient beneath Snæfellsjökull is best explained by a combined magma differentiation mechanism, namely assimilation-fractional crystallisation with less than 10% of crustal assimilation. Therefore a clear link between the mode of silicic magma formation and the thermal state of the Icelandic crust appears to exist.

Acknowledgement We greatly appreciate the guidance and help from Serge Fourcade at "Géosciences Rennes" during the O-isotope analysis, in addition to his pertinent remarks on an earlier version of this manuscript. Analytical assistance from Delphine Auclair, Chantal Bosq, Karine David and Mhammed Benbakkar is also gratefully acknowledged. Fruitful discussions on trace element modelling with Hervé Martin were very useful. We could not have had better company than Gudrun Larsen and Bergrun A. Oladottir during the fieldwork at Hrafntinnusker. Efficient English corrections by Fran Van Wyk de Vries were much appreciated. Constructive comments from two anonymous reviewers improved the manuscript. Finally, a grant from the Icelandic Science Foundation and the French-Icelandic Jules Verne collaboration program is acknowledged.

References

Bachmann O, Bergantz GW (2004) On the origin of crystal-poor rhyolites: extracted from batholithic crystal mushes. J Petrol 45:1565–1582

Breddam K, Kurz MD, Storey M (2000) Mapping out the conduit of the Iceland mantle plume with helium isotopes. Earth Planet Sci Lett 176:45–55

Brooks CK, Henderson P, Ronsbo JG (1981) Rare earth element partitioning between allanite and glass in the obsidian of Sandy Braes, northern Ireland. Miner Mag 44:157–160

Cantagrel F, Pin C (1994) Major, minor and rare-earth element determinations in 25 rock standards by ICP-atomic emission spectrometry. Geostand Newslett 18:123–138

Carmichael ISE (1964) The petrology of Thingmuli, a tertiary volcano in Eastern Iceland. J Petrol 5:435–460

Carpentier M (2003) Variabilité géochimique des basaltes Holocène islandais. Université Blaise Pascal, Clermont Ferrand, p 53

Christiansen EH, McCurry M (2006) Contrasting origins of Cenozoic silicic volcanic rocks from the Western Cordillera of the United States. Bull Volcanol (in press)

Clayton RN, Mayeda TK (1963) The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis. Geochim Cosmochim Acta 27:43–52

Condomines M, Tanguy JC, Kieffer G, Allègre CJ (1982) Magmatic evolution of a volcano studied by 230 Th–238 U disequilibrium and trace elements systematics: the Etna case. Geochim Cosmochim Acta 46:1397–1416

Deer WA, Howie RA, Zusmna J (1963) Rock-forming minerals, vol Longmans. Green and co LTD, London

Eiler J (2005) Oxygen isotope variations of basaltic lavas and upper mantle rocks. In: Walley JW, Cole DR (eds) Reviews in Mineralogy and geochemistry, vol 43. The mineralogical society of America, Washington, pp 319–364

Eysteinsson H, Gunnarsson K (1995) Maps of gravity, bathymetry and magnetics for Iceland and surroundings. In: vol. Orkustofnun, Reykjavik, 05–95055/JHD–07

Flovenz OG, Saemundsson K (1993) Heat flow and geothermal processes in Iceland. Tectonophysics 225:123–138

Flude S, Burgess R, McGarvie DW (2004) Volcanism at Ljósufjöll central volcano, western Iceland: chemostratigraphy and geochronology. In: IAVCEI, vol, Pucón-Chile

Förster H-J (1998) The chemical composition of the REE-Y-Th-U-rich accessory minerals in peraluminous granites of the Erzgebirge-Fichtelgebirge region, Germany, PartI: The monazite-(Ce)-brabantite solid solution series. Am Mineral 83:259–272

Fujimaki H, Tatsumoto M, Aoki K (1984) Partition coefficients of Hf, Zr, and REE between phenocrysts and groundmasses. J Geophys Res 89:662–672

Furman T, Frey FA, Meyer PS (1992) Petrogenesis of evolved basalts and rhyolites at Austurhorn, Southeastern Iceland: the role of fractional crystallisation. J Petrol 33:1405–1445

Gautason B, Muehlenbachs K (1998) Oxygen isotopic fluxes associated with high-temperature processes in the rift zones of Iceland. Chem Geol 145:275–286

Gudmundsdóttir LS, Sigmarsson O (2006) Highest measured strontium isotope ratios in Icelandic rocks: Rb and Sr systematics in Ljósufjöll volcanics, Snæfellsnes peninsula. Natural Science Symposium, Reykjavik

Gunnarsson B (1987) Petrology and petrogenesis of silicic and intermediate lavas on a propagating oceanic rift. The Torfajökull and Hekla central volcanoes south-central Iceland. Johns Hopkins University, Baltimore, Maryland, p 432

