



Mantle heterogeneity during the formation of the North Atlantic Igneous Province: Constraints from trace element and Sr-Nd-Os-O isotope systematics of Baffin Island picrites

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[1] Sr-Nd-Os-O isotope and major and trace element data from ~62 Ma picrites from Baffin Island constrain the composition of mantle sources sampled at the inception of North Atlantic Igneous Province (NAIP) magmatism. We recognize two compositional types. Depleted (N-type) lavas have low $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.702990–0.703060) and $^{187}\text{Os}/^{188}\text{Os}_i$ (0.1220–0.1247) and high $^{143}\text{Nd}/^{144}\text{Nd}_i$ (0.512989–0.512999) and are depleted in incompatible elements relative to primitive mantle. Enriched (E-type) lavas have higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.703306–0.703851) and $^{187}\text{Os}/^{188}\text{Os}_i$ (0.1261–0.1303), lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ (0.512825–0.512906), and incompatible element concentrations similar to, or more enriched than, primitive mantle. There is also a subtle difference in oxygen isotope composition; E-type lavas are marginally lower in $\delta^{18}\text{O}_{\text{olivine}}$ value (5.16–4.84‰) than N-type lavas (5.15–5.22‰). Chemical and isotopic variations between E- and N-type lavas are inconsistent with assimilation of crust and/or subcontinental lithospheric mantle and appear to instead reflect mixing between melts derived from two distinct mantle sources. Strontium-Nd-O isotope compositions and incompatible trace element abundances of N-type lavas suggest these are largely derived from the depleted upper mantle. The $^{187}\text{Os}/^{188}\text{Os}_i$ ratios of N-type lavas can also be explained by such a model but require that the depleted upper mantle had γOs of approximately -5 to -7 at 62 Ma. This range overlaps the lowest γOs values measured in abyssal peridotites. Baffin Island lava compositions are also permissive of a model involving recharging of depleted upper mantle with ^3He -rich material from the lower mantle (Stuart et al., *Nature*, 424, 57–59, 2003), with the proviso that recharge had no recognizable effect on the lithophile trace element and Sr-Nd-Os-O isotope composition. The origin of the enriched mantle component sampled by Baffin Island lavas is less clear but may be metasomatized and high-temperature-altered recycled oceanic lithosphere transported within the proto Iceland plume. Differences between Baffin Island lavas and modern Icelandic basalts suggest that a range of enriched and depleted mantle sources have been tapped since the inception of magmatism in the province. Similarities between Baffin Island lavas erupted and those of similar age from East and West Greenland also suggest

that the enriched component in Baffin Island lavas may have been sampled by lavas erupted over a wide geographic range.

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1. Introduction

[2] A goal of many studies of continental flood basalts (CFB) and other magmatic rocks associated with the interaction between mantle plumes and continental lithosphere has been characterization of their mantle sources [e.g., Carlson *et al.*, 1981; Ellam *et al.*, 1992; Baker *et al.*, 1996; Hart *et al.*, 1997; Fitton *et al.*, 1997; Schaefer *et al.*, 2000; Kempton *et al.*, 2000; Ellam and Stuart, 2000; Stuart *et al.*, 2000, 2003; Peate *et al.*, 2003]. Despite the obscuring effects of crustal or lithospheric contamination, flood basalts provide unique information about the composition of upwelling asthenosphere beneath continental lithosphere and may allow better access to the magmatic products of upwelling “heads” of starting mantle plumes than many ocean island basalt (OIB) suites [Campbell and Griffiths, 1990; Schaefer *et al.*, 2000].

[3] The North Atlantic Igneous Province (NAIP), one of the most extensively studied magmatic provinces on earth, is an ideal location to examine the mantle sources of CFB owing to the thick sequences of mantle-derived lavas present over a wide geographic area and the preservation of material erupted from Palaeogene time through to the present [Saunders *et al.*, 1997]. The earliest recognized phases of volcanism are the ~60–62 Ma sequences now outcropping on Baffin Island, in West, East and Southeast Greenland, in the British Tertiary Province (BTP) [Saunders *et al.*, 1997]. This volcanism has been previously attributed to the earliest phase of activity of the same hot spot, or upwelling mantle “plume,” believed responsible for anomalously high rates of magmatism at the Icelandic portion of the mid-Atlantic ridge [Brooks, 1973; White and

McKenzie, 1989; Saunders *et al.*, 1997; Pedersen *et al.*, 2002].

[4] NAIP lavas of all ages show evidence of derivation from isotopically and chemically heterogeneous mantle sources [O’Nions *et al.*, 1975; Zindler *et al.*, 1979; Hémond *et al.*, 1993; Thirwall *et al.*, 1994, 2004; Thirwall, 1995; Hanan *et al.*, 2000; Kent and Fitton, 2000; Chauvel and Hémond, 2000; Kempton *et al.*, 2000; Stuart *et al.*, 2000, 2003; Skovgaard *et al.*, 2001; Peate and Stecher, 2003]. This is best documented in studies of recent Icelandic lavas [O’Nions *et al.*, 1975; Zindler *et al.*, 1979; Hémond *et al.*, 1993; Hards *et al.*, 1995; Thirwall, 1995; Fitton *et al.*, 1997; Chauvel and Hémond, 2000; Hanan *et al.*, 2000; Skovgaard *et al.*, 2001; Thirwall *et al.*, 2004]. In the older portions of the NAIP, particularly those erupted onto continental crust, the obscuring effects of assimilation of continental crust and lithospheric mantle and of melting beneath thick lithosphere complicate interpretations of this compositional diversity [Holm *et al.*, 1993; Lightfoot *et al.*, 1997; Saunders *et al.*, 1997; Hansen and Nielsen, 1998; Tegner *et al.*, 1998; Ellam and Stuart, 2000; Schaefer *et al.*, 2000; Peate *et al.*, 2003], but chemical and isotopic variations that are plausibly attributed to mantle source variations are evident in several suites [e.g., Holm *et al.*, 1993; Thirwall *et al.*, 1994; Lightfoot *et al.*, 1997; Saunders *et al.*, 1997; Kempton *et al.*, 2000; Schaefer *et al.*, 2000; Stuart *et al.*, 2000, 2003; Bernstein *et al.*, 2001; Peate and Stecher, 2003]. A key question, and one with important ramifications for the structure and dynamics of upwelling mantle plumes, is whether the same range of chemical and isotopic compositional components recognized in modern Icelandic lavas is also represented in Palaeogene sequences

[Thirwall *et al.*, 1994; Kempton *et al.*, 2000; Stuart *et al.*, 2000; Peate and Stecher, 2003]. In addition, some recent studies have also examined the possible roles for ancient oceanic crust and lithosphere (recycled via subduction) in the generation of the range of depleted and enriched chemical and isotopic signatures in Iceland lavas [Chauvel and Hémond, 2000; Skovgaard *et al.*, 2001; Thirwall *et al.*, 2004].

[5] In this paper we present new Sr, Nd, Os, and O isotope and trace element data for a suite of eight picritic lavas from Baffin Island, Canada. Variations in the major and trace element and isotopic compositions of these lavas show that they were not significantly contaminated during their passage through the subcontinental lithospheric mantle (SCLM) and continental crust and thus can be used to investigate the composition and origin of their mantle source regions. Note that by the terms “enriched” and “depleted” we refer to melts or mantle sources with incompatible element concentrations greater or less than the estimated composition of the bulk silicate earth, and radiogenic isotopic signatures consistent with long-term enrichment or depletion in incompatible elements.

2. Geological Background and Samples

[6] The picritic lavas that are the subject of this study are from Baffin Island, Canada and were erupted during the opening of the Davis Strait, which now separates Greenland from Baffin Island. White and McKenzie [1989] proposed that Palaeogene lavas and associated intrusive rocks from Baffin Island, East and West Greenland, the Faeroe Islands, and the British Tertiary province, are products of the initial phases of melting of the mantle plume currently located beneath Iceland. This conclusion has been largely accepted by subsequent workers [e.g., Robillard *et al.*, 1992; Chalmers *et al.*, 1995; Kent, 1995; Fitton *et al.*, 1997; Saunders *et al.*, 1997; Storey *et al.*, 1998; Kent and Fitton, 2000]. Although Gill *et al.* [1995] suggested that the Baffin Bay and West Greenland magmas formed from a separate short-lived mantle plume, subsequent authors [Chalmers *et al.*, 1995; Kent, 1995; Saunders *et al.*, 1997] have demonstrated that a single plume head, with temperatures up to 1500°C over a diameter of 2000 km was likely active during initiation of the NAIP. Although the age of the Baffin Island sequence is not precisely known, hyaloclastites and lavas outcropping on Baffin Island are normally magnetized [Deutsch *et al.*, 1971] and are correlated with

the earliest exposed phase of the West Greenland flood basalts (the Anaanaa Member of Pedersen *et al.* [2002]), which erupted at ~60–62 Ma [Storey *et al.*, 1998].

[7] The geological setting and petrography of the samples included in our study are described by Francis [1985] and Robillard *et al.* [1992]. Palaeogene lavas and associated hyaloclastites on Baffin Island occur in a series of outcrops along the eastern coast of Baffin Island and smaller associated islands and overlie Precambrian gneisses and Tertiary clastic sediments. Our samples are from eight picritic pillow lava flows that outcrop in the lower portion of a 750 m-thick section dominated by submarine-erupted pillow breccias and pillow lavas at the northeast end of Padloping Island (Lat. 67°10'N and Long. 62°25'W), just off the coast of Baffin Island [Clarke and Upton, 1971].

[8] The chemical compositions of whole rocks and matrix glasses from Baffin Island lavas, which are amongst the most MgO-rich Phanerozoic lavas known, have been extensively discussed previously [Clarke, 1970; Clarke and Upton, 1971; O'Nions and Clarke, 1972; Francis, 1985; Clarke *et al.*, 1988; Gill *et al.*, 1992; Robillard *et al.*, 1992; Herzberg and O'Hara, 2002; Stuart *et al.*, 2003]. The samples we have analyzed are a subset of those described by Francis [1985] and Robillard *et al.* [1992] and are taken from the porphyritic glassy margins of individual cooling units. These consist primarily of olivine phenocrysts (15–33 volume%) plus minor plagioclase and clinopyroxene microclites set in a glass matrix. Red alteration of the glass is present along fractures [Robillard *et al.*, 1992].

[9] Estimates of the conditions of melting for Baffin Island picrites vary, but overall suggest that primary melts formed at relatively deep levels and from large degrees of melting: Francis [1985] suggested that the Baffin Island picrites are the result of ~15–20% melting of spinel lherzolite at depths of ~80 km; Gill *et al.* [1992] estimated that Baffin Island and West Greenland picrites formed from slightly higher degrees of melting (24–30%) and at depths between 60–90 km, with melting capped by the lithosphere at ~60 km; and Herzberg and O'Hara [2002], from a forward model of lava sample 10014, suggested melt fractions between 10–11% (for a depleted source) and 23–28% (enriched source), and average melting depths between 70–120 km. Volcanic activity on Baffin Island and West Greenland correlates closely with regions of thinning lithosphere associated with the initial stages of formation of the Davis

Straight and, possibly, the Labrador Sea [Hill, 1991; Gill *et al.*, 1992, 1995] as well as with older regions of lithospheric extension and basin formation [Gill *et al.*, 1992, 1995; Larsen *et al.*, 1992]. For this reason it is widely considered that Baffin Island and West Greenland lavas largely represent melting of upwelling asthenospheric mantle, rather than of the continental lithosphere [Gill *et al.*, 1992, 1995; Larsen *et al.*, 1992; Francis, 1985; Herzberg and O'Hara, 2002]. On the basis of the most magnesian olivine phenocrysts, primary melts contained ~14–20 wt.% MgO [Francis, 1985; Herzberg and O'Hara, 2002], and the compositions of matrix glasses can be derived by fractional crystallization of 32–35 wt.% olivine from such a parental magma with 18 wt.% MgO [Francis, 1985]. Robillard *et al.* [1992] suggested from examination of trace element systematics that neither fractional crystallization, partial melting, nor crustal contamination produced the chemical and isotopic differences between N- and E-type magmas and that the differences between these magma types was most likely due to heterogeneity within the mantle sources of the picritic magmas. Stuart *et al.* [2003] also show that Baffin Island lavas (including samples from the same location as for this study, but not the exact same samples) have the highest magmatic $^3\text{He}/^4\text{He}$ ratios yet recorded, up to 49.5 R_A (where R_A is the atmospheric $^3\text{He}/^4\text{He}$ ratio; 1.39×10^{-6}), and that high $^3\text{He}/^4\text{He}$ ratios correlate with depleted chemical and isotopic signatures (e.g., low $^{87}\text{Sr}/^{86}\text{Sr}$, high $^{143}\text{Nd}/^{144}\text{Nd}$ and low [La/Sm]_N).

3. Analytical Methods

3.1. Major and Trace Element Analyses

[10] Robillard *et al.* [1992] reported the concentrations of major and some trace element in bulk separates of matrix glass from the same Baffin Island samples we have studied here. However, to ensure consistency with our database of melt inclusion analyses [Kent *et al.*, 1998; A. J. R. Kent, manuscript in preparation, 2004] and to supplement existing analyses with data for additional elements (Sc, V, Co, Rb, Nb, Ba, Pr, Hf, Ta, Pb, Th, U), we have analyzed the major and trace element compositions of matrix glasses using electron microprobe analysis (EMPA), secondary ion mass spectrometry (SIMS), and laser ablation ICP-MS (LA-ICP-MS) techniques. These in-situ analysis techniques also allowed us to avoid phenocrysts, microlites and altered and devitrified regions in glasses. Analyses were performed on hand-picked glass fragments mounted in epoxy

and polished. Prior to analysis, glasses were examined by optical microscopy and backscattered electron imagery to identify and avoid microlites and areas of altered and devitrified glass. Our LA-ICP-MS and SIMS measured concentrations are in generally good agreement with the ICP solution measurements of Robillard *et al.* [1992], although the LA-ICP-MS-measured concentrations are consistently 5–10% higher, consistent with dilution of the glass separates by microlites and/or phenocrysts.

