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ORIGINAL ARTICLE

Petrogenetic study of Mesoproterozoic volcanic rocks of North Delhi fold belt, NW Indian shield: implications for mantle conditions during Proterozoic

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Abstract Mesoproterozoic North Delhi fold belt of NW Indian shield comprises three volcano-sedimentary basins viz. Bayana, Alwar and Khetri aligned parallel to each other from east to west. Each basin contains excellent exposures of mafic volcanic rocks. Major, trace and rare earth element abundances of volcanic rocks of the three basins are significantly diverse. Bayana and Alwar volcanics are tholeiites bearing close similarity with low Ti-continental flood basalts. However, Bayana volcanics are characteristically enriched in incompatible trace elements and REEs while Alwar volcanics display least enriched incompatible trace element abundances and flat REE patterns. The Khetri volcanics exhibit a transitional composition between tholeiite and calc-alkaline basalts. REE based source modeling suggests that Bayana suite was formed from the melts derived from 1 % to 10 % (avg. 4 %) of the partial melting of a spinel lherzolite source giving a residual mineralogy of 56 % Olv, 25 % Opx and 19 % Cpx. Whereas Alwar suite evolved through 12 %–20 % (avg. 15 %) partial melting of the same source with a residual mineralogy 61 % Olv, 25 % Opx and 14 % Cpx. Khetri volcanics are exposed at two localities Kolihan and Madhan–Kudhan. The Kolihan volcanics were derived from 1 % to 6 % (avg. 4 %) partial melting with residual mineralogy 56 % Olv, 25 % Opx and 19 % Cpx whereas the magma of Madhan Kudhan volcanic suite was generated by 15%–30 % partial melting of the same source leaving behind 64 % Olv, 25 % Opx and 11 % Cpx as residual mineralogy. This source modeling proves that melts of Bayana and Alwar tholeiites were generated by partial melting of a common source within the spinel stability field under the influence of mantle plume. During the course of ascent, Bayana melts were crustally contaminated but Alwar melts remained unaffected. There was two tier magma production in Khetri region, one from the partial melting of the mantle wedge overlying the subducted oceanic plate which formed Kolihan suite and two the melting of the subducted plate itself generating Madhan–Kudhan volcanics. It is interpreted that during Mesoproterozoic (1,800 Ma), the continental lithosphere of NW Indian shield suffered stretching, attenuation and fracturing in response to a rising plume. Consequently, differential crustal extension coupled with variable attenuation brought the asthenosphere to shallower setting which led to the production of tholeiitic melts. These melts enroute to the surface suffered variable lithospheric contamination depending upon the thickness of traversed crust. The Khetri basin attained maturity which resulted in the generation of true oceanic crust and its subsequent destruction through subduction. The spatial existence of three suites of mafic volcanics of diverse chemical signatures is best example of subduction–plume interaction. It is therefore, proposed that the Mesoproterozoic crust of NW Indian shield has evolved through the operation of a complete Wilson cycle at about 1,832 Ma, the age of mafic volcanics of Khetri basin.

Keywords Geochemistry · Source modeling · North Delhi fold belt · Mesoproterozoic volcanism · Indian shield · Aravalli craton

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1 Introduction

Aravalli-Delhi fold belt of northwestern Indian shield is a unique terrain where rock-records covering the whole Proterozoic era are well preserved (Heron 1953; Roy 1988). In this region, along with different types of metasediments, there exists wide spread occurrences of mafic magmatic rocks that provide important clues regarding plate boundaries during the Proterozoic (Ahmad and Tarney 1991; Khan and Raza 1993; Raza and Khan 1993; Sinha-Roy 2000; Abu-Hamatteh et al. 1994; Raza et al. 2007). These volcano-sedimentary sequences were emplaced on an Archean basement (3,300–2,500 Ma; Gopalan et al. 1990) popularly known as banded gneissic complex (BGC; Heron 1953).The entire Aravalli–Delhi belt is divisible into two segments occurring to the south and north of Ajmer city (Fig. 1). The southern part comprises two major groups of supracrustal sequences i.e. Aravalli

