Contrib Mineral Petrol (2002) 142: 476–484
DOI 10.1007/s004100100301

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Compositional recycling structure of an Archean super-plume: Nb–Th–U–LREE systematics of Archean komatiites and basalts revisited

Received: 5 April 2001 / Accepted: 24 July 2001 / Published online: 4 September 2001

Abstract The volcanic stage of the 2.7-Ga Abitibi greenstone belt, Canada, is dominated by bimodal arc magma series and komatiite–basalt sequences. The latter represents an aerially extensive oceanic plateau erupted from an anomalously hot super-plume. Komatiites define a linear array of Nb/Th vs. Nb/U, extending from Nb/Th = 8–20, and Nb/U = 26–58, whereas basalts plot on a separate, but overlapping, field extending to higher Th/U but lower Nb/Th values. Inter-element ratios of Th, U, Nb, and LREE of komatiites and basalts plot with Phanerozoic and modern ocean plateau basalts. Th, U, Nb, and LREE are fractionated in subduction zones into low Nb/Th, Nb/U, and Nb/LREE arc crust, and complementary high Nb/Th, Nb/U, and Nb/LREE residual slab. Accordingly, the Archean komatiite–basalt association may be explained by a plume that likely originated from the core–mantle boundary with komatiites erupted from a hot axis containing recycled oceanic crust, and basalts erupted from the plume annulus that entrained upper mantle containing recycled oceanic and continental crust. High Nb/Th and Nb/U of plume-related volcanic sequences documented in Abitibi, Yilgarn, and Baltic Archean greenstone belts suggest that the extraction and recycling of continental crust may have occurred early in the Archean.

Introduction

Hofmann and coworkers pioneered a new trace element approach for constraining the inter-related questions of the source of oceanic basalts and crust–mantle evolution. Their approach focused on the concentrations and ratios of highly incompatible elements such as Nb/U and Ce/Pb (Hofmann et al. 1986). This led initially to a two-stage model for crust–mantle evolution, and then a multistage model, in which the depleted upper mantle and continental crust were viewed as broadly complementary reservoirs (Hofmann et al. 1986; Hofmann 1997a). The highly incompatible elements, specifically Th–Nb–Ce systematics, were also used by Saunders et al. (1988) to develop a multistage model of recycling continental and oceanic crust into the mantle source of mid-ocean ridge basalts (MORB) and ocean island basalts (OIB).

Jochum et al (1991) extended this approach to the Th–Nb–La systematics of komatiites and basalts, to assess the mantle source of komatiites, and the secular evolution of mantle composition and crustal growth. From their results, Jochum et al. (1991) concluded that (1) many Archean komatiites are characterized by normalized depletions of Nb and Th relative to LREE, and accordingly were derived from a mantle source that evolved independently from the upper mantle; and (2) basalts show a secular increase in Nb/Th because of progressive depletion of the upper mantle complementary to growth of continental crust.

Analyses of 3.0- to 2.7-Ga tholeiitic basalts, spatially associated with komatiites, from greenstone belts of Western Australia, Finland, and Canada, have yielded results different from those of Jochum et al. (1991; see also Sylvester et al. 1997; Puchtel et al. 1998; Kerrich et al. 1999). There are large ranges of Nb/Th, Nb/La, and Th/La, independent of absolute concentrations, about the primitive mantle values, which are consistent with recycling of continental and oceanic crust into the mantle source of the plumes from which the komatiite–basalt sequences erupted (Kerrich et al. 1999). Given these results for basalts, new Th, U, Nb, and REE analyses are reported for associated komatiites. There are no high precision data for Th, U, and Nb at low concentrations present in Abitibi komatiites, excepting two analyses in Jochum et al. (1991). We use the data to test the composite Archean mantle plume model of

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Editorial responsibility: T.L. Grove

Campbell et al. (1989) – "Heads its basalts, tails its komatiites" – and evaluate deep versus shallow recycling into the Archean mantle.

Geological setting

Komatiites and spatially associated basalts in this study are from the southern volcanic zone (SVZ) of the Abitibi greenstone belt. A review of the geology of this belt is given by Jackson and Fyon (1992), and details of geological settings of the samples can be found in Xie et al. (1993) and Kerrich et al. (1999). A brief summary of the geological setting is presented here.

The SVZ is a composite, tectonically imbricated, terrane of plume-related komatiite–basalt sequences, and bimodal tholeiitic to calc-alkaline magmatic arc sequences with associated trench turbidites, intruded by synvolcanic granitoid plutons (Jackson and Fyon 1992). In the SVZ, where the samples were collected for this study, komatiites and spatially associated basalts extend along the strike, in a semi-continuous outcrop, for more than 300 km, from Timmins in the west to Val d'Or in the east. Volcanism in the SVZ extends from 2,750 to 2,700 Ma, with a preserved area of $\sim$200 $\times$ 500 km (Corfu 1993). Accordingly, aerially extensive komatiites and tholeiitic basalts are associated in space and time from outcrop scale exposure, to large scale map patterns, with precise age constraints (Arndt and Nesbitt 1982; Jackson and Fyon 1992; Corfu 1993; Wyman et al. 1999). Komatiite and basalt volcanism of this age originally extended over 1,000 km along the strike, west into the formerly contiguous Wawa Subprovince (Percival et al. 1994).

