

# The origins of carbonatites and related rocks from the Grønnedal-Íka Nepheline Syenite complex, South Greenland: C-O-Sr isotope evidence

N. J. G. PEARCE<sup>1</sup>, M. J. LENG<sup>2</sup>, C. H. EMELEUS<sup>3</sup> AND C. M. BEDFORD<sup>3,4</sup>

<sup>1</sup> Institute of Earth Studies, University of Wales, Aberystwyth, Ceredigion, SY23 3DB, UK

<sup>2</sup> NERC Isotope Geosciences Laboratory, Keyworth, Nottingham, NG12 5GG, UK

<sup>3</sup> Department of Geological Sciences, University of Durham, Durham, DH1 3LE, UK

<sup>4</sup> Present address: Radley College, Abingdon, Oxford, OX14 2HR, UK

## Abstract

The Grønnedal-Íka ring complex ( $1299 \pm 17$  Ma) in the Gardar province, South Greenland is composed of a range of layered nepheline syenites which were intruded at a late stage by xenolithic syenite and a plug of carbonatite. The complex was subsequently intruded by a variety of basic dykes, including olivine dolerites, kersantites, vogesites, spessartites, camptonites and an alnöite, and then extensively faulted. The nepheline syenite magmas, produced by fractional crystallisation of basic magmas, show a range in  $\delta^{13}\text{C}$  ( $-3.86$  to  $-7.57\text{‰}$ ) and  $\delta^{18}\text{O}$  ( $8.27$  to  $15.12\text{‰}$ ), distinctly different to the carbonatites which form a tight group with average  $\delta^{13}\text{C} = -4.31 \pm 0.22 \text{‰}$ , (1 s.d.) and average  $\delta^{18}\text{O} = 7.18 \pm 0.41\text{‰}$  (1 s.d.). Initial  $^{87}\text{Sr}/^{86}\text{Sr}$  isotope ratios (typically 0.703) suggest the syenites and carbonatites have not assimilated crustal rocks, and therefore the C and O isotope variation within each group is a result of isotopic evolution during fractional crystallisation. A suite of lamprophyre dykes ( $\delta^{13}\text{C} -3.86$  to  $-7.86\text{‰}$  and  $\delta^{18}\text{O} 9.12$  to  $10.81\text{‰}$ ) form a coherent group whose stable isotope compositions overlap part of the syenite field, and again are distinctly different from the carbonatites. A single alnöite has  $\delta^{13}\text{C} = -3.32\text{‰}$  and  $\delta^{18}\text{O} = 12.34\text{‰}$ . C and O isotope ratios are consistent with origins of syenitic and lamprophyric magmas from a similar source. Despite geochemical evidence which suggests a genetic link between nepheline syenites and carbonatites, C and O isotopic evidence shows that they are not related directly by liquid immiscibility. Comparisons are made between similar rock types from Grønnedal-Íka and from the Gardar Igaliko Dyke Swarm. The possible role of F in controlling  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  during crystallisation of calcite from carbonatite magmas is discussed.

KEYWORDS: carbonatite, nepheline syenite, Grønnedal-Íka complex, Gardar, Greenland.

## Introduction

MAGMATISM in the Mid-Proterozoic Gardar Province, South Greenland was related to repeated episodes of continental rifting which generated large volumes of alkali and transitional basic magmas. Fractional crystallisation of basic parents, with or without crustal contamination, produced residual magmas of alkali granite and nepheline syenite compositions. The geology of the Gardar Province has been summarised by Emeleus and Upton (1976) and Upton and Emeleus (1987). Carbonatite occurs within the Gardar Province as a late-stage plug associated with Early Gardar nepheline syenite at Grønnedal-Íka (Emeleus, 1964; Bedford, 1989), as

dykes within the Late Gardar Igaliko Nepheline Syenite Complex (Emeleus and Harry, 1970; Pearce, 1988; Pearce and Leng, 1996), and as mantle-derived eruptive rocks at Qassiarssuk (previously Qagssiarssuk; Stewart, 1964, 1970; Hayward and Jones, 1991; Andersen, 1997). Emeleus (1964) gave the first detailed description of the geology of the Grønnedal-Íka nepheline syenite complex. He concluded that a gradual build-up of  $\text{CO}_2$  with fractionation in the syenites culminated in the forceful emplacement of a carbonatite plug at a late stage. Figure 1 shows a simplified geological map of the complex, based on the mapping of Emeleus (1964). Gill (1972) studied the geochemistry of the silicate rocks of the Grønnedal-Íka complex, but not

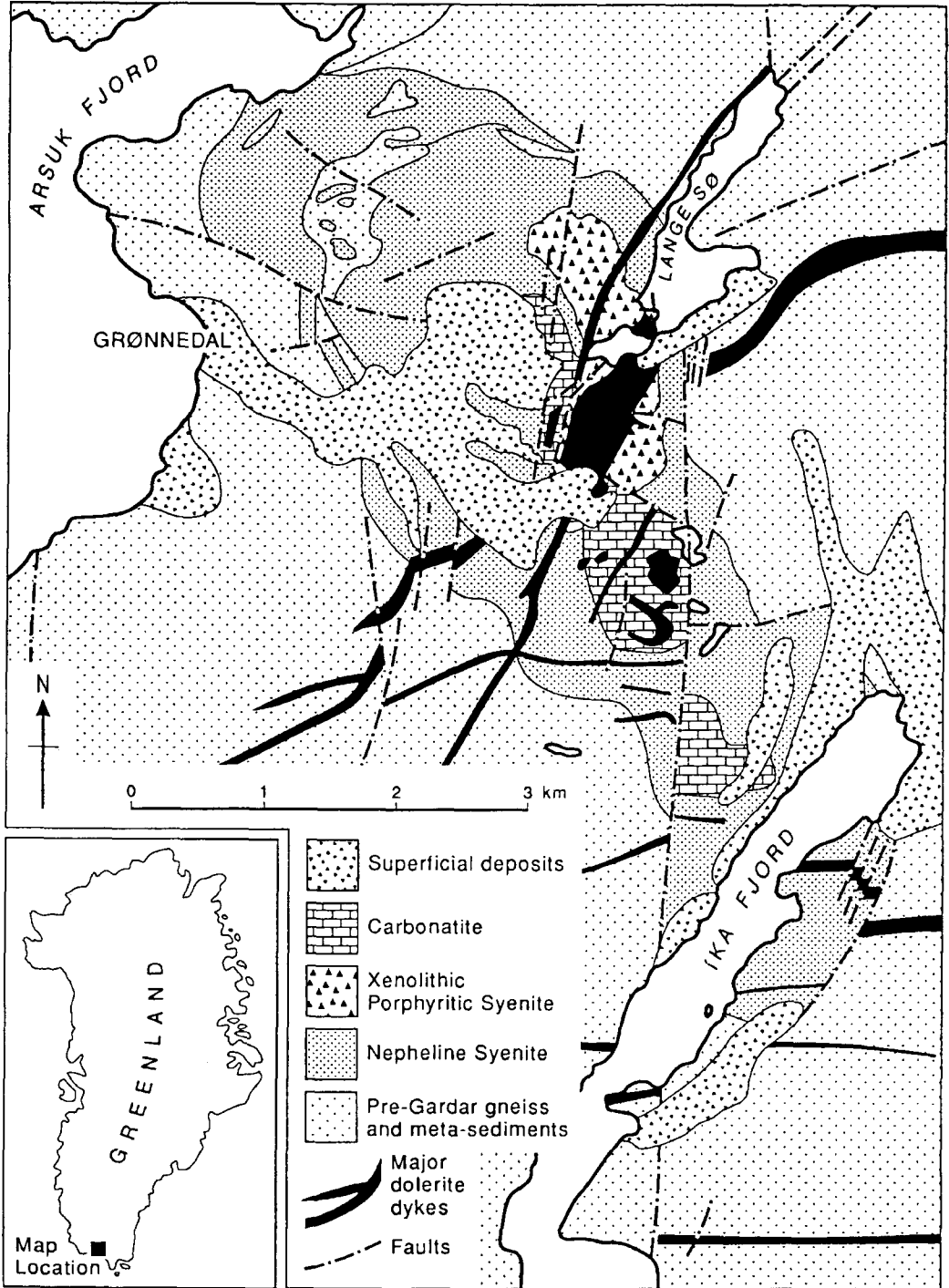


FIG. 1. Geological sketch map of the Grønnedal-Íka nepheline syenite complex based on Emeleus (1964).

