Archean zircons from the mantle: The Jormua ophiolite revisited

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ABSTRACT

The Jormua ophiolite represents seafloor from an Early Proterozoic ocean to continent transition zone that mainly consisted of Archean subcontinental lithospheric mantle. These mantle rocks were exhumed as a result of extreme crustal thinning and detachment faulting in association with the opening of the Svecofennian Sea. At the prerift stage of continental breakup, residual lithospheric peridotites became intruded by alkaline melts that formed a diverse suite of clinopyroxenite and hornblendite-garnetite dikes and veins. These Proterozoic dikes contain Archean zircon xenocrysts inherited from deeper sources of the continental mantle. The relatively large spread of ^{207}Pb/^{206}Pb ages between 3106 and 2718 Ma suggests that the zircons are derived from a variety of source rocks. Some xenocrysts have U and Th abundances comparable to zircon in common alkali basalts, whereas a population of Archean high-U and high-Th zircons is similar to those described from mica-amphibole-rutile-ilmenite-diopside–bearing mantle xenoliths. These are the oldest zircons found from upper-mantle rocks anywhere and imply that the Fennoscandian cratonic root was relatively cool and strongly metasomatized by ca. 3.1 Ga.

Keywords: zircon, mantle origin, Archean, ophiolite, rifting, Fennoscandian Shield, Finland.

INTRODUCTION

The Jormua ophiolite was the first Early Proterozoic or older mafic-ultramafic complex that was recognized as an ophiolite (Fig. 1; Kontinen, 1987). For a long time it was considered a typical slow-spreading-ridge ophiolite, consisting of an oceanic crustal unit and remnants of an oceanic mantle diapir, that formed in an embryonic ocean basin ca. 1950 Ma (on the basis of U-Pb zircon ages of gabbros and plagiogranites). More recently, ideas of the geotectonic setting of the Jormua ophiolite have undergone significant revision, because detailed field mapping of its mantle section has brought out major differences between the individual blocks of the complex (Peltonen et al., 1998). The eastern and northern blocks still closely resemble typical ophiolite with enriched mid-oceanic-ridge basalt (E-MORB) sheeted dikes, variably textured gabbros, pillow lavas, hyaloclastites, and mantle tectonites with podiform chromitites (Table 1). However, the western block turned out to be truly different. There, oceanic crustal rocks are absent, but instead the mantle tectonites are extensively veined by a diverse suite of alkaline dikes that are strikingly similar to those found within orogenic lherzolite massifs (e.g., Bodinier et al., 1987) and mantle xenoliths (e.g., Irving, 1980) elsewhere. These observations led Peltonen et al. (1998) to suggest that the Jormua ophiolite is a composite body that contains fragments of both Archean Karelian subcontinental mantle (western block) and Early Proterozoic oceanic lithosphere (eastern and northern blocks) that formed in a tectonic setting related to continental breakup. The idea of the presence of continental lithospheric mantle in Jormua gained strong support from the study of Tsuru et al. (2000), who demonstrated that chromite separated from residual peridotites yielded very depleted present-day ^{187}Os/^{188}Os with an average calculated γ_{Os} (1950 Ma) of -5.1 ± 0.8. Such a negative value requires that the peridotites were strongly depleted at least 1 b.y. before the time of the formation of the gabbros and lavas ca. 1.95 Ga.

In this paper we make a further attempt to constrain the age of the mantle rocks exposed at Jormua by dating zircons from two alkaline mantle dikes.

SAMPLES

The alkaline dike suite that intrudes the western and central block peridotites consists of (1) ''dry'' clinopyroxenites, (2) oceanic-island basalt (OIB)–type ultramafic lampro-

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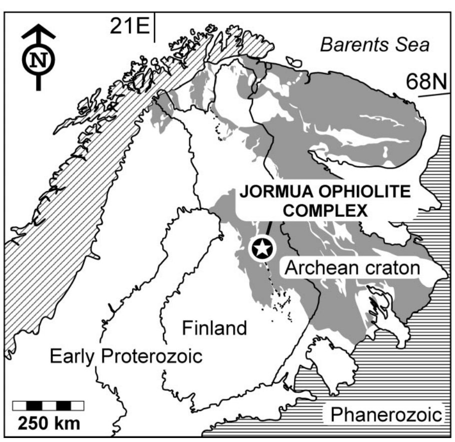

Figure 1. Generalized geological map of Fennoscandian Shield showing location of Jormua ophiolite in eastern Finland. Shaded area is Archean orthogneiss and migmatite; white area is post-Archean, mainly Svecofennian mobile belt. Caledonian orogenic belt is represented with diagonal lines.

