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Palaeoproterozoic Mafic magmatism in the Indian Shield: Petrologic, Geochemical and Thermal constraints on Proterozoic Mantle

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In this paper we have synthesized the published and unpublished geochemical data on the Palaeoproterozoic mafic magmatism in the Indian Shield. Palaeoproterozoic mafic magmatism is widespread in the Indian Shield; it mostly emplaced as dyke intrusions within the cratons and south Indian granulite region and as intrusives/traps in the intra-cratonic basins and the Eastern Ghat Mobile Belt. Integration of the U-Pb ages with palaeomagnetic results identified at least four discrete Palaeoproterozoic igneous events at 2.36-2.37 Ga, 2.1-2.2 Ga, 2.0-2.1 Ga 1.89-1.99 Ga and probably two other events at about 2.4 Ga and 1.8 Ga. The Palaeoproterozoic magmatism across the Indian cratons seems geochemically monotonous and exclusively mafic and sub-alkalic tholeiitic basalt/basaltic andesite in composition with typical enrichment of large ion lithophile and light rare earth elements. Fractional crystallization is the dominant mechanism controlling the geochemical spectrum of the Indian Palaeoproterozoic magmas with little indications of crustal assimilation. Asthenosphere mantle is the major supplier of material for the Palaeoproterozoic igneous activity. Sub-Continental Lithosphere Mantle (SCLM) seems to be the major contributor for the enriched characteristics of the Palaeoproterozoic mafic magmas erupted in the Indian Shield. Thermal energy for the initiation of melting is likely contributed by mantle plumes although a passive rifting is not ruled out. Geochemistry of the Palaeoproterozoic mafic magmas in some way appears to have similarities to that of the end-Cretaceous Deccan basaltic magmas and do reflect plume-lithosphere interaction. The SCLM beneath the Indian Shield was possibly enriched by addition of fluids/

melts of deep mantle origin. We suggest that the major segments of the Indian SCLM were generated at around 3 Ga coinciding with a major crustal building activity in the Indian shield.

Introduction

Mafic igneous activity is an important feature of the Proterozoic Earth particularly in the stabilized Archaean cratons and may be the only significant geological event at about 2 Ga. The genesis of each pulse of igneous activity clearly constitutes a major thermal event affecting the Earth's mantle. The advent of precise U-Pb dating (e.g. LeCheminant and Heaman, 1989) is remarkable in the sense that it indicated the emplacement of Proterozoic magmatic activity took place over a very restricted time interval, which allowed drawing parallel with the Phanerozoic continental flood basalt magmatism. This implies that a large amount of thermal energy was available for melting the mantle, but it was apparently delivered quickly and periodically.

The magmas of the continental igneous units pass through thick Archaean crust before emplacement. There is always the question of the extent to which the magma compositions have been modified by processes like fractional crystallization *en route*, assimilation with fractional crystallization (AFC) or ponding of magmas near the Moho with melting, assimilation, storage and homogenization before emplacement. At the same time, the compositional variations in the continental igneous lithologies have potentially been explained in terms of partial melting, dynamic melting, mantle lithosphere enrichment processes or heterogeneous mantle reservoirs among others. Despite this apparent conceptual clarity, our understanding of the origin of mantle compositional heterogeneity, in particular the origin of the enriched geochemical signatures inferred from within plate igneous lithologies, remains somewhat incomplete (Hofmann and White, 1982; Sobolev et al., 2005 and 2007; Pilet et al., 2008 and 2011; Herzberg, 2011; Niu et al., 2012; Trela et al., 2015 and references therein). These enriched mantle sources are believed to have been developed by injection of melts derived from oceanic crust

or sediment through subduction zone processes or by infiltration of low degree partial melts or fluids from lower mantle.

This contribution is oriented to summarize (1) nature of Proterozoic mafic magmatism in the Indian shield and (2) prominent petrological features of this magmatism, and then try to account for these features in the context of our present understanding of mantle evolution. Some comparisons will also be drawn with Phanerozoic continental flood basalt provinces, where relevant in view of some similarities between these two magmatic events. New precise U-Pb ages have started appearing from the Indian shield, as summarized in the following section, to indicate the occurrence of mafic magmas at different intervals during the Proterozoic; they provide a useful window to monitor mantle evolution, particularly the sub-continental lithosphere, which appears to be the dominant source for enriched compositions of the most mafic igneous occurrences. Thermal energy needed for these igneous events may lie within mantle plumes. The conclusions reached in this study may contribute to the thermal evolution and mantle plume debate (e.g. Campbell, 2005; Davies, 2005; Foulger, 2005; Foulger et al., 2005; Niu, 2005; Campbell and Davies, 2006) although the debate cannot be resolved using the geochemistry alone.

Geological Features of Mafic Igneous Activity

The Indian shield comprises vast terrains of Archaean cratons,

Proterozoic basins and mobile belts. The Dharwar, Bastar, Singhbhum, Bundelkhand and Aravalli cratons in India are among the oldest cratons of the world. The Cuddapah, Gwalior, Bijawar and Vindhyan are the prominent Palaeoproterozoic sedimentary basins formed on the cratonic margins. Granulite terrain in South India and the Eastern Ghats Belt towards the east coast constitute the prominent Precambrian mobile belts.

The Proterozoic magmatism has significant presence in all the cratons, sedimentary basins and mobile belts. It occurs more frequently in the form of mafic dykes as has been observed in many other cratons of the world (Ernst, 2014), but are never seen extending into the sedimentary basins. The dyke intrusions in the Dharwar craton and adjoining granulite terrain have been the subject of extensive multidisciplinary studies. Profuse occurrence of these dykes is seen on western margin of the Cuddapah Basin and is grouped into three/four strike trends. These dykes appear to extend all along the Dharwar craton up to the west coast, but are distributed as individual clusters in different areas all across the craton and also extend in south into the northern parts of adjoining granulite region of south India (Fig. 1). Mafic dyke intrusions in the Bundelkhand and Bastar cratons appear to be more continuous without much intermittent discontinuity. Mafic dykes represent the main Proterozoic mafic igneous activity in Singhbhum and Aravalli cratons too. Palaeoproterozoic mafic dykes and lava flows are known from Eastern Ghats Belt and adjoining regions (Fig. 1). Mafic magmatism is known from the Dongargarh belt, the Bastar craton and as Dalma traps from the Singhbhum craton. The former is unusually crustal-contaminated

Figure 1: (a) Map illustrating broad geological sketch of the Indian shield with locations of the Palaeoproterozoic mafic igneous emplacements. The mafic igneous activity is manifested as dyke intrusions within the Dharwar, Bundelkhand, Bastar and Singhbhum cratons. Locations of the Gwalior, Bijawar and Kondapalli-Kandra emplacements are marked as GW, BJ and KK respectively; (b) more detailed geological sketch of the south India showing locations of studied dyke intrusions as rectangles Khammam-Peddavur (KP), Mahabubnagar-Wanaparti (MW), Anantapur-Gooty (AG) and Tirupati-Raychoti (TR) are around the Cuddapah Basin; Kolar (KR), Harohalli (HH), Kunigal (KU) are in the Eastern Dharwar craton away from the Cuddapah Basin; Udupi-Mangalore (UM), Piriapatna (PP) and Karwar (KW) are in the Western Dharwar craton; Tiruvannamalai (TM), Dharmapuri (DP), Agali-Coimbatore (AC) and north Kerala (NK) are in the Southern Granulite Terrain. Location of the Cuddapah traps/intrusives is demarcated as CT.