the carbonatites. He ascribed the presence of the carbonatite within the syenites to the evident crustal weakness in the area, where several major faults intersect, a feature which could have assisted emplacement of the carbonatite. In addition he tentatively suggested that the carbonatite may have been genetically unrelated to the syenites, based on the then widely recognised association between carbonatite and feldspar-poor or feldspar-free silicates. Bedford (1989) re-described the complex, producing a large number of new analyses, including *REE* data. He concluded from the partitioning of *REE* and other trace elements, notably Hf and Ta, that the carbonatite formed by liquid immiscibility from a nepheline syenite magma, the late-stage Xenolithic Porphyritic Syenite being the conjugate silicate liquid (see Fig. 1). This model is similar to that proposed by Pearce (1988) for the origins of carbonatite from the predominantly phonolitic Igaliko Dyke Swarm.

In a study of rocks from the Igaliko Dyke Swarm, compositionally similar to Grønndal-Åka, Pearce and Leng (1996) described the isotopic and chemical evolution of carbonatites and related rocks. From similarities in C and O isotopes, they identified a close spatial, chemical and isotopic relationship between the carbonatite and aillikites (carbonate-bearing ultramafic lamprophyre *sensu* Rock, 1987; silicocarbonatites *sensu* Woolley *et al.*, 1996) dykes of Igaliko, either by liquid immiscibility, or a result of increasing degrees of partial melting from a fertile CO<sub>2</sub>-rich mantle source. However, this contradicted earlier geochemical evidence, particularly from *REE* concentrations, which had strongly suggested an immiscible relationship between carbonatites and phonolites of the Igaliko Dyke Swarm (see Pearce, 1988; Hamilton *et al.*, 1989).

Here we present a brief summary of the geological evolution of the Grønndal-Åka complex along with new stable and radiogenic isotope data, which are used to interpret the origins of the carbonatite magma. These data are also compared to the isotopic data from rocks of similar composition from the Igaliko Dyke Swarm.

### Geological setting and evolution

The Grønndal-Åka intrusion (1299 ± 17 Ma; Blaxland *et al.*, 1978) consists predominantly of foyaitic nepheline syenites and carbonatites cutting early Proterozoic gneisses and metamorphosed sediments (Emeleus, 1964; Bedford, 1989). Marginal syenites show complex field relations and grade inwards to laminated Upper and Lower Series nepheline syenites (abbreviated to USS and LSS in figures and tables) containing cumulus nepheline, alkali feldspar and, less commonly, aegirine-augite,

with variable quantities of intercumulus biotite, apatite and opaque oxides. Crystal lamination in these units is often slumped, a result of magmatic currents and not compaction of the cumulus pile (Bedford, 1989). Bodies of mafic and porphyritic microsyenites (MMS, PMS) and syenitic ring dykes (GSA, GSB) intrude the main layered series. Postcumulus growth on pyroxenes extends compositions from salitic cores in some of the Marginal Syenites to almost pure acmite whilst zoning in the layered rocks of the Upper and Lower Series is less extreme from about Ac<sub>50</sub> to Ac<sub>90</sub>.

The Grønndal-Åka complex shows some striking mineralogical differences when compared with other Gardar centres, being notably poor in both olivine and amphibole. These features indicate a magma with high *f*<sub>O<sub>2</sub></sub> and low *P*<sub>H<sub>2</sub>O</sub>. Additionally, the onset of enrichment of Na and Fe<sup>3+</sup> in pyroxenes at relatively high Mg contents also indicates a relatively oxidising magma. Fractionation of these magmas was controlled by the crystallisation of alkali feldspar, nepheline and pyroxene, with minor phases (notably apatite and zircon) exerting an important control on the *REE* concentrations. The occurrence of zircon in all units of this syenite is unusual, and may be a result of late or post-magmatic processes, which may also have been responsible for much of the observed variation in trace elements (Bedford, 1989). There is a gradual increase in the CO<sub>2</sub> content of the syenitic magmas expressed in the increasing abundance of cancrinite with time.

At a late stage in the evolution of the complex, a plug of xenolithic porphyritic syenite (XPS) was forcefully intruded, and this, in turn, was followed by the emplacement of a central plug of calcio-carbonatite which had entrained fragments of the earlier nepheline syenites. The carbonatite consists predominantly of calcite, with siderite, magnetite, apatite, sphalerite, pyrite and Sr-rich barite (Bedford, 1989; Emeleus, 1964). Emeleus (1964) notes that the carbonatite becomes progressively more sideritic towards its centre.

Patchy, late-stage metasomatism, has affected parts of all units. Rocks immediately adjacent to the carbonatite are highly altered, and the occurrence of zircon, alkali ferromagnesian minerals, sodalite veins and recrystallisation in feldspars has been attributed to the effects of peralkaline, Cl and CO<sub>2</sub>-rich fluids emanating from both the carbonatite and syenites. G238 (USS) shows signs of this metasomatic alteration with clouded alkali feldspar and late-stage veins of carbonate.

Subsequent to the emplacement of the carbonatite, the complex was dissected by several phases of faulting and intruded by a variety of dykes. Emeleus (1964) reconstructed the initial form of the intrusion, to show an oval body comprising multiple nested

intrusions younging towards the centre. Faulting and dyke emplacement overlapped. Dykes include a series of lamprophyres and an alnöite which have been described by Emeleus (1964); some were analysed by Gill (1972). These include kersantite (GGU27049, GGU27109), vogesite (GGU27021), spessartite (GGU27061), camptonite (GGU31814) and a dyke of alnöitic affinities (GGU31838). Samples prefixed 'G' were collected by CMB and CHE in July and August 1987, the lamprophyres and alnöite were collected by CHE in the late 1950s and early 1960s, their sample numbers referring to the collections of the Geological Survey of Greenland (GGU).

### Analytical methods

Major and trace elements in nepheline syenites and carbonatites were analysed using a Philips PW1400 XRF in the Department of Geological Sciences in the University of Durham. Major elements were determined from samples dissolved in lithium borate-metaborate fused glass discs (Norrish and Hutton, 1969), calibrated against certified reference materials. Trace elements in silicate rocks were determined from pressed powder briquettes, calibrated against certified reference materials, and corrected for mass absorption coefficients calculated from the major element composition (see Pearce, 1988). To overcome the problems of extreme trace element concentrations expected in the Grønneal-Åka carbonatites, and to match matrices, trace elements in the carbonatites were calibrated against carbonatites analysed by Pearce (1988) following methods developed by Hodgson (1985).

REE, U, Th, Cr, Sc, Cs, W, Hf and Ta were determined in nepheline syenites and carbonatites by Instrumental Neutron Activation Analysis at the Department of Geological Sciences, University of Durham, following the methods of Henderson and Williams (1981), Potts *et al.* (1981) and Potts (1987). Two lamprophyres (GGU27021 and GGU31814) and the alnöite (GGU31838) were analysed for selected trace elements by ICP-MS at the Institute of Earth Studies, University of Wales, Aberystwyth, using methods described in Pearce *et al.*, 1997. Analyses of all major and trace elements were monitored against reference materials or previously analysed samples and gave acceptable results. Samples were selected for isotopic analysis based on their geochemistry and petrography. Samples were chosen to show no signs of metasomatism (with one exception – G238) and to cover the range of rock types within the intrusion.