TABLE 1. CHARACTERISTICS OF THE JORMUA OPHIOLITE

Rock type	Western block	Central block	Eastern block	Northern block
Crustal units				
Extrusive unit	–	–	++	+
Sheeted-dike complex	–	++	++	+
Gabbros and plagiogranites	–	–	++	+
Ultramafic cumulates	–	–	?	–
Intrusive to mantle tectonites				
Tholeiitic dike suite				
E-MORB-dikes	?	++	++	+
Gabbroic feeder dikes	–	+	++	+
Chromitite pods	–	–	+	–
Alkaline dike suite				
Clinopyroxenite mantle dikes	++	+	–	–
Hornblendite mantle dikes	++	–	–	–
OIB-type dikes	?	+	–	–
Garnetite veins	+	–	–	–
Carbonatitic veins	+	–	–	–
Mantle tectonites				
Lherzolites (>3 wt% Al_2O_3)	+	–	+	?
Depleted lherzolites (1 < Al_2O_3 < 3 wt%)	++	++	++	+
Harzburgites (<1 wt% Al_2O_3)	–	+	+	+

Note: (+) present; (++) abundant; (–) absent. OIB is oceanic-island basalt; E-MORB is enriched mid-ocean-ridge basalt.

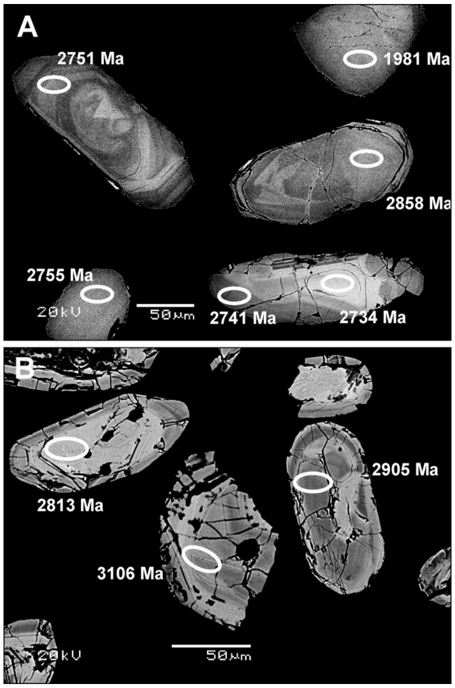

Figure 2. Backscattered-electron images of zircons from clinopyroxenite mantle dikes of Jormua ophiolite. A: Dike 24F. B: Dike 23B. Ion microprobe sample spots and derived $^{207}Pb/^{206}Pb$ ages are indicated.

phyre dikes, and (3) hornblendites, which include garnetite veins and carbonatitic segregations. We separated zircons from two dry clinopyroxenites.

Clinopyroxenite dike sample 23B was taken from the western block, where alkaline mantle dikes are particularly common (Table 1). It comes from the central part of a 2-m-wide medium-grained, equigranular, and homogeneous cumulus-textured dike. The second clinopyroxenite sample 24F comes from a similar but somewhat thinner (0.8 m wide) dike within the central block peridotites. Both 23B and 24F have sharp intrusive contacts against the peridotite, and their orientations are broadly parallel with the chromite-defined mantle foliation. They lack clear modal and textural variations across the dike, and Peltonen et al. (1998) suggested that filter pressing due to concomitant mantle deformation probably depleted these dikes of intercumulus liquid. Both of the dated samples consist of cumulus diopside crystals (diameter, ϕ = 1–2 mm) and abundant subhedral ilmenite, which probably is also a cumulus phase. The amount of intercumulus material is variable. Sample 23B is mesocumulate and contains only 9 ppm Zr. Sample 24F is orthocumulate, containing 460 ppm Zr. Such a high Zr content, together with high TiO_2 for 24F, suggests that it formed from more evolved melt than dike 23B.