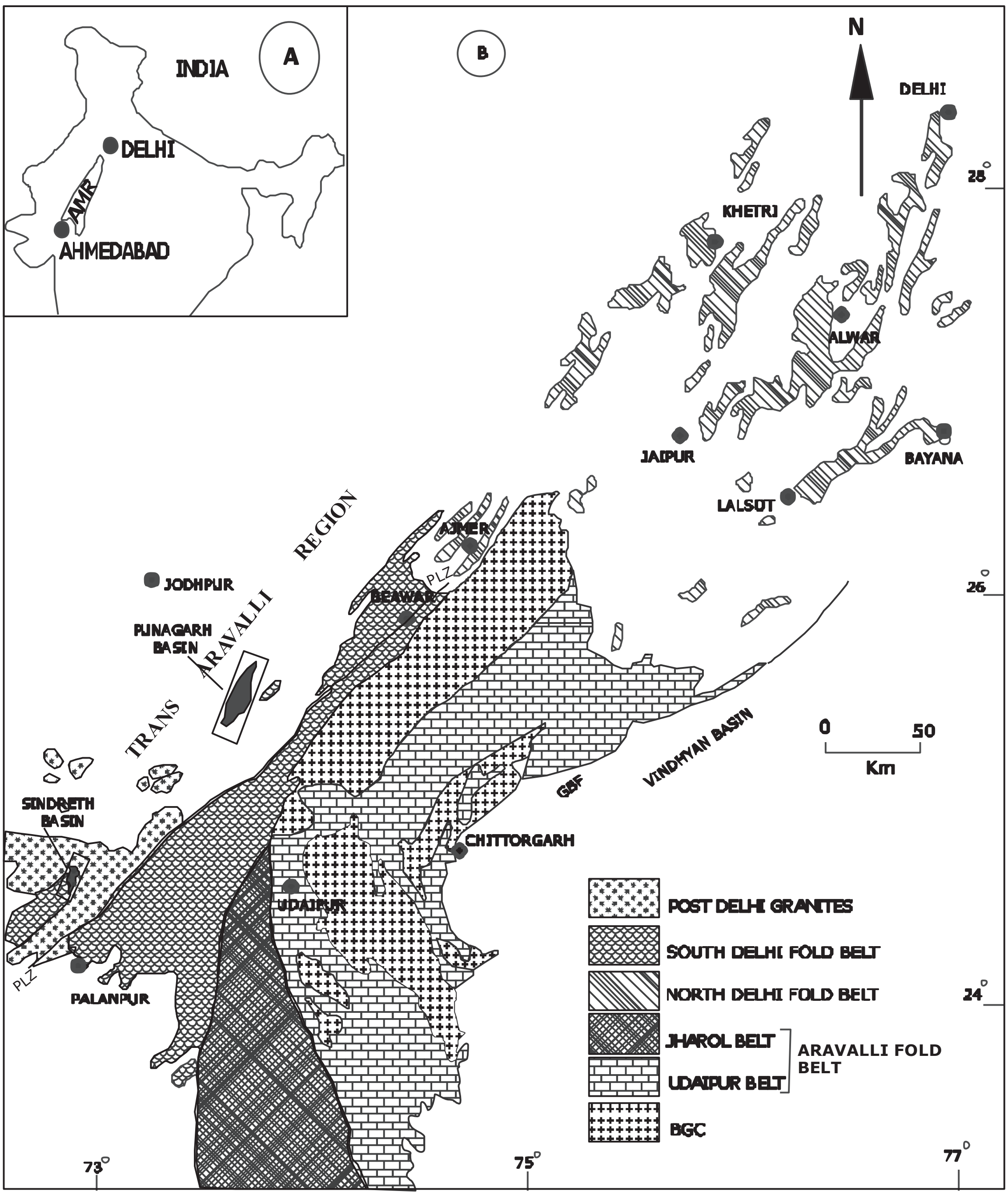

Fig. 1 *a* Outline map of India showing position of Aravalli mountain range. *b* Simplified geological map of Aravalli mountain range of Aravalli craton (after Gupta et al. 1997) showing distribution of various lithological domains. (*PLZ* Phulad lineament zone, *GBF* great boundary fault)

(Palaeoproterozoic) and Delhi (Mesoproterozoic) supergroups. The northern segment, referred to as North Delhi fold belt (NDFB), is entirely composed of the Mesoproterozoic rocks included in Delhi Supergroup. The NDFB is broadly constituted by three sedimentational domains (Fig. 1) which are from east to west: the Bayana basin, the Alwar basin and the Khetri basin (Singh 1982). Each of these domains is characterized by the occurrence of well preserved volcanic sequences.

Since a close relationship between tectonism and magmatism is well established (Condie 1982; Pearce 1983; Watters and Pearce 1987), the study of mafic rocks provide excellent opportunity to understand Proterozoic tectonics, crustal growth and conditions of the subjacent mantle at the time of their formation in this part of India shield.