Isotopic analysis was performed at NIGL, Keyworth. Strontium was separated by conventional ion-exchange techniques, and Sr isotope ratios were measured on a VG 354 mass spectrometer. Analytical

uncertainties are estimated at 0.005% for  $^{87}\text{Sr}/^{86}\text{Sr}$  and 0.5% for  $^{87}\text{Rb}/^{86}\text{Sr}$ . Accuracy was checked using the NBS 987 Sr isotope standard and was within acceptable limits. For C and O isotope analysis of the carbonate, samples were reacted with anhydrous phosphoric acid *in vacuo* overnight, at a constant 25°C, using essentially the method of McCrea (1950). The clean CO<sub>2</sub> thus liberated was separated from water vapour and collected for analysis. Measurements were made on a VG Optima mass spectrometer. The data were corrected for isobaric effects using the method of Craig (1957) and the results expressed as  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  per mil (‰) relative to PDB and SMOW. Measured  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values were normalised through the laboratory standards. Overall analytical reproducibility was better than 0.02‰ (2 sigma).

### Geochemistry of carbonatite and related rocks

It is not intended here to discuss the overall chemical evolution of the Grønneal-Åka complex, already fully described by Gill (1972) and Bedford (1989), but only to highlight aspects particularly relevant to the origins of the carbonatite.

Chemically the carbonatites all classify as sövites or ferro-carbonatites adopting the scheme of Woolley (1982) and are illustrated in Fig. 2. The Grønneal-Åka carbonatites (Table 1) are chemically very similar to the carbonatite dykes from Igaliko (Pearce, 1988; Pearce and Leng, 1996). The carbonatites are all poor in alkalis ( $\text{Na}_2\text{O} + \text{K}_2\text{O} < 1.3$  wt.%). SiO<sub>2</sub> reaches 15.2 wt.%, and roughly correlates with Al<sub>2</sub>O<sub>3</sub> (reaching 4.91 wt.%), a reflection of the content of either or both alkali feldspar and alkali pyroxene in the carbonatites. P<sub>2</sub>O<sub>5</sub> reaches 3.64 wt.%, reflecting the high apatite content of some samples. MnO reaches 2.93 wt.% in whole rocks, present in solid solution in both calcite (between 0.44 and 2.48 wt %), and in siderite (between 1.74 and 4.87 wt.%). In the carbonatite calcites, MnO often exceeds FeO (which reaches 1.91 wt.% FeO), even where calcite and siderite coexist. MgO only reaches 1.26 wt.% in the carbonatites, where it is present in alkali pyroxenes and in solid solution in calcites (between 0.03 to 0.23 wt.%) and in siderite (between 2.19 and 4.73 wt.%). The small increase in whole rock MgO in the carbonatites correlates very strongly with increasing Fe<sub>2</sub>O<sub>3</sub> (maximum 16.01 wt.%) as does the increase in MnO.

Of those elements generally regarded as trace components of igneous rocks, Sr reaches 32,000 ppm, Ba 20,000 ppm, Nb 894 ppm and Th 270 ppm in the carbonatites. Zr is low to not detectable in most samples, and Rb reaches only 56 ppm. REE, Y, Ni and Sr all show strong positive correlations ( $r > \sim 0.8$ ) with Fe<sub>2</sub>O<sub>3</sub> in the carbonatites.

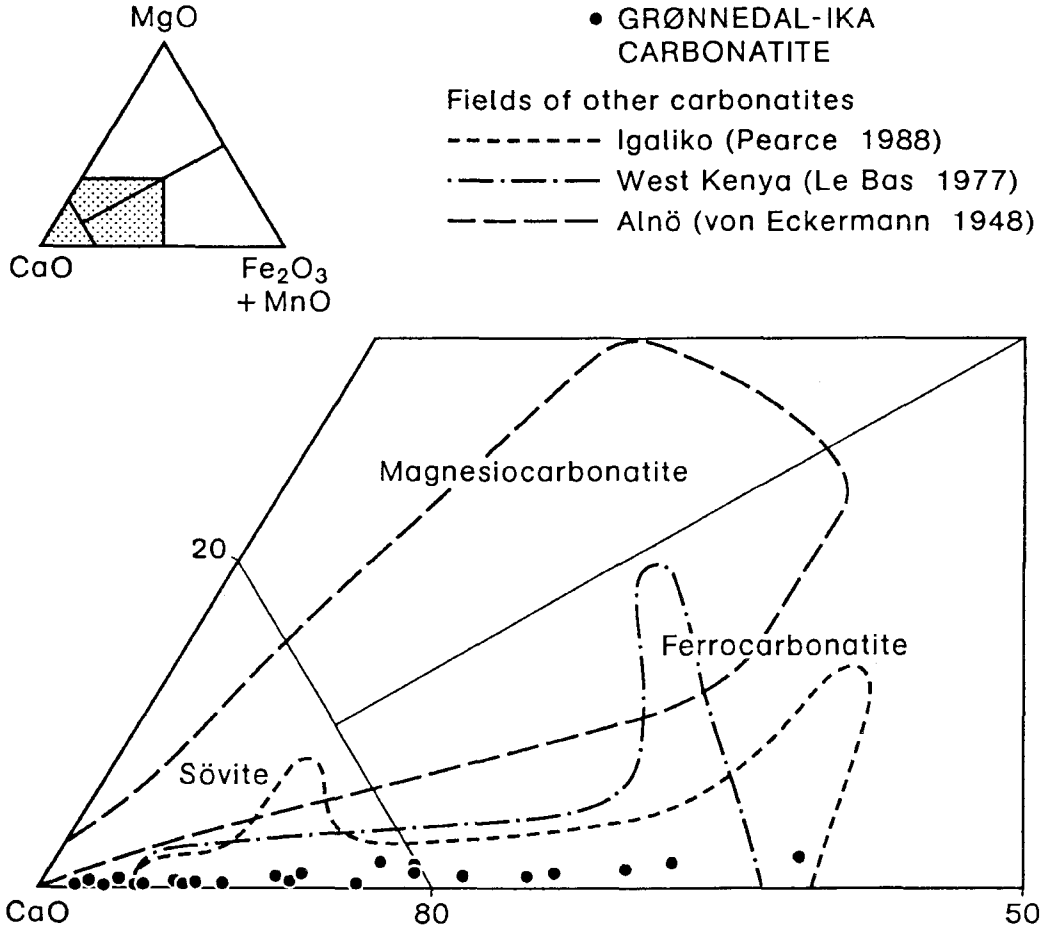


FIG. 2. Chemical classification of carbonatites from the Grønnedal-Íka complex, based on the scheme of Woolley (1982). Analyses from Bedford (1989). Fields of other carbonatites are shown for comparison.

All nepheline syenites and carbonatites show *LREE* enrichment (Fig. 3). The carbonatites and late-stage syenites (including the Xenolithic Porphyritic Syenite) show no Eu anomalies. The remaining syenitic rocks show only very small negative Eu anomalies, despite clear chemical (low K, low Sr) and petrographic evidence for feldspar fractionation driving magmatic evolution in the silicates. The Grønnedal-Íka carbonatites have similar *REE* concentrations to those from Igaliko. The *LREE* enrichment shown by the syenites is consistent with fractional crystallisation from a *LREE* enriched basic parent, probably derived by a small degree of partial melting from a fertile garnet-bearing mantle source. The lack of Eu anomalies is consistent with high  $f_{O_2}$  in the magmas. There is a distinct gap in the *REE* compositions separating the

nepheline syenites from the carbonatites, with the carbonatites showing approximately 10 times higher *REE* than the syenites. Hf and Ta, and the Hf/Ta ratio are higher in the syenites than the carbonatites (Bedford, 1989).