[11] Major elements were analyzed using a JEOL 833 electron microprobe at Lawrence Livermore National Laboratory following the procedures outlined by Kent and Elliott [2002]. Analyses of BHVO-1 glass, an in-house standard made from the USGS BHVO-1 rock powder standard, were performed during each analysis session to provide a long-term monitor of accuracy and precision. The average of 114 BHVO-1 glass analyses (Table 1) is within 1σ (the standard deviation of the mean derived from the repeated analyses) of the accepted values for this standard for all elements. Analytical uncertainties, based on replicate analyses, are estimated at (2σ) <3% for SiO_2 , Al_2O_3 , and Na_2O ; <5% for CaO , FeO , MgO , and TiO_2 ; <10% for K_2O ; and <50% for Cr_2O_5 and MnO .

[12] Trace element concentrations in matrix glasses were analyzed by LA-ICP-MS in the W.M. Keck Collaboratory for Plasma Spectrometry, Oregon State University using a NewWave DUV 193 μm ArF Excimer laser and VG PQ ExCell Quadrupole ICP-MS. Analysis conditions are shown in Table 2. Ablation was performed using He as the carrier gas, and He plus ablated particulates were mixed with the Ar nebulizer gas flow immediately prior to the plasma torch. Background count rates were measured for 30 s prior to ablation and subtracted directly from count rates measured during ablation. Plasma torch conditions were optimized so that oxide production (estimated from measured ThO/Th ratios) were <1–2% and separate measurement of LREE-doped synthetic glasses confirms that contributions from REE oxides on MREE and HREE peaks were negligible ($\ll 1\%$) for the measured compositions.

[13] Each individual analysis represents 45 s of data acquisition during ablation (equating to a lateral beam translation distance of ~225 microns). The average depth of ablation along the trench produced by progressive ablation was about ~5 microns. Total count time per mass peak during ablation was ~2 seconds s. Abundances of individual trace

Table 1. Major and Trace Element and Sr-Nd Composition of Matrix Glasses, Os Isotopic Compositions of Pillow Margins and O-Isotope Composition of Olivine and Glass From Baffin Island Lavas^a

	Sample										
	10004	10013	10014	10019	10021	10024	10029	10064	BHVO- 1G ^b ±1σ ^b	BHVO- 2G ^c ±1σ ^c	BCR- 2G ^d ±1σ ^d
	wt. %										
SiO ₂	49.49	48.73	49.29	49.09	49.50	48.99	49.26	49.72	50.13	1.00	
TiO ₂	1.01	1.04	1.34	1.02	1.08	1.15	1.12	1.11	2.7	0.11	
Al ₂ O ₃	15.52	15.91	14.83	16.04	15.47	15.43	15.94	15.78	14.07	0.35	
Cr ₂ O ₃	0.04	0.06	0.05	0.05	0.05	0.05	0.05	0.06	0.05	0.02	
FeO	9.70	9.56	10.58	9.25	9.62	9.82	9.35	9.98	10.96	0.36	
CaO	14.31	14.64	13.93	14.13	14.23	13.39	13.85	13.87	11.19	0.38	7.12
MnO	0.17	0.17	0.19	0.17	0.19	0.19	0.16	0.16	0.15	0.06	
MgO	8.59	9.05	7.89	9.35	8.63	8.35	9.21	8.58	7.05	0.30	
Na ₂ O	2.00	1.85	2.14	1.86	1.98	2.15	1.84	1.95	2.26	0.06	
K ₂ O	0.15	0.05	0.17	0.05	0.16	0.16	0.05	0.10	0.48	0.04	
Total	101.08	101.89	101.09	101.57	101.83	100.75	101.26	101.06			
Mg#	63.70	65.23	59.64	66.70	63.99	62.75	66.12	63.02			
	ppm										
Sc	52.6	51.1	53.0	48.7	51.9	48.5	44.3	45.1			31.8 0.4 33
Ti	7403	7033	8960	6826	7557	7655	7197	7198			15698 45 13500
V	308	307	337	303	324	357	336	309			303 3.1 416
Cr	358	365	354	361	359	364	392	339			277 5.13 18
Mn	1458	1484	1661	1375	1473	1539	1423	1442			1354 29 1520
Co	42.7	45.9	47.4	44.0	43.4	49.4	48.5	46.1			43 0.6 37
Rb	3.93	1.02	3.42	1.09	3.98	3.96	1.43	2.50			8.98 0.69 48
Sr	148	130	175	130	149	157	142	150			404 3.4 346
Y	26.5	24.9	28.6	23.7	25.8	23.7	20.7	22.3			27.5 0.3 37
Zr	73.1	62.9	91.7	60.5	72.3	65.6	58.5	65.5			173 1.4 188
Nb	6.19	2.07	7.59	2.09	6.38	6.7	2.51	4.51			19 0.3 13.6
Ba	60.3	15.1	66.8	14.5	62.1	79.3	17.7	41.8			130 1.5 674
La	5.44	2.74	6.69	2.53	5.36	5.20	2.74	4.15			15.6 0.2 25.3
Ce	12.1	7.83	15.5	7.28	12.3	13	8.59	10.8			37.5 0.2 53.8
Pr	1.88	1.46	2.46	1.32	1.87	1.86	1.43	1.68			5.49 0.07 6.88
Nd	9.67	7.38	12.1	7.16	9.20	9.11	7.33	8.42			25.2 0.7 28.8
Sm	2.79	2.53	3.52	2.57	2.93	3.00	2.48	2.72			6.28 0.27 6.57
Eu	1.10	1.00	1.29	1.00	1.12	1.17	1.00	1.06			2.07 0.06 1.95
Gd	3.43	3.42	4.33	3.40	3.65	3.34	2.87	3.30			6.14 0.3 6.79
Dy	4.50	4.27	4.98	4.17	4.59	4.32	3.60	3.99			5.49 0.04 6.45
Er	2.82	2.60	3.08	2.58	2.81	2.50	2.15	2.35			2.68 0.05 3.72
Yb	2.63	2.36	2.77	2.41	2.45	2.40	2.15	2.26			2.00 0.09 3.34
Hf	1.99	1.61	2.32	1.60	1.82	1.67	1.50	1.72			4.31 0.11 4.8
Ta	0.43	0.15	0.47	0.13	0.38	0.37	0.15	0.27			1.32 0.06 0.86

Table 1. (continued)

	Sample										
	10004	10013	10014	10019	10021	10024	10029	10064	BHVO-1G ^b ±1σ ^b	BHVO-2G ^c	BCR-2G ^d ±1σ ^c
Pb	0.56	0.38	0.81	0.45	0.63	1.08	0.58	0.65		1.78	0.05 11
Th	0.60	0.19	0.59	0.16	0.67	0.46	0.14	0.35		1.27	0.04 6.20
U	0.13	0.04	0.14	0.03	0.13	0.11	0.04	0.09		0.40	0.01 1.69
K/Nb	202	183	182	188	207	197	166	183			
[La/Sm]N	1.23	0.68	1.19	0.62	1.15	1.09	0.69	0.96			
K ₂ O/	0.15	0.04	0.12	0.05	0.15	0.14	0.04	0.09			
TiO ₂											
ΔNb	0.26	-0.11	0.19	-0.09	0.27	0.34	-0.04	0.15			
δ ¹⁸ O _{olivines}	5.04, 5.04	5.21, 5.22	5.00, 5.10		5.10, 5.16	5.04, 5.07	5.15, 5.17	4.84			
‰											
δ ¹⁸ O _{glass}	5.31		5.07	5.42		5.09		5.02			
‰											
[Os], ppt		949	382	2004	1761		1343	1089			
[Re], ppt		514	268	466	482		587	381			
¹⁸⁷ Re/ ¹⁸⁸ Os ^e		2.561	3.321	1.1	1.294		2.067	1.655			
¹⁸⁷ Os/ ¹⁸⁸ Os ^e		0.1273(8)	0.1327(7)	0.1237(5)	0.1316(13)		0.1241(10)	0.1278(4)			
¹⁸⁷ Os/ ¹⁸⁸ Os _i		0.1247	0.1293	0.1226	0.1303		0.1220	0.1261			
γOs _i		-3.1	0.5	-4.7	1.3		-5.2	-2.0			
⁸⁷ Sr/ ⁸⁶ Sr ^e	0.703917(13)	0.703024(13)	0.703355(13)	0.703081(11)	0.703879(12)	0.703283(12)	0.703015(12)	0.703646(13)			
⁸⁷ Sr/ ⁸⁶ Sr _i	0.703851	0.703005	0.703306	0.703060	0.703813	0.70322	0.70299	0.703604			
εSr _i	-8.2	-20.2	-15.9	-19.4	-8.7	-17.2	-20.4	-11.7			
¹⁴³ Nd/ ¹⁴⁴ Nd ^e	0.512919(9)	0.513086(8)	0.512964(9)	0.513086(9)	0.512906(8)	0.512989(9)	0.513075(12)	0.512988(16)			
¹⁴³ Nd/ ¹⁴⁴ Nd _i	0.512845	0.512999	0.512890	0.512994	0.512825	0.512905	0.512989	0.512906			
εNd _i	5.6	8.6	6.5	8.5	5.2	6.8	8.4	6.8			

^a All initial ratios calculated at t = 62 Ma.

^b Major element compositions and 1σ errors for BHVO-1G represent the average and standard deviation for 114 EMPA analyses of this glass.

^c Trace element compositions and 1σ errors for BHVO-2G represent the average and standard deviation for 5 LA-ICP-MS analyses of this glass.

^d Trace element compositions for BCR-2G used for calibration of LA-ICP-MS analyses (data from *Fisk and Kelley* [2002], *Kent et al.* [2004], and USGS recommended values).

^e Measured ratios. Errors shown in brackets are 2σ and in the last decimal places shown (e.g., 0.512919(9) = 0.512919 ± 0.000009).

Table 2. LA-ICP-MS Analysis Conditions

	Description
Instrumentation	VG ExCell and NewWave DUV 193 ArF Excimer laser
<i>Analyzer Conditions</i>	
Aerosol carrier gas flow	0.75 l/min (He)
Nebulizer gas flow	0.95 l/min (Ar)
Outer (cool) gas flow	13.00 l/min (Ar)
Detector mode	Dual (pulse counting and analogue)
RF power	1350 W
Vacuum pressure	8.0×10^{-7} mbar (analyzer)
	1.6 mbar (expansion chamber)
Mass Table	^{43}Ca , ^{45}Sc , ^{47}Ti , ^{51}V , ^{53}Cr , ^{55}Mn , ^{59}Co , ^{85}Rb , ^{88}Sr , ^{89}Y , ^{90}Zr , ^{93}Nb , ^{137}Ba , ^{139}La , ^{140}Ce , ^{141}Pr , ^{146}Nd , ^{147}Sm , ^{153}Eu , ^{157}Gd , ^{163}Dy , ^{166}Er , ^{172}Yb , ^{178}Hf , ^{181}Ta , ^{208}Pb , ^{232}Th , ^{238}U
Dwell time/mass/scan	10 ms
Total scan time	~300 ms
<i>Laser Conditions</i>	
Wavelength	193 nm
Frequency	5 hz
Pulse duration	15 ns
Spot diameter	50 μm
Lateral translation rate	5 $\mu\text{m/s}$ (total distance ~225 μm)
Ablation duration	45 s
Output energy	200 mJ at 193 nm (~15 J/cm ²)
<i>Standardization</i>	
Internal standard	^{43}Ca
Calibration standard	BCR-2G
Secondary standard	BHVO-2G

elements were calculated relative to the USGS glass standard BCR-2G, which was analyzed under identical conditions throughout the analysis session (see Table 1 for the values used for calibration). ^{43}Ca was used as an internal standard in conjunction with CaO contents measured by electron probe. USGS glass standard BHVO-2G was also analyzed to monitor accuracy and precision. Analyses for BHVO-2G and Baffin Island samples shown in Table 1 represent the average of 5 separate analyses per sample. Relative errors in concentrations calculated from the external standard deviation from multiple analyses typically correlate positively with $(\sqrt{C_i})/C_i$ (where C_i = concentration of element i) consistent with counting statistics being the dominant source of error. Although dependant upon elemental abundances in individual glass samples, external errors calculated from multiple analyses were typically $\leq 5\%$ for most elements (Sc, Ti, V, Cr, Mn, Co, Ni, Rb, Sr, Y, Zr, Nb, Ba, LREE, Er); $\leq 10\%$ for Dy, Gd, Yb, Hf, Ta, Pb; and $\leq 15\%$ for U and Th (at 2σ). Measured concentrations from LA-ICP-MS analyses also agree well with those determined for a subset of these elements by secondary ion mass spectrometry (SIMS) for the same glass chips, measured using a modified Cameca 3f ion microprobe at Lawrence

Livermore National Laboratory following the techniques outlined by *Kent and Elliott* [2002]. These data are given in Appendix A. Although LA-ICP-MS and SIMS data agree well, the higher counts rates associated with LA-ICP-MS typically provide more precise data, so we exclusively refer to the LA-ICP-MS data in the discussion of our analyses.