Based on intercalated komatiites, boninite series volcanic flows, and arc tholeiitic basalts, Wyman and coworkers interpreted this volcanic stage of the Abitibi belt as jamming of an ocean plateau into an intra-oceanic arc (Wyman 1999). The volcanic terranes were intruded by syntectonic tonalites between 2,695 to 2,675 Ma. Metamorphism was syntectonic with prevalently greenschist facies (Jackson and Fyon 1992). There is a general agreement that komatiite–basalt volcanic sequences in Archean greenstone belts, and Phanerozoic komatiites, represent liquids segregated from anomalously hot mantle plumes (Campbell et al. 1989; Kerr et al. 1996). Alternatively, Grove et al. (1997) concluded from field observations that Barberton spinifex units crystallized from cooler, hydrous liquid as sills or dykes, not flows. Based on experimental results, Parman et al. (1997) also concluded that Barberton komatiites were hydrous magmas intruded at mid-crustal levels. The arguments for wet versus dry komatiites have been reviewed by Arndt et al. (1998).

Analytical methods

Major elements were determined using X-ray fluorescence spectrometry (XRF). Trace elements were analyzed using inductively coupled plasma-mass spectrometry (ICP-MS; VG PQ3 S-Option) at the Great Lakes Institute of Environmental Research, University of Windsor. The VG PQ3 S-Option offers high sensitivity, which is required to determine Th and U precisely and accurately, as these elements are at extremely low levels in most komatiites. Accordingly, some samples previously analyzed by a Perkin–Elmer Elan 5000 ICP-MS (Xie et al. 1993) were reanalyzed using the PQ3 S-Option, in order to obtain more precise and accurate Th and U data.

Sample preparations was performed in a class-100 clean room environment. The calibration involved standard additions, internal standardization, and external calibration (Jenner et al. 1990; Longerich et al. 1990; Xie et al. 1994). The Standard Reference Material (SRM), BIR-1, and BHVO-1, are used to monitor the data quality. Over all, for element levels in the komatiites analyzed, the precision and accuracy is estimated as follows: 1–3% for REE's and Zr, and 5–7% for Nb, Hf, Th, and U, based on analysis of SRMs and sample duplicates (Table 1).

Results

Selected major and trace element data for the Abitibi komatiites are presented in Table 1. The majority of komatiites presented in this study are Al-undepleted with MgO = 20–30%, Al_2O_3/TiO_2 = 16–19% (cf. Arndt 1994), depleted LREE, but flat HREE, and variable positive Zr and Hf anomalies (Fig. 1). Al-depleted komatiites from Boston township are characterized by MgO = 22–29% and Al_2O_3/TiO_2 = 4.5–5.4, with enriched LREE but depleted HREE, and negative Zr and Hf anomalies (Fig. 1), interpreted as the presence of residual majorite garnet during deep melting of a mantle plume (Xie and Kerrich 1994). Tholeiites are spatially associated with komatiites at each location; they are characterized by MgO = 7–13%, with flat REE patterns for those associated with Al-undepleted komatiites, but positively fractionated patterns in Boston tholeiitic basalts (Xie and Kerrich 1994; Kerrich et al. 1999).

Collectively, the komatiites define a linear array on a Nb/U vs. Nb/Th plot (Fig. 2a, Table 1), extending from Nb/Th = 7–20, and Nb/U = 26–58 (Table 1, Fig. 2a; r = 0.93), with relatively lower Th/U ratios averaging 3.4 ± 0.3 (Table 1). In contrast, associated tholeiites plot in a field extending to higher Nb/U ratios than komatiites at given Nb/Th ratios, and at relatively higher Th/U ratios averaging 4.2 ± 0.5 (Fig. 2a; Kerrich et al. 1999). Abitibi komatiites and basalts plot with ocean plateau basalts such as Iceland and Ontong Java, and ocean island basalts (OIB) on incompatible element ratio diagrams, but separate from arc basalts (Fig. 2a, b). Komatiites have lower concentrations of incompatible elements than basalts, and consequently plot separately from OIB in Nb/Th vs. Nb co-ordinates (Jochum et al. 1991), as is also the case for Gorgana komatiites and basalts with a heterogeneous mantle (Kerr et al. 1996; Arndt et al. 1997).

Table 1 Major and trace elements of komatiites from the Archean Abitibi greenstone belt, Canada