#### Isotope data

##### *C and O isotopes*

$\delta^{13}C$  and  $\delta^{18}O$  data for selected Grønnedal-Íka carbonatites, syenites, lamprophyres and the alnöite are given in Table 2, and presented graphically in Fig. 4. The carbonatites all have a very similar isotope composition with  $\delta^{13}C$  values between  $-4.7$  to  $-4.0$  (average  $\delta^{13}C = -4.31 \pm 0.22$ ), and  $\delta^{18}O$  covering a slightly greater range between 6.7 to 7.9

TABLE 1. Major and trace element geochemistry of selected rocks from the Grønneidal-Åka complex

Sample Number	G28	G30	G47	G55	G101	G102	G193	G219	G20	G238	G169	G172	G202a	G226	G37	G44	G145	G21	27109	27049	31838	31814
Rock Type	CBT	CBT	CBT	CBT	CBT	CBT	LSS	LSS	USS	USS	GSA	GSA	GSB	GSB	PMS	MMS	XPS	MAFIC	KERS	KERS	ALNO	CAMP
Major elements, weight %																						
SiO <sub>2</sub>	1.64	1.54	0.00	0.97	0.78	0.00	51.44	49.98	54.08	51.00	57.75	55.12	54.50	54.70	56.53	50.61	54.00	51.05	48.60	42.67	n.a.	n.a.
Al <sub>2</sub> O <sub>3</sub>	0.28	0.05	0.00	0.39	0.47	0.17	24.79	21.92	19.96	27.73	17.83	17.17	21.89	20.63	19.69	18.52	18.69	12.59	15.05	14.80		
Fe <sub>2</sub> O <sub>3</sub>	1.17	9.03	13.39	16.01	5.59	0.89	4.51	7.45	8.80	4.24	7.53	8.68	6.46	5.35	8.60	7.23	8.07	16.72	12.38	11.85		
MgO	0.22	0.57	0.58	0.75	0.29	0.00	0.07	0.36	0.71	0.19	0.26	0.78	0.57	0.24	0.20	0.42	0.40	1.92	3.50	7.63		
CaO	52.42	45.49	44.72	40.32	48.41	54.29	37.31	3.17	2.06	1.23	1.71	2.46	1.31	1.91	0.66	1.09	2.42	5.75	6.32	7.06		
Na <sub>2</sub> O	0.18	0.18	0.21	0.19	0.18	0.16	11.26	10.76	5.53	4.97	7.66	7.82	7.34	9.08	6.17	8.45	7.12	4.91	4.44	2.60		
K <sub>2</sub> O	0.29	0.00	0.00	0.22	0.22	0.00	5.37	4.53	5.59	5.48	5.10	4.50	5.17	5.59	6.88	5.39	4.84	3.72	3.62	5.31		
TiO <sub>2</sub>	0.03	0.02	0.00	0.05	0.08	0.02	0.12	0.12	0.29	0.18	0.44	0.48	0.19	0.15	0.16	0.25	0.22	0.71	2.21	2.91		
MnO	0.94	1.78	2.93	1.54	1.46	0.10	0.09	0.23	0.24	0.10	0.19	0.20	0.16	0.18	0.10	0.18	0.27	0.37	0.21	0.31		
P <sub>2</sub> O <sub>5</sub>	0.00	0.13	0.04	1.99	0.99	0.00	0.00	0.20	0.03	0.18	0.09	0.20	0.18	0.13	0.06	0.11	0.35	0.70	1.40	0.15		
LOI	41.97	40.34	36.56	35.39	38.57	43.02	2.15	1.45	2.81	4.77	1.57	2.70	2.35	2.16	1.06	8.10	3.77	1.64	4.08	6.79		
Total	99.14	99.13	98.43	97.82	97.04	98.65	100.17	100.17	100.10	100.07	100.13	100.11	100.12	100.12	100.11	100.35	100.15	100.08	101.81	102.08		
Trace elements, ppm																						
Method(s)	XRF	INAA	XRF	XRF	INAA	XRF	INAA	XRF	XRF	INAA	XRF	INAA	XRF	XRF	XRF	XRF	INAA	XRF	XRF	ICP-MS	ICP-MS	ICP-MS
Li																				172	31.8	59.3
Be																				16.3	0.82	2.39
Sc																				19	20.6	21.2
V	0	0	0	0	15	0	6	5	3	5	15	21	4	6	7	5	16	20	129	178	170	170
Cr	1	5	4	5	0	6	6	6	5	6	7	7	6	7	6	5	7	8	225	86.5	313	313
Co																			38.2	57.5	51.5	51.5
Ni	12	21	21	20	12	13	9	8	9	8	12	11	9	9	8	10	13	12	108	121	212	212
Cu	0	0	0	0	0	0	18	18	18	21	20	21	17	18	18	18	7	20	43	47.7	175	175
Zn	326	143	69	899	101	2029	90	107	134	150	172	114	111	111	77	19	201	225	385	168	198	198
Ga	0	2	0	1	0	0	45	45	40	37	36	31	37	28	38	41	40	27	27	14.1	23	23
Rb	46	41	34	32	34	28	165	132	171	159	166	190	197	150	162	202	133	141	183	30.1	60.2	60.2
Sr	7174	6114	15277	21680	16727	13861	238	196	753	1296	342	944	572	534	406	150	1253	297	1105	1650	1340	1602
Y	157	374	372	393	235	240	18	17	22	19	71	48	9	29	0	4	19	44	65	27	27	27.9
Zr	0	0	0	0	0	0	684	497	749	387	960	1142	385	541	189	444	1262	1010	1643	98.3	176	176
Nb	37	161	23	41	29	23	138	54	179	75	203	279	105	118	89	109	223	153	317	17.6	48	48
Cs		3.95					2.36	1.49	1.36		2.33	2.9				2.98			2.86	1.55	4.2	4.2
Ba	617	166	123	778	19778	2435	153	151	370	368	366	379	335	303	142	214	689	297	1093	643	1507	1507
La	950	882	954	977	1152	214	29.9	48	51	40	68	101	23.2	67	14	37	41.9	109	68	119	28.2	61.1
Ce	1850	2852	2108	1945	2176	753	74	135	100	74	170	190	61.7	108	23	61	109	207	216	57.1	126	126
Pr																			27.3	7.8	7.8	16.8

TABLE 1. (contd.)

Sample Number	G28	G30	G47	G55	G101	G102	G193	G219	G20	G238	G169	G172	G202a	G236	G37	G44	G145	G21	27109	27049	31838	31814
Rock Type	CBT	CBT	CBT	CBT	CBT	CBT	LSS	LSS	USS	USS	GSA	GSA	GSB	GSB	PMS	MMS	XPS	MAFIC	KERS	KERS	ALNO	CAMP
Nd	1126	2129	1827	1371	1305	836	30.4	69.7	58	46	87	98	35.8	73	30	44	55	129	101	101	31.8	69
Sm		248.7				117.4	4.8	10.6			12.5		5.4				9.56	17.3	7.03	7.03	13	
Eu		66.8				35	1.13	1.86			3.1		1.3				2.52		4.56	4.56	2.43	4.08
Gd		40.8				12.7	5.37	1.76			1.98		0.72				3.77		16	16	6.64	11
Tb		27.6				18.7	0.65	1.14			1.72		0.5				0.92		2.18	2.18	0.98	1.34
Dy																			11.1	11.1	5.73	6.66
Ho		103.9				36.7	0.68	1.46			5.4		0.43				1.71		2.01	2.01	0.99	1.04
Er																			4.88	4.88	2.96	2.53
Tm																			0.63	0.63	0.39	0.32
Yb		10.3					6.11	2.29			5.7		0.9				2.61		2.87	2.87	2.33	1.69
Lu		0.84					0.5	0.4			0.82		0.2				0.53		0.2	0.2	0.19	0.13
Hf		4.6					0.34	17.2			22.1		11.1				26		16.1	16.1	2.61	3.98
Ta		6.17					8.6	7.9			12.2		4.4				18.5		27.1	27.1	0.35	2.27
Pb	2	2	2	2	3	2	12	10	13	11	14	14	12	12	11	12	3	12	16	16	6.68	7.46
Th	66	61	45	205	270	65	9	9	10	9	10	11	8	10	8	8	3	10	18.9	18.9	0.75	4.57
U	11	9	16	20	17	15	2	1	2	4	3	4	2	3	2	2	9	2	5.8	5.8	0.53	1.12

Abbreviations used CBT - carbonatite; LSS - Lower Series Syenite; USS - Upper Series Syenite; GSA - GSA Ring Dyke; GSB - GSB Ring Dyke; PMS - Porphyritic Microsyenite; MMS - Mafic Microsyenite; XPS - Xenolithic Porphyritic Syenite; MAFIC - Mafic Syenite; KERS - Kersantite; Voges - Vogesite; ALNO - Alnöite; Spess - Spessartite; CAMP - Camptonite.