3.2. Os, Sr, Nd, and O Isotopic Analyses

[14] Sr and Nd isotopic compositions of matrix glasses were analyzed at the University of Melbourne, Australia. Glass chips for isotopic analysis were selected by hand-picking (avoiding mineral and altered materials) washed in ultra-pure water, leached with hot acid (4N HCl) for 1 hr, and then rinsed with ultrapure water prior to dissolution in HF + HNO₃. Strontium and Nd were separated by standard ion exchange methods and loaded on single Ta (Sr) or double Re-Ta (Nd) filament assemblies. Isotope measurements were made on a Finnigan MAT262 mass spectrometer operated in static multicollection mode and normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Measurements of the NIST SRM-987 Sr standard gave $^{87}\text{Sr}/^{86}\text{Sr} = 0.710230$; the La Jolla Nd Stan-

dard gave $^{143}\text{Nd}/^{144}\text{Nd} = 0.511860$. Precision on individual ratios is reported as twice the standard error. Measured $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios have been corrected for decay of ^{147}Sm and ^{87}Rb in the 62 Ma since eruption using the measured Sm, Nd, Rb, and Sr concentrations of matrix glasses (Table 1). The magnitude of this correction is between 0.003–0.009% for $^{87}\text{Sr}/^{86}\text{Sr}$ and 0.014–0.018% for $^{143}\text{Nd}/^{144}\text{Nd}$ ratios.

[15] Osmium isotopic ratios and Re and Os abundances were measured at the University of Copenhagen following the techniques detailed by *Larsen et al.* [2003]. Hand picked, fresh chips from the chilled margins of six lava samples (consisting of matrix glass plus olivine phenocrysts) were chosen for analysis and then powdered in an agate mortar. Powders were weighed and spiked with an ^{190}Os - ^{187}Re -enriched solution and dissolved in reverse aqua regia (14N HNO_3 :10N $\text{HCl} = 3:1$) in carius tubes at 230°C for 1 week. Osmium was distilled from aqua regia directly into 8N HBr and purified following the techniques of *Roy-Barman and Allegre* [1994]. Osmium isotopic ratios were analyzed on a VG Sector 54 thermal ionization mass spectrometer. Rhenium was purified using the liquid extraction technique of *Cohen and Waters* [1996] and analyzed on an Axiom MC-ICP-MS using Ir-doped solutions for controlling mass fractionations. Procedural blanks were <30 pg and 3 pg for Re and Os. Measured $^{187}\text{Os}/^{188}\text{Os}$ ratios were corrected for in-situ ^{187}Re decay over 62 Ma using measured Re/Os ratios and the magnitude of this correction is between 1–2%. The blank $^{187}\text{Os}/^{188}\text{Os}$ composition during the period of analysis was 0.131 ± 0.005 (1σ , $n = 3$) and the sample $^{187}\text{Os}/^{188}\text{Os}$ ratios were correspondingly corrected for this blank. γOs values have been calculated using modern-day primitive mantle values of $^{187}\text{Os}/^{188}\text{Os} = 0.1296$, $^{187}\text{Re}/^{188}\text{Os} = 0.428$, and $\lambda_{\text{Re}} = 1.666 \text{ e}^{-11}$ [*Meisel et al.*, 2001; *Smoliar et al.*, 1996].

[16] Oxygen isotope analyses of olivine and glass were conducted at the California Institute of Technology following techniques outlined by *Eiler et al.* [2000a]. Olivine and glasses were hand picked to be free of alteration and inclusions. Analyses were normalized to analyses of standard UWG-2. The estimated analytical precision is $\leq 0.05\%$.

4. Results and Discussion

[17] The results of major element, trace element and Sr-Nd-Os-O isotopic analyses are listed in Table 1.

4.1. Compositional Variations in Baffin Island Lavas

[18] Two geochemically distinct lava types (labeled E- and N-type by analogy with E- and N-MORB by *Robillard et al.* [1992]) occur in the Baffin Island sequence (Figure 1). Although the compositional boundary between them is arbitrary, *Robillard et al.* [1992] defined N-type lavas as those with matrix glasses having $\text{K}_2\text{O}/\text{TiO}_2 < 0.08$ and $[\text{La}/\text{Sm}]_{\text{N}} < 0.7$, and we adopt this definition. In this study N-type lavas have $[\text{La}/\text{Sm}]_{\text{N}}$ from 0.62–0.69 and $\text{K}_2\text{O}/\text{TiO}_2$ from 0.04–0.05 and E-type lavas have $[\text{La}/\text{Sm}]_{\text{N}}$ from 0.96–1.23 and $\text{K}_2\text{O}/\text{TiO}_2$ from 0.09–0.15.

[19] Our data shows that there are also differences in the radiogenic isotope ratios of E- and N-type lavas. E-type lavas have distinctly higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than N-type lavas (Table 1; Figure 2), and the Sr, Nd, and Os isotopic compositions of E- and N-type lavas are well correlated (Figure 2). Radiogenic isotope compositions are also strongly correlated (positively for $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$ and negatively for $^{143}\text{Nd}/^{144}\text{Nd}$) with the degree of incompatible element enrichment (Figure 3).

[20] Baffin Island lavas define a negatively sloped array on a plot of $^{87}\text{Sr}/^{86}\text{Sr}_i$ versus $^{143}\text{Nd}/^{144}\text{Nd}_i$ (Figure 4). Data from our study overlap those of *Stuart et al.* [2003], also from Baffin Island, although they tend to slightly lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ at a given $^{87}\text{Sr}/^{86}\text{Sr}_i$. Baffin Island lavas also have higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ and lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios than the estimated modern “plume-free” Mid-Atlantic Ridge MORB composition of *Taylor et al.* [1997], although they project toward the calculated composition of this component at 62 Ma (Figure 4a). Strontium and Nd isotopic compositions also overlap those of uncontaminated lavas from West Greenland [*Holm et al.*, 1993; *Lightfoot et al.*, 1997]. The range of $^{187}\text{Os}/^{188}\text{Os}_i$ ratios for Baffin Island lavas overlap those reported for uncontaminated basalts from ~60–62 Ma sequences from East and West Greenland [e.g., *Brooks et al.*, 1999; *Schaefer et al.*, 2000] but extend to significantly lower values of $^{187}\text{Os}/^{188}\text{Os}_i$ than previously reported for nonalkalic lava suites from East and West Greenland lavas (0.127 [*Brooks et al.*, 1999; *Schaefer et al.*, 2000]), the British Tertiary province (0.129 [*Parkinson et al.*, 2001]) and for Iceland (0.127 [*Martin*, 1991; *Skovgaard et al.*, 2001; *Smit et al.*, 2001]). Alkalic lavas from West Greenland (Maníðtlát Member) have $^{187}\text{Os}/^{188}\text{Os}_i$ ratios and γOs_i values that

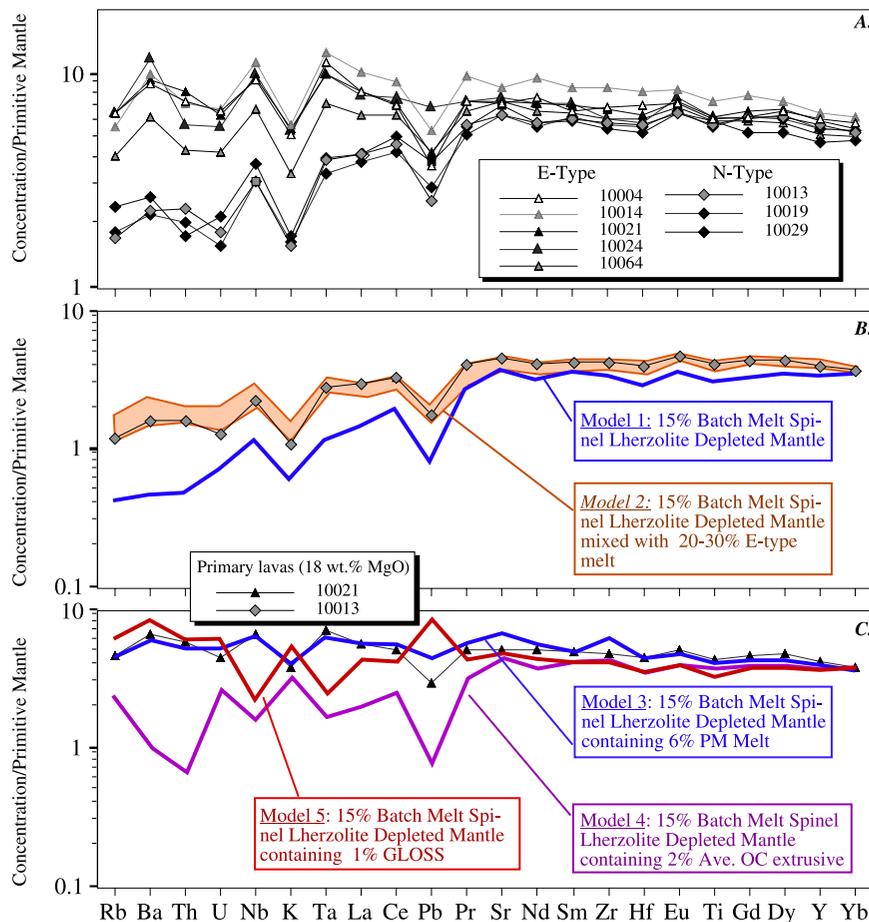


Figure 1. (a) Primitive mantle-normalized trace element concentrations for matrix glasses from Baffin Island lavas measured by LA-ICP-MS. Primitive mantle values for normalizing and melt modeling are from *Sun and McDonough* [1989]. Symbols are as per legend. (b) Results of batch modal melting calculations for depleted melts. Primary magma compositions for representative N- and E-type primary melts in Figures 1b and 1c were estimated from samples 10013 and 10021 and calculated by incrementally adding equilibrium olivine (assuming all trace elements shown to be perfectly incompatible in olivine) to matrix glass compositions until MgO reached the estimated primary magma MgO content of 18 wt.% from *Francis* [1985]. Melt models were calculated using the study of *Francis* [1985] as a guide to source mineralogy and degree of melting. All melts were calculated using 15% batch modal melting of spinel lherzolite (olivine, 58%; orthopyroxene, 13%; clinopyroxene, 26%; spinel, 3%) and partition coefficients compiled by *Gurenko and Chaussidon* [1995] and *Rollinson* [1993]. Depleted mantle (DM) starting compositions are from *Gurenko and Chaussidon* [1995] and W. F. McDonough (personal communication, 2003). Model 1 shows the calculated melt composition produced by 15% melting of depleted mantle. Model 2 shows calculated melt compositions produced by mixing this melt with a primary E-type melt (based on the composition of sample 10021). (c) Results of batch modal melting calculations for enriched melts. Melts were calculated using the same method as described above. Model 3 shows the 15% melting of a depleted mantle source that also contains 7% of a metasomatic melt, itself derived from 6% melting of a primitive mantle composition spinel lherzolite source. Model 4 shows 15% melting of a depleted mantle source that also contains 2% average ocean crust extrusive. Model 5 shows 15% melting of a depleted mantle source that also contains 1% GLOSS average subducted sediment [*Plank and Langmuir*, 1998]. The sources, proportions, and compositions of metasomatic melts that produce the best fit shown by Model 3 were determined by varying the amount and composition of the metasomatic melt added to the DM source to minimize the sum of the squares of the residuals (SSR) in comparison to the 10013 compositions shown. The final SSR for the composition shown was 1.3. It was not possible to produce a final melt with low SSR (<5) using a metasomatic melt derived from a depleted mantle composition source.

extend to values that are lower than those of N-type lavas from Baffin Island (Figure 5), with $^{187}\text{Os}/^{188}\text{Os}_i$ as low as 0.10643 [Larsen *et al.*, 2003]. Baffin Island lavas also show ranges of Os and Re concentrations (382–2004 and 268–587 ppt, respectively) similar to those of uncontaminated West and East Greenland picrites [Brooks *et al.*, 1999; Schaefer *et al.*, 2000].

[21] Major element compositions of E- and N-type glasses, corrected for olivine fractionation to $\text{MgO} = 8$ wt.% (by incremental addition of equilibrium olivine) also show distinct differences between E- and N-type lavas (Figure 3), although correlation coefficients for major elements shown in this figure are generally lower than for those between incompatible trace elements and isotopic

compositions. Overall, E-type lavas have higher Na_2O , FeO_T , and SiO_2 contents and lower CaO and Al_2O_3 contents than N-type lavas.