The geochemical characteristics of primary magmas are governed by various factors such as composition of their mantle sources, degree of partial melting and physico-chemical conditions at the time of their generation (Bertrand 1991). However, the processes like fractional crystallization, magma mixing and crustal contamination modify the primitive composition of the magmas which is often manifested by diverse rock types occurring in a region. These controls vary from one tectonic setting to another, and thus mafic rocks are evolved with distinct chemical characteristics in specific tectonic environment. Therefore, the geochemical studies of mafic Proterozoic sequences can contribute significantly in understanding the nature of mantle source(s) and in turn magma genesis.

The aim of this study is to constraint the petrogenetic history of NDFB volcanics with the help of hypothetical source modeling and the effect of degree of partial melting experienced by the melts using their geochemical attributes.

2 Geological setting

Detailed account of the geology of the Aravalli mountain belt and study are given in Raza et al. (2001, 2007). In brief, Aravalli mountain belt of north western India shield (Fig. 1) runs for more than 700 km between Delhi in the north and Ahmadabad in south (Roy 1988). This belt fringes the northwestern margin of the Indian shield and is constituted by lithologic association of various ages ranging in age from Archean to Neoproterozoic. Its main lithoconstituents are the Archean banded gneissic complex (BGC), Paleoproterozoic Aravalli supergroup, Mesoproterozoic Delhi supergroup and Neoproterozoic Vindhyan supergroup. BGC forms the basement for the Proterozoic supracrustal sequences of Aravalli, Delhi and Vindhyan supergroups.

BGC is a composite, structurally complex gneissic terrain showing great lithological similarity to the Archean gneissic complexes occurring in other parts of the Indian shield (Naqvi and Rogers 1987). This is an ensemble of varied rock types including TTG, K-granite, granitic gneisses, myrmekites, pegmatites and amphibolites with minor amount of mafic–ultramafic rocks and sediments, amongst which granitoid gneisses of different compositions and amphibolites constitute the bulk of this basement.

The Aravalli Supergroup is exposed in Udaipur and Jharol belts. The Udaipur belt is a komatiite–tholeiite bearing carbonatic sedimentary shelf facies whereas the Jharol belt is made up of carbonate-free deep water facies in association with ultramafic rocks (Roy and Paliwal 1981).

The Mesoproterozoic Delhi supracrustals are the major constituent of Aravalli mountain belt. These rocks are exposed all along the length of Aravalli mountain range in a linear belt. Based upon outcrop style and other lithological attributes, the Delhi belt has been divided into two sub belts viz. North Delhi fold belt (NDFB) and South Delhi fold belt (SDFB) across Jaipur–Dausa transcurrent fault (Sinha Roy 1985). The SDFB occurs to the east of Jharol belt and consists of highly folded and deformed rocks exhibiting polyphase deformation. Its western margin is marked by the occurrence of Phulad ophiolitic mélange (Gupta et al. 1980). Geochemical studies of the mafic volcanics of Phulad ophiolite advocate that it is a fragment of island arc (Khan et al. 2005) which developed at the time of closure of Delhi basin.

NDFB, the volcanics of which are the subject of present study, comprises three sub parallel basins (Fig. 1). These are, from east to west, the Bayana basin, the Alwar basin and the Khetri basin. Their volcano-sedimentary infills have been divided into two groups namely Alwar and Ajabgarh representing lower and upper divisions (Heron 1953).The Bayana basin consists of 3,000 m thick volcano-sedimentary sequence comprising quartzite, conglomerate, shale and lava flows (Fig. 2). The lava flows are confined only to its basal part. The volcano-sedimentary infill of Alwar basin contains volcanics along with sandstone, siltstone and minor conglomerate. The lithlogic assemblage as a whole is estimated to be about 5,000 m thick in Alwar basin (Banerjee and Singh 1977). The Khetri basin which hosts Khetri Copper belt is constituted by schists, phyllites with interlayered quartzite, metagrewackes, marbles, calc-silicate rocks and ortho-amphibolites (Dasgupta 1968). The amphibolites are of two generations. The older amphibolite occurs as sheet in metasediments whereas the younger amphibolites have cross cut relationship with the major structures.