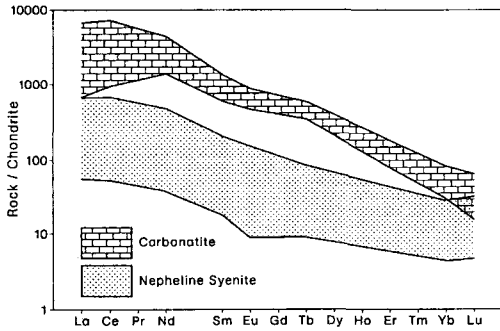


FIG. 3. Chondrite normalised REE concentrations for carbonatites and nepheline syenites from the Grønnedal-Íka intrusion. Note the general lack of Eu anomalies in both carbonatites and syenites. Data from Bedford (1989). Normalisation factors from Boynton (1984).

between 9.1 and 12.3 (average  $\delta^{18}\text{O} = 10.30 \pm 0.63$ ). This field overlaps part of the syenite field. The rock of alnöitic affinities has the highest  $\delta^{13}\text{C}$  of samples analysed from the Grønnedal-Íka area ( $-3.32$ ) and higher  $\delta^{18}\text{O}$  (12.34) than the lamprophyres and carbonatites.

The syenites (with one exception) form a coherent group, with  $\delta^{13}\text{C}$  in the range  $-7.6$  to  $-3.9$  and  $\delta^{18}\text{O}$  between 8.7 and 15.1, with a strong positive correlation between  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  (see below). The exception, G238, lying away from the main trend is a metasomatised Upper Series Syenite containing late-stage veins of carbonate and has an isotope composition close to the carbonatites. The petrography of this sample shows it has clearly been invaded by fluids derived from the late carbonatite plug.

The C and O isotope compositions of the lamprophyres and the alnöite overlap the syenites, again forming a group isotopically distinct from the carbonatites, with no obvious continuum.

(average  $\delta^{18}\text{O} = 7.18 \pm 0.41$ ). These data define a linear array with a positive correlation (see below). The five lamprophyres also form a relatively coherent group, ranging in  $\delta^{13}\text{C}$  between  $-7.9$  and  $-3.3$  (average  $\delta^{13}\text{C} = -5.34 \pm 1.44$ ) and in  $\delta^{18}\text{O}$

*Sr isotopes*

The Sr isotope data from representatives of the carbonatites, syenites, lamprophyres and the alnöitic

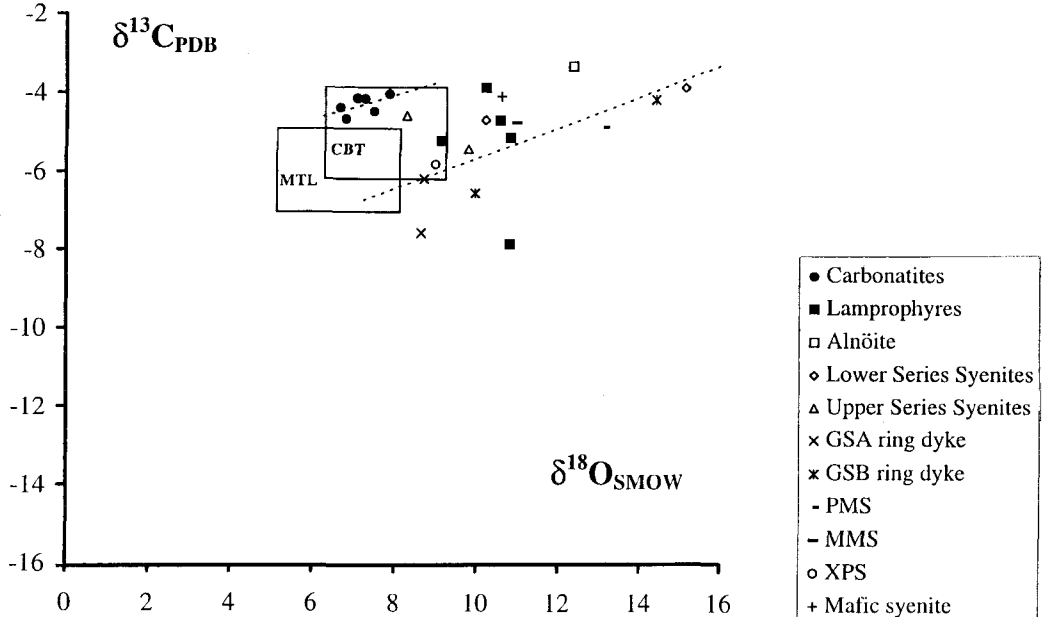


FIG. 4. C and O isotope data showing relationships between the carbonatite and associated rocks from Grønnedal-Íka. 'MTL' is the box for mantle-sourced silicate magmas (Des Marias and Moore, 1984; Kyser *et al.*, 1982). 'CBT' box encloses the expected C and O isotopic composition for mantle-derived carbonatite magma (Deines, 1989; Clarke *et al.*, 1993). Dashed lines are best fits through the nepheline syenites and the carbonatites. Abbreviations in key as Table 1.



dyke are given in Table 2. The Rb-Sr system for the carbonatites does not define an isochron. This has been noted in Rb-Sr systems for several carbonatites (cf. Pearce and Leng, 1996) and is ascribed to differences in  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios, that are not correlated with the  $^{87}\text{Rb}/^{86}\text{Sr}$  ratios, over a limited spread of Rb/Sr (Andersen, 1987).

Initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (Sr(i)) have been recalculated to 1300 Ma, and for most carbonatites are between 0.7029 and 0.7031 (average =  $0.7030 \pm 0.0001$ ), with the exception of G30 which has Sr(i) of 0.7307. The syenites have similar Sr(i) to the carbonatites and spread from 0.7020 to 0.7039 (average =  $0.7032 \pm 0.0008$ ). The lamprophyres have slightly lower Sr(i) than both the carbonatites and the syenites, ranging between 0.7026–0.7029. The alnöite has a low Sr(i) = 0.7024.

### Discussion

Sr(i) from the majority of carbonatites and syenites are uniform at  $0.7032 \pm 0.0006$ . The data show that the carbonatite and syenite magmas either belong to a single group, or were generated from a similar source, characterised by a broadly bulk earth Sr isotope composition at the time of melt generation, giving an average time-corrected  $\epsilon\text{Sr}$  of about 2.5. The lamprophyres and alnöite, which have slightly lower Sr(i) than the averages for carbonatites and syenites, may have been generated from the same or a slightly depleted source.