Group	Renault		Malartic								Tisdale				Munro				Boston					
Sample	93-303	-304	-338A	-339	-340	-341	-342	-343	-344	-345	t-1	t-7	t-8	92-176	mp-1	mp-4	mp-6	mp-5	92-42	92-44	92-41	93-43[a]	BIR-1[a]	BIR-1[b]
SiO_2[c]	47.17	44.61	43.21	44.84	44.13	46.29	46.36	44.91	45.90	47.40	48.24	49.37	42.41	48.47	45.47	46.22	46.16	44.62	41.62	45.16	41.87	42.08		
TiO_2	0.64	0.62	0.20	0.37	0.20	0.32	0.37	0.26	0.34	0.34	0.56	0.52	0.47	0.51	0.33	0.23	0.41	0.31	0.43	0.81	0.43	0.46		
Al_2O_3	10.72	10.39	5.25	6.83	4.79	6.67	6.64	5.58	6.70	6.09	7.38	8.48	7.55	7.25	6.05	4.34	7.89	5.67	1.91	3.62	2.18	2.43		
Fe_2O_3	13.48	14.07	13.74	13.79	13.32	12.45	10.32	11.98	11.09	10.23	13.89	13.40	13.28	13.09	11.26	10.89	12.37	11.05	19.97	18.39	19.68	19.39		
MnO	0.18	0.19	0.20	0.18	0.21	0.19	0.18	0.19	0.18	0.17	0.30	0.28	0.23	0.22	0.19	0.23	0.20	0.19	0.30	0.26	0.30	0.34		
MgO	21.20	23.12	34.57	27.69	34.11	27.73	30.65	32.83	29.98	29.76	20.05	15.93	26.56	21.65	30.21	32.94	24.12	31.89	32.98	24.41	33.12	32.28		
CaO	5.96	6.83	2.75	5.98	3.19	6.14	5.33	4.18	5.73	5.83	9.23	10.35	9.32	8.74	6.07	5.09	7.85	5.70	2.73	7.17	2.36	2.95		
K_2O	n.d.[d]	n.d.	n.d.	0.07	n.d.	0.04	n.d.	n.d.	n.d.	0.02	0.07	0.01	0.01	0.01	0.07	0.01	0.20	0.06	0.01	0.03	0.04	0.04		
Na_2O	0.61	0.13	0.10	0.22	0.08	0.19	0.15	0.10	0.11	0.17	0.22	1.60	0.12	0.01	0.31	0.01	0.76	0.47	0.01	0.08	n.d.	n.d.		
P_2O_5	0.05	0.05	n.d.	0.03	n.d.	n.d.	0.02	n.d.	n.d.	n.d.	0.05	0.05	0.04	0.04	0.03	0.03	0.04	0.03	0.04	0.06	0.03	0.03		
Mg#	0.78	0.78	0.85	0.82	0.85	0.83	0.87	0.86	0.86	0.86	0.76	0.72	0.81	0.78	0.86	0.87	0.81	0.86	0.78	0.75	0.79	0.79		
LOI	6.25	7.05	8.90	5.95	8.5	.8	6.95	7.9	6.55	6.6	4.47	5.08	10	5.45	4.65	8.95	3.95	4.90	9.2	5.6	8.65	7.9		
Sc[e]	25.0	20.2	23.9	10.3	25.0	20.1	10.3	25.3	8.1	12.9	9.2	35.1	23.7	19.1	9.7	9.0	7.9	9.4	8.5	6.7	5.0	6.1	47.7	44.0
V	168	234	67	119	64	105	70	7	72	78	129	223	189	184	92	72	133	109	69	102	60	67	361	313
Ta	0.192	0.151	0.011	0.039	0.011	0.034	0.053	0.033	0.037	0.039	0.117	0.090	0.086	0.104	0.034	0.024	0.054	0.031	0.392	0.761	0.385	0.471	0.054	0.040
Nb	1.44	1.95	0.31	0.40	0.34	0.43	0.55	0.41	0.47	0.53	1.53	0.98	0.88	1.21	0.40	0.29	0.45	0.34	3.94	10.74	3.49	3.88	0.59	0.60
Zr	41.7	38.7	13.4	12.1	12.6	14.4	10.7	17.6	11.3	9.4	24.8	30.3	25.8	31.1	10.3	7.8	15.0	12.2	12.9	46.0	5.8	10.2	16.0	15.5
Hf	1.26	1.15	0.42	0.39	0.39	0.43	0.29	0.47	0.39	0.27	1.00	0.86	0.86	0.95	0.58	0.40	0.83	0.62	0.42	1.08	0.32	0.47	0.639	0.60
Th	0.207	0.141	0.033	0.038	0.033	0.035	0.040	0.169	0.037	0.038	0.085	0.072	0.078	0.078	0.039	0.035	0.037	0.042	0.257	0.576	0.230	0.193	0.031	0.030