3 Field occurrences of mafic volcanics in NDFB

The mafic volcanic rocks in Bayana basin (BYV) occur in the form of flows, agglomerates, volcanic breccia and tuffs. They attain a maximum thickness of about 1,000 m including sedimentary interbeds.The variation in the

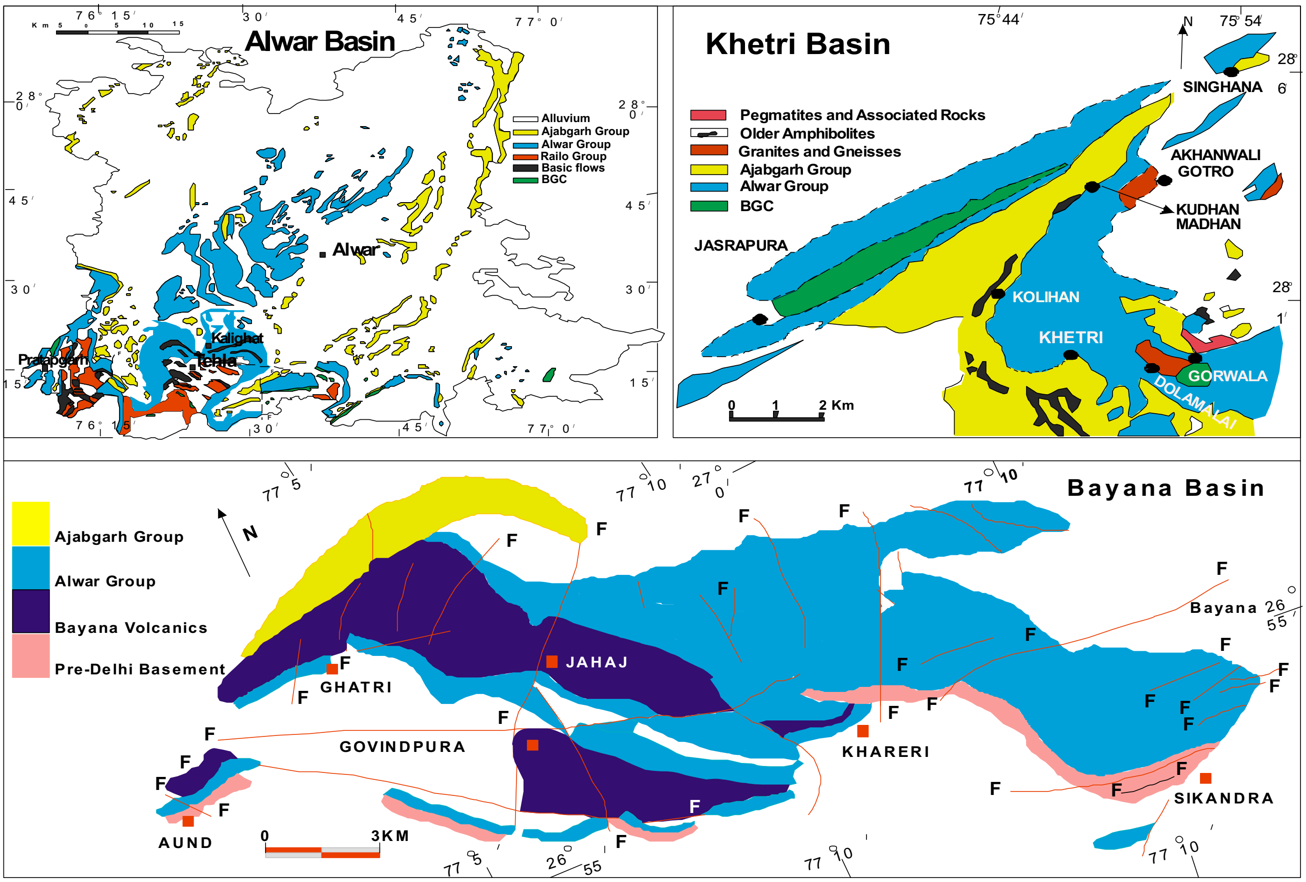

Fig. 2 Simplified geological maps of the basins of North Delhi fold belt

thickness of individual flow is from 1.2 to 123 m. Total number of eighteen flows identified in the area have been classified into lower, middle and upper groups comprising seven, three and eight flows respectively (Banerjee and Singh 1977). The flows are generally fine to medium grained, dark grey to at places greenish black in colour with well preserved vesicles and amygdules. Vesicles are generally circular and up to 5 mm but rarely 3 cm in diameter. Vesicularity decreases from top to bottom in each flow. Texturally, these rocks are generally porphyritic showing sub-ophitic and occasionally hyalo-ophitic texture. Mineralogically these volcanics are predominantly composed of cacic augite and plagioclase (bytownite–labradorite) followed by opaques and minor glass.

The Alwar volcanics (ALV) occur in the form of massive to vesicular lava flows, pyroclastic breccia, spatter, chert beds and tuffs. This volcano-sedimentary sequence attains maximum thickness of 5,000 m with cumulative thickness of volcanics up to 1,700 m. A maximum of sixteen flows have been identified (Banerjee and Singh 1977). The visual and petrographic characteristics of these volcanics are almost similar to those of BYV.