[22] Oxygen isotope ratios of glass and olivine from Baffin Island lavas are given in Table 1, and plotted against Nd and Os isotopic compositions in Figures 2a and 6. Differences in $\delta^{18}\text{O}$ between glass and olivine range from 0.27 to -0.03 (where these pairs were measured) and are not consistent with the known fractionation between olivine and basaltic melt at magmatic temperatures (typically 0.5‰ [Eiler, 2001]). Such disequilibrium could reflect assimilation of crustal and/or hydrothermally altered volcanic material by mafic melts after olivine crystallization [Garcia *et al.*, 1998], although in our subsequent discussion we show from the compositions of matrix glasses that crustal contamination in these magmas appears minimal. We suggest that the most likely source of O-isotope disequilibrium is low-temperature hydration or alteration during submarine eruption or in the ensuing 62 Ma. Many hand samples of Baffin Island lavas show thin red ($< \sim 5$ mm) alteration selvages along fractures (although every effort was made to avoid this material during selection of glass for analysis). Although alteration could potentially also affect the compositions of other labile elements, such as Na, K, Rb, Sr and Pb, correlations between these elements and $\delta^{18}\text{O}_{\text{glass}}$ and between Sr isotopic composition and $\delta^{18}\text{O}_{\text{glass}}$

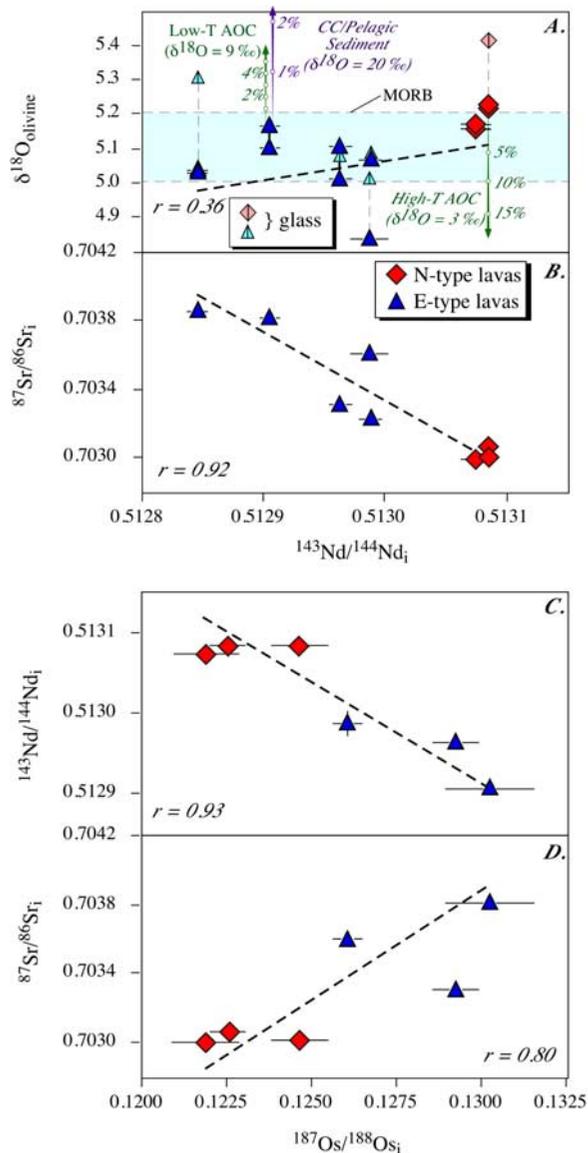


Figure 2. $^{143}\text{Nd}/^{144}\text{Nd}_i$ versus (a) $\delta^{18}\text{O}_{\text{olivine}}$ and $\delta^{18}\text{O}_{\text{glass}}$, and (b) $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{187}\text{Os}/^{188}\text{Os}_i$ versus (c) $^{87}\text{Sr}/^{86}\text{Sr}_i$ and (d) $^{143}\text{Nd}/^{144}\text{Nd}_i$ for Baffin Island lava samples. Symbols are as per legend. All isotopic ratios are calculated for an eruption age of 62 Ma. Thin dashed lines join glass and olivine from the same lava sample, and solid lines join olivine from the sample. Blue field bounded by dashed lines in Figure 2a shows the range of MORB $\delta^{18}\text{O}_{\text{olivine}}$ values from Eiler *et al.* [2000b]. Thick dashed lines show best fit regressions for all Figures, and the value of the correlation coefficient r is given in italics. Error bars represent 2σ . With the exception of Figure 2a, all correlations are significant at the 80% or greater confidence limit [Rollinson, 1993]. Tapered arrows in Figure 2a show the calculated effect on $\delta^{18}\text{O}_{\text{olivine}}$ of adding low-temperature altered ocean crust (Low-T AOC) with $\delta^{18}\text{O} = 9\text{‰}$ and continental crust (CC) or pelagic sediment with $\delta^{18}\text{O} = 20\text{‰}$ to the melt from which the N-type type olivine with the highest measured $\delta^{18}\text{O}_{\text{olivine}}$ crystallized, and high-temperature-altered oceanic crust (High-T AOC) with $\delta^{18}\text{O} = 3\text{‰}$ to the melt from which the E-type olivine with the highest measured $\delta^{18}\text{O}_{\text{olivine}}$ crystallized. Proportions of added material are as per labels.

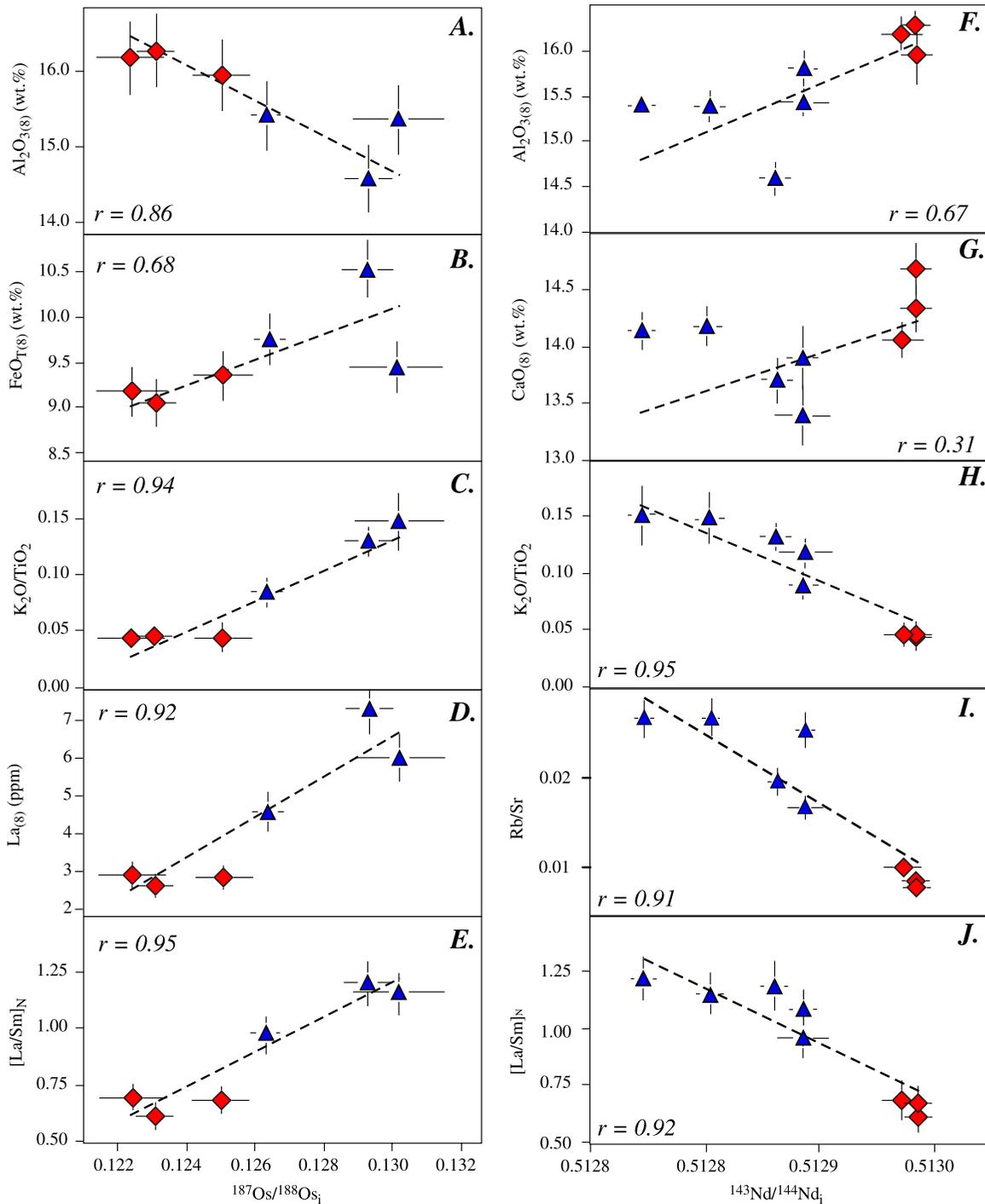


Figure 3. Comparison between the major and trace element compositions and isotopic compositions of Baffin Island lavas showing $^{187}\text{Os}/^{188}\text{Os}_i$ versus (a) $\text{Al}_2\text{O}_{3(s)}$, (b) $\text{FeO}_{T(s)}$, (c) $\text{K}_2\text{O}/\text{TiO}_2$, (d) $\text{La}_{(s)}$, and (e) $[\text{La}/\text{Sm}]_N$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$ versus (f) $\text{Al}_2\text{O}_{3(s)}$, (g) CaO , (h) $\text{K}_2\text{O}/\text{TiO}_2$, (i) Rb/Sr , and (j) $[\text{La}/\text{Sm}]_N$. Symbols are the same as for Figure 2, and error bars represent 2σ errors. Major and trace element concentrations are corrected to $\text{MgO} = 8$ wt.% to account for differences in olivine fractionation (although the differences between E- and N-type lavas are still apparent if this correction is not made). Thick dashed lines represent best fit regressions in all Figures, and the value of the correlation coefficient r is given in italics. With the exception of Figures 3b and 3g, all correlations are significant at the 80% or greater confidence limit [Rollinson, 1993].

are poor (with the correlation coefficient r^2 typically $\ll 0.3$), while correlations between labile elements and more immobile incompatible elements (LREE, Nb, Ti) are good (typically with $r^2 \gg 0.7$), suggesting that alteration of $\delta^{18}\text{O}$ in glasses has not significantly effected trace elements concentrations and Sr isotopes. However, we have little confidence that the $\delta^{18}\text{O}$ values of glasses reflect those of pre-eruptive magmas and focus our attention instead on the $\delta^{18}\text{O}$ values of olivine phenocrysts. There are also small differences between E- and N-type lavas from Baffin Island in terms of $\delta^{18}\text{O}_{\text{olivine}}$ (Figure 2): Olivines from N-type lavas having $\delta^{18}\text{O}$ values of 5.15–5.22‰, whereas $\delta^{18}\text{O}_{\text{olivine}}$ in E-type lavas are mostly 5.05–5.16‰ (one olivine sample from 10064 has $\delta^{18}\text{O}_{\text{olivine}} = 4.84\text{‰}$). Although these differences are outside of analytical error ($\pm 0.05\text{‰}$), they are small, and more work will be required to establish whether they are significant. The key point, however, is that the olivines from all but one sample

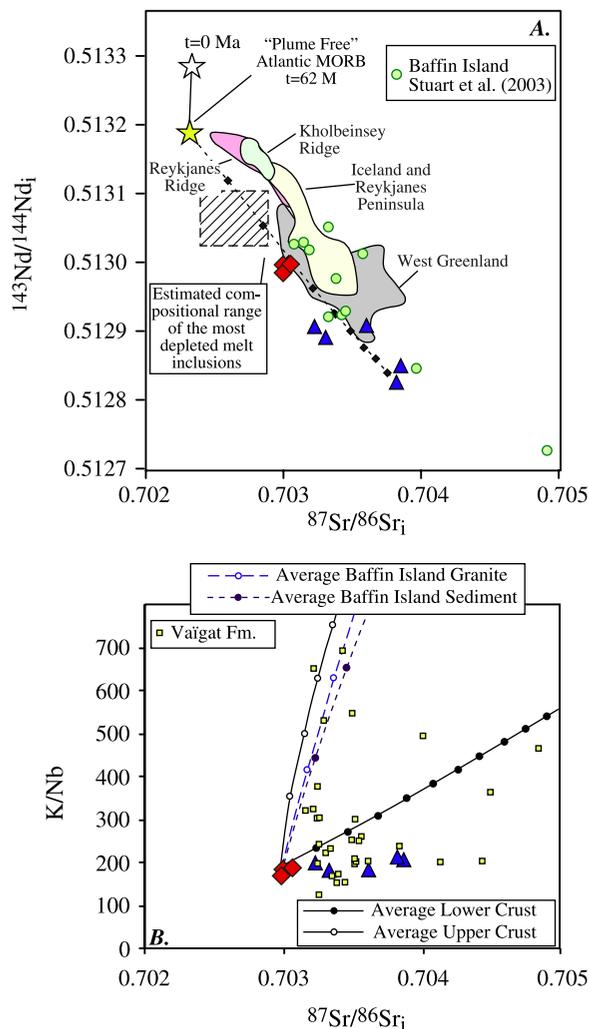


Figure 4. (a) $^{87}\text{Sr}/^{86}\text{Sr}_i$ versus $^{143}\text{Nd}/^{144}\text{Nd}_i$ for lavas from Baffin Island from this study and from *Stuart et al.* [2003]. Also shown are fields representing data from uncontaminated West Greenland lavas from the Vaigat Formation and equivalents [*Holm et al.*, 1993] with the exception of samples 278559 and 181031/II, which are considered crustally contaminated (L. M. Larsen, personal communication, 2004), Iceland, and the Kholbeinsey and Reykjanes Ridges from *Thirwall et al.* [2004]. The estimated modern depleted mantle and “plume-free” Atlantic depleted upper mantle composition is from *Taylor et al.* [1997], and the 62 Ma composition ($^{87}\text{Sr}/^{86}\text{Sr} = 0.70230$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51319$) has been calculated using depleted mantle values of $\text{Rb}/\text{Sr} = 0.00315$ and $\text{Sm}/\text{Nd} = 0.328$. All data from Baffin Island have been age corrected for post-eruption Rb and Sm decay to 62 Ma; data from Iceland are not age corrected. The data of *Holm et al.* [1993] have been corrected to 60 Ma. Symbols for Baffin Island lavas are as for Figure 3 or as shown in the legend. Hatched square shows the range of estimated depleted melt compositions for Baffin Island calculated using the lowest measured $[\text{La}/\text{Sm}]_N$ and $\text{K}_2\text{O}/\text{TiO}_2$ values measured in melt inclusions [*Kent et al.*, 1998; A. J. R. Kent et al., manuscript in preparation, 2004] and the correlation observed between Sr and Nd isotope composition and $[\text{La}/\text{Sm}]_N$ and $\text{K}_2\text{O}/\text{TiO}_2$ observed in matrix glasses (Figure 2). The mixing line shown represents mixing between melts derived from 15% batch melting of depleted mantle (Model 4 in Figure 1) with $\text{Sr} = 72$ and $\text{Nd} = 3.8$ ppm and Sr and Nd isotope composition equivalent to that calculated for DM at 62 Ma (see above) and a parental Baffin Island E-type melt with $\text{Sr} = 174$ and $\text{Nd} = 9.7$ ppm and $^{87}\text{Sr}/^{86}\text{Sr} = 0.703813$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512825$. Points on the mixing line represent 10% increments. (b) K/Nb versus $^{87}\text{Sr}/^{86}\text{Sr}_i$ for Baffin Island lavas from this study and for lavas from the Vaigat Formation, West Greenland from *Holm et al.* [1993] and *Lightfoot et al.* [1997]. Symbols are the same as for Figure 3 or as shown in the accompanying legend. Lines show calculated mixing trends between an N-type melt ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7030$, $\text{K}/\text{Nb} = 190$, $\text{K} = 417$ ppm, $\text{Sr} = 142$ ppm) and (1) average upper crust ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710$, $\text{K}/\text{Nb} = 2430$, $\text{K} = 34000$ ppm, $\text{Sr} = 350$ ppm); (2) average lower crust ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710$, $\text{K}/\text{Nb} = 3518$, $\text{K} = 6000$ ppm, $\text{Sr} = 347$ ppm); (3) average Baffin Island granitoid composition ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710$, $\text{K}/\text{Nb} = 4100$, $\text{K} = 27500$ ppm, $\text{Sr} = 302$ ppm); and (4) average Baffin Island metasediment composition ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710$, $\text{K}/\text{Nb} = 2190$, $\text{K} = 21900$ ppm, $\text{Sr} = 140$ ppm). Circles along mixing lines represent mixing in 2% increments. Average Baffin Island sediment and granite compositions are calculated from the data of *Thériault et al.* [2001]; average crustal compositions are based on *Taylor and McLennan* [1985].

are within the range expected for olivine in equilibrium with N-MORB melts (see the band in Figure 2a [Eiler, 2001]), and the exception is only slightly lower than this range.