RESULTS

Clinopyroxene cumulate 24F yielded abundant zircons. These can be subdivided into two types: (1) prismatic, pale brown grains with bipyramidal grain tips, and (2) prismatic, colorless, and bright or somewhat turbid grains with rounded tips or more rounded anhedral forms (Fig. 2A). We dated 14 grains, all yielding concordant ages with bimodal age distribution (Table 2; Fig. 3). Prismatic growth-zoned type 1 grains all yielded Archean ages. The concordia age calculated for seven of these grains is 2747 ± 8 Ma (mean square of weighted deviates, MSWD = 3.8), whereas one grain records an older age of 2858 ± 7 Ma. Type 2 grains record younger ages and a wider range of ages between 2037 and 1942 Ma. The second clinopyroxenite sample, 23B, yielded only ~30 euhedral pale to dark brown, translucent or turbid, prismatic grains. They have much higher U and Th abundances than the 24F grains, and most of them are strongly corroded and could not be analyzed. Only a few grains contain fresh domains larger than the beam size of the ion microprobe (Fig. 2B). They yielded concordant Archean $^{207}Pb/^{206}Pb$ ages mainly between 2832 and 2800 Ma; two older grains yielded ages of 3106 ± 3 Ma and 2905 ± 9 Ma.

DISCUSSION AND CONCLUSIONS

Clinopyroxene cumulate dikes contain both Archean and Proterozoic zircons with ages either much older or close to that of the oceanic crustal sequence of the Jormua ophiolite (ca. 1.95 Ga). Such a bimodal distribution of $^{207}Pb/^{206}Pb$ ages warrants two distinct interpretations: (1) the clinopyroxene cumulate dikes are Archean in age, and some zircon recrystallized during Early Proterozoic events; or (2) the dikes are Proterozoic, and the Archean zircon population is inherited from deeper lithospheric source(s). Sm-Nd isotope data published by Peltonen et al. (1998) provide important constraints for this problem. Clinopyroxenites yield ε_{Nd}(1950 Ma) values between −1.0 and +0.8 with depleted mantle values (T_{DM}) of ca. 2 Ga. An Archean mantle origin would require initial ε_{Nd} values close to a conceivable mantle composition, e.g., ε_{Nd} of ~+2, but the calculated ε_{Nd} values for the clinopyroxenites at 2.8 Ga range from −6 to +6. Thus, the Sm-Nd isotope data clearly sug-

TABLE 2. ION MICROPROBE U-Pb DATA ON JORMUA ZIRCONS

Sample spot	Derived ages (Ma) $^{207}Pb/^{206}Pb$	±σ	$^{207}Pb/^{235}U$	±σ	$^{206}Pb/^{238}U$	±σ	Calibrated ratios $^{207}Pb/^{206}Pb$	±σ (%)	$^{207}Pb/^{235}U$	±σ (%)	$^{206}Pb/^{238}U$	±σ (%)	Rho	Elemental data U (ppm)	Th (ppm)	Pb (ppm)
24F-02a	2734	5	2760	18	2795	42	0.1891	0.28	14.148	1.88	0.5428	1.86	0.99	296	267	230
24F-02b	2741	8	2767	18	2803	43	0.1899	0.49	14.257	1.93	0.5446	1.86	0.97	137	58	98
24F-03a	2750	12	2776	19	2813	42	0.1909	0.76	14.399	2.00	0.5471	1.85	0.93	35	43	29
24F-06a	2758	5	2744	18	2726	41	0.1918	0.30	13.921	1.88	0.5264	1.85	0.99	220	227	172
24F-09a	2718	12	2746	19	2784	43	0.1872	0.73	13.941	2.02	0.5402	1.89	0.93	41	38	32
24F-11a	1981	12	1968	18	1956	33	0.1217	0.65	5.946	2.07	0.3545	1.96	0.95	141	85	65
24F-12a	2858	7	2873	19	2895	44	0.2039	0.46	15.936	1.92	0.5668	1.87	0.97	102	26	74
24F-13a	2751	8	2768	19	2792	43	0.1910	0.51	14.274	1.94	0.5419	1.87	0.97	93	70	70
24F-14a	2755	5	2756	18	2757	42	0.1914	0.30	14.090	1.88	0.5338	1.85	0.99	274	115	191
24F-24a	1942	12	1968	17	1993	32	0.1190	0.69	5.947	1.99	0.3624	1.86	0.94	85	35	38
24F-26a	1984	11	1970	17	1957	31	0.1219	0.63	5.961	1.96	0.3548	1.85	0.95	117	63	53
24F-26b	1998	5	2001	17	2005	32	0.1228	0.28	6.178	1.87	0.3648	1.85	0.99	452	278	215
24F-27a	2014	19	2005	19	1997	32	0.1239	1.07	6.206	2.14	0.3632	1.86	0.87	83	27	36
24F-41a	2037	6	2040	17	2043	33	0.1256	0.36	6.458	1.90	0.3729	1.86	0.98	348	190	167
23B-01a	2832	5	2810	15	2780	34	0.2007	0.30	14.923	1.55	0.5392	1.52	0.98	369	202	250
23B-03a	3106	3	3072	16	3020	39	0.2379	0.22	19.604	1.61	0.5976	1.59	0.99	581	313	482
23B-04a	2800	3	2836	15	2886	36	0.1968	0.15	15.323	1.56	0.5647	1.55	1.00	2193	664	1587
23B-05a	2905	9	2787	15	2627	33	0.2099	0.54	14.560	1.62	0.5031	1.52	0.94	100	47	69
23B-06a	2813	2	2798	15	2777	36	0.1984	0.15	14.725	1.58	0.5384	1.57	1.00	1719	558	1196