The Khetri volcanics (KHV) occur as thick concordant layers and lenses of amphibolites which are considered to have been emplaced as basic sills during or just after the sedimentation (Roy Chowdhary and Dasgupta 1965; Dasgupta 1968). These mafic rocks have been studied in detail by several workers (Mehta et al. 2000) who observed volcanic features within these rocks. These rocks show uniform textural relationship and mineral assemblages. Major mineral constituents of these rocks are amphiboles followed by plagioclase, quartz and chlorite with accessory epidote, apatite and calcite. Some relatively less altered samples contain relict pyroxene showing porphyritic and sub-ophitic textures. The amphibolites of Khetri basin have their best preservation at Kolihan and Madhan–Kudhan localities (Fig. 2) and referred to herein as Kolihan volcanics (KLV) and Madhan–Kudhan volcanics (MKV) respectively for the reasons discussed in following section.

4 Geochemistry

4.1 Analytical techniques

Major, trace and rare earth elements analyses of whole rock samples of volcanic rocks of Bayana, Alwar and Khetri basins of NDFB belt are given in Table 1. Major and some

Table 1 Geochemical composition of volcanic rocks of Bayana, Alwar and Khetri basins of North Delhi fold belt, NW Indian shield (Data reproduced from Raza et al. 2007)

Sample code	Bayana volcanics							
	K-1	K-6	KR-5	K1-6	K1-13	J-13	By13	By15
SiO_2	50.76	50.13	50.94	50.62	50.6	58.34	52.75	51.04
TiO_2	1.33	1.34	1.46	1.06	1.59	1.51	1.83	1.96
Al_2O_3	11.95	14.84	9.88	11.06	9.96	12.26	10.42	10.18
FeO	14.17	11.69	14.16	14.53	14.71	13.17	12.66	13.96
MnO	0.22	0.23	0.21	0.22	0.23	0.23	0.2	0.19
MgO	8.7	7.33	10.26	9.71	10.87	2.06	11.02	7.03
CaO	8.61	7.08	8.04	7.66	10.47	1.62	7.76	8.29
Na_2O	1.64	2.05	1.28	1.64	1.69	3.5	2.14	1.49
K_2O	0.9	0.56	0.52	1.55	0.74	4.71	1.86	1.02
P_2O_5	0.16	0.12	0.13	0.23	0.16	0.22	0.18	0.17
Total	98.44	95.37	96.88	98.28	101.02	97.62	100.82	95.33
Mg no.	57	57	61	59	61	55	53	53
CaO/Al_2O_3	0.72	0.48	0.81	0.69	1.05	0.13	0.74	0.81
Ni	80	75	65	91	45	137	88	47
Cr	374	203	217	397	280	134	186	110
Co	30	27	29	32	30	58	31	27
Rb	26	6	17	25	17	23	59	28
Sr	221	107	229	222	256	196	148	176
Ba	151	83	111	NA	220	248	527	358
Zr	55	61	50	88	59	103	83	189
Y	18	17	19	22	19	26	25	25
Nb	6	9	9	6	9	14	16	13
Th	1.45	1.37	1.56	2.01	1.51	2.63	2.61	2.61
U	0.3	0.28	0.28	0.93	0.48	1.56	nd	nd
Ta	0.75	0.95	0.92	0.75	0.92	1.84	1.3	1.25
La	11.07	10.02	12.96	12.12	16.87	43.19	17.21	16.73
Ce	22.68	20.76	22.98	23.64	36.08	80.27	34.91	32.64
Pr	2.96	2.81	3.13	3.06	4.72	8.89	4.37	3.25
Nd	13.76	12.47	13.95	14.36	19.91	31.37	18.14	13.3
Sm	3.02	2.73	3.55	3.52	4.96	5.36	5.2	3.91
Eu	1.23	1.08	1.14	1.22	1.45	1.41	1.61	1.16
Gd	3.18	2.79	3.55	3.4	4.38	4.59	5.22	3.14
Tb	0.55	0.54	0.55	0.57	0.78	0.66	0.84	0.41
Dy	0.34	2.9	3.17	3.33	4.59	3.94	4.56	2.73
Ho	0.63	0.61	0.71	0.72	0.92	0.83	nd	nd
Er	1.7	1.62	1.91	1.99	2.5	2.45	2.35	1.39
Tm	0.26	0.22	0.26	0.3	0.37	0.43	nd	nd
Yb	1.22	1.35	1.29	1.51	1.84	2.17	2.05	1.06
Lu	0.15	0.18	0.18	0.17	0.28	0.31	0.29	0.14
(La/Yb)n	6.51	5.32	7.21	5.76	6.58	14.28	6.02	11.32
(La/Ce)n	1.26	1.25	1.46	1.32	1.21	1.39	1.27	1.32
(La/Nd)n	1.59	1.58	1.83	1.66	1.67	2.71	1.87	2.48
(Ce/Nd)n	1.26	1.27	1.26	1.26	1.38	1.95	1.47	1.87
(Ce/Yb)n	5.16	4.27	4.95	4.35	5.45	10.28	4.73	8.55
La/Nb	1.85	1.11	1.44	2.02	1.87	3.09	1.08	1.29
La/Sm	3.67	3.67	3.65	3.44	3.4	8.06	3.31	4.28