(Sensarma et al., 2002) and therefore do not permit source characterization. In case of the Dalma volcanics, precise ages or palaeomagnetic constraints are not available and are considered to show oceanic signatures (Yellur, 1977; Roy et al., 2002). Thus these two occurrences are excluded for the present discussion.

The mafic igneous activity is manifested in the form of dykes probably because fractures can propagate rapidly and opening fractures can be quickly filled with fluid magma. The abundant dyke population clearly reflects significant extension and the cause of the extension may be related to mantle processes, even though the tectonic environment differs. Now we know that dyke intrusions can propagate laterally for hundreds of kilometers, and conceptually it is difficult to imagine why magma should penetrate laterally for such huge distances (Tarney, 1992). Consequently, the clusters over a large area in the Dharwar craton may have genetic linkages and the volume of magma involved in each dyke event must be fairly large. For example, the volume of magma involved in the 1250 Ma old Mackenzie Swarm centered around Muskox intrusion is estimated at 90000 km³ (Fahrig, 1987). Such volumes are comparable with some Phanerozoic Large Igneous Provinces (LIPs) represented by continental flood basalt provinces. However, the Phanerozoic flood basalt eruptions, in contrast to Proterozoic mafic igneous activity, occur as large surface eruptions with relatively limited number of genetically linked dyke intrusions. Erosion may be argued as one of the causes for the lack of surface eruptions corresponding to Proterozoic dykes, however, the explanation is not conclusive as the volcanic units of Archaean age and some volcanic units of Proterozoic age are well preserved. In this context, the trap rocks in the Cuddapah basin, south India and the Gwalior and Bijawar volcanics in the sedimentary basins in Central India stand as best Indian examples.

The mafic igneous units in the form of surface eruptions are very small in volume and are always present within the lower stratigraphic sections of the Palaeoproterozoic basins. In south India, these rocks are exposed along the western margin in the lower Cuddapah succession (Fig. 1), unconformably overlying the granite-gneiss complex of the Eastern Dharwar craton. These igneous units are variably described as basaltic flows, mafic/ultramafic sills and ash fall tuffs (Srikantia, 1984; Nagaraja Rao et al., 1987; Anand et al., 2003). The Gwalior and Bijawar traps occur in the Palaeoproterozoic Gwalior and Bijawar basins respectively in central India (Fig. 1). The Gwalior traps occur as basaltic lavas underlying the upper Vindhyan Supergroup (Crawford and Compton, 1970) or as sills (Samom et al., 2017). The Bijawar mafic igneous rocks comprise both volcanic flows and gabbro units. The Kondapalli-Kandra Palaeoproterozoic mafic rocks occur as dykes, sills and flows within the Nellore Schist Belt-Eastern Ghat Belt tectonic milieu and represent plate-margin igneous activity (Fig. 1).

A few observations are worth making here. In spite of propagation to such large lateral extent, it is puzzling that these dyke intrusions like their counterparts in the world do not have the silicic compositions. Similarly, the dykes although were emplaced into the high-grade granulite and amphibolite facies gneisses, the country rock never shows signs of melting. No central igneous complexes (magma chambers) feeding the Palaeoproterozoic dyke intrusions are noticed. Only on the southwestern segment of the Cuddapah Basin, presence of a high-density lopolitic magmatic body that may have fed the dyke intrusions was postulated (Krishna Brahmam and Dutt, 1992; Rambabu, 1993; Mishra and Tiwari, 1995; Anand et al., 2003; Radhakrishna et al., 2007).

Chronology, Palaeomagnetism and Periodicity

Precise dating of mafic igneous units has been a difficult task for long time. The initial K-Ar and the Rb-Sr whole rock isochron methods yielded a spectrum of ages covering almost the whole of Proterozoic Eon. However, Halls (1987) compilation on Proterozoic mafic magmatism shows discrete events separated by periods of inactivity. Advances of U-Pb zircon and baddeleyite dates (LeCheminant and Heaman, 1989) of mafic igneous units yielded ages with much higher precision and improved our understanding of temporal controls on mafic magmatic activity. Palaeomagnetic characteristic magnetization results provide an added advantage to identify discrete igneous events when integrated with high precision U-Pb geochronology. There have been reports of a few U-Pb baddeleyite Palaeoproterozoic ages on igneous units from the Dharwar craton of the Indian shield (Halls et al., 2007; French et al., 2008; French and Heaman, 2010; Kumar et al., 2012 and 2015) and Sm-Nd whole rock-mineral isochron ages (Zachariah et al., 1995; Dash et al., 2013). Similarly two U-Pb baddeleyite Palaeoproterozoic ages are available from the igneous units of the Bastar craton (French et al., 2008; Gregory et al., 2018). A U-Pb zircon Palaeoproterozoic date is reported from mafic dyke intrusion of the Bundelkhand craton (Pradhan et al., 2012) and Southern Granulite terrain (Pivarunas et al., 2018). Precise U-Pb age of 1.85 Ga was obtained for a sheeted dyke from the Kandra ophiolite complex in SE India (Vijaya Kumar et al., 2010). Although there are some re-equilibration problems with U-Pb zircon or rutile dates, judicious use of these dates may identify discrete events. Simultaneously, large number of palaeomagnetic results was also obtained recently from Palaeoproterozoic mafic dyke intrusions in these cratons and adjacent granulite region (Radhakrishna and Mathew, 1996; Radhakrishna et al., 2003; Halls et al., 2007; Meert et al., 2011; Pradhan et al., 2012; Radhakrishna et al., 2013a; Radhakrishna et al., 2013b; Pivarunas et al., 2018). Integration of the U-Pb ages with the palaeomagnetic results identified at least four discrete Palaeoproterozoic igneous events at 2.36-2.37 Ga, 2.1-2.2 Ga, 2.1-2.0 Ga 1.99-1.89 Ga and probably two other events at about 2.4 Ga and 1.8 Ga. Although the igneous activity in the Indian shield comprises distinct phases of igneous emplacements, the proportion of mafic magmatic units in each phase is not known. Recent focus of U-Pb baddeleyite dating in Singhbhum craton have yielded ages close to the 2.2-2.1 Ga and 1.8 Ga groups (Shankar et al., 2014; Srivastava et al., 2019) and also unraveled presence of older Neoarchaean (2.75-2.80 Ga) mafic intrusive pulses (Kumar et al., 2017) that are outside the scope of this work.