Note: Errors are at 1 sigma level. The U-Pb analyses were run by using the Cameca IMS 1270 ion microprobe at the Nordsim Laboratory, Swedish Museum of Natural History, Stockholm. The spot diameter for the 4 nA primary O_2^- ion beam was ~ 30 µm, and oxygen flooding in the sample chamber was used to increase the ionization of Pb. Four counting blocks including total 12 cycles of the Zr, Pb, Th, and U species were measured in each spot. The mass resolution (M/ΔM) was ~5400 (10%). The raw data were calibrated against a zircon standard and corrected for background at mass 204.2 and modern common lead (T=0; Stacey and Kramers, 1975). For a detailed description of the analytical procedure see Whitehouse et al. (1999).

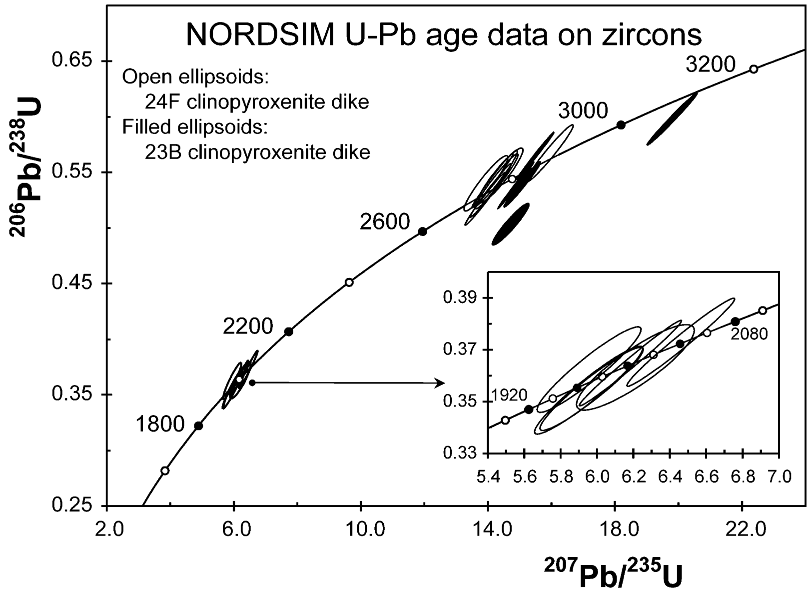

Figure 3. Concordia diagram showing U-Pb age data on zircons from two clinopyroxenite mantle dikes of Jormua ophiolite (Finland).

gest that the light rare earth element fractionation and the crystallization age of the dikes are Proterozoic. We think that this discrepancy between Archean zircon ages and Proterozoic Sm-Nd model ages shows that Archean zircons occur as xenocrysts in Proterozoic dikes.