4.2. Origin of Chemical and Isotopic Variations in Baffin Island Lavas

[23] Robillard *et al.* [1992] concluded that differences in trace element composition between N- and E-type lavas at Baffin Island are unlikely to be the result of variations in the conditions of partial melting of a single homogenous mantle source and/or low-pressure crystallization-differentiation processes. Our new isotopic and trace element

data strongly support this conclusion. Both the range of isotopic compositions observed in Baffin Island lavas and the correlations between Sr, Nd, and Os isotope ratios on the one hand and trace and major element compositions on the other are unlikely to have resulted from melting of a homogeneous source and/or crystallization-differentiation processes alone. Our results suggest that the formation of the Baffin Island lavas involved some process of mixing between at least two components of contrasting isotopic and chemical composition.

[24] We consider three hypotheses for explaining the compositional and isotopic variations of Baffin Island lavas: (1) contamination of mantle-derived magmas by continental crust; (2) assimilation of SCLM material by ascending magmas; and (3) derivation from a sublithospheric mantle source

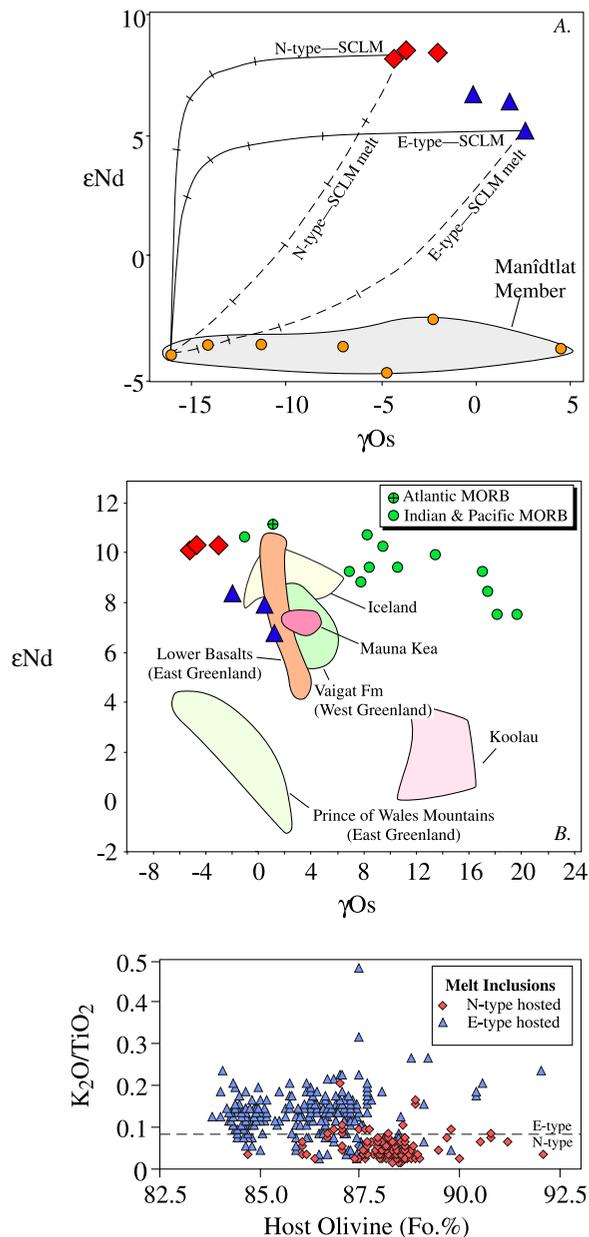


Figure 5. γOs versus ϵNd for Baffin Island lavas and other samples. (a) Comparison of Baffin Island lavas and alkalic Maniðtlat Member lavas from West Greenland [Larsen *et al.*, 2003]. Models for mixing between representative E- and N-type melts (samples 10013 and 10021) and Maniðtlat lava sample 264093 from Larsen *et al.* [2003] (that with the lowest γOs) were used to constrain the composition of the SCLM in this region. Mixing is shown both between melts (dashed lines) and between Baffin Island lavas and bulk SCLM (solid lines). Ticks on mixing lines represent 20% increments by weight. The composition of SCLM used is $[\text{Os}] = 9.2$ ppb, $[\text{Nd}] = 1$ ppm, $^{187}\text{Os}/^{188}\text{Os}_i = 0.1065$, $^{143}\text{Nd}/^{144}\text{Nd}_i = 0.51236$; SCLM melt, $[\text{Os}] = 0.9$ ppb, $[\text{Nd}] = 10$ ppm, $^{187}\text{Os}/^{188}\text{Os}_i = 0.1065$, $^{143}\text{Nd}/^{144}\text{Nd}_i = 0.51236$. E- and N-type lava compositions are from Table 1. Circles represent Maniðtlat member lava compositions. Other symbols are the same as for Figure 1. (b) Comparison between Baffin Island lavas and the compositions of MORB glasses and fields representing the range of compositions reported for lavas from Iceland, Hawaii (Mauna Kea and Koolau), uncontaminated Vaigat Formation picrites from West Greenland (~60–62 Ma), picrites from the Lower Basalts of East Greenland (~60–62 Ma), and post-plateau alkalic and tholeiitic lavas from the Prince of Wales Mountains (53–55 Ma). Data are from Schiano *et al.* [1997], Lassiter and Hauri [1998], Hansen and Nielsen [1998], Brooks *et al.* [1999], Schaefer *et al.* [2000], Skovgaard *et al.* [2001], and Peate *et al.* [2003]. (c) Measured $\text{K}_2\text{O}/\text{TiO}_2$ ratios in olivine-hosted melt inclusions from E- and N-type Baffin Island lavas versus composition of the olivine host crystals. Melt inclusions from both lavas types hosted in olivines with composition $>\text{Fo}_{90}$ have compositions within the range observed for inclusions in less-fosteritic hosts. Data are from A. J. R. Kent (unpublished data, 2003).

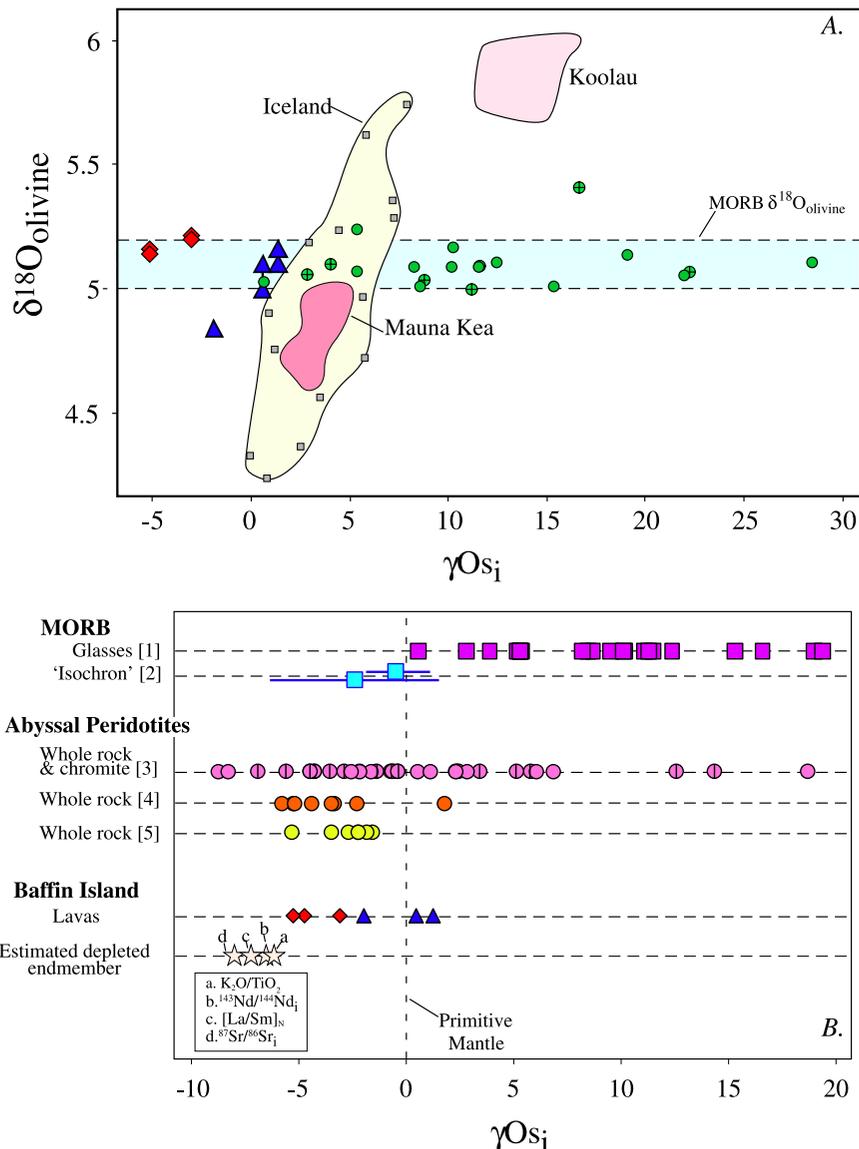


Figure 6. (a) $\delta^{18}\text{O}_{\text{olivine}}$ versus γOs_i for Baffin Island lavas and MORB glasses and for lavas from Iceland and Hawaii (Mauna Kea and Koolau). Symbols are as for Figure 1 and Figure 5. Data sources are the same as for Figure 5 with additional oxygen isotope data from *Eiler et al.* [1996, 2000b]. Blue field bounded by dashed lines shows the accepted range of MORB $\delta^{18}\text{O}_{\text{olivine}}$ values [*Eiler et al.*, 2000b]. (b) Comparison of γOs_i for Baffin Island lavas, MORB, and abyssal peridotites. Also shown as stars are four estimates of the γOs_i for the depleted mantle source for Baffin lavas based on best fit lines calculated from least squares linear regressions for correlations between $^{187}\text{Os}/^{188}\text{Os}_i$ and (a) $\text{K}_2\text{O}/\text{TiO}_2$, (b) $^{143}\text{Nd}/^{144}\text{Nd}_i$, (c) $[\text{La}/\text{Sm}]_N$, and (d) $^{87}\text{Sr}/^{86}\text{Sr}_i$, in Baffin Island lavas (Figures 2 and 3), and either the estimate of Sr and Nd isotope composition of the depleted North Atlantic MORB source at 62 Ma (see Figure 4) or the calculated $[\text{La}/\text{Sm}]_N$ (0.39) and $\text{K}_2\text{O}/\text{TiO}_2$ (0.01) for the 15% batch melt of a depleted mantle shown in Figure 1 (see caption for details). Data sources for MORB are [1] glasses from *Schiano et al.* [1997] and [2] initial ratios calculated from “isochrons” generated from MORB samples by *Gannoun et al.* [2004]. Note that MORB glasses with γOs_i greater than 20 (four samples) are not shown. Data sources for abyssal peridotites are [3] whole rock (solid circle) and chromite (solid circle with vertical line) from *Standish et al.* [2002], [4] whole rock analyses from *Snow and Reisberg* [1995], and [5] whole rock analyses from *Brandon et al.* [2000]. Error bars shown for [2] represent the reported errors in initial ratios calculated from “isochron” intercepts [*Gannoun et al.*, 2004].

that contains chemically and isotopically distinct components.

4.2.1. Interaction With Continental Crust or the SCLM

[25] Throughout the NAIP, including Iceland, there is considerable evidence for modification of mantle-derived lavas by interaction with the crust during magma transport and storage [e.g., *Condomines et al.*, 1983; *Nicholson et al.*, 1991; *Sigmarsson et al.*, 1992; *Holm et al.*, 1993; *Lightfoot et al.*, 1997; *Saunders et al.*, 1997; *Gee et al.*, 1998; *Hansen and Nielsen*, 1998; *Eiler et al.*, 2000a; *Schaefer et al.*, 2000; *Stuart et al.*, 2000, 2003; *Peate and Stecher*, 2003; *Peate et al.*, 2003; *Yaxley et al.*, 2004]. *Yaxley et al.* [2004] suggest that assimilation of continental crust had a marked effect on the compositions of matrix glasses and melt inclusions from at least some of the lavas exposed on Baffin Island (although they did not study the same samples examined here), and much of the Baffin Island sequence did indeed erupt through and onto early Proterozoic gneissic and migmatitic basement rocks [*Jackson and Taylor*, 1972; *Robillard et al.*, 1992].