U	0.054	0.045	0.011	0.011	0.011	0.011	0.011	0.022	0.009	0.011	0.038	0.020	0.020	0.025	0.012	0.011	0.011	0.012	0.082	0.184	0.066	0.071	0.010	0.010
Y	14.3	15.6	5.0	7.5	5.1	6.9	7.1	6.7	7.6	7.6	8.1	10.8	9.2	8.4	6.7	5.8	8.3	6.3	4.1	6.9	3.9	4.5	15.1	16.0
La	3.46	2.73	0.42	0.32	0.39	0.33	0.23	0.39	0.25	0.29	1.29	0.99	0.87	0.92	0.34	0.14	0.39	0.27	3.85	7.31	3.51	3.27	0.627	0.62
Ce	8.77	6.96	0.98	0.99	0.98	1.05	0.75	1.15	0.79	0.88	3.63	2.85	2.64	3.01	0.98	0.45	1.20	0.82	8.70	16.22	7.39	6.90	1.913	1.95
Pr	1.25	1.02	0.18	0.18	0.17	0.19	0.14	0.21	0.14	0.15	0.58	0.47	0.45	0.51	0.17	0.09	0.22	0.15	1.16	2.11	1.04	0.98	0.363	0.38
Nd	5.48	4.89	0.83	1.14	0.95	1.17	0.88	1.09	0.88	0.94	3.22	2.60	2.48	2.81	1.02	0.58	1.36	0.90	4.93	10.04	4.46	4.39	2.337	2.50
Sm	1.73	1.54	0.44	0.54	0.43	0.54	0.40	0.49	0.42	0.44	1.10	0.99	0.96	0.96	0.46	0.33	0.62	0.39	1.04	1.89	0.95	0.96	1.099	1.10
Eu	0.58	0.50	0.21	0.19	0.22	0.20	0.17	0.21	0.15	0.17	0.41	0.42	0.22	0.33	0.18	0.09	0.24	0.16	0.28	0.48	0.21	0.24	0.521	0.54
Gd	2.51	2.16	0.56	0.88	0.60	0.85	0.65	0.78	0.70	0.70	1.56	1.44	1.30	1.30	0.73	0.58	0.99	0.67	0.98	1.82	0.93	0.91	1.824	1.85
Tb	0.41	0.38	0.11	0.18	0.12	0.17	0.13	0.14	0.14	0.14	0.28	0.27	0.23	0.24	0.15	0.12	0.20	0.13	0.14	0.28	0.14	0.14	0.354	0.36
Dy	2.72	2.51	0.79	1.22	0.83	1.19	0.90	1.01	0.96	0.96	1.84	1.80	1.53	1.54	1.00	0.87	1.35	0.90	0.84	1.59	0.79	0.81	2.555	2.50
Ho	0.58	0.53	0.18	0.27	0.20	0.26	0.19	0.26	0.21	0.21	0.38	0.38	0.33	0.32	0.22	0.18	0.29	0.20	0.16	0.31	0.15	0.15	0.575	0.57
Er	1.66	1.53	0.57	0.79	0.56	0.75	0.57	0.75	0.62	0.62	1.07	1.11	0.90	0.90	0.64	0.51	0.83	0.57	0.43	0.83	0.40	0.42	1.680	1.70
Tm	0.26	0.22	0.09	0.11	0.09	0.11	0.08	0.11	0.09	0.09	0.15	0.16	0.13	0.13	0.10	0.07	0.13	0.08	0.06	0.11	0.05	0.06	0.255	0.26
Yb	1.62	1.37	0.59	0.74	0.55	0.72	0.53	0.73	0.55	0.58	0.91	1.00	0.83	0.77	0.62	0.46	0.82	0.55	0.39	0.70	0.35	0.35	1.636	1.65
Lu	0.23	0.19	0.10	0.10	0.11	0.10	0.07	0.12	0.08	0.08	0.12	0.14	0.12	0.11	0.09	0.07	0.11	0.08	0.05	0.10	0.05	0.05	0.246	0.26
Nb/U	26	43	28	37	31	38	49	19	50	47	40	48	44	49	34	26	42	28	48	58	53	55		
Nb/Th	6.9	13.9	9.3	10.6	10.3	12.4	13.9	2.4	12.7	13.8	17.9	13.7	11.4	15.4	10.3	8.1	12.2	8.0	15.3	18.6	15.1	20.1		
Th/Nb	0.14	0.07	0.11	0.09	0.10	0.08	0.07	0.41	0.08	0.07	0.06	0.07	0.09	0.06	0.10	0.12	0.08	0.11	.07	0.05	0.07	0.05		
Ce/Nb	6.09	3.56	3.18	2.45	2.90	2.42	1.36	2.79	1.71	1.67	2.38	2.91	2.98	2.50	2.47	1.56	2.67	2.67	2.21	1.51	2.12	1.78		
Th/U	3.80	3.12	3.00	3.47	3.00	3.07	3.49	7.78	3.93	3.40	2.26	3.51	3.86	3.15	3.27	3.15	3.40	3.51	3.14	3.13	3.50	2.73		
(La/Sm)cn	1.29	1.14	0.61	0.38	0.60	0.40	0.37	0.52	0.39	0.42	0.75	0.64	0.59	0.62	0.48	0.28	0.41	0.44	2.39	2.50	2.38	2.19		