Table 1 continued

Sample code	Bayana volcanics							
	K-1	K-6	KR-5	K1-6	K1-13	J-13	By13	By15
Ce/Nb	3.78	2.31	2.55	3.94	4.01	5.73	2.18	2.51
Nb/La	0.54	0.9	0.69	0.5	0.53	0.32	0.93	0.78
Nb/Ce	0.26	0.43	0.39	0.25	0.25	0.17	0.46	0.4
Sm/Yb	2.48	2.02	2.75	2.33	2.7	2.47	2.54	3.69
Rb/Sr	0.12	0.06	0.07	0.11	0.07	0.12	0.4	0.16
Zr/Y	3.06	3.59	2.63	4	3.11	3.96	3.32	7.56
Al_2O_3/TiO_2	8.98	11.07	6.77	10.43	6.26	8.12	5.69	5.19
CaO/TiO_2	6.47	5.28	5.51	7.23	6.58	1.07	4.24	4.23
Y + Zr	73	78	69	110	78	129	108	214
$TiO_2 \times 100$	133	134	146	106	159	151	183	196
Ti %	0.8	0.8	0.88	0.64	0.95	0.91	1.1	1.18
Ti ppm	7,973	8,033	8,753	6,355	9,532	9,053	10,971	11,750
Ti/Y	442.96	472.55	460.67	288.85	501.69	348.17	438.83	470.01
Nb/Nb*	0.55	0.91	0.80	0.30	0.83	0.86	–	–
Rb/Ba	0.17	0.07	0.15	–	0.08	0.09	0.11	0.08

Sample code	Alwar volcanics							
	Al 2	Al 3	Al 4	Al 5	Al 6	Al 7	Al 9	Al 10
SiO_2	49.15	48.24	48.50	48.47	49.48	50.37	49.54	48.26
TiO_2	1.02	0.98	0.98	0.96	0.98	1.27	0.97	1.32
Al_2O_3	12.58	12.73	12.13	12.47	12.79	12.45	12.50	12.81
FeO	13.54	13.42	14.00	13.84	13.56	14.14	13.19	15.46
MgO	9.61	8.26	9.11	8.68	7.67	6.85	8.04	6.17
CaO	10.09	9.71	9.25	8.95	9.68	8.35	9.30	6.75
Na_2O	2.76	2.99	3.27	3.43	3.35	3.48	3.37	4.49
K_2O	0.11	0.20	0.13	0.14	0.16	0.16	0.13	0.22
P_2O_5	0.11	0.10	0.10	0.10	0.10	0.12	0.10	0.13
MnO	0.19	0.17	0.22	0.22	0.21	0.18	0.21	0.21
Total	99.16	96.80	97.69	97.26	97.98	97.37	97.35	95.82
Mg no.	56	53	54	53	50	47	52	42
Ni	45	61	49	53	46	19	37	24
Cr	138	227	127	142	141	92	137	108
Co	49	102	60	63	76	99	72	129
Rb	1	2	2	3	3	3	3	3
Sr	118	117	92	108	106	62	74	54
Ba	0	1	2	2	1	1	0	1
Zr	60	58	59	61	61	71	87	76
Y	20	20	23	31	21	24	20	24
Nb	15	14	13	11	11	22	19	23
Th	0.40	1.20	1.50	1.90	1.02	0.80	0.40	0.82
Ta	0.52	0.98	0.95	0.88	0.87	0.86	0.55	0.85
La	4.19	3.90	4.15	4.05	2.73	4.20	2.71	3.89
Ce	11.97	10.34	11.88	10.32	7.60	10.89	7.51	10.15
Pr	1.42	1.25	1.39	1.37	1.07	1.49	1.09	1.36
Nd	8.83	7.84	8.75	8.39	6.79	8.05	6.86	8.82
Sm	2.47	2.25	2.30	2.22	2.11	2.35	2.14	2.25
Eu	0.77	0.87	0.78	0.84	0.72	0.80	0.71	0.89