[26] Despite this, *Robillard et al.* [1992] argued that assimilation of continental crust appears to be minor for the lava suites studied herein and, importantly, that contamination was unable account for the compositional differences between E- and N-type lavas. Our data support this conclusion. Olivines from E- and N-type lavas, with one exception, lie within the narrow range in $\delta^{18}\text{O}$ range expected for olivine in equilibrium with melts derived from normal MORB-source mantle peridotite [*Eiler*, 2001]. Addition of more than a few percent of continental crust (with $\delta^{18}\text{O}$ of $\geq \sim 10\text{‰}$) would move $\delta^{18}\text{O}_{\text{olivine}}$ outside this range (Figure 2a). E-type lavas also tend to have slightly lower $\delta^{18}\text{O}_{\text{olivine}}$ values than N-type lavas (Figures 2a and 6), which is inconsistent with addition of high $\delta^{18}\text{O}$ crust to produce the more incompatible element enriched compositions evident in E-type lavas.

[27] K/Nb ratios of matrix glasses, which are a sensitive indicator of addition of the crustal material to basaltic melts [e.g., *Hansen and Nielsen*, 1998; *Kent et al.*, 2002], also argue against significant assimilation of crustal material. Samples from the Vaigat Formation from West Greenland (Figure 4b), which includes a number of lavas that are clearly crustally contaminated [*Lightfoot et al.*, 1997; *Schaefer et al.*, 2000] show a large range of K/Nb ratios (up to ~ 1800). However, in our Baffin

Island samples, K/Nb ratios of the matrix glasses (166–207), are low relative to continental crust (~ 1100 – 2500 [*Taylor and McLennan*, 1985]), and have values similar to typical MORB [*Sun and McDonough*, 1989]. Mixing calculations show that although addition of $\sim 8\%$ of average upper or lower crust material (Figure 4b) to an N-type melt would increase the $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio to 0.704, equivalent to that of the most enriched E-type melts, this would be accompanied by an increase in K/Nb to ~ 350 (lower crust) or 800 (upper crust), well outside the range of values measured in E- and N-type lavas.

[28] Another possibility is that the compositional differences between E- and N-type lavas result from interaction between asthenosphere-derived melts and SCLM material during magma ascent. Although melting to produce the primary melts for Baffin Island and West Greenland lavas is considered to have occurred within the asthenospheric mantle beneath regions of thinning lithosphere [*Francis*, 1985; *Gill et al.*, 1992, 1995; *Larsen et al.*, 1992; *Herzberg and O'Hara*, 2002], these melts could also have interacted with the overlying lithosphere during magma ascent. Several studies have shown that incorporation of ancient lithosphere components can alter the Os isotope signature of basaltic melts [e.g., *Ellam et al.*, 1992; *Horan et al.*, 1995; *Widom et al.*, 1999; *Peate et al.*, 2003]; however, at Baffin Island the observed trend from high $^{187}\text{Os}/^{188}\text{Os}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ in Baffin Island E-type lavas to low $^{187}\text{Os}/^{188}\text{Os}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ and high $^{143}\text{Nd}/^{144}\text{Nd}$ in N-type lavas (Figure 2) is difficult to reconcile with lithospheric contamination. Ancient SCLM is generally characterized by subchondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios, as a result of their being residues of ancient partial melting events, and also by relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$, reflecting long-term enrichment in Rb/Sr and LREE due to metasomatism that appears to have more strongly affected the most melt-depleted (and therefore lowest $^{187}\text{Os}/^{188}\text{Os}$) parts of the SCLM [*Ellam et al.*, 1992; *Olive et al.*, 1997]. This trend is evident in many ancient cratonic mantle xenolith suites [e.g., *Walker et al.*, 1989; *Carlson and Irving*, 1994; *Pearson et al.*, 1995a, 1995b; *Olive et al.*, 1997] and CFB lavas from the Karoo and Siberian Traps that show similar relations have also been inferred to be contaminated by SCLM material [*Ellam et al.*, 1992; *Horan et al.*, 1995]. Furthermore, the alkalic lavas of the Manitlat Member of the Vaigat Formation of West Greenland, which are thought to represent melts of the SCLM in this region (and thus can

be inferred to sample lithosphere of similar composition to that through which Baffin Island melts must also have passed) range to very low $^{187}\text{Os}/^{188}\text{Os}_i$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios (shown as γ_{Os} as low as -16 and ϵ_{Nd} of ~ -4 in Figure 5a) [Larsen *et al.*, 2003]. Thus the trend from high $^{187}\text{Os}/^{188}\text{Os}_i$ and $^{87}\text{Sr}/^{86}\text{Sr}_i$ and low $^{143}\text{Nd}/^{144}\text{Nd}_i$ in Baffin Island E-type lavas to low $^{187}\text{Os}/^{188}\text{Os}_i$ and $^{87}\text{Sr}/^{86}\text{Sr}_i$ and high $^{143}\text{Nd}/^{144}\text{Nd}_i$ in N-type lavas is difficult to produce by mixing either E- or N-type primary melt with SCLM material. This is demonstrated in Figure 5a, in which we used estimates of the composition of the SCLM and of partial melts of this mantle component based on Maniitlat Member lava compositions to calculate mixing models between the SCLM and the sources of Baffin Island lavas. Such models cannot explain the Baffin Island trends, assuming that the Maniitlat Member lavas provide an accurate estimate of the SCLM composition. The trace element patterns of N-type lavas also argue against addition of melts from the lithosphere (to produce low $^{187}\text{Os}/^{188}\text{Os}$ ratios in N-type melts), as the Maniitlat Member lavas show that lithospheric melts in this region have trace element patterns dominated by melting of amphibole + apatite \pm phlogopite [Larsen *et al.*, 2003]. The low trace element abundances in N-type melts would make them particularly susceptible to addition of these trace element-rich melts.

[29] We also note that although deformed forsterite-rich ($\text{Fo}_{90}\text{-Fo}_{92}$) olivines in Baffin Island lavas have been interpreted as xenocrysts by Francis [1985] and thus could represent disaggregated lithospheric material [Widom *et al.*, 1999; Peate *et al.*, 2003], these olivines contain melt inclusions with compositions indistinguishable from those in the unshered and less magnesian olivines ($\text{Fo}_{89}\text{-Fo}_{85}$) that dominate the phenocryst populations (Figure 5c). This suggests that these olivines are simply derived from deformed cumulates from the same magmatic systems responsible for the erupted magmas, as is generally believed to be the case for many forsterite-rich and deformed olivines from Hawaiian lavas [e.g., Helz, 1987].

4.2.2. Heterogeneity Within Sublithospheric Mantle Sources

[30] Given that the chemical and isotopic trends in Baffin Island lavas are inconsistent with assimilation of known continental crust and/or SCLM materials, we suggest that the trend defined by E- and N-type lava compositions represents mixing between chemically and isotopically enriched and

depleted components located in mantle source regions beneath the continental lithosphere. Overall, the minimal crustal and SCLM contamination and the simple trends between isotopic and trace element composition shown by Baffin Island lavas provide a relatively clean case for discriminating the nature and origin of the mantle sources involved in the initial stages of the NAIP.

[31] The simplest interpretation of the correlations between Sr, Nd and Os isotope ratios and between these and trace elements compositions (Figures 2 and 3) is that the Baffin Island lavas contain contributions from both “depleted” (low $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$, high $^{143}\text{Nd}/^{144}\text{Nd}$, and incompatible element depleted) and “enriched” (high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$, low $^{143}\text{Nd}/^{144}\text{Nd}$, and incompatible element-enriched) sublithospheric mantle sources. In the following sections we choose the simplest interpretation, which is that individual lavas represents mixtures of melts derived from the two different sources, but we note that they could also represent melting of mixtures of enriched and depleted mantle material. Regardless, the interbedding of E- and N-type lavas within the Baffin Island sequence [Robillard *et al.*, 1992] suggests that both depleted and enriched source materials existed in close proximity in the mantle regions where these lavas were generated.

[32] Our data also suggest that the N-type source component may have been depleted in FeO_T and Na_2O and enriched in CaO and Al_2O_3 relative to the E-type source component (e.g., Figure 2). However, precise definition of the differences in major element composition of the mantle sources is not straightforward for melts formed during polybaric mantle melting, where potential temperature, melting mechanisms, and melt separation parameters can strongly control melt major element compositions, and major element compositions of melts are unlikely to directly relate to their source compositions [e.g., Hirschmann *et al.*, 1999; Herzberg and O’Hara, 2002]. For this reason we have concentrated on using the isotopic and trace element compositions to constrain the ultimate sources of the components that contribute to E- and N-type lavas.

4.2.2.1. Origin of the Depleted Component in Baffin Island Lavas

[33] Previous studies of the North Atlantic region and existing ideas about the fluid dynamics of upwelling mantle plumes suggest two potential sources for the chemically and isotopically depleted

mantle material sampled by NAIP and Icelandic lavas. One is depleted material from the upper mantle (referred to here as depleted upper mantle and akin, but not necessarily identical, to that which melts to form MORB) entrained by an ascending mantle plume [Campbell and Griffiths, 1990; Robillard et al., 1992; Fitton et al., 1997; Kent and Fitton, 2000; Stuart et al., 2003]. Alternatively, depleted material, generally considered to be portions of subducted oceanic lithosphere, may also be transported within the plume itself (referred to here as depleted plume material) [Thirwall, 1995; Hards et al., 1995; Chauvel and Hémond, 2000; Kempton et al., 2000; Skovgaard et al., 2001; Fitton et al., 2003; Thirwall et al., 2004]. Distinction between these two potential sources is important for understanding the dynamics of plume upwelling and transport [e.g., Kempton et al., 2000] and for Baffin Island, is also critical for constraining the origin of the unusual high $^3\text{He}/^4\text{He}$ signature in depleted lavas [Stuart et al., 2003].

[34] Previous workers have argued that depleted lavas from Baffin Island largely sample the depleted upper mantle [Robillard et al., 1992; Stuart et al., 2003]. In particular, Stuart et al. [2003] suggested that Sr and Nd isotope compositions close to North Atlantic MORB and negative values of ΔNb in depleted Baffin Island lavas (Table 1) show that these melts largely derived from depleted upper mantle (negative ΔNb shows that these samples fall on the array defined by North Atlantic MORB on plots of $\log[\text{Nb}/\text{Y}]$ versus $\log[\text{Zr}/\text{Y}]$, and are distinct from a separate array defined by Iceland samples [Fitton et al., 1997, 2003]).

[35] In the following section we use our new data to further test this hypothesis. Osmium and O isotope data also provide a potential means to distinguish between depleted upper mantle and depleted plume sources, as hydrothermal alteration and long term melt depletion within the recycled oceanic lithosphere (the most likely candidate for depleted plume components), is expected to produce Os and O isotope compositions that are distinctive from the depleted upper mantle [e.g., Lassiter and Hauri, 1998; Skovgaard et al., 2001; Schaefer et al., 2002].

[36] Simple calculations show that both the Sr and Nd isotope compositions and the trace element abundances of Baffin Island lavas are consistent with these representing mixtures of melts derived from a depleted upper mantle source, and those derived from an enriched source. The array defined by $^{87}\text{Sr}/^{86}\text{Sr}$; and $^{143}\text{Nd}/^{144}\text{Nd}$; (Figure 4a) projects

toward the estimated “plume free” composition of North Atlantic depleted upper mantle at 62 Ma [Taylor et al., 1997], and a model for mixing of melts derived from this source and the most enriched of our E-type lavas (see figure caption for details) suggest that N-type melts are the product of mixing between melts of a depleted upper mantle source and $\sim 20\text{--}30\%$ enriched melt (of the same composition as the most enriched Baffin Island lava). As the E-type end-member is probably more enriched than the most enriched E-type basalt, the amount of enriched melt required to mix with depleted melts to produce N-type lava compositions is considered a maximum.

[37] This model can also reproduce the trace element compositions of N-type lavas, where a melt derived from 15% batch melting of a depleted mantle spinel lherzolite source [Francis, 1985] is mixed with 20–30% of a primary melt based on the most enriched E-type lava composition (Figure 1c). Melt mixing is also consistent with the observed compositions of olivine-hosted melt inclusions in Baffin Island lavas [Kent et al., 1998; A. J. R. Kent et al., manuscript in preparation, 2004]. Inclusions hosted in olivine from both E- and N-type lavas range to more depleted compositions than N-type lavas, with $\text{K}_2\text{O}/\text{TiO}_2$ and $[\text{La}/\text{Sm}]_{\text{N}}$ ratios as low as 0.01 and 0.39 respectively (e.g., Figure 5c), suggesting that melts more depleted than N-type lavas were indeed present within the Baffin Island melting system. If the correlations between Nd and Sr isotope ratios and $\text{K}_2\text{O}/\text{TiO}_2$ and $[\text{La}/\text{Sm}]_{\text{N}}$ in matrix glasses (Figures 3c, 3e, 3h, and 3j) are used to estimate the isotope composition of the most depleted melt inclusions, this equates to isotopic compositions which approach those estimated for depleted upper mantle at 62 Ma, although these still require mixing with $\sim 10\text{--}20\%$ of enriched melt (Figure 4).

[38] Oxygen isotope compositions of olivine phenocrysts from Baffin Island N-type lavas, all lie within the range expected for olivine formed from melts formed in equilibrium with a N-MORB source (Figure 2) [Eiler, 2001], again consistent with derivation of the melts from which olivine crystallized from depleted upper mantle. Importantly, normal mantle-like $\delta^{18}\text{O}_{\text{olivine}}$ values also argue against significant contributions from recycled hydrothermally altered oceanic lithosphere or crust within a depleted plume component, which are expected to have oxygen isotope compositions outside the typical mantle range [e.g., Gregory

and Taylor, 1981; Eiler, 2001]. In particular, the $\delta^{18}\text{O}_{\text{olivine}}$ values of N-type lavas limits contributions from high-temperature hydrothermally altered portions of the depleted oceanic lithosphere (with $\delta^{18}\text{O}$ of +3) to at least less than 10 wt.% (Figure 2a). We note that $\delta^{18}\text{O}_{\text{olivine}}$ values below those of typical MORB values have been reported from lavas from Iceland and Hawaii and have been argued to derive from recycled high-temperature-altered oceanic lithosphere or crust within upwelling mantle plume sources [Lassiter and Hauri, 1998; Skovgaard et al., 2001]. However, other studies suggest that low $\delta^{18}\text{O}$ may instead reflect interaction between mantle-derived melts and hydrothermally altered crust in shallow magmatic environments [e.g., Gee et al., 1998; Eiler et al., 2000a; Wang et al., 2003], and at present the origin of the low $\delta^{18}\text{O}$ signal in ocean island basalt suites remains uncertain.