[a]This study
[b]Recommended, Govindaraju 1994
[c]Major elements in wt%
[d]Not detected
[e]Trace elements in ppm; all re-calculated to an anhydrous basis

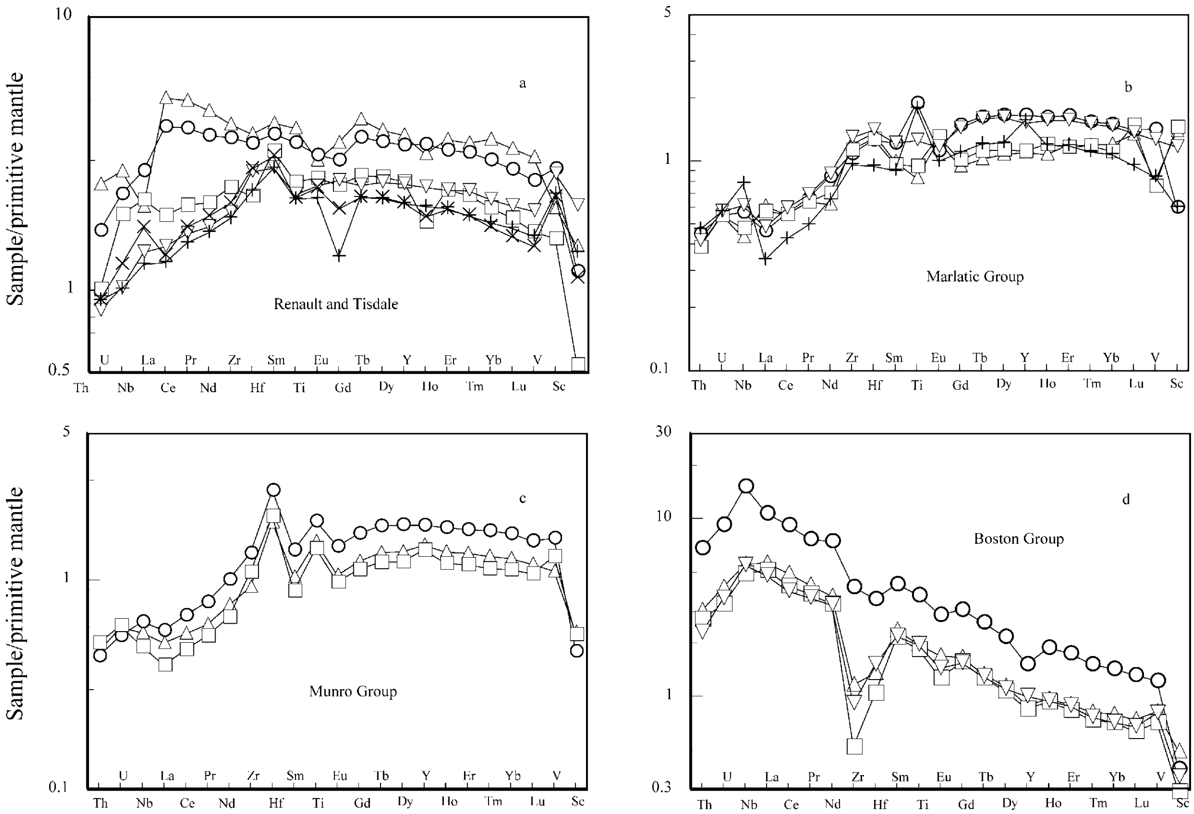

Fig. 1 Concentration of trace elements in representative Abitibi komatiites normalized to primitive mantle values (Geochemical Earth Reference Model, GERM, www.earthref.org)

Discussion

Screening for alteration or crustal contamination

Komatiites for this study, and spatially associated basalts reported in previous studies, were carefully selected for minimal alteration (see Discussion in Kerrich et al. 1999). In the field, volcanic textures are well preserved. In thin section igneous textures are present. The majority of samples have a LOI < 7 wt%, no secondary carbonate, and zero or minor Eu anomalies. There is consensus that the elements least sensitive to mobility during seafloor alteration and metamorphism are Al, Ti, the high field strength elements (HFSE: Th, Nb, Ta, Zr, Hf), REE (excepting Ce and Eu), Y, Sc, and V (Sun and Nesbitt 1979; Ludden et al. 1982; Jochum et al. 1991; Lahaye et al. 1995). Aluminum-undepleted komatiites have chondritic Ti/Zr, Zr/Y, Al_2O_3/TiO_2 ratios; and the compliance to olivine control line plots is considered as evidence of lack of modification of MgO and the elements in question (Arndt 1994). Only minor mobility of U is confirmed by correlated Nb/U and Nb/Th ratios, as well as smooth primitive mantle normalized patterns through Th–U–Nb (Figs. 1 and 2a), although there is more scatter on Nb/U vs. Nb than Nb/Th vs. Nb plots. Citing concerns on U mobility, Puchtel et al. (1998) used U–Pb isotopes to recalculate U and Th concentrations in Baltic komatiites. Their calculated Nb/U and Nb/Th ratios are shown in Fig. 2a for comparison. Although the Abitibi komatiite data are more scattered than the recalculated Baltic data, the Nb/U and Nb/Th ratios of Abitibi komatiites show a good correlation and are collinear with the Baltic data, suggesting that U was not strongly affected by secondary alteration.

The volcanic stage of the Abitibi greenstone belt is considered to have evolved largely in an intra-oceanic setting. There is geological, trace element, and isotopic evidence for the absence of crustal contamination of these komatiites and spatially associated basalts (Xie et al. 1993; Kerrich et al. 1999). The volcanic sequences have interflow deep-water marine sediments (Jackson and Fyon 1992), and there is no geological evidence for older basement in the areas sampled, as is the case in the 2.7 Ga Norseman–Wiluna belt of Western Australia (Campbell and Hill 1988; Sun et al. 1989). No xenocrystic zircons have been found in these Abitibi komatiites and basalts. There are no trends of increasing SiO_2, or LREE with decreasing MgO and Ni, nor development of negative Nb and Ti anomalies, as in the crustally contaminated komatiites of Western Australia (Sun et al. 1989). Some komatiites have both Nb/U and Nb/Th ratios as high as modern depleted upper mantle, which rules out any crustal contamination (Fig. 2a). The trend from depleted mantle towards continental crust in Nb/Th vs. Nb/U co-ordinates could be interpreted as crustal contamination (Fig. 2a). However, the komatiites and majority of basalts plot to lower Ce/Nb than