Table 1 continued

Sample code	Alwar volcanics							
	Al 2	Al 3	Al 4	Al 5	Al 6	Al 7	Al 9	Al 10
Gd	3.39	3.21	3.40	3.23	3.04	3.35	3.06	3.01
Tb	0.60	0.57	0.61	0.58	0.53	0.56	0.57	0.59
Dy	3.43	3.43	3.45	3.41	3.30	3.45	3.39	3.53
Er	2.37	2.41	2.40	2.28	2.35	2.38	2.36	2.26
Yb	2.10	2.14	2.11	2.08	2.09	2.07	2.10	2.10
Lu	0.31	0.32	0.30	0.30	0.30	0.32	0.32	0.31
(La/Yb)n	1.43	1.31	1.41	1.40	0.94	1.46	0.93	1.33
(La/Ce)n	0.90	0.97	0.90	1.01	0.93	1.00	0.93	0.99
(La/Nd)n	0.94	0.98	0.93	0.95	0.79	1.03	0.78	0.87
(Ce/Nd)n	1.03	1.01	1.04	0.94	0.85	1.03	0.84	0.88
(Ce/Yb)n	1.58	1.34	1.56	1.38	1.01	1.46	0.99	1.34
La/Nb	0.28	0.28	0.32	0.37	0.25	0.19	0.14	0.17
La/Sm	1.70	1.73	1.80	1.82	1.29	1.79	1.27	1.73
Ce/Nb	0.80	0.74	0.91	0.94	0.69	0.50	0.40	0.44
Nb/La	3.58	3.59	3.13	2.72	4.03	5.24	7.01	5.91
Nb/Ce	1.25	1.35	1.09	1.07	1.45	2.02	2.53	2.27
Sm/Yb	1.18	1.05	1.09	1.07	1.01	1.14	1.02	1.07
Rb/Sr	0.01	0.02	0.02	0.03	0.03	0.05	0.04	0.06
Zr/Y	3.00	2.90	2.57	1.97	2.90	2.96	4.35	3.17
Al_2O_3/TiO_2	12.33	12.99	12.38	12.99	13.05	9.80	12.89	9.70
CaO/TiO_2	9.89	9.91	9.44	9.32	9.88	6.57	9.59	5.11
Y + Zr	80	78	82	92	82	95	107	100
$TiO_2 \times 100$	102	98	98	96	98	127	97	132
Ti %	0.61	0.56	0.59	0.58	0.59	0.76	0.58	0.79134
Ti ppm	6,115	5,575	5,875	5,755	5,875	7,613	5,815	7,913
Ti/Y	305.75	278.77	255.44	185.65	279.77	317.24	290.76	329.73
Nb/Nb*	3.90	2.18	1.76	1.34	2.22	4.04	6.15	4.34
Rb/Ba		2	1	1.5	3	3		3

Sample code	Khetri volcanics							
	Kolihan section				Madhan–Kudhan section			
	KH 2	KL6	KL7	KH 5	KH 12	KH 13	KH 14	KH 15
SiO_2	42.6	44.14	50.6	50.09	44.06	42.63	47.7	42.86
TiO_2	0.18	0.35	1	0.84	0.12	0.15	0.12	0.14
Al_2O_3	8.31	8.42	12.34	11.94	9.94	7.75	5.34	10.78
FeO	16.52	18.58	11.94	11.44	15.89	19.89	13.24	16.81
MgO	11.17	12.22	8.33	11.21	8.53	9.45	11.98	9.5
CaO	14.12	11.29	9.59	9.21	12.64	11.56	11.44	14.11
Na_2O	2.92	2.38	3.03	3.43	1.41	2.95	3.7	2.4
K_2O	0.12	0.9	0.7	0.47	0.25	0.26	0.42	0.41
P_2O_5	0.14	0.15	0.16	0.2	0.15	0.15	0.16	0.22
MnO	0.22	0.25	0.28	0.21	0.45	0.43	0.31	0.39
Total	96.3	98.68	97.97	99.04	93.44	95.22	94.41	97.62
Mg no.	55	54	56	64	49	46	62	50
CaO/Al_2O_3	1.7	1.34	0.78	0.77	1.27	1.49	2.14	1.31
Ni	36	234	71	17	17	15	12	16
Cr	60	111	192	210	29	45	25	24