Emeleus (1964) has postulated that during evolution of the nepheline syenites by fractional crystallisation, a gradual build up of carbonate, as evidenced by increasing primary cancrinite, led to the late-stage forceful emplacement of a plug of residual carbonatite at the centre of the intrusion. Within the carbonatite further evolution was marked by an increase inward of Fe, a result of calcite fractionation leaving a sideritic residue. This also caused increases in MgO, MnO, Sr, Y, REE towards the centre of the carbonatite mass. Bedford (1989), in a detailed geochemical study of the Grønndal-Íka complex, suggested an origin for the carbonatites by liquid immiscibility from a  $\text{CO}_2$ -rich nepheline syenite, leaving the Xenolithic Porphyritic Syenite unit as the conjugate silicate liquid. This origin is consistent with experimental data from Kjarsgaard and Hamilton (1988, 1989) where, at low temperatures, relatively alkali poor carbonatites can be produced by immiscibility from nepheline syenite (phonolitic) melts. Bedford (1989) suggested immiscibility occurred at temperatures well below  $1100^\circ\text{C}$  and pressures of about 0.5 GPa. REE and trace element partitioning experiments between immiscible carbonatite and phonolite melts (Hamilton *et al.*, 1989) are consistent with the observed distributions of these elements. This enrichment in REE could also be

a result of extreme evolution of the nepheline syenites also producing the build-up of  $\text{CO}_2$  simply by fractional crystallisation (cf. Emeleus, 1964). However, the higher Hf and Ta contents, and the higher Hf/Ta ratio, in the syenites are not consistent with a fractionation model for the origin of the carbonatite, but are similar to concentrations predicted in immiscibility experiments (Hamilton *et al.*, 1989). The immiscibility model proposed by Bedford (1989) for the origin of the Grønndal-Íka carbonatites is similar to that proposed by Pearce (1988) for the carbonatites of the Igaliko Dyke Swarm, by immiscibility from  $\text{CO}_2$ -rich phonolites. Recently however isotopic studies by Pearce and Leng (1996), have shown the Igaliko carbonatites to be unrelated to phonolite magmas, but to have much closer affinities with the carbonate-bearing ultramafic lamprophyres (*sensu* Rock, 1987, *viz.* aillikites), with field and geochemical evidence suggesting an origin involving liquid immiscibility.

The  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  isotope composition of the Grønndal-Íka carbonatites are similar to many other magmatic carbonatites, sitting in the upper left corner of the 'carbonatite box' which encompasses expected compositions for mantle derived carbonatites (Deines, 1989; Clarke *et al.*, 1993). A linear correlation ( $r = 0.576$ ) exists between the  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$ , having a slope of about 0.3. This trend parallels the trend displayed in C and O isotopes of many other carbonatite complexes which show slopes between 0.2 and 0.4 and appears to reflect a fundamental process in carbonatite genesis (Deines, 1989), related to the fractionation of a carbonate mineral (Deines, 1970).

A good correlation also exists between  $\text{Fe}_2\text{O}_3$  and  $\delta^{18}\text{O}$  in the six Grønndal-Íka carbonatites analysed ( $r = -0.648$ , see Fig. 5), and in those carbonatites with  $>5$  wt.%  $\text{Fe}_2\text{O}_3$ , the correlation between increasing Fe content and decreasing  $\delta^{18}\text{O}$  is almost perfect ( $r = -0.98$ ). The decrease in  $\delta^{18}\text{O}$  appears to be the result of calcite fractionation within the carbonatite producing a more Fe-rich,  $^{16}\text{O}$ -rich residue. Whilst the correlation between  $\delta^{13}\text{C}$  and Fe content is not as good as for  $\delta^{18}\text{O}$ , a strong negative correlation is evident for the four most Fe-rich carbonatites ( $r = -0.784$ ), indicating an increase in  $^{12}\text{C}$  in residual liquids during calcite fractionation. The decrease observed in  $\delta^{18}\text{O}$  with fractionation is unusual. Nielsen and Buchardt (1985) and Reid and Cooper (1992) both report increasing  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  with fractionation in carbonatites. Knudsen and Buchardt (1991) however describe decreasing  $\delta^{13}\text{C}$  during evolution of a carbonatite with no systematic variation in  $\delta^{18}\text{O}$ .

Whilst most of the carbonatites have Sr(i) about 0.703, one (G30) has an unusually high initial ratio of 0.7307. In common with the other carbonatites, G30

Table 2. Isotope data from Grønmedal-fika

	$\delta^{13}\text{C}_{\text{PDB}}$	$\delta^{18}\text{O}_{\text{SMOW}}$	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}(t)$	$\epsilon\text{Sr}(t)$
Carbonatite	G28	-4.47	7.47	46	0.0185	0.70342	0.70308	2
Carbonatite	G30	-4.15	7.25	41	0.0194	0.73110	0.73074	396
Carbonatite	G47	-4.66	6.79	34	0.0062	0.70317	0.70305	2
Carbonatite	G55	-4.38	6.65	32	0.0043	0.70303	0.70295	0
Carbonatite	G101	-4.03	7.85					
Carbonatite	G102	-4.14	7.06	28	0.0058	0.70304	0.70293	
Vogesite	27021	-5.12	10.81					
Kersantite	27049	-5.20	9.12	183	0.3209	0.70888	0.70290	-1
Spessartite	27061	-3.86	10.21					
Kersantite	27109	-4.68	10.55					
Camptonite	31814	-7.86	10.81	60.2	0.1087	0.70465	0.70262	-4
Alnoite	31838	-3.32	12.34	30.1	0.065	0.70363	0.70242	-7
Lower Series Syenites	G193	-3.86	15.12					
Lower Series Syenites	G219	-4.67	10.20					
Upper Series Syenites	G20	-5.41	9.78					
Upper Series Syenites	G238	-4.57	8.27	159	0.355	0.71029	0.70368	11
GSA Ring Dyke	G169	-7.57	8.64	166	1.3071	0.72773	0.70289	-1
GSA Ring Dyke	G172	-6.18	8.71					
GSB Ring Dyke	G202a	-4.16	14.38					
GSB Ring Dyke	G226	-6.55	9.96	150	0.8135	0.71719	0.70203	-13
Porphoritic Microsyenite	G37	-4.84	13.10	162	1.1564	0.72503	0.70349	8
Mafic Microsyenite	G44	-4.73	10.96					
Xenolithic Porphyritic Syenite	G145	-5.81	8.99	133	0.3072	0.70965	0.70393	14
Mafic Syenite	G21	-4.08	10.59					

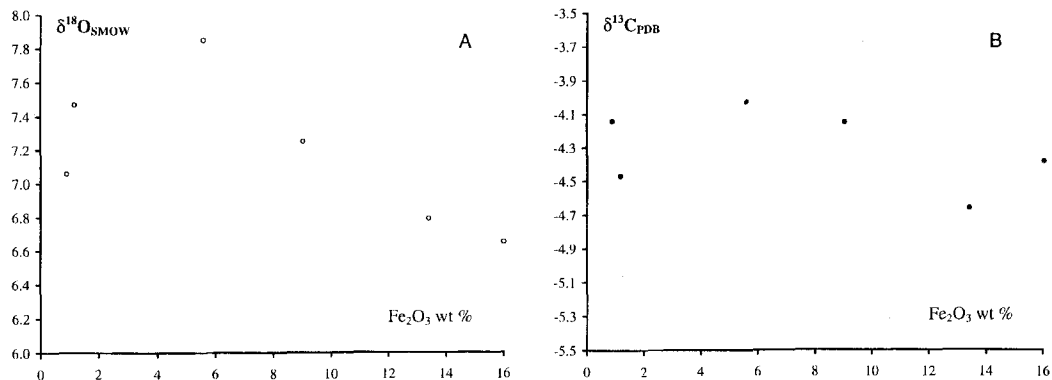


FIG. 5. (A)  $\delta^{18}\text{O}$  vs.  $\text{Fe}_2\text{O}_3$  wt.% and (B)  $\delta^{13}\text{C}$  vs.  $\text{Fe}_2\text{O}_3$  wt.% in carbonatites from Grønnedal-Íka.