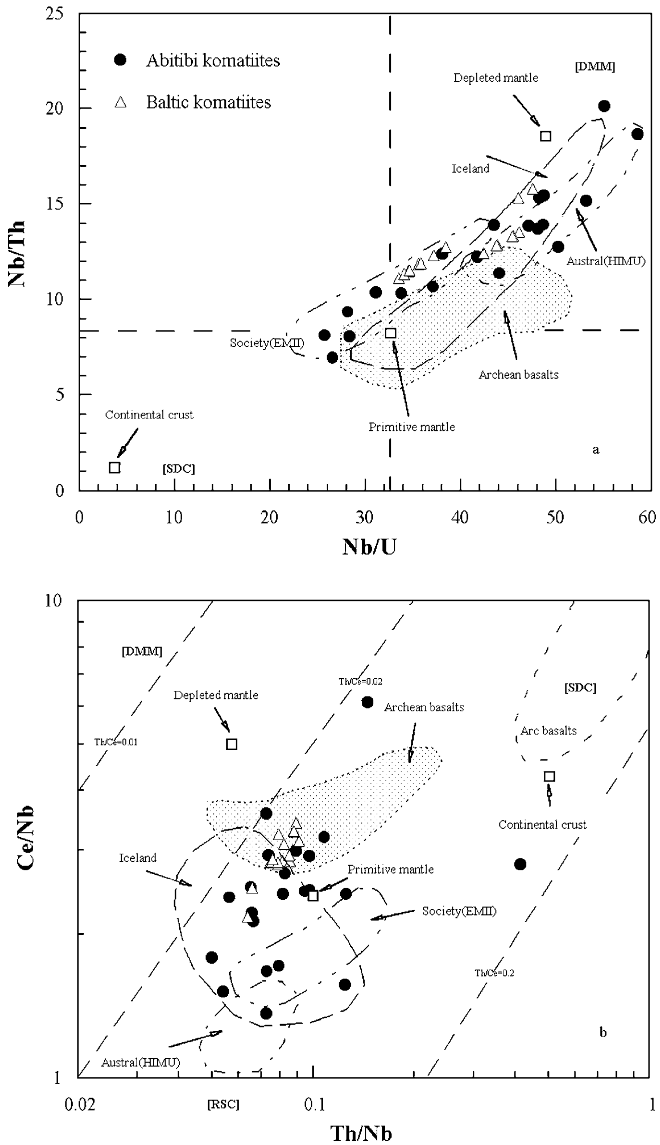

Fig. 2 **a** Nb/Th vs. Nb/U of Archean komatiites and basalts. The precision of the ratios is indicated by the *error bars*. Data sources: Abitibi komatiites (this study); Baltic komatiites (Puchtel et al. 1998, 1999); Abitibi basalts (Xie and Kerrich 1994; Kerrich et al. 1999); Yilgarn basalts (Sylvester et al. 1997); Iceland (Hemond et al. 1993); Society (EMII) and Austral (HIMU; Hemond et al. 1994); primitive mantle, depleted upper mantle and continental crust (Geochemical Earth Reference Model, GERM, www.earth-ref.org). **b** Ce/Nb vs. Th/Nb plot. The precision of ratios is shown by *error bars*. *DMM* Depleted MORB source mantle; *RSC* residual slab component; *SDC* subduction derived component, after Saunders et al. (1988). *Symbols* and data sources as in **a**

depleted mantle and continental crust on Fig. 2b. Abitibi komatiites and basalts all have depleted Sr, Nd, Pb, and Hf isotope compositions, suggesting the absence of crustal contamination (Kerrich et al. 1999, and references therein). Similar sets of criteria have been used to select the least altered suites of Archean basalts from the western Australian (Sylvester et al. 1997) and the Baltic cratons (Puchtel et al. 1998, 1999), and the Superior Province, Canada (Polat et al. 1998; Kerrich et al. 1999; Polat and Kerrich 2000) to interpret primary Th–U–Nb–LREE relationships.

Crustal recycling

Based on Th–Nb–Ce systematics, Saunders et al. (1988) developed a model for ocean plateau and ocean island basalts, and MORB, in terms of mixing between three end members: (1) depleted MORB source mantle (DMM); (2) a residual ocean slab component (RSC) processed through a subduction zone and recycled; and (3) a recycled subduction derived arc component (SDC), complementary to the residual ocean slab. During processing of oceanic lithosphere through a subduction zone, Th, U, and LREE are preferentially lost from the slab in fluids and/or melts to enrich the sub-arc mantle and crust, whereas Nb and Ti are not, leaving a complementary depleted residual slab with positive Nb and Ti anomalies. In keeping with the model of Saunders et al. (1988), the ranges of Nb/Th and Nb/U in the Archean tholeiitic basalts have been interpreted in terms of prior recycling of variable quantities of continental (low Nb/Th) and oceanic (high Nb/Th) crust into the mantle source of the plumes from which the basalts subsequently erupted at 2.7 Ga (Kerrich et al. 1999). In contrast, the komatiites extend to higher Nb/Th and Nb/U ratios, but lower Ce/Nb ratios, with a pronounced recycled slab component (Fig. 2a, b).

Slab recycling to account for komatiites with positive primitive mantle normalized Nb anomalies is similar in many respects to the model for HIMU ocean island basalts (OIB) of White (1993) and Hofmann (1997b, and references therein). HIMU OIB are characterized by (1) positive normalized anomalies of Nb relative to Th and La, (2) negative anomalies of Pb relative to Ce, and (3) low $^{87}Sr/^{86}Sr$ ratios compared with EM I and EM II type OIB, in addition to high μ. White (1993) and Hofmann (1997b, and references therein) account for the positive Nb anomaly by preferential loss of Th and LREE relative to Nb from the slab to the arc, leaving a residual slab with a complementary positive Nb anomaly; for loss of Pb from the slab relative to Th and U generating a complementary negative Pb anomaly with high U/Pb; and for loss of Rb over Sr leaving a low Rb/Sr slab. These proposed element fractionations are in accord with experimental evidence for relative solubility in aqueous fluids (Pearce and Peate 1995). Recycling of the residual slab, and its incubation in the mantle creates a mantle source with the characteristics of HIMU OIB.

Independently, McDonough (1991) and Rudnick et al. (2000) have argued for recycled slabs in the mantle characterized by positive Nb and Ti anomalies, in order to mass balance the Nb and Ti budgets in the Earth's silicate reservoirs. They attributed the anomalies to fractionation of Th and LREE from Nb and Ti, retained in rutile, during dehydration of ocean crust in a subduction zone.