There are reports of U-Pb Neoproterozoic zircon ages from one mafic dyke in each of the two Cratons (1025.6 ± 3.8 Ma on the western margin of Cuddapah Basin in the Dharwar craton; Pradhan et al., 2010; 1113 ± 7.4 Ma on one mafic dyke of the Bundelkhand craton; Pradhan et al., 2012). However, sufficient palaeomagnetic directions that characterize this Neoproterozoic igneous activity are not available at present.

Geochemistry and Petrological Features

A large body of major and trace element geochemical data has accumulated for Proterozoic igneous suites across the cratons in the Indian shield see Table 1. The igneous units in south India are also

Table 1: Average major and trace element compositions of the Papaeoproterozoic mafic rocks from different provinces of the Indian Shield.

oxides (wt.%)	E. Dharwar			W. Dharwar			SGT			Bastar			Bundelkhand			Cuddapah Basin			Gwalior Basin			Eastern Ghats Belt			Aravalli			
	AVG	SD	N	AVG	SD	N	AVG	SD	N	AVG	SD	N	AVG	SD	N	AVG	SD	N	AVG	SD	N	AVG	SD	N	AVG	SD	N	
SiO ₂	49.74	1.82	247	49.98	1.70	58	50.90	2.05	96	50.13	1.80	68	51.25	3.07	79	49.92	2.24	46	49.90	1.1	76	51.2	3.05	13	50.30	4.22	25	
TiO ₂	1.11	0.54	247	1.04	0.27	58	1.28	0.41	96	1.16	0.65	68	1.06	0.49	79	1.05	0.55	46	1.61	0.32	76	1.46	0.7	13	1.52	0.78	25	
Al ₂ O ₃	13.26	1.67	247	13.15	2.32	58	13.54	1.19	96	13.39	1.16	68	13.73	1.01	79	13.13	2.69	46	14.6	1	76	14.1	2.3	13	13.1	1.22	25	
Fe ₂ O ₃	14.02	2.24	247	13.35	1.30		14.50	1.93	96	13.65	1.89	68	12.73	2.29	79	13.69	2.27	46	13.03	1.39	76	13.9	3.25	13	13.1	2.68	25	
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CaO	9.60	1.09	247	10.25	1.11	58	9.80	1.11	96	9.49	1.21	68	8.87	1.75	79	8.22	1.75	46	8.71	0.65	76	8.24	1.75	13	9.87	2.07	25	
MgO	8.06	2.69	247	8.07	3.00	58	6.48	2.09	96	7.06	1.80	68	7.16	1.34	79	10.16	6.73	46	5.61	1.15	76	7.49	4.56	13	7.42	1.91	25	
MnO	0.19	0.03	247	0.19	0.02	58	0.20	0.02	96	0.18	0.02	68	0.18	0.02	79	0.19	0.02	46	0.17	0.01	76	0.19	0.05	13	0.24	0.06	25	
Na ₂ O	2.14	0.48	247	2.02	0.53	58	1.96	0.38	96	2.06	0.46	68	2.43	0.64	79	2.13	1.15	46	2.59	0.27	76	1.52	0.65	13	2.07	1.16	25	
K ₂ O	0.85	0.47	247	0.45	0.26	58	0.74	0.26	59	0.81	0.47	68	1.16	0.66	79	0.98	0.80	46	1.01	0.21	76	0.58	0.47	13	0.99	0.90	25	
P ₂ O ₅	0.16	0.13	247	0.11	0.05	58	0.18	0.15	59	0.17	0.19	68	0.12	0.11	79	0.14	0.09	46	0.15	0.03	76	0.14	0.1	11	0.20	0.22	25	
LOI	0.42	0.46	78	0.78	0.40	58	0.64	0.76	76	1.74	0.89	33	1.89	1.21	45	2.25	0.82	46	1.66	0.86	76				1.23	1.30	25	
Total	99.19	1.22	247	99.38	0.44	58	100	0.98	96	98.95	0.68	68	99.67	1.28	79	99.78	0.74	46	99.08	0.64	76	98.9	1.69	13	100	1.09	25	
Mg. No.	0.52	0.10	247	0.53	0.09	58	46.5	8.43	96	0.52	0.09	68	0.54	0.07	79	62.99	13.84	46	45.4	5.37	76	48.6	15.5	13	52.8	9.69	25	
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Trace elements (ppm)																												
Rb	43	32	201	16	9	21	25	9		38	25	53	47	37	79	26.46	17.38	46	34.2	10.8	71	21.9	34.8	13	31.2	40.45	25	
Sr	199	98.1	201	188	68	21	198	72		185	126	53	154	120	79	151.6	63.41	46	261	97.9	71	223	171	13	212	164.25	25	
Ba	198	137	196	124	63	21	257	89		303	489	53	265	257	79	319.24	324.7	46	249	28.4	71	206	108	13	211	291.60	25	
Th	3.7	4.3	197	2.95	2.02	21	2.58	1.63	87		2.02	1.52	3.6	2.6	40	3.68	1.62	46	3.51	1.12	71	1.86	1.16	12				
Pb	15.93	44.01	88	3.87	2.54	21	28.34	49.90	36		3.04	0.48	20	2.2	64	5.70	5.43	46	6.46	2.13	71	5.88	2.07	7				
U	0.83	0.61	130	0.93	1.12	21	0.73	0.22	9		0.46	0.46	50	0.80	137	0.86	0.56	46	0.82	0.36	71	0.59	0.44	12				
V	260	97	97	303	62	21	321	70	60		314	79	27	268	55	79	202.80	49.82	46	314	72.1	71	310	149	13	283	94.15	25
Cr	370	388	175	278	280	21	233	271	78		271	276	53	281	177	79	773.85	999.7	46	87.5	41.6	71	479	566	13	254	193.41	25
Ni	151	107	187	84	65	21	133	91	81		120	62	53	111	68	79	266.30	310.5	46	65.4	22.3	71	220	178	13	102	81.83	25
Co	55	28	75	60	12	21	66	25	54		54	8	20	40	8	79	62.71	22.74	46	51.3	6.54	71	66.9	30.5	13			
Sc	38	27	80	35	6	21	37.5	9	36		46	4	20	37	9	79	34.79	5.80	46	32.3	4.19	71	37.9	12.9	13			
Cu	120	57	86	150	67	21	173	84	34		130	43	20	102	54	77	83.63	37.69	46	43.6	17.6	71	157	303	13			
Zn	101	24	88	113	23	21	112	84	60		76	14	20	94	30	79	91.27	29.02	46	115	29.1	71	160	180	13	629	1413.44	20
Ga	18	4	107	18	4	21	19.9	7	41		18	4	50	17	3	79	11.91	3.61	44	21.3	2.47	71	18.5	5.21	13	20.6	3.98	25
Nb	6.24	5.15	196	4.2	2.0	21	4.65	2.1	86		6.6	6.1	53	6.9	2.5	79	4.82	2.58	46	12.3	4.54	71	10	6.06	13	9.64	5.90	25
Zr	93	40	190	134	123	21	68.5	35	93		84	52	53	114	75	79	80.61	39.10	46	126	27.5	71	60.4	50.6	13	140	109.35	25
Y	24	10	201	32	13	21	27.6	8	66		21	8	53	24	7	79	19.71	8.31	46	24.6	4.67	71	24.3	14.4	13	33.2	16.06	25
Hf	4.40	6.62	122	4.06	3.76	21	1.92	0.83	36		2.06	1.03	50	2.25	1.01	21	2.44	1.06	46	2.93	0.66	7	1.56	1.07	13			
Ta	8.76	37.59	100	0.30	0.15	21	0.4	0.23	36		0.61	0.52	50	0.59	0.36	21	0.67	0.30	46	0.88	0.28	7	5.89	13.9	13			
La	13.91	12.92	123	13.67	6.07	21	11.8	5.20	43		14.00	10.87	44	17.32	18.23	32	12.00	5.41	46	15.2	3.5	71	13	5.91	13	15.7	10.57	20
Ce	27.75	20.03	122	29.30	12.06	21	24.7	10.61	43		29.99	21.33	44	32.50	32.04	32	26.55	10.94	46	33.5	7.58	71	29	12.8	13	31.8	23.54	25
Pr	2.90	4.59	124	3.09	1.21	21	3.15	1.49	34		4.68	4.13	50	2.82	1.47	21	3.06	1.42	46	4.29	0.95	71	3.47	1.49	12			
Nd	14.54	9.05	123	16.99	6.29	21	14.5	6.26	39		15.76	10.15	50	17.47	13.47	32	12.78	6.09	46	18.4	3.98	71	16.6	7.14	13	20.2	13.50	25
Sm	3.90	2.68	124	3.90	1.30	21	3.29	1.33	36		4.10	2.15	50	3.85	2.17	32	3.09	1.47	46	4.45	0.92	71	3.85	1.6	13	5.55	3.45	25
Eu	1.18	0.99	124	1.20	0.36	21	1.07	0.39	36		1.79	1.61	50	1.21	0.60	32	1.13	0.52	46	1.37	0.27	71	1.28	0.43	13	1.7	0.83	25
Gd	4.37	2.36	124	5.41	1.83	21	3.51	1.31	34		4.42	2.18	50	4.46	1.99	32	3.98	1.83	46	4.6	0.96	71	4.26	2.27	13	5.99	3.14	25
Tb	0.40	5.70	117	0.88	0.31	21	0.61	0.23	33		0.88	0.41	50	0.74	0.33	21	0.62	0.28	46	0.74	0.15	71	0.72	0.39	12			
Dy	4.51	2.46	124	4.91	1.75	21	3.76	1.26	33		4.25	1.88	50	4.11	1.45	32	3.88	1.70	46	4.26	0.81	71	4.09	2.31	13	5.99	2.88	25
Ho	0.86	3.02	117	1.05	0.40	21	0.81	0.37	17		1.02	0.46	50	1.20	0.57	32	0.82	0.35	46	0.88	0.17	71	0.87	0.52	12			
Er	3.20	2.86	124	3.29	1.31	21	2.21	0.77	34		2.75	1.17	50	2.79	1.03	21	2.43	1.02	46	2.35	0.45	71	2.4	1.54	13	3.49	1.63	25
Tm	0.53	0.56	117	0.56	0.24	21	0.33	0.14	16		0.51	0.27	50	0.38	0.14	21	0.34	0.15	46	0.35	0.07	71	0.33	0.23	12			
Yb	2.84	1.59	118	3.14	1.36	21	2.28	0.73	36		2.53	1.16	50	2.29	0.82	32	2.26	0.93	46	2.18	0.43	71	2.14	1.41	13	3.35	1.81	25
Lu	0.46	0.25	118	0.47	0.21	21	0.35	0.13	36		0.46	0.27	50	0.34	0.12	32	0.33	0.14	46	0.31	0.06	71	0.33					