[39] Low $^{187}\text{Os}/^{188}\text{Os}$ ratios have been argued to reflect long-term melt depletion in recycled oceanic lithosphere sources within mantle plumes [e.g., Pearson et al., 1995a; Lassiter and Hauri, 1998; Skovgaard et al., 2001; Schaefer et al., 2002], and Baffin Island N-type lavas have $^{187}\text{Os}/^{188}\text{Os}_i$ ratios that are distinctly lower than the estimated composition of the primitive mantle at the time of eruption ($^{187}\text{Os}/^{188}\text{Os}_{\text{PM}} = 0.1286$, assuming a modern primitive mantle composition of 0.1296 and Re/Os ratio of 0.079 [Schaefer et al., 2000; Meisel et al., 2001]). However, the key is to be able to determine whether the $^{187}\text{Os}/^{188}\text{Os}$ ratio estimated for the depleted source can be distinguished from depleted upper mantle, which is also likely to have $^{187}\text{Os}/^{188}\text{Os}$ ratios lower than that of the primitive mantle [e.g., Schaefer et al., 2000, 2002].

[40] To make such a comparison, we must first estimate the Os isotope composition of the depleted material sampled by Baffin Island N-type lavas. If we initially assume that the depleted reservoir sampled by Baffin Island lavas has the same Sr and Nd and trace element composition as the depleted upper mantle at 62 Ma (as argued above), we can estimate the Os isotope composition of this end-member using best fit straight lines for the correlations we observe between $^{187}\text{Os}/^{188}\text{Os}_i$ and $^{87}\text{Sr}/^{86}\text{Sr}_i$, $^{143}\text{Nd}/^{144}\text{Nd}_i$, $[\text{La}/\text{Sm}]_N$ and $\text{K}_2\text{O}/\text{TiO}_2$ (using the latter two as indices of the variations in enrichment in incompatible elements) in Baffin Island lavas (Figures 2 and 3), coupled with the $^{87}\text{Sr}/^{86}\text{Sr}_i$, $^{143}\text{Nd}/^{144}\text{Nd}_i$ and $[\text{La}/\text{Sm}]_N$ and $\text{K}_2\text{O}/\text{TiO}_2$ ratios estimated for the depleted

end-member (see Figure 6 caption for further details). This suggests that the depleted source had $^{187}\text{Os}/^{188}\text{Os}_i$ at 62 Ma between 0.1174 and 0.1205 (γOs_i between -7.2 and -5.5). Thus, if Baffin Island lavas do sample a depleted upper mantle reservoir, this must have had significantly lower $^{187}\text{Os}/^{188}\text{Os}$ (and γOs) than the primitive mantle at 62 Ma. We note that we could also get broadly similar estimates for the Os isotope composition of the Baffin Island depleted component from the intercept of arrays defined by Baffin Island and MORB samples in Os versus Nd (Figure 6a) and Os versus Sr isotope space, although recent results shed some uncertainty on the significance of the extension of the MORB array to highly radiogenic $^{187}\text{Os}/^{188}\text{Os}$ isotope compositions (high γOs) [Schiano et al., 1997; Gannoun et al., 2004]. In Figure 6b (using γOs to minimize age-related differences between Baffin Island and younger samples) we have compared our estimates of the composition of the Baffin Island depleted component to published compositions of MORB and abyssal peridotites, which should both sample the depleted upper mantle [e.g., Dick et al., 1984; Snow and Reisberg, 1995; Brandon et al., 2000]. Both MORB and abyssal peridotites range to high γOs values (these may relate to recent ingrowth, seawater alteration and/or assimilation of altered material [e.g., Snow and Reisberg, 1995; Standish et al., 2002; Gannoun et al., 2004]), but abyssal peridotites also extend to significantly lower γOs than MORB (including those estimated from the “isochron” approach of Gannoun et al. [2004]). The γOs estimated for the Baffin Island depleted component are lower than the lowest values MORB, but are within the uncertainty for the lowest MORB initial compositions reported by Gannoun et al. [2004] and also overlap with the lowest γOs from abyssal peridotites (Figure 6a). If the lowest compositions measured for MORB glasses are representative of the depleted upper mantle then Baffin Island lavas sample a depleted reservoir with lower γOs . However, if the Os isotope compositions of abyssal peridotites represent the depleted upper oceanic mantle composition, then the depleted component sampled by Baffin Island lavas overlaps that of the depleted upper mantle. Overall, we believe that at present we cannot distinguish between the estimated Os isotope composition of the Baffin Island depleted component and that of the depleted upper mantle. Clarification of this will rely on further investigation of the relationship between Os isotope compositions of abyssal peridotites and MORB

and better definition of depleted upper mantle Os compositions.

[41] From the above we conclude that the Sr-Nd-Os-O isotope and trace element compositions of Baffin Island N-type lavas are consistent with the depleted source sampled by Baffin Island melts being depleted upper mantle material. However, one remaining problem with this interpretation is the high $^3\text{He}/^4\text{He}$ ratios (up to $49.5 R_A$) measured in Baffin Island lavas [Stuart *et al.*, 2003]. The depleted upper mantle is generally considered to have the same $^3\text{He}/^4\text{He}$ ratio as MORB, which is currently $\sim 8 \pm 2 R_A$ [Graham, 2002]. Further, the correlation between low $^{87}\text{Sr}/^{86}\text{Sr}$, high $^{143}\text{Nd}/^{144}\text{Nd}$, low $[\text{La}/\text{Sm}]_N$ and high $^3\text{He}/^4\text{He}$ suggest that it is the depleted lavas that sample the highest $^3\text{He}/^4\text{He}$ mantle domains [Stuart *et al.*, 2003]. Although we have not analyzed He isotope composition of the samples from this study, if the correlations between $[\text{La}/\text{Sm}]_N$ and $^{143}\text{Nd}/^{144}\text{Nd}$, and $^3\text{He}/^4\text{He}$ reported by Stuart *et al.* [2003] also apply to our samples then they suggest that N type lavas would have $^3\text{He}/^4\text{He}$ of $\sim 40 R_A$ (and for E-type $\sim 15\text{--}30 R_A$). Stuart *et al.* [2003] suggest that the apparent paradox of high $^3\text{He}/^4\text{He}$ ratios in melts sourced largely from the depleted mantle may be resolved if depleted upper mantle material in the sources of Baffin Island lavas mixes, prior to melting, with a small amount of volatile-rich, ^3He -bearing, lower mantle material, although this process is at present poorly constrained. If such a flux did occur, then our data suggest that this flux did not significantly effect the Sr-Nd-Os-O and lithophile trace element composition of the depleted upper mantle reservoir involved.

4.2.2.2. Origin of the Enriched Component in Baffin Island Lavas

[42] Isotopically and chemically enriched compositions are common features of many mantle-plume-derived lava suites [e.g., Hofmann, 1997], and contributions from recycled ocean crustal components (primarily pelagic sediments and/or basaltic crust) are widely invoked to account for these features [Zindler and Hart, 1986; Hofmann, 1988, 1997; Hauri and Hart, 1993]. Although Baffin Island E-type lavas have distinctly higher $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ than N-type lavas, they do not extend to the high Os and Sr isotope compositions found in some OIB suites, particularly many with the so-called “HIMU” and “EM” signatures [Zindler and Hart, 1986; Hofmann, 1997; Hauri and Hart, 1993]. E-type lavas also have $\delta^{18}\text{O}_{\text{olivine}}$ that range

from similar to, to just slightly lower, than those of N-type lavas (Figures 2a and 6). This limits the amount of recycled pelagic sediment ($\delta^{18}\text{O} \sim 20\%$) to negligible levels, at most less than $\sim 1\%$, and for low-temperature altered oceanic crust material ($\delta^{18}\text{O} \sim 9\%$) to less than 2–3% (assuming that the difference between the E-type lava with the highest $\delta^{18}\text{O}_{\text{olivine}}$ and the N-type lava with the lowest $\delta^{18}\text{O}_{\text{olivine}}$ is due to addition of high- $\delta^{18}\text{O}$ material; Figure 2a). Simple batch melting models suggest that addition of such small amounts of these materials to a depleted mantle source followed by 15% melting [Francis, 1985] will not reproduce the Nb-Ta contents and LREE characteristic of E-type lavas (Figure 1c).

[43] Another potential source for the enriched component in Baffin Island lavas is the deeper, peridotitic parts of the recycled oceanic lithosphere that have experienced high-temperature hydrothermal alteration. This could explain the slightly lower $\delta^{18}\text{O}_{\text{olivine}}$ in E-type lavas, as studies of ophiolites show that deeper portions of the oceanic lithosphere that have been hydrothermally altered at relatively high temperatures can have low $\delta^{18}\text{O}$ ($\sim +3$ [Gregory and Taylor, 1981; Cocker *et al.*, 1982; Eiler, 2001]), and in this case the maximum difference between E- and N-type lavas in $\delta^{18}\text{O}_{\text{olivine}}$ is consistent with addition of up to $\sim 10\text{--}15\%$ of altered oceanic lithosphere with $\delta^{18}\text{O}$ of +3. However, such material is also the residue remaining after MORB melt extraction, and by itself would be unlikely to be associated with radiogenic isotope compositions and enriched chemical signatures that characterize E-type lavas. Enriched compositions might be generated by melting of mixtures of high-temperature-altered peridotite, and relatively small degree melts generated during subduction that remain trapped within the oceanic lithosphere [Niu *et al.*, 2002; Niu and O’Hara, 2003]. Such a process should have little impact on oxygen isotope compositions [Eiler *et al.*, 2000b], but will strongly alter incompatible trace element abundances. It is possible to generate trace element compositions close to E-type lavas from batch melting of a depleted mantle composition spinel lherzolite source containing a significant proportion (up to 7 wt.%) of an enriched melt (Figure 1c), although other combinations of these parameters are also possible. The best fit is obtained when the trapped melt is derived from a primitive mantle source, and not from depleted mantle, as might be more generally expected for metasomatic melts generated within oceanic lithosphere during subduction (Figure 1). The isotope signatures of an enriched source produced in this

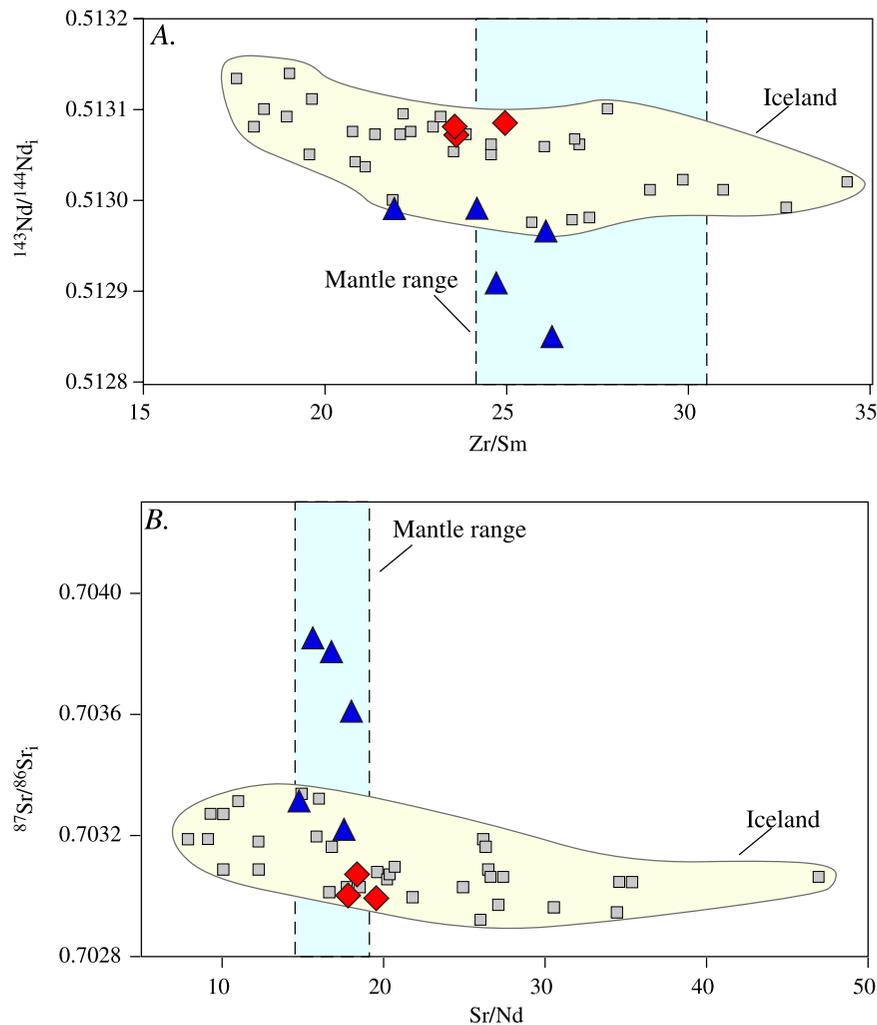


Figure 7. Plots of (a) Zr/Sm versus $^{143}\text{Nd}/^{144}\text{Nd}$ and (b) Sm/Nd versus $^{87}\text{Sr}/^{86}\text{Sr}$ for lavas from Baffin Island and Iceland. Iceland data are taken from Chauvel and Hémond [2000] and Kempton *et al.* [2000]. Symbols for Baffin Island lavas are as for Figure 1. Blue areas labeled “mantle range” in both figures show the range of Sr/Nd and Zr/Sm ratios produced by 15% batch melting of depleted and primitive mantle composition spinel lherzolites.

manner is difficult to constrain precisely, as the initial isotopic composition and age of the metasomatizing melt is unknown. However, in general, addition of incompatible element enriched metasomatic melt would result in increased Rb/Sr and Re/Os and decreased Sm/Nd , and with time such a source would evolve to higher Sr and Os and lower Nd isotope ratios.