The time constants for deep recycling of a slab into a plume that rises from the core–mantle boundary to erupt

HIMU basalts is estimated at 1–1.5 Ga (Vidal 1992; Hofmann 1997b); recycling rates in the Archean are unconstrained, but are considered to be faster (Fyfe 1978). Basalts have a shallow recycled slab and a subduction-derived arc component or continental crust, and hence their upper mantle source entrained into the plume has a distinct signature from the slab-dominated plume axis (Figs. 2b). Like their Superior Province counterparts, basalts from the Yilgarn and Baltic cratons plot to lower Nb/Th and Nb/U, but higher Ce/Nb, than coexisting komatiites (Fig. 2a, b).

The different trends of komatiites and basalts to respectively lower and higher Th/U (Fig. 2b) can be explained as follows. If the thermal structure of a subduction zone is such that slab melting occurs, then Th is more incompatible than U in the presence of residual garnet (Hauri et al. 1994a); tonalitic liquids from slab melting will then have high Th/U, and the residual slab low Th/U, but high Nb/Th. In a cooler subduction zone dominated by slab dehydration-peridotite wedge melting, U is more incompatible than Th. Consequently, arc magmas will have lower Th/U ratios, and the residual slab higher Th/U but lower Nb/Th (Pearce and Peate 1995; Hawkesworth et al. 1997; Mata 1998).

Tonalite–trondhjemite–granodiorite (TTG) suites characterized by high Al, high La/Yb are abundant in Archean terranes (Martin 1999). There is a general consensus that Archean TTG's form by slab melting beneath arcs leaving residual eclogite, whereas, in Phanerozoic arcs, basalts form by slab dehydration-wedge melting, and evolve by crustal assimilation–fractional crystallization to granitoids characterized by low Al and low La/Yb (Drummond et al. 1996, and references therein). Support for the existence of a low Th/U residual slab after tonalite extraction comes from experimental D's (Hauri et al. 1994a), and direct evidence that Archean TTG and Archean sediments that sample average upper continental crust, possess higher Th/U than Phanerozoic counterparts (Taylor and McLennan 1985; Drummond et al. 1996).

The observed range of Nb/U, Nb/Th, and Nb/LREE in Abitibi komatiites could be controlled by residual mantle phases such as garnet during melting. For example, Al-depleted Boston komatiites have negative Zr and Hf anomalies consistent with a majorite garnet signature (Fig. 1; Xie and Kerrich 1994). However, Al-undepleted komatiites (i.e., Munro, Renault and Marlatic Groups) do not show either a majorite or pyrope garnet signature, given flat HREE, and melted at 5–7 GPa (Herzberg 1995). Hence their trace element systematics cannot be explained by the involvement of garnet (Table 1, Fig. 1). Furthermore, Al-depleted and Al-undepleted komatiites are colinear in Nb/Th–Nb/U co-ordinates (Fig. 2a).

There are few papers on greenstone belt komatiites and basalts that report both Nd-isotope and high precision incompatible element data. Epsilon Nd values of circa 2.7 to 2.8 Ga basalts and komatiites in the Superior Province extend from +1 to +4 (Shirey and Hanson 1986; Henry et al. 1998). From their compilation, Shirey and Hanson (1986) noted that the range of Epslon Nd values differs little from the Neo- to Paleo-Archean. They interpreted this result in terms of short-term recycling of crust into the shallow mantle. In one study of the 2.7 Ga Wawa belt, formerly contiguous with the Abitibi, Polat et al. (1999) showed that Nb/Th ratios correlate with Epslon Nd values, endorsing dispersion of both parameters as mixing between mantle end members.

In summary, the hot axis of the plume that generated Archean komatiites appears to contain a deeply recycled oceanic slab component (low Th/U), ejected perhaps from the core–mantle boundary, whereas the plume annulus that produced associated basalts entrained upper mantle that has variable mixtures of shallow recycled oceanic and continental crust (high Th/U; Fig. 3). Subduction recycling of light continental crust into the mantle has been controversial. However, there are several possible mechanisms including sediment subduction, slab failure, and subduction-erosion (Von Huene and Scholl 1993; Hildebrand and Bowring 1999).

Zoned plumes

There is evidence from seismic tomography, geochemistry, numerical modeling, and experimental fluid dynamics that some mantle plumes are zoned. A low velocity anomaly has been detected below Iceland, extending from near the surface to >400 km (Bijwaard and Spakman 1999). Breddam et al. (2000) detected a gradient of $^3He/^4He$ from 8 to 20 R_A in central Iceland. The 100-km diameter R_A high coincides with a gravity minimum and seismic velocity low in the mantle inferred to define the Iceland plume axial conduit at depth. $^3He/^4He$ correlates negatively with Nb/Th.

Radial zonation has also been proposed for the Society Islands and Galapagos plumes based on the spatial distribution of compositional and isotopic systematics (Hauri et al. 1994b; Hanan and Graham 1996; Hoernle et al. 2000). Campbell (1998) proposed a model for Archean komatiites involving melting in the hot axis of a mantle plume that originates in the CMB layer, and includes recycled ocean lithosphere. In summary, there are numerous empirical lines of evidence that, backed by theoretical considerations, mantle plumes are radially zoned in composition and temperature, and the results of this study provide evidence that an Archean superplume was also zoned.