DATA SOURCES: Eastern Dharwar Craton: Radhakrishna et al. (unpublished data); French et al., 2010; Damodara Reddy et al., 1995; Srivastava et al., 2014; Western Dharwar Craton: Radhakrishna et al. (unpublished data). Southern Granulite Terrain: Radhakrishna et al., 1995 and

fairly characterized chronologically and palaeomagnetically (Zachariah et al., 1995; Radhakrishna and Mathew, 1996; Dash et al., 2013; Halls et al., 2007; French et al., 2008; French and Heaman, 2010; Kumar et al., 2012; Radhakrishna et al., 2003, 2013a and 2013b; Gregory et al., 2018). Trace element (Radhakrishna et al., 1995; Radhakrishna and Mathew, 1998; Anand et al., 2003; authors unpublished data and as detailed in figure 2 caption), and Sr-Nd and/or O-isotopic compositions are known from some mafic intrusions (Radhakrishna et al., 1995; Radhakrishna and Mathew, 1998; Pandey et al., 1997) and for the Cuddapah trap rocks (Anand et al., 2003).

The clusters of igneous units, represented either in the form of dykes or volcanic eruptions/sills in all cratons/basins, have restricted mineralogical and geochemical compositional ranges. Augite, plagioclase and titanomagnetite $\pm$ minor hypersthene constitute the mineralogy of the Palaeoproterozoic mafic magmas. Most of the mafic rocks are oversaturated quartz tholeiites, but range to olivine tholeiites. In the total alkali-silica diagram of Le Maitre (2002; Fig. 2), these igneous units are sub-alkalic tholeiitic basalts and rarely range into basaltic andesites. Co-magmatic suites of andesitic or dacitic character are virtually absent. They show a clear Fe-enrichment trend, and this is matched by a wide range of REE and trace element abundances. This trend appears to be common for all the Palaeoproterozoic igneous rocks worldwide. Conversely, alkaline magma types are not present in Early Proterozoic igneous suites, as observed on almost every craton across the globe (e.g. North America, Condie et al., 1987; Greenland, Nielsen, 1987; Antarctica, Sheraton et al., 1990; see other compilations in Halls and Fahrig, 1987). Magmas of true alkaline nature appear to become much more significant in the later Proterozoic and particularly after 1.5 Ga (Leelanandam et al., 2006). The trace element geochemistry of the Palaeoproterozoic igneous suites is summarized in the chondrite-normalized REE plots (Fig. 3) and mantle-normalized multi-elemental plots (Fig. 4). The variations within individual suite of rocks are controlled by fractional crystallization, a feature marked by lowering of Mg number with increasing REE abundances. The shapes of the patterns show moderate enrichment in the LREE and lithophile elements, distinct negative Nb (and Ta) anomalies and smaller negative Sr anomalies; some of the samples have small negative Ti anomalies.