Gunnarsson B, Marsh BD, Taylor JHP (1998) Generation of Icelandic rhyolites: silicic lavas from the Torfajökull central volcano. J Volcanol Geotherm Res 83:1–45

Harris C, Smith HS, Le Roex AP (2000) Oxygen isotope composition of phenocrysts from Tristan da Cunha and Gough Island lavas: variation with fractional crystallization and evidence for assimilation. Contrib Mineral Petrol 138:164–175

Jakobsson SP (1972) Chemistry and distribution pattern of recent basaltic rocks in Iceland. Lithos 5:365–386

Jóhannesson H, Flores RM, Jónsson J (1982) A short account of the Holocene tephrochronology of the Snaefellsjökull central volcano, Western Iceland. Jökull 31:23–30

Jónasson K (1994) Rhyolite volcanism in the Krafla central volcano, North-east Iceland. Bull Volcanol 56:516–528

Le Bas MJ, Le Maitre R, Streckeisen A, Zanettin B (1986) A chemical classification of volcanic rocks based on the total alkali-silica diagram. J Petrol 22:745–750

Macdonald R, McGarvie DW, Pinkerton H, Smith RL, Palacz ZA (1990) Petrogenetic evolution of the Torfajökull volcanic complex, Iceland I. Relation between the magma types. J Petrol 31:461–481

Mahood G, Hildreth W (1983) Large partition coefficients of trace elements in high-silica rhyolites. Geochim Cosmochim Acta 47:11–30

McGarvie DW (1984) Torfajökull: a volcano dominated by magma mixing. Geology 12:685–688

Miyashiro A (1978) Nature of alkalic volcanic rock series. Contrib Mineral Petrol 66:91–104

Montel J-M (1986) Experimental determination of the solubility of Ce-monazite in SiO_2-Al_2O_3-K_2O-Na_2O melts at 800°C, 2 kbar, under H_2O-saturated conditions. Geology 14:659–662

Nash WP, Crecraft HR (1985) Partition coefficients for trace elements in silicic magmas. Geochim Cosmochim Acta 49:2309–2322

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

Oskarsson N, Sigvaldason GE, Steinthorsson S (1982) A dynamic model of rift zone petrogenesis and the regional petrology of Iceland. J Petrol 23:28–74

Paster TP, Schauwecker DS, Haskin LA (1974) The behavior of some trace elements during solidification of the Skaergaard layered series. Geochim Cosmochim Acta 38:1549–1577

Prestvik T, Goldberg S, Karlsson H, Grönvold K (2001) Anomalous strontium and lead-isotope signatures in the off-rift Öraefajökull central volcano in south-east Iceland, Evidence for enriched endmember(s) of the Iceland mantle plume? Earth Planet. Sci Lett 190:211–220

Sigmarsson O, Condomines M, Fourcade S (1992a) A detailed Th, Sr and O isotope study of Hekla: differentiation processes in an Icelandic volcano. Contrib Mineral Petrol 112:20–34

Sigmarsson O, Condomines M, Fourcade S (1992b) Mantle and crustal contribution in the genesis of recent basalts from off-rift zones in Iceland: constraints from Th, Sr and O isotopes. Earth Planet Sci Lett 110:149–162

Sigmarsson O, Hémond C, Condomines M, Fourcade S, Oskarsson 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 269:28–28

Sisson TW, Bacon CR (1999) Gas-driven pressing in magmas. Geology 27:613–616

Stecher O, Carlson RW, Gunnarsson B (1999) Torfajokull: a radiogenic end-member of the Iceland Pb-isotopic array. Earth Planet Sci Lett 165:117–127

Stormer JC, Nicholls J (1978) XLFRAC: a program for interactive testing of magmatic differentiation models. Comput Geosci 87:51–64

Sun SS, McDonough WF (1989) Chemical and isotopic systematics of oceanic basalts: Implications for the mantle composition and processes. In: Magmatism in the ocean basin, vol 42. Geol. Soc. Sp. Publi., pp 313–345

Thordarson T, Larsen G (2006) Volcanism in Iceland in historical time: volcano types, eruption styles and eruptive history. J Geodyn. DOI 10.1016/j.jog.2006.09.005

Walker GPL (1966) Acid rocks Iceland. Bull Volcanol 29:375–406

Watson EB (1979) Zircon saturation in felsic liquids: experimental results and applications to trace element geochemistry. Contrib Mineral Petrol 70:407–419

Wolfe CJ, Bjarnason IT, VanDecar SC, Solomon SC (1997) Seismic structure of the Iceland mantle plume. Nature 385:245–247