Implications

Aluminum-depleted komatiites are characterized by negative anomalies of Zr and Hf relative to MREE (Xie and Kerrich 1994; Lahaye et al. 1995; Polat et al. 1999). This characteristic can be interpreted as deep melting in the presence of majorite garnet at depths of 400–670 km (Xie and Kerrich 1994). If this interpretation is correct, then the recycled slab component of komatiites must

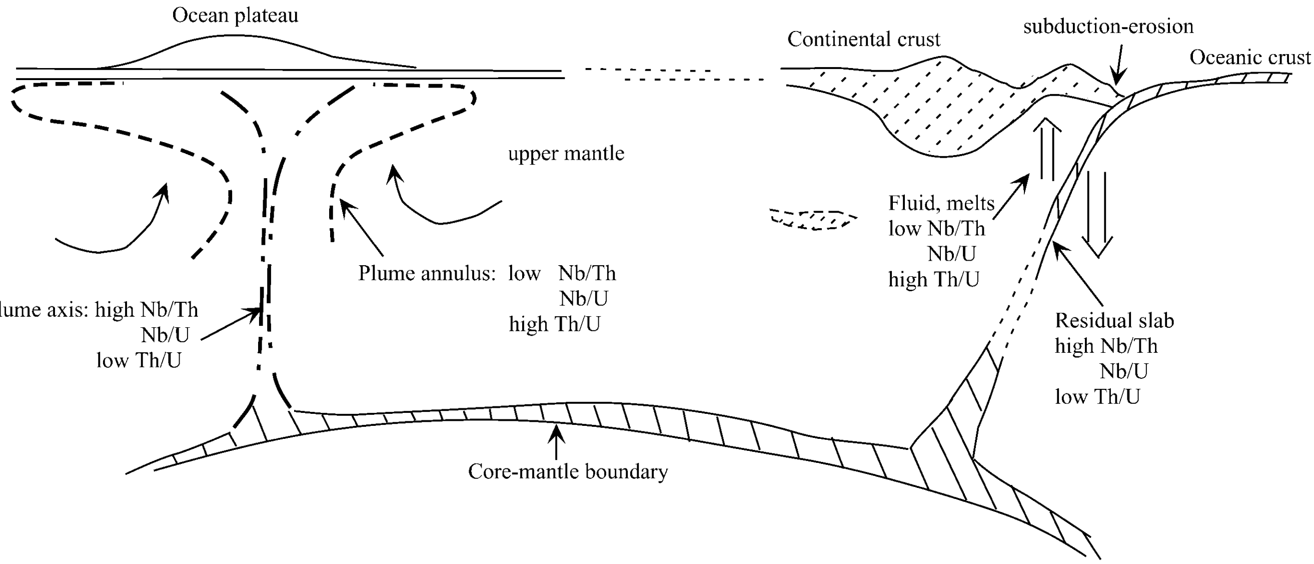

Fig. 3 Starting plume model for Archean komatiite–basalt sequences (cf. Campbell et al. 1989). Plume "head" generates basalts with upper mantle entrainment and mixture of recycled subduction derived arc component and continental crust. In contrast, the plume "tail" contains deep recycled residual slab components

have descended to the 670 km discontinuity, and possibly to the core–mantle boundary, before being incorporated into the mantle source of the plume from which the komatiites erupted.

These results support the model of Campbell et al. (1989) in which komatiites represent melting of the hot core of a mantle plume (the "tail"), which originated at the core–mantle boundary and includes recycled slab. Associated basalts are melts of the cooler plume "head" that entrained the upper mantle (Fig. 3; cf. Campbell et al. 1989). There is clear evidence of the composite compositional nature of mantle plumes in the Pb isotope and trace element variability of the ocean island basalts (Saal et al. 1998; Fig. 2a, b).

Variably high Nb/Th and Nb/U ratios have now been identified in ocean plateau basalts from 3.0 to 2.7-Ga greenstone belts of the Yilgarn, Baltic, and Superior cratons (Sylvester et al. 1997; Puchtel et al. 1998, 1999; Kerrich et al. 1999), and now in komatiites (Xie et al. 2000). If this was a global phenomenon, and assuming complementary low Nb/Th and Nb/U arc crust, then the implication is that the continental crust grew very early, with recycling, as in the model of Fyfe (1978) and Armstrong (1991). A further implication is that given Nb/Th and Nb/U in depleted mantle was as variable in the Archean as in Phanerozoic ocean plateau or island basalts, such as Iceland, Society and Austral (Fig. 2), then the corollary is that progressive crustal growth models of McCulloch and Bennett (1994) and Collerson and Kamber (1999) based on an apparently complementary relationship between continents and depleted mantle are unconstrained, because unknown quantities of recycled continental and oceanic crust reside in the mantle (Rudnick et al. 2000). Finally, as suggested by Rudnick et al. (2000), incorporation of recycled oceanic lithosphere into mantle plumes would generate a range of Nb/Ta in OIB, e.g., HIMU with high Nb/Ta and EMI with low Nb/Ta. Investigation of Nb–Ta relationships in Archean komatiite–basalts sequences would provide further constraints on the deep recycling of oceanic lithosphere and its incorporation into mantle plumes.

Acknowledgements We thank Brian Fryer of the Great Lakes Institute of Environmental Research, University of Windsor, for allowing us to use the VG PQ3 ICP-MS, and Andrew Toms for his help during the analyses. P. Hollings is thanked for his early input in the project. The manuscript was greatly improved by the critical review and suggestions of the two journal reviewers, R. Carlson and an anonymous reviewer. The ICP-MS lab at the University of Saskatchewan is supported in part by an NSERC MFA grant and the University. R.K. acknowledges the George McLeod endowment.

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