The isotope data of the samples exhibit a moderate range in $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios (0.70322–0.7063 for mafic intrusions in granulite region; 0.7052–0.7081 for Traps/intrusive basalts in Cuddapah Basin). Their $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios vary between 0.5097 and 0.5102; ϵNd values correspondingly range from +1 to -10. Sr and Nd isotopic values of similar range are also reported from 1.88 Ga BD2 dykes in Bastar craton (Srivastava et al., 2009) and Bijawar igneous units (Pandey et al., 2012). In a Sr-Nd isotope plot (figure not shown), the Proterozoic mafic samples strikingly differ from the mantle components defined by Hart and Zindler (1989). The oxygen isotope data provide an important insight to the petrogenetic problem. Available whole rock

Figure 2. Total alkali-silica (TAS) diagram (after Le Maitre, 2002) for Palaeoproterozoic mafic igneous rocks from the Indian shield. The samples show tholeiitic basalt to basaltic andesite composition. Some samples marginally spread to outside the field probably due to some alteration affects. Dashed line separates tholeiitic and alkaline fields. Total 708 sample data are plotted. Data source: as given in Table 1.

$\delta^{18}\text{O}$ values for the Indian Proterozoic igneous intrusives (Fig. 5) are rather uniform and mostly ranging between +5 and +6‰, relative to standard mean ocean water (SMOW). By comparison, the mean $\delta^{18}\text{O}$ value for typical MORB glasses is $+5.5 \pm 0.2$ ‰ (e.g. Eiler 2001; Cooper et al., 2004). For fresh whole-rock samples of ocean island basalt, values of $\delta^{18}\text{O}$ generally are $< +5.8$ ‰ (Kyser et al., 1990; Eiler, 2001).

Most of the whole population of Palaeoproterozoic igneous rocks from India have Mg numbers < 0.60 and correspondingly MgO values are < 9.0 wt% (down to 4 wt%) with Cr and Ni values as low as < 200 ppm and < 100 ppm respectively indicating variable degrees of fractionation. Only some mafic to ultramafic sill samples in the Cuddapah Basin have their MgO contents ranging up to 28 wt% due to olivine accumulation. A plot between Mg# and $(\text{La}/\text{Lu})_N$ (Fig. 6) clearly indicates the dominant role played by fractional crystallization in the compositional variation of the Palaeoproterozoic magmas in the Indian shield. Only a few samples show effects of assimilation fractionation. Vectors for partial melting, fractional crystallization and assimilation fractional crystallization are shown in the Figure 6. It seems Proterozoic magmas from the Bastar and Bundelkhand cratons are formed by higher degrees melting compared to those from Gwalior and Cuddapah basins as illustrated by lower $(\text{La}/\text{Lu})_N$ ratios at similar Mg#. Alternatively, the mantle sources for Proterozoic magmas from the Bastar and Bundelkhand cratons could be more enriched compared to those for Gwalior and Cuddapah basins. The fractional crystallization involved olivine + clinopyroxene $\pm$ plagioclase $\pm$ orthopyroxene. It is interesting that the fractionation trends are Fe-rich tholeiitic and calc-alkaline trends are virtually absent. This appears true for the early Proterozoic rocks worldwide. Even the Scourie dykes (Weaver and Tarney, 1981) with variable amounts of primary hornblende ($\pm$ biotite) that reflect high $p\text{H}_2\text{O}$ conditions during crystallization never follow calc-alkaline fractionation trend with the implication that high $p\text{H}_2\text{O}$ is not the only factor determining fractionation trends; these conditions are suggested to have been inherited at source region. The mantle sources

Figure 3. Chondrite-normalized rare earth element patterns showing variation within the Palaeoproterozoic mafic igneous occurrences from cratons and sedimentary Basins of the Indian shield. Chondrite values are after Sun and McDonough (1989). For discussion see the text. Legend Abbreviations: SGT = Southern Granulite Terrain; EGB = Eastern Ghats Belt; EDC = Eastern Dharwar Craton; WDC = Western Dharwar Craton.

Region-wise details are as follows: Agali-Coimbatore mafic dykes (South Indian Granulite Terrain); Anantapur-Gooty mafic dykes (Eastern Dharwar craton on the margins of Cuddapah Basin); Aravalli, Bundelkhand and Bastar cratons in Northern and Central India; Cuddapah and Gwalior are traps/sills within the Cuddapah Basin in south India and Gwalior Basin in northern India; Kondapalli-Kandra mafic rocks are from the Eastern Ghats Belt; Udupi-Mangalore mafic dykes (Western Dharwar craton).

Figure 4. Primitive mantle-normalized multi-elemental plots showing variation within the Palaeoproterozoic mafic igneous occurrences from the Indian shield. Primitive mantle values are after Sun and McDonough (1989). Abbreviations and regional-wise details are as given in figure 3. For discussion see the text.

Figure 5. Whole-rock $\delta^{18}\text{O}$ values from mafic igneous intrusive samples of south India compared with the values of Continental and Oceanic mafic lavas (Kyser, 1990). Values from late Cretaceous (65–70 Ma) dolerites of Central Kerala are also plotted for comparison. Figure modified after Radhakrishna et al. (2004).

Figure 6: $\text{Mg\#}$ versus $(\text{La/Lu})_N$ plot for the Palaeoproterozoic mafic igneous rocks from the Indian shield. Geochemical variation in the Palaeoproterozoic magmas in the Indian shield can be explained by fractional crystallization wherein for a large decrease in $\text{Mg\#}$ s there is only a little increase in $(\text{La/Lu})_N$ ratios. Only few samples show possible effects of assimilation fractional crystallization. It seems Proterozoic magmas from Bastar and Bundelkhand cratons are formed by higher degrees melting compared to those from Gwalior and Cuddapah basins as illustrated by lower $(\text{La/Lu})_N$ ratios at similar $\text{Mg\#}$ s (shown in dotted squares). Vectors for fractional crystallization (F), assimilation fractional crystallization (AFC) and partial melting (PM) are shown. Legend Abbreviations as in figure 3

beneath the old Archaean cratons seem to be quite reduced (Daniels and Gurney, 1991).

The compositions (LREE and LIL elemental enrichment and negative Nb anomalies) of the Proterozoic continental mafic igneous rocks are argued to be the result of massive crustal contamination of magma while some others prefer, in sharp contrast, contamination of the mantle source itself. Incompatible element ratios and isotopic ratios are used to negate significant crustal contamination of the Proterozoic mafic dyke intrusions in south India (Radhakrishna et al., 1995; Radhakrishna and Mathew, 1998) and in Bundelkhand craton (Radhakrishna et al., 2019). Furthermore, there are no field evidences of crustal melting associated with any of the Proterozoic igneous occurrences in India. It is demonstrated in many Proterozoic igneous units that crustal contamination has no significant role in petrogenesis (for example, Scourie dykes: Weaver and Tarney, 1981 and Waters et al. 1990; Greenland norite dykes: Hall and Hughes, 1990; East Antarctic Proterozoic dykes: Sheraton et al., 1990). Available $\delta^{18}\text{O}$ values of the Indian Palaeoproterozoic igneous units are not in conformity with any significant crustal contamination. A comparison between $1/\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 7) of all available data from different igneous units in India is inconsistent with a simple mixing model between crust and mantle-derived melts although it is impossible to rule out non-systematic mixing with the present data set. However, Anand et al. (2003) argued for significant contamination of mantle melts with local crust in some samples of the lavas and intrusives of the Cuddapah Basin.