Where did these Archean grains come from? A characteristic feature of the data set is the relatively large spread of the Archean ages. Although most of the grains in dike 24F are ca. 2750 Ma, one grain has a $^{207}Pb/^{206}Pb$ age of 2858 Ma. Dike sample 23B is similar but has a larger spread of Archean ages. Three grains have $^{207}Pb/^{206}Pb$ ages of ca. 2810 Ma, but one grain has an age of 2905 $\pm$ 9 Ma, and another grain has an age of 3106 $\pm$ 3 Ma. Such a spread of ages is thought to reflect lithologic heterogeneity of the source of these zircon xenocrysts. The discovery of Archean zircons in Early Proterozoic mantle dikes implies that they were incorporated into the clinopyroxene cumulate–forming magma from older Archean dikes or veins at deeper levels of the Karelian subcontinental mantle. Their postulated presence is not surprising in view of the antiquity of Karelian cratonic roots and the multistage igneous history. The oldest upper and lower crustal rocks in the western part of the Karelian craton record U-Pb zircon ages to 3.5 Ga, and a full range of 2.7–3.2 Ga igneous rocks is exposed within the Archean terrain. All these magmatic events have left their traces in the continental mantle in the form of veins and dikes that may have supplied zircon to alkaline dike–forming magmas. It is interesting to note that the U and Th abundances of 24F and 23B dike zircons are distinct. Archean zircons in 24F contain an average of 150 ppm U and 106 ppm Th (Table 2), which are typical abundances for zircons found in common mafic alkalic igneous rocks (e.g., Belousova et al., 1999). In contrast, four of the five zircons from dike 23B contain 1216 ppm U and 434 ppm Th, on average (Table 2), values that are more typical for zircons in mica-amphibole-rutile-ilmenite-diopside–bearing mantle xenoliths (e.g., Konzett et al., 1998) and a zircon inclusion found in diamond (Kinny and Meyer, 1994).

Worldwide, zircons from upper-mantle rocks have rarely been dated. Most U-Pb–dated grains have been recovered from kimberlites, where zircons occur as large millimeter- to centimeter-sized megacrysts (e.g., Davis et al., 1976; Kinny et al., 1989; Griffin et al., 2000; Peltonen and Mänttäri, 2001). In most cases the ages of kimberlitic zircons approximate the emplacement age of the kimberlite host, implying that the origin of zircon megacrysts is somehow related to that of the kimberlite magmatism or, alternatively, that the isotopic clock of zircon is continuously reset at ambient mantle temperatures. By far the oldest kimberlitic zircons are the older-generation zircons of the Permian Jwaneng DK2 kimberlite (Botswana), which yield $^{207}Pb/^{206}Pb$ ages between 2805 Ma and 2125 Ma. Kinny et al. (1989) argued that these grains record an Archean episode of kimberlite magmatism in the mantle. However, Belousova et al. (1999)—on the basis of trace element composition of these zircons—came to the conclusion that these grains were more likely derived from a crustal source and not from the mantle. More reliable mantle chronology is obtained by dating zircons that occur in continental mantle samples in situ. In addition to the mantle dikes of the Jormua ophiolite, such an approach has been applied to metasomatized mantle xenoliths and their hydrous veins (Kinny and Dawson, 1992; Rudnick et al., 1999; Konzett et al., 1998, 2000), diamond inclusions (Kinny and Meyer, 1994), and mantle dikes exposed in orogenic lherzolite massifs (Sánchez-Rodríguez and Gebauer, 2000; Ordóñez et al., 2001). None of these studies reports Archean zircons from the mantle, and, to our knowledge, the 3.1 Ga xenocrystic zircons in the Jormua dikes are the oldest zircons found from Earth's mantle.

Zircon dating of the clinopyroxene cumulate dikes confirms the earlier views that most—if not all—mantle peridotites of the Jormua ophiolite are pieces of the old Archean Karelian continental mantle and do not represent the 1.95 Ga oceanic mantle diapir. However, because gabbros and plagiogranites from the oceanic crustal unit of the complex have yielded high-precision Early Proterozoic crystallization ages of ca. 1950 Ma, the Jormua ophiolite is not entirely Archean. Instead, the mantle section consists of two distinct components: (1) Archean (older than 3.1 Ga) subcontinental mantle peridotites and (2) a suite of convective mantle–derived alkaline igneous rocks (clinopyroxene cumulate dikes and hornblendite-garnetite-carbonatite veins) that intruded the lithospheric peridotites before or during the earliest stages of the ca. 1.95 Ga continental rifting (Fig. 4). During the more advanced stages of rifting, the Archean continental mantle became exposed to the seafloor as a result of detachment faulting and was intruded by E-MORB magmas from the oceanic mantle diapir that gave rise to the ca. 1950 Ma basaltic lid on top of the continental mantle rocks.