also has high Sr content (see Table 1). This high Sr content would easily mask any contamination by an isotopically evolved, but Sr-poor basement, although the contaminant material would have to have a very high Sr content, with exceptionally high  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio to produce Sr(i) of 0.7307. Thus, in the case of G30, which in all other respects appears to be normal carbonatite, contamination seems an unlikely cause of the high  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio as C and O isotopes are unaffected. There is a possibility that preferential enrichment of radiogenic  $^{87}\text{Sr}$  may have occurred (Blaxland *et al.*, 1975), producing the elevated  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio in this sample. Signs of contamination can only be confidently shown in samples with relatively low Sr concentrations, namely the syenites, which here show Sr(i) typically between 0.702 and 0.704, and appear to be largely uncontaminated. Many of the other Si-undersaturated Gardar centres show mantle type Sr isotopic ratios (Sr(i) = 0.702 to 0.704) whereas the Si-oversaturated complexes generally record higher Sr(i) (typically >0.707), probably the result of some crustal assimilation (Blaxland *et al.*, 1978). The character of any crustal rocks with which the Grønnedal-Íka carbonatite may have interacted is unknown. Locally, the country rocks are Ketilidian gneisses and meta-sediments, although the boundary of the Archaean craton is only some 5 km to the north west of the intrusion. Pearce and Leng (1996) have discussed some of the possible sources of isotopically evolved Archaean and Proterozoic crustal material in relation to the contamination seen in the Igaliko Dyke Swarm. Similar material may be present in the Grønnedal-Íka area.

In the syenites, the possibility of loss of radiogenic Sr by weathering, or post magmatic Rb-metasomatism, may account for the low Sr(i) reported in G266 (0.70203). The O isotope composition of this

sample however does not indicate weathering, which would produce high  $\delta^{18}\text{O}$ , this appearing to be one of the most isotopically primitive of syenites.

$\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  in the syenitic rocks form a linear trend ( $r = 0.757$ ) with a slope of 0.39, (excluding G238) with the most isotopically primitive samples close to the 'mantle box' (Des Marias and Moore, 1984; Kyser *et al.*, 1982), indicative of a mantle source for the carbonate in these rocks. This is sub-parallel to the trend shown by the C and O isotope evolution in the carbonatites.

The carbonate in both the syenites and lamprophyre dykes from Grønnedal-Íka show a range of  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values, extending to heavy O values at light C. This can be explained by either, or a combination, of two processes, viz. (i) as a result of having had a protracted period of crustal residence, during which C and O exchanged with the country rocks or with surface derived, low-temperature waters, or (ii) oxygen isotope fractionation accompanying fractional crystallisation of feldspars, feldspathoids, clinopyroxenes and oxides during magmatic evolution. The second process concurs with Sr isotope data (see above and below). In addition, no evidence of contamination by basement-derived material can be recognised from the petrography or geochemistry of these samples. The majority of the carbonatites, lamprophyres and syenitic rocks from Grønnedal-Íka also show no C and O isotope evidence of contamination, which would produce and increased  $\delta^{18}\text{O}$  with a wide scattering of  $\delta^{13}\text{C}$  (evident in phonolite dykes from the Igaliko Dyke Swarm, Pearce and Leng, 1996, see below). The trend in C and O isotopes in syenites from Grønnedal-Íka is thus interpreted as a fractional crystallisation trend.

Whilst the trace element geochemistry of the Grønnedal-Íka nepheline syenites and carbonatites is

consistent with a petrogenetic model involving liquid immiscibility (see above), the difference in the  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  isotopes between these groups strongly contradicts this. The carbonatites form an isotopically distinct group, markedly separated from both the nepheline syenites and the lamprophyres/almöite, suggesting that these groups are genetically unrelated. Only one sample (G238) has C and O isotopes which plot between the carbonatite and syenite fields. This sample shows clear petrographic evidence (veins) of secondary carbonate, presumably precipitated from late-stage metasomatic fluids emanating from the carbonatite (see above). The remaining syenites appear unaffected (both petrographically and isotopically) by metasomatism.

Field evidence suggests a relatively small time span between the emplacement of the syenites and the forceful emplacement of the carbonatite (Emeleus, 1964). The oldest age of the carbonatite is constrained by the nepheline syenites at  $1299 \pm 17$  Ma, and the youngest possible age by the emplacement of the Mid-Gardar 'brown dykes' ('dolerite' in Fig. 1, see Berthelsen and Henriksen, 1975), which are older than the near-by Kûngnât complex at  $1219 \pm 16$  Ma (Patchett *et al.*, 1978). In the light of C and O isotope evidence, the close association of carbonatite with nepheline syenite at Grønnedal-Íka may be coincidental. Elsewhere in the Gardar province carbonatite is commonly associated with  $\text{CO}_2$ -rich ultramafic lamprophyres (Stewart, 1970; Craven, 1985; Upton and Fitton, 1985; Upton and Emeleus, 1987; Pearce and Leng, 1996), although globally nepheline syenite-carbonatite associations are not uncommon (e.g. Nemegosenda Lake, Ontario, Woolley, 1987). None of the lamprophyre samples, nor the almöite, analysed here have stable isotope compositions which overlap the carbonatites. These data are insufficient to show or refute any postulated relationship. However, swarms of 'lamprophyre dykes' were emplaced locally early in the Mid-Gardar, including a suite of biotite-rich ultramafic lamprophyres in the vicinity of Kûngnât and Ivigut (Berthelsen and Henriksen, 1975), and were post-dated by the more abundant olivine dolerite 'brown dykes' (Upton and Emeleus, 1987). The possibility exists, then, that the carbonatite at Grønnedal-Íka is not related to the nepheline syenites, but may be associated with early Mid-Gardar carbonate-rich ultramafic lamprophyre magmatism in the area, an association common elsewhere in the Province.

#### Comparison with similar rocks from the Igaliko Dyke Swarm.

There are several notable differences between the isotope compositions and geochemistry of carbonatites and phonolites from the Igaliko Dyke Swarm

and carbonatites and nepheline syenites from Grønnedal-Íka. Sr(i) values from uncontaminated samples (carbonatites and phonolites/syenites) from both suites are typically around 0.703, indicative of a depleted mantle source. Some phonolites from Igaliko show Sr(i) as high as 0.711, indicative of some crustal contamination, a feature not recorded from the Grønnedal-Íka syenites.

Figure 6 shows the relationship between the C and O isotope ratios for carbonatites, ultramafic and alkaline lamprophyres (*sensu* Rock, 1987) and phonolites from the Igaliko Dyke Swarm (Pearce and Leng, 1996), and carbonatites, nepheline syenites, lamprophyres and almöite from the Grønnedal-Íka complex. Chemically, the Grønnedal-Íka carbonatites are notably poorer in F and have higher sulphate contents (reflected in the relatively common occurrence of barite) than the Igaliko rocks. In contrast to the tight grouping of the Grønnedal-Íka carbonatites, the Igaliko carbonatites all have lower  $\delta^{13}\text{C}$  (ranging between  $-4.9$  and  $-7.9$ ), and all but three samples are unusual in showing extremely low values in  $\delta^{18}\text{O}$  (between 2.5 to 6). This is much more depleted in  $^{18}\text{O}$  than is typical for carbonatites (Deines, 1989), whilst the Grønnedal-Íka carbonatites are more typical of other carbonatite complexes.