Majority of the Palaeo-proterozoic rocks from the Indian shield fall in the field for sub-continental lithospheric mantle derived basalts except a few samples (particularly Kondapalli-Kandra) falling within the field for sub-lithosphere mantle in the plot defined by Th/Yb versus

Ta/Yb variation (Fig. 8A; Pearce, 2003). Some samples from Bastar, Bundelkhand extend into the sub-lithosphere field. For comparison, Phanerozoic Deccan flood basalts are shown in the figure. The Deccan basalts are grouped into Plume-derived and Lithosphere-derived based on geochemical and isotopic signatures (Vijaya Kumar et al., 2018). Majority of the Plume-derived basalts are restricted to the sub-lithosphere array. However, some of the plume-derived basalts plot above the field for melts derived from sub-lithosphere mantle sources (Fig. 8A). Such a variation is interpreted in terms of overprinting on original sub-lithospheric mantle sources (Pearce, 2003) or crustal contamination of deeper mantle derived magmas (Arculus, 1987). The Deccan basalts with strong lithosphere component exclusively plot above the field for melts derived from the sub-lithospheric mantle sources (Fig. 8A). It is clear that none of the Palaeoproterozoic mafic magmas and Deccan igneous rocks illustrate true

Figure 7: $^{143}\text{Nd}/^{144}\text{Nd}$ versus $1/\text{Nd}$ plot for the Palaeoproterozoic mafic magmas from the Indian shield. Samples from mafic intrusives of granulite region south of Nilgiri massif (Agali; Radhakrishna et al., 1995), mafic flows/sills in the Cuddapah Basin (Anand et al., 2003), MB Nagar mafic dykes on the margin of Cuddapah Basin (Pandey et al., 1997), Bijawar igneous rocks in Central India (Pandey et al., 2012) and Raigarh region in Bastar (Pandey et al., 2012) are plotted. Depleted mantle (DM) of Miller and O'Nions (1985) and Tonalite Gneiss and Granite Gneiss values of Peucat et al. (1989) are also plotted. These compositions differ sharply from the compositions of Neoarchaean boninite units of the Bastar region (Srivastava et al., 2009).

NMORB signatures alone (Fig. 8A). We suggest that a dominant lithosphere component for most of the mafic rocks of the Palaeoproterozoic and Phanerozoic igneous provinces of India.

Similar results are obtained based on Th/Nb and Zr/Nb relationship (Fig. 8B), which demonstrates that very few mafic rocks from Bastar, Bundelkhand cratons have lower Th/Nb (<0.15) ratios – geochemical characteristics similar to sub-lithospheric mantle sources (Fig. 8B). Typical depleted mantle derived NMORB-type melts are located remotely from the Palaeoproterozoic mafic magmas from the Indian shield. We suggest that mafic magmas with Th/Nb ratios >0.4 (Fig. 8B) represent melts exclusively derived from or with significant contribution from sub-continental lithospheric sources. Mafic magmas with Th/Nb ratios between 0.15 and 0.4 overlap, which we interpret in terms of lithosphere-plume interaction.

Mantle Sources

Sources responsible for enriched geochemistry (subducted oceanic crust vs SCLM)

The above arguments point out that the mantle source of the Indian Proterozoic mafic igneous units is actually enriched in incompatible elements than the primitive mantle (PM: Sun and McDonough, 1989; Lyubetskaya and Korenaga, 2007). Such enriched sources are now known in many continental basalt provinces and the ocean island basalts. However, the processes of enrichment significantly vary as reflected by abundances and ratios of incompatible elements, as well as radiogenic isotopes. The heterogeneities resulting from addition of distinctly different metasomatic agents are generally accepted due to plate and plume tectonics.

Figure 8: Th/Yb versus Ta/Yb (A) and Zr/Nb versus Th/Nb (B) variations in the Palaeoproterozoic mafic igneous rocks from the Indian shield (the colour and symbol codes are as given in figure 6). Plume- and Lithosphere-derived basalts from the Phanerozoic Deccan flood basalt province are shown for comparison. Fields for sub-lithospheric mantle, island arc and active continental margin are after Pearce (2003). The compositions for NMORB (normal mid-ocean ridge basalt), EMORB (enriched mid-ocean ridge basalt, PM (primitive mantle) and OIB (ocean island basalt) are after Sun and McDonough (1989); composition of UCC (upper continental crust) is after Rudnick and Gao (2003). Vectors for addition of subducting component (S), upper crustal contamination (UC) and partial melting (M) are shown.

Three major hypotheses were proposed for the enriched component(s) in the source in terms of incompatible trace elements (see for details Niu et al., 2012):

- (1) The hypothesis of recycled oceanic crust wherein the subducted ocean crust returns to upper mantle via core-mantle boundary as proposed by Hofmann and White (1982). However, the incompatible trace element chemistry is not in conformity with the proposed recycled oceanic crust contribution. Relatively more incompatible trace element patterns as in the early Proterozoic magmas cannot be reproduced from the recycled oceanic crust either theoretically or experimentally (Hirschmann et al., 2003; Kogiso et al., 2003; Dasgupta et al., 2007; Pilet et al., 2008). The model

subsequently underwent some revision. According to the revised model, pyroxenite sources develop by melt-rock reaction (reaction of the eclogite-derived melts with peridotite; Sobolev et al., 2005 and 2007) or solid state diffusion reaction between the eclogite (recycled oceanic crust) and harzburgite (depleted mantle) (Herzberg, 2011). These models developed by using the geochemical database of Hawaiian volcanism. Sobolev et al. (2007) very elegantly reinforced the idea to quantify the recycled ocean crust and pyroxenite content under different continental/oceanic basalt regions of different lithosphere thicknesses. For example, 10-20% recycled oceanic crust in the sources for the MORB, up to 60% for the Hawaii OIB and 100% for the Siberian basalts. However, both the recycled oceanic crust (represented by eclogite) and the depleted mantle (represented by harzburgite) are depleted in incompatible trace elements and cannot produce enriched compositions of the Indian Palaeoproterozoic igneous rocks or other similar Phanerozoic basalt eruptions. For this reason and other reasons explained by Niu et al (2012), there are experimental and physical difficulties to support the reactions proposed in these models.