4.2.2.3. Implications for NAIP Mantle Sources

[44] Recognition of components having long-lived and compositionally distinct components within the sources of NAIP lavas has important implications for the chemical structure and dynamics of the Iceland mantle plume through time. Over the last decade there has been considerable debate about the origin and evolution of mantle sources within the

NAIP and, in particular, within Icelandic lavas [Thirwall, 1995; Hards *et al.*, 1995; Fitton *et al.*, 1997, 2003; Saunders *et al.*, 1997; Taylor *et al.*, 1997; Hanan *et al.*, 2000; Chauvel and Hémond, 2000; Ellam and Stuart, 2000; Kempton *et al.*, 2000; Kent and Fitton, 2000; Peate *et al.*, 2003; Peate and Stecher, 2003; Thirwall *et al.*, 2004]. Although it has been clearly demonstrated that the mantle sources for NAIP lavas have varied substantially over space and time since the inception of NAIP magmatism at ~ 62 Ma [Saunders *et al.*, 1997; Fitton *et al.*, 1997; Kempton *et al.*, 2000], many workers have sought to connect components recognized in Icelandic lavas to those identified within the older NAIP magmas [e.g., Thirwall, 1995; Fitton *et al.*, 1997; Saunders *et al.*, 1997; Ellam and Stuart, 2000; Kempton *et al.*, 2000; Kent

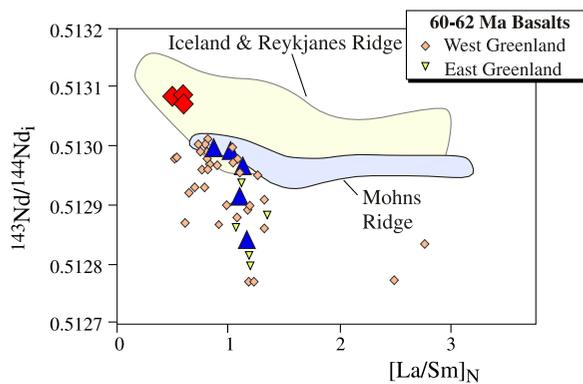


Figure 8. $[La/Sm]_N$ versus $^{143}Nd/^{144}Nd_i$ for Baffin Island lavas and 62–60 Ma basalts from East and West Greenland. Fields showing the compositional range of basalts from Iceland and the Reykjanes Ridge and Mohns Ridge are also shown for comparison (data from Mertz and Haase [1997], Chauvel and Hémond [2000], and Kempton *et al.* [2000]). Symbols are as per the legend and Figure 1. Data are also shown for lavas from the Lower Basalt sequence of East Greenland [Hansen and Nielsen, 1998] and the Vaigat formation of West Greenland [Holm *et al.*, 1993; Lightfoot *et al.*, 1997]. Both East and West Greenland data sets have been filtered to minimize the effects of crustal contamination by removing samples with $La/Nb > 1$.

and Fitton, 2000; Bernstein *et al.*, 2001; Peate *et al.*, 2003; Peate and Stecher, 2003].

[45] Our data show that there are distinct differences in the chemical and isotopic compositions of enriched and depleted lavas erupted at Baffin Island and those from Iceland. This suggests that there are also differences in the composition of mantle sources sampled at the inception of NAIP magmatism and those sampled by the modern day plume. Depleted N-type lavas from Baffin Island have lower $^{187}Os/^{188}Os_i$ ratios and γOs_i values than the lowest yet measured for Iceland (Figure 5). Depleted Icelandic lavas also have lower $\delta^{18}O_{olivine}$ than Baffin Island lavas (Figure 6) [Skovgaard *et al.*, 2001], although the origin of the low $\delta^{18}O$ signal remains unclear (see above). Baffin Island lavas also do not exhibit the distinctive large variations in Sr/REE, Pb/Ce and Zr/Sm evident in Iceland. As shown in Figure 7, Baffin Island lavas have relatively restricted range of Sr/Nd and Zr/Sm ratios, broadly compatible with ratios expected from large degree melts of depleted or primitive composition mantle, and they do not extend to the high and low values shown by Icelandic rocks even though E-type lavas have significantly higher $^{143}Nd/^{144}Nd_i$ and higher

$^{87}Sr/^{86}Sr_i$ than the most enriched Iceland samples. Finally, Baffin Island lavas also have characteristically low $^{143}Nd/^{144}Nd_i$ at a given $[La/Sm]_N$ (Figure 8; they also have higher $^{87}Sr/^{86}Sr_i$ although this is not shown in the figure).

[46] Although differences in the melting processes are likely to influence trace element signatures, they are unlikely to produce the large differences evident in ratios of elements (Zr/Sm, Nd/Sm, Pb/Ce) with similar incompatibility during mantle melting. Chauvel and Hémond [2000] argue that the distinctive Sr/Nd and Zr/Sm ratios of picritic (high Sr/Nd, low Zr/Sm) and alkalic (low Sr/Nd and high Zr/Sm) lavas from Iceland reflect contributions from melting of recycled oceanic crust gabbroic (for picrites) and basaltic components (for alkali basalts), and Chauvel and Hémond [2000] and Skovgaard *et al.* [2001] argue that these are long-lived ($> \sim 2.5$ Ga) compositional features of the mantle source for Icelandic lavas. Alternatively, Thirwall *et al.* [2004] instead suggest that represent relatively young ($< \sim 170$ Ma) features, possibly related to residues from earlier partial melting events. Regardless of their exact origin, the compositional differences between Baffin Island lavas and Iceland suggest that both the enriched and depleted sources sampled by Baffin Island lavas are compositionally different from those sampled by the modern Iceland plume. Inferred differences in the nature of the depleted source are also consistent with suggestions that lavas from Iceland sample depleted material from within the plume, rather than the surrounding depleted upper mantle [Thirwall, 1995; Hards *et al.*, 1995; Chauvel and Hémond, 2000; Kempton *et al.*, 2000; Skovgaard *et al.*, 2001; Fitton *et al.*, 2003; Thirwall *et al.*, 2004].

[47] The enriched component we have identified at Baffin Island may also have been sampled by lavas erupted in East and West Greenland lavas during the initial stages of NAIP magmatism. Lavas of ~ 60 –62 Ma age from West and East Greenland also show the characteristically low $^{143}Nd/^{144}Nd_i$ and high $^{87}Sr/^{86}Sr_i$ at a given $[La/Sm]_N$ ratio that distinguishes enriched lavas from Baffin Island and Iceland (Figure 8). These similarities suggest that the enriched mantle source sampled at Baffin Island may have been sampled by lavas erupted over a large geographic range, and if so, represented a volumetrically significant mantle source. At present it is uncertain whether this component persisted beyond the time of continental breakup and voluminous volcanism at ~ 56 Ma. Peate *et*

Table A1. SIMS Measurements of the Trace Element Compositions of Baffin Island Matrix Glasses

	10004	10013	10014	10019	10021	10024	10029	10064	BVHO-1G ^a	±1σ ^a
	<i>Measured Concentrations, ppm</i>									
Sr	157	140	180	135	162	153	154	161	390	12
Y	26	25	28	24	26	26	24	24	27	0.3
Zr	74	64	94	63	79	72	71	73	180	8.1
Nb	6.5	2.2	8.3	2.2	7.9	7.5	3	4.8	19.5	2
Ba	63	16	75	15	75	80	20	45	133	10
La	5.6	2.8	7.2	2.7	6.3	5.4	3.4	4.4	15.5	0.8
Ce	13.8	8.4	16.9	7.9	13.7	13.2	9.5	11.7	38	2.1
Pr	2.0	1.5	2.6	1.5	2.1	1.9	1.6	1.8	5.5	0.3
Nd	9.5	7.6	12.9	7.6	10.2	9.0	8.7	9.4	24.7	1.4
Sm	3.0	2.8	3.6	2.6	3.3	3.0	3.1	2.9	6.2	0.4
Eu	1.1	1.1	1.3	1.0	1.2	1.2	1.2	1.2	2.1	0.1
Dy	4.6	4.1	5.2	4.3	4.8	4.6	4.4	4.4	5.3	0.4
Er	2.8	2.5	3.2	2.6	3.1	2.9	2.7	2.6	2.6	0.2
Yb	2.0	1.9	2.3	1.9	2.3	2.2	2.1	2.0	2.0	0.2
	<i>SIMS/LA^b</i>									
Sr	1.06	1.07	1.03	1.04	1.08	0.98	1.09	1.08		
Y	0.98	1.00	0.98	1.01	1.01	1.10	1.16	1.08		
Zr	1.01	1.02	1.03	1.04	1.09	1.10	1.21	1.11		
Nb	1.05	1.06	1.09	1.05	1.24	1.12	1.20	1.06		
Ba	1.04	1.06	1.12	1.03	1.21	1.01	1.13	1.08		
La	1.03	1.02	1.08	1.07	1.18	1.04	1.24	1.06		
Ce	1.14	1.07	1.09	1.09	1.11	1.02	1.11	1.08		
Pr	1.06	1.03	1.06	1.14	1.12	1.02	1.12	1.07		
Nd	0.98	1.03	1.07	1.06	1.11	0.99	1.19	1.12		
Sm	1.08	1.11	1.02	1.01	1.13	1.00	1.25	1.07		
Eu	1.00	1.10	1.01	1.00	1.07	1.03	1.20	1.13		
Dy	1.02	0.96	1.04	1.03	1.05	1.06	1.22	1.10		
Er	0.99	0.96	1.04	1.01	1.10	1.16	1.26	1.11		
Yb	0.76	0.81	0.83	0.79	0.94	0.92	0.98	0.88		

^a Average and standard deviation from 5 separate measurements made over the analysis session.

^b Represents the ratio of elemental concentration measured by SIMS over that measured by LA-ICP-MS.

al. [2003] and *Bernstein et al.* [2001] have argued that the Iceland plume has sampled similar composition enriched sources since breakup, and if this is the case then the clear differences between Icelandic enriched lavas and Baffin Island E-type lavas suggests that the enriched mantle sources sampled at the inception of magmatism differ from those sampled after continental breakup. Such a change may have coincided with a marked increase in melt productivity at this time [*Peate and Stecher*, 2003].

5. Conclusions

[48] 1. Picrites from Baffin Island, Canada, that erupted during the initial phases of formation of the North Atlantic Tertiary Province show a range of isotopic and chemical compositions, and correlations between Sr-Nd-Os isotope compositions, indices of trace element enrichment, and major element oxides. These compositions are best explained by mixing between melts

derived from distinct depleted (i.e., low $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$, high $^{143}\text{Nd}/^{144}\text{Nd}$, and lower concentrations of highly incompatible elements relative to primitive mantle) and enriched (high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$, low $^{143}\text{Nd}/^{144}\text{Nd}$, and more elevated concentrations of highly incompatible elements) mantle components. All lavas represent plausible admixtures of melts derived from these two sources, and contributions from crust and SCLM appear minor.

[49] 2. The Sr-Nd-Os-O isotope and trace element composition of the depleted end-member appears similar to that projected for the ambient North Atlantic depleted upper mantle at the time of eruption. Osmium isotope compositions are lower than primitive mantle, but overlap with the lowest compositions estimated for MORB from internal “isochrons” and those measured for abyssal peridotites and thus cannot at present be distinguished from depleted upper mantle. High $^3\text{He}/^4\text{He}$ in Baffin Island depleted lavas may reflect recharge of depleted upper mantle by ^3He -rich flux from

the lower mantle. Oxygen isotopes suggest that contributions from depleted lithosphere recycled within the proto Iceland mantle plume appear minor.

[50] 3. The origin of the enriched mantle component is less clear. Lower $\delta^{18}\text{O}_{\text{olivine}}$ measured in enriched lavas rule out substantial contributions from recycled pelagic sediments or low-temperature altered basaltic crust. The enriched source may be plausibly derived from high-temperature-altered components of recycled oceanic lithosphere within the upwelling plume, if this also contained a small proportion of melt derived from a more enriched source.

[51] 4. Compositional differences between Baffin Island and modern Icelandic lavas suggest that both depleted and enriched mantle sources sampled by Baffin Island lavas differ from those tapped by recent Icelandic lavas, consistent with changes in both depleted and enriched sources for the NAIP magmas since the inception of magmatism at ~ 62 Ma. The enriched source for Baffin Island E-type lavas may also have been sampled more widely by 60–62 Ma lavas from East and West Greenland.

Appendix A: SIMS Trace Element Analysis of Matrix Glasses

[52] Trace element concentrations of matrix glass chips from Baffin Island measured by SIMS are given in Table A1 and are presented for comparison with LA-ICP-MS data given above. Analyses were conducted at Lawrence Livermore National Laboratory using techniques outlined by Kent and Elliott [2002]. SIMS and LA-ICP-MS measurements were made on the same set of hand-picked glass chips. For SIMS analysis NIST 612 glass was used as a primary calibration standard and analysis of BHVO-1G basalt glass standard (an in-house standard made from the USGS BHVO-1 rock powder standard) used to monitor accuracy and precision are also given.

[53] SIMS and LA-ICP-MS measurements typically agree extremely well, generally to $\ll 10\%$. Larger variations are seen for some elements, particularly those present at low abundances, and are generally still within analytical precision. N-type samples, particularly 10029 also exhibit larger deviations for some elements, possible reflecting increased uncertainties relating to exceptionally low abundances of some elements

in these depleted glasses. For Yb low SIMS-measured concentrations are also evident in some samples.

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