The two Igaliko carbonatites with the lowest  $\delta^{13}\text{C}$  values are unusually rich in F. GGU326395 has  $\delta^{13}\text{C} = -7.9$ ,  $\delta^{18}\text{O} = 5.3$  and  $F = 13.7$  wt.%, GGU325910 has  $\delta^{13}\text{C} = -7.2$ ,  $\delta^{18}\text{O} = 5.0$  and  $F = 9.45$  wt.%. Two samples taken from a single Igaliko dyke

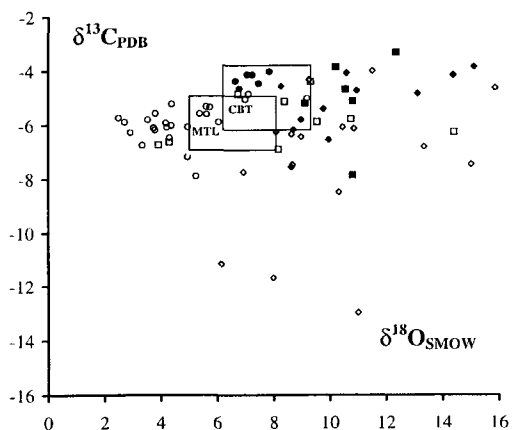


FIG. 6. C and O isotope data showing relationships between rocks from the Grønnedal-Íka Complex (filled symbols) and Igaliko Dyke Swarm (open symbols). Circles = carbonatites; squares = lamprophyres and almöite; diamonds = nepheline syenites/phonolites.

(GGU325908 and GGU325910) shows a marked variation in F with increasing height along strike. The stratigraphically lower sample, GGU325908 (with  $\delta^{13}\text{C} = -5.6$ ,  $\delta^{18}\text{O} = 3.8$  and  $F = 0.47$  wt.%), grades upwards into GGU325910 (data above) with gradually increasing fluorite and decreasing calcite. In this dyke  $\delta^{18}\text{O}$  increases with increasing F, probably a result of calcite fractionation, and  $\delta^{13}\text{C}$  decreases markedly. High F content thus appears to retain  $^{12}\text{C}$  and  $^{18}\text{O}$  in the carbonatite melt, and is possibly the cause of the strong enrichment in light C in more evolved examples. In Grønndal-Íka, with very low F contents, calcite fractionation causes a decrease in both  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$ .

The Igaliko carbonatite with highest  $\delta^{18}\text{O}$  (GGU58254) is unusual in that it contains barite, more akin to the Grønndal-Íka samples. If the unusually high F (see above) and the high S samples from Igaliko are discounted, the Grønndal-Íka carbonatites sit at one end of a linear trend of  $^{13}\text{C}$  and  $^{18}\text{O}$  through the Igaliko carbonatite data. These 'extreme' Igaliko carbonatites however give some indication as to the direction in which isotopes evolve during fractionation. Although F is much lower in the Grønndal-Íka rocks than those from Igaliko, the emplacement of the Ivigtut granite/cryolite body at  $1222 \pm 25$  Ma (Blaxland *et al.*, 1978) is testament to high F contents in magmas generated in the vicinity of Grønndal-Íka towards the end of Mid-Gardar activity.

The Igaliko ultramafic lamprophyres (*sensu* Rock, 1987) show similar  $\delta^{13}\text{C}$  ( $-4.4$  to  $-6.9$ ) and wider range of  $\delta^{18}\text{O}$  (3.9 to 14.4) when compared to the Igaliko carbonatites with which they are closely associated (Pearce and Leng, 1996). The alkaline lamprophyres from Igaliko (mostly camptonites) have isotope ratios which range between  $-4.4$  and  $-4.9$  for  $\delta^{13}\text{C}$  and 7.4 to 11.3 for  $\delta^{18}\text{O}$ . The five Grønndal-Íka lamprophyres (both alkaline and calc-alkaline using the classification of Rock, 1987) have similar  $\delta^{18}\text{O}$  to the Igaliko alkaline lamprophyres, but have higher  $\delta^{13}\text{C}$ .

The C and O isotopic compositions of carbonate from the phonolites and oversaturated rocks from the Igaliko Dyke Swarm show no systematic relationship but scatter with a range of  $\delta^{13}\text{C}$  from  $-4.0$  to  $-14.1$  and  $\delta^{18}\text{O}$  from 6.1 to 15.9. By contrast the Grønndal-Íka syenites are more tightly grouped, occupying the top portion of the field covered by the evolved Igaliko silicate dykes. This is consistent with no crustal contamination of the Grønndal-Íka carbonatites, a process also detected in both the stable and radiogenic isotope ratios of the Igaliko silicates (Pearce and Leng, 1996). The C and O isotope compositions displayed by the Grønndal-Íka syenites probably represent the typical range of mantle derived silicate magmas from the Gardar

province, which evolved by fractional crystallisation from basic parents, but prior to any interaction with crustal rocks. Petrographic and geochemical evidence indicate that these distributions in both Igaliko and Grønndal-Íka cannot be attributed to secondary processes such as deuteric or metasomatic alteration.

One further notable difference between the two suites of samples is that the syenites at Grønndal-Íka crystallised under higher  $f_{\text{O}_2}$  which suppresses the formation of Eu anomalies during feldspar fractionation. With the exception of two cancrinite phonolites at Igaliko, all other evolved silicate rocks show depletion in Eu as a result of feldspar fractionation. The presence of aegirine in the Grønndal-Íka carbonatites may also attest to a relatively high  $f_{\text{O}_2}$  with some Fe present as  $\text{Fe}^{3+}$ . Aegirine is absent from the Igaliko carbonatites, which also contain no siderite.

### Conclusions

The carbonatites of Grønndal-Íka are have a stable isotope composition distinct from the carbonate in the nepheline syenites of the complex. Whilst field relations suggest a similar age for the syenites and the carbonatites, and trace element partitioning experiments are consistent with an immiscible relationship between these rock types, the stable isotope data shows that the carbonatites and syenites cannot be genetically related. C and O isotopes from the Grønndal-Íka carbonatite form a continuum with carbonatites from the Igaliko Dyke Swarm. These carbonatites are different ages and thus are not directly related, but similar processes, including derivation from a similar mantle source followed by broadly similar evolution, may have led to the similarity in isotopic composition. Notable differences between the two groups of carbonatites however are evident, with the Igaliko carbonatites having much higher F, which may be an important factor in governing the behaviour of  $^{13}\text{C}$  and  $^{18}\text{O}$  during magmatic evolution. High F appears to cause a lowering of  $^{13}\text{C}$  with increasing  $^{18}\text{O}$  in residual melts during calcite extraction, whilst in a low F magma, calcite crystallisation produces an increase in both  $^{13}\text{C}$  and  $^{18}\text{O}$ . The behaviour of C and O isotopes during fractionation of the Grønndal-Íka carbonatite appears different from several other carbonatite bodies (e.g. Gardiner Complex – Nielsen and Buchardt, 1985; Qaqarssuk Complex – Knudsen and Buchardt, 1991; Dicker Willem complex – Reid and Cooper, 1992).

If the Grønndal-Íka carbonatites are unrelated to the nepheline syenites, their origin needs to be explained. A common association in the Gardar province is carbonatite with ultramafic lamprophyre. Ultramafic lamprophyre dykes occur in the Ivigtut

area, emplaced at the at the onset of Mid-Gardar magmatic activity. These predate the olivine dolerite 'brown dykes' which cut the Grønndal-Íka complex and may possibly be related to the carbonatite. Unfortunately, data from this study are not sufficient to indicate whether relationship exists between the Grønndal-Íka carbonatites and any lamprophyric rocks from the area.

Once the carbonatite at Grønndal-Íka was emplaced as a late-stage plug, it evolved isotopically and chemically by fractionation of calcite. Isotopic evolution within the Grønndal-Íka carbonatite differs from normal evolutionary trends (Deines, 1989). Syenitic rocks from the Grønndal-Íka complex also show isotopic fractionation caused most probably by extraction of alkali feldspar. Neither C, O nor Sr isotope compositions show any evidence for crustal contamination of the syenites or carbonatites. One carbonatite with unusually high Sr(i) may have been enriched in radiogenic Sr by metasomatic fluids within the intrusion.

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