- (2) Some authors have suggested that subducted terrigenous sediments (upper continental crust derived material) may be responsible for enriched signatures of mantle sources in terms of both incompatible trace elements and isotopes (e.g., Chauvel et al., 1992; White and Duncan, 1996; Hofmann, 1997; Jackson et al., 2007). In light of plate tectonics and mantle circulation models and high P experimental studies that show enriched incompatible trace element signature of sources can be attributed to recycled K-hollandite-bearing continental sediments to Transition Zone (App et al., 2008). However, subducted terrigenous sediment signatures may have been diluted in the mantle source regions and continental crust-like signatures may have been largely “smoothed out”. The Palaeoproterozoic igneous units also have subtle Sr depletions (Figure 4) in contrast to its enrichment in the subducted sediments (Figure 4 of Niu et al., 2012). Hence, it may be difficult to conceive subducted sediment to be the major source material for mantle metasomatism and the possibility requires further investigation.
- (3) It has been recognized that mantle sources of the intra-plate continental magmatism, such as the Palaeoproterozoic igneous rocks in the Indian shield, are enriched in incompatible elements relative to the PM, e.g., $[La/Sm]_{EM} > [La/Sm]_{PM}$ (e.g., Sun and McDonough, 1989; McKenzie and O’Nions, 1995; Niu et al., 2002; Niu and O’Hara, 2003; Prytulak and Elliott, 2007; Humphreys and Niu, 2009). Sub-continental lithospheric mantle (SCLM) is postulated as a potential enriched source for many continental basalts (McKenzie and O’Nions, 1995). The geochemically enriched component in the SCLM must be of metasomatic origin as evidenced by the presence of hydrous minerals (e.g., amphibole, phlogopite) and vein lithologies (e.g., garnet pyroxenite, pyroxenite and hornblende) and the idea gained wider acceptance based on mantle xenolith data (Frey and Green, 1974; Frey et al., 1978; Menzies and Murthy, 1980; Menzies, 1983; Menzies and Hawkesworth, 1987; O’Reilly and Griffin, 1988; Takazawa et al., 2000; Agashev et al., 2013; Hughes et al., 2014; Kuku^{3a} et al., 2015; Ionov et al., 2015; Puziewicz et al. 2015 and

many others) and massif peridotites (Frey et al., 1985; Pilet et al., 2005). Although, the involvement of metasomatized SCLM is accepted, the mechanism of metasomatism in the SCLM that is, how the metasomatized materials enter source regions is not clearly known. However, the mechanism is suggested to be conceptually similar to that in the growing oceanic lithosphere (Niu et al., 1999, 2002; Niu and O’Hara, 2003, Niu, 2009; Humphreys and Niu, 2009)

Mechanism of SCLM Development

To begin with, the primitive mantle or a depleted mantle (as a result of previous melt removal) may be regarded as pre-metasomatic mantle sources; their metasomatism has been ascribed to infiltration by a low-*F* melts that are enriched in volatiles (e.g., H₂O and CO₂) and incompatible elements. The experimental works by White and Wyllie (1992) and Baker and Wyllie (1992) stressed the importance of incipient partial melting in the presence of both H₂O and CO₂ in generating the metasomatic agents which can modify both oceanic and continental lithospheres. Some investigators have reiterated the original idea of Wyllie proposal that the Low Velocity Zone (LVZ) may be entirely or at least largely caused by CO₂-generated incipient melts (e.g., Yaxley et al., 1998; Presnall and Gudfinnsson, 2005, 2008; Dasgupta et al., 2007). These results coupled with the observation that free volatile components available in the upper mantle must be CO₂ or dominated by CO₂ over H₂O (Keppler et al., 2003) suggest that CO₂-rich melt or carbonatite melt may be the metasomatic agent in the mantle.

In order to test the possible carbonatite melt metasomatism, the more primitive magma compositions of the Palaeoproterozoic (in order to eliminate the effects of fractional crystallization or assimilation fractional crystallization, if any) in India are compared (Fig.9) with the carbonatite melts of deep transition zone origin (Walter et al., 2008) and the compositional variants of the Indian carbonatites (Ackerman et al., 2017). The Palaeoproterozoic magmas, despite having some similarities, have significant deviations from the carbonatite melts of deep transition zone, natural carbonatites and the carbonatite melts simulated in high pressure experiments (Brey et al., 2008). These differences in the near primary melts indicate that carbonatite melts alone cannot be the metasomatic agents but low H₂O-CO₂-rich fluids or silicate melts are necessary for the formation of SCLM. Thus, silicate melt compositions when plotted in Figure 9 (not shown), it is evident that enrichment through addition of these liquids would enhance the LILE, but would also produce considerably more than the observed levels of less incompatible elements (LREE, Sr, P, Zr and Nb). Therefore, the observed enrichment pattern (LILE>LREE>HFSE) in the Palaeoproterozoic magmas suggests metasomatism by an incompatible element enriched fluid phase rather than by small degree partial melts. We note that Cawood et al (2018) recognised the role of fluids/melts for the enriched lithosphere mantle, but linked the process to subducted slabs that are refuted above.

These enriched components are believed to reside in the accessory mineral phases. Amphibole, apatite and phlogopite have an important role as accessory phases in producing LREE and LILE enriched melts. Judging from the abundances of LILE, P and LREE, apatite and phlogopite are not the enriched phases of the SCLM and amphibole seems to be more appropriate as the metasomatic phase. The higher $Kd_{amph-liq}$ for Nb in mantle amphibole (Ionov and Hofmann, 1995) and the observation that amphibole can produce large negative Nb

anomalies (Chazot et al., 1994) also do indicate the presence of amphibole as metasomatising mineral phase in the SCLM source under the Indian shield. Experimental results, showing that amphibole is stable up to about 75 km depth (Ringwood, 1975) and the observation that mantle sources of Hawaiian-type basalts may contain up to 10% amphibole (Hammond, 1986) are in agreement with the possible presence of amphibole in the source mantle.

Role of SCLM (melting vs interaction with asthenosphere melts)

At this stage, it is more relevant to assess the role of SCLM in the generation of Palaeoproterozoic magmas in the Indian shield. Gallagher and Hawkesworth (1992) argued that significant volume of melts can be produced from hydrated lithospheric mantle and the melts would have compositions controlled by amphibole. Arndt and Christensen (1992) argued that any thermal perturbation in lithospheric mantle cannot result in large melts generation, even if the lithosphere is wet; the Nb troughs can be related to a reaction of asthenosphere-derived magmas with lithosphere. The relative depletion of HFSE and the negative Nb anomalies in some flood basalts are suggested to have resulted by means of magma-mantle interaction (Kelemen et al., 1990). Similarly, the incompatible element enrichment with low Nb in the Proterozoic dykes of Sweden is ascribed to interaction of asthenosphere-derived melt-lithosphere mantle (Patchett et al., 1994). The present data on Proterozoic Indian igneous units may not suggest that they are entirely lithosphere mantle melts, but indicate possible interaction of lithosphere with plume/asthenosphere melts (see Fig. 8).