The revised ophiolite stratigraphy of the Jormua ophiolite bears striking similarities to the seafloor from modern nonvolcanic passive margins. One such location is the Cretaceous West Iberia continental margin, where the ocean to continent transition zone between the rifted and thinned continental crust and true oceanic crust has been studied in detail (e.g., Whitmarsh et al., 2001). There, seismic studies have identified an ~100-km-wide zone of partially serpentinized lithospheric peridotites exposed at the seafloor. Sampling of this zone has indicated that scarce basaltic rocks, locally pillowed, were deposited—as in Jormua—on a substratum of continental lithospheric peridotites that enclose older, strongly sheared and metamorphosed gabbro intrusions and alkaline pyroxenites (Cornen et al., 1999). Phanerozoic analogies for the Jormua ophiolite include the Northern Apennines ophiolites, which were formed in conjunction with the opening of the Jurassic Tethys Ocean (Rampone and Piccardo, 2000), and Zabargad Island (e.g., Bonatti

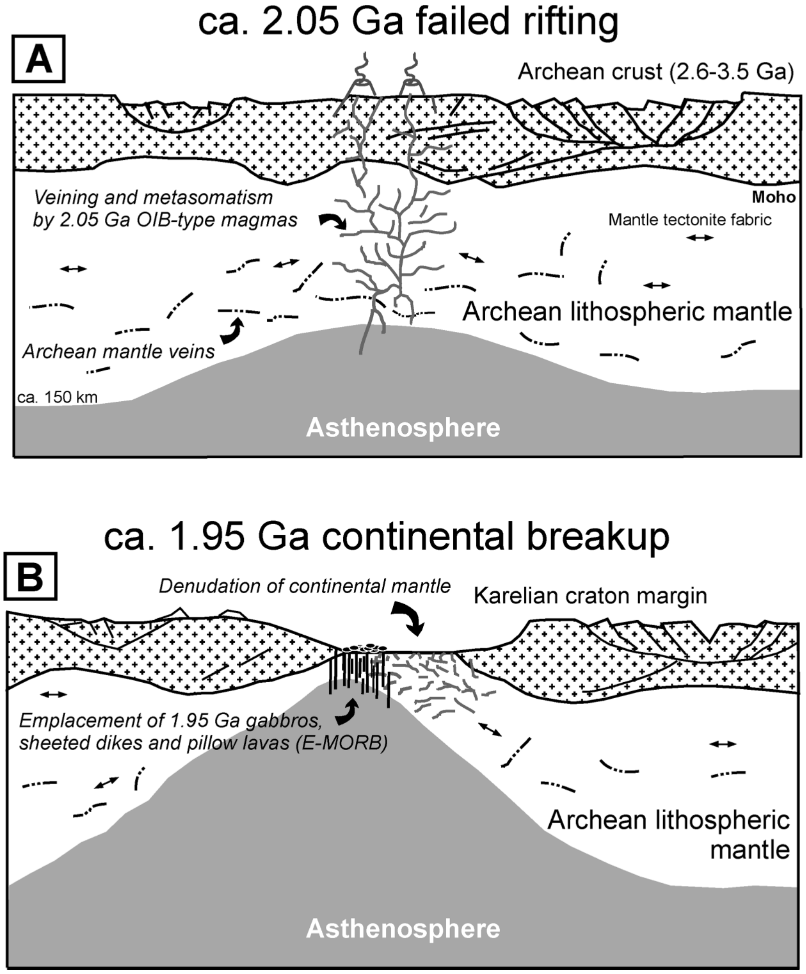

Figure 4. Lithosphere-scale model for origin of Jormua ophiolite. A: Early extension event ca. 2.05 Ga initiates low-degree partial melting of asthenosphere, resulting in emplacement of alkaline magmas and formation of clinopyroxenitic and hornblenditic dikes within subcontinental mantle. These magmas carry zircon xenocrysts from Archean mantle veins. B: Breakup of craton and exposure of veined subcontinental lithospheric mantle at seafloor by asymmetric extension. Enriched mid-ocean ridge basalts (E-MORB) are emplaced through and within stretched subcontinental mantle to form gabbros, sheeted dikes, and pillow lavas. OIB—oceanic-island basalt.

et al., 1981) of the Red Sea, where continental mantle became exhumed owing to extreme crustal thinning and detachment faulting during the final stages of continental breakup.

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Notes