Timing of SCLM development

Having established the contribution of the SCLM in the petrogenesis of the Palaeoproterozoic magmas, the timing of its formation through metasomatism is an important petrological aspect. Its formation could be near contemporaneous with the generation of these magmas or alternatively, it might have inherited similar enriched elemental characteristics during the early stages of crustal building activity in the Indian shield. The isotopic data for one of the mafic dyke suites (Radhakrishna et al., 1995), which gave Sm-Nd whole rock "isochron" of ca. 3.0 Ga, is remarkably in

Figure 9: Least fractionated Palaeoproterozoic mafic igneous compositions (A) in comparison with (B) compositional variants of Indian carbonatites (Ackerman et al., 2017) and (C) transition zone carbonatite composition (Walker et al., 2008). Values normalized to primordial mantle values (Sun and McDonough, 1989).

agreement with crustal residence ages (T_{DM} model ages) of 2.87 ± 0.08 Ga estimated from the Palaeoproterozoic igneous units in south India and also with the model ages reported for the ~2.1 Ga old Gwalior traps and for the 2.7 Ga BD1 dykes in Bastar craton. The only exception is the model ages from 1.88 Ga BD2 dyke intrusions that

have yielded significantly lower model ages from rest of the data. These results are perfectly consistent with the suggestion that this age (~2.9 Ga) marks the timing of processes associated with the formation and emplacement (underplating) of the SCLM that is some 0.6 to 1.0 Ga prior to the Palaeoproterozoic igneous activity. Coincidence of this age with many residence ages near 3.0 Ga from the granulite crustal rocks of the Nilgiri massif (Peucat et al 1989) and 3000-2900 Ma trondhjemite plutonic activity in India (for example, Naqvi and Rogers, 1987; p. 19; Radhakrishna and Ramakrishnan, 1993) suggest that the SCLM formation under the Indian shield is synchronous with a major crustal stabilization event. Similarly, simultaneous formation of SCLM and basement stabilization much prior to the Palaeoproterozoic igneous activity is known from Precambrian terrains elsewhere (the Kaapvaal craton in South Africa: Marsh et al., 1989; Lewisian complex in Scotland: Waters et al., 1989). Furthermore, this age is marked broadly by the timing of stabilization of lithosphere and significant changes in Earth behavior (Cawood et al., 2018).

A voluminous hydrous lithospheric mantle across the cratons appears to have developed during the Archaean (~3 Ga), and it has contributed significantly to the Palaeoproterozoic magmas. Despite broad similarities with tholeiitic composition, there exist some subtle differences in the enriched characteristics and some of the incompatible element ratios. For example, at least two subgroups of magmas were identified in the Palaeoproterozoic mafic dyke samples of the Bundelkhand (Radhakrishna et al., 2019). However, more striking is the absence of Fe-rich noritic magmatism that marked the significant Palaeoproterozoic magmas of the Scourie dykes in Lewisian complex (Weaver and Tarney, 1981) and the Ferro-picrite occurrences, which are widespread in many Archaean cratons during Neoarchaeon, in the large volumes of Palaeoproterozoic magmas in the Indian shield. Does it anyway relate to the supercontinental configurations during that time or does it indicate heterogeneities were distributed throughout the early Proterozoic-Late Archaean mantle (for example, Milidragovic et al., 2014), or such Fe-rich domains, if any, under the Indian shield were exhausted during the Neoarchaeon melting as suggested by Milidragovic et al. (2016).

Thermal Constraints

Two significant features of Palaeoproterozoic mafic igneous activity- the widespread distribution indicating large volumes of magma generation and apparent occurrence of each magma pulse within a narrow time bracket- require enough thermal energy to generate the mantle melts and to turn-on and turn-off the thermal tap very quickly. It has been common to appeal to mantle plumes to supply this energy for melting in spite of divergent views on the mechanism (White and McKenzie, 1989; Griffiths and Campbell, 1990; Richard et al., 1989). It may be noted that the Palaeoproterozoic magmas of the Indian shield and elsewhere share common geochemical characteristics (enrichment of LILE and LREE; Nb and Ta depletion) with the Phanerozoic continental flood basalts that are mantle plume derived. The main distinction is that the Phanerozoic continental flood basalts occur as widespread extrusives whereas the Palaeoproterozoic magmas emplaced mainly as dyke intrusions without large associated extrusives. Although giant dyke swarms are known to be feeders for many plateau basalts, the absence of extrusive pulses during the Palaeoproterozoic magmatism appears real and not an artifact of erosion. The differences between mode of emplacement of

Phanerozoic (with large extrusive pulses) and Palaeoproterozoic (with little or no extrusive magmas) mafic magmas may have some implications for the depths of origin of the mantle plumes. The mantle plumes for the continental flood basalts might have their origin at greater depths, at the CMB (core mantle boundary) and developed large head diameters (>1000 km; White and McKenzie, 1989) on arrival at the base of the lithosphere resulting in large-flood volcanism. On the other hand, the Palaeoproterozoic magmas may have formed from plumes of shallow origin at 670 km depth and lack sufficiently large thermal energy and caused only lithosphere attenuation and the crustal fractures filled with magmas that are manifested largely as dyke intrusions. Such an interpretation is consistent with the experimental results of Griffith and Campbell (1990). However, the mantle potential temperatures estimated by the geochemical modeling of Cuddapah igneous rocks (Anand et al., 2003) do not indicate major thermal anomaly supporting a passive rifting model. Resolution of these issues becomes the focus for future studies.

Conclusions

1. Palaeoproterozoic mafic magmatism is widespread in the Indian Shield and mostly emplaced as dyke intrusions and seldom as traps in the Proterozoic sedimentary basins. It is seen as almost exclusive mafic magmatism with little felsic rock association.
2. U-Pb ages coupled with palaeomagnetic results identified at least four discrete Palaeoproterozoic igneous events at 2.36-2.37 Ga, 2.1-2.2 Ga, 2.1-2.0 Ga 1.99-1.89 Ga and probably two other events at about 2.4 Ga and 1.8 Ga.
3. All the Palaeoproterozoic magmas are sub-alkaline tholeiitic in composition with enrichment of large ion lithophile and light rare earth elements. Their alkaline counterparts are virtually absent.
4. The geochemical variation in the Palaeoproterozoic mafic magmas can be explained by simple fractional crystallization. Effects of crustal assimilation are shown by very few samples.
5. Sub-Continental Lithosphere Mantle (SCLM) seems to be the major contributor of the enriched geochemistry of the Palaeoproterozoic mafic magmas emplaced in the Indian Shield. Raising mantle plumes may have supplied the material and thermal energy for the initiation of melting; however, passive rifting is an alternative explanation that needs to be examined further.
6. SCLM contribution as an overprint on asthenosphere melts is very evident in the geochemistry of the Palaeoproterozoic magmas.
7. The SCLM beneath the Indian Shield was possibly enriched by addition of fluids/melts of deep mantle origin. Amphibole seems to be the major host for the enriched components.
8. We speculate that the major segments of the Indian SCLM were generated in the early part of the earth history at around 3 Ga and are linked to the major crustal building activity in the Indian shield.

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