Table 1 continued

Sample code	Khetri volcanics							
	Kolihan section				Madhan–Kudhan section			
	KH 2	KL6	KL7	KH 5	KH 12	KH 13	KH 14	KH 15
Co	21	97	27	18	8	12	10	7
Rb	2	8	13	10	1	13	9	1
Sr	18	34	16	50	32	6	10	21
Ba	19	51	172	90	11	16	12	10
Zr	50	72	53	96	11	41	6	11
Y	25	14	10	17	16	29	11	7
Nb	6	18	9	6	2	4	2	2
Th	9.88	8.62	3.16	9.88	3.97	3.44	2.1	1.6
U	0.28	10.24	0.7	nd	0.51	0.45	nd	nd
Ta	0.65	0.71	0.54	0.56	1.52	1.1	0.37	0.29
La	14.72	32.13	14.28	15.89	18.5	25.48	21.47	18.52
Ce	29.69	49.98	27.62	30.72	32.23	34.48	20.96	25.46
Pr	3.56	4.64	3.34	3.55	3.55	4.39	2.84	3.17
Nd	12.97	17.05	14.24	14.17	13.69	16.91	15.05	11.62
Sm	2.59	4.36	2.84	3.04	3.51	4.85	3.21	2.88
Eu	0.75	0.86	0.79	0.69	1.67	3.71	1.43	3.05
Gd	4.16	4.15	2.25	3.38	3.57	5.3	3.8	2.18
Tb	0.66	0.66	0.31	0.51	0.51	0.92	0.52	0.29
Dy	3.09	3.05	1.92	2.76	2.14	4.24	2.23	1.49
Ho	nd	0.58	0.34	nd	nd	0.67	nd	0.25
Er	1.32	1.43	0.82	1.16	0.83	1.49	0.99	0.5
Tm	nd	0.24	0.13	nd	nd	0.22	nd	0.08
Yb	1.02	1.43	0.71	0.91	0.5	0.75	0.48	0.37
Lu	0.14	0.17	0.12	0.11	0.06	0.08	0.05	0.05
(La/Yb)n	10.35	16.12	14.43	12.53	26.54	24.37	32.08	35.9
(La/Ce)n	1.28	1.66	1.34	1.34	1.48	1.91	2.65	1.88
(La/Nd)n	2.24	3.71	1.98	2.21	2.66	2.97	2.81	3.14
(Ce/Nd)n	1.75	2.24	1.48	1.65	1.8	1.56	1.06	1.67
(Ce/Yb)n	8.09	9.71	10.81	9.38	17.91	12.77	12.13	19.11
La/Nb	2.45	1.79	1.59	2.65	9.25	6.37	10.74	9.26
La/Sm	5.68	7.37	5.03	5.23	5.27	5.25	6.69	6.43
Ce/Nb	4.95	2.78	3.07	5.12	16.12	8.62	10.48	12.73
Nb/La	0.41	0.56	0.63	0.38	0.11	0.16	0.09	0.11
Nb/Ce	0.2	0.36	0.33	0.2	0.06	0.12	0.1	0.08
Sm/Yb	2.54	3.05	4	3.34	7.02	6.47	6.69	7.78
Rb/Sr	0.11	0.24	0.81	0.2	0.03	2.17	0.9	0.04
Zr/Y	2	5.14	5.3	5.65	0.69	1.41	0.55	1.57
Al_2O_3/TiO_2	46.17	24.06	12.34	14.21	82.83	51.67	44.5	77
CaO/TiO_2	78.44	32.26	9.59	10.96	105.33	77.07	95.33	100.79
Y + Zr	75	86	63	113	27	70	17	18
$TiO_2 \times 100$	18	35	100	84	12	15	12	14
Ti %	0.11	0.21	0.6	0.5	0.07	0.09	0.07	0.08
Ti ppm	1,079	2,098	5,995	5,036	719	899	719	839
Ti/Y	43.16	149.88	599.5	296.22	44.96	31.01	65.4	119.9
Nb/Nb*	0.50	0.12	0.48	–	–	0.11	0.20	–
Rb/Ba	0.11	0.16	0.08	0.11	0.09	0.81	0.75	0.10

higher abundance trace elements (Ni, Cr, Zr, Y and Nb) were determined by X-ray fluorescent spectrometry (XRF) at Wadia Institute of Himalayan Geology, Dehradun following methods given in Bhat and Ahmad (1990). Analyses of REE and other trace elements were performed by inductively coupled plasma-mass spectrometry (ICP-MS) at N.G.R.I., Hyderabad following the procedure given in Balaram (1991) and Balaram et al. (1996). Details of the analytical precision and accuracy of the geochemical data are given in Raza et al. (2001, 2007).

4.2 Geochemical characteristics

BYV show small range of variation in their SiO_2 content (50.13 %–52.75 %) but the variation in Al_2O_3 is large (9.88 %–14.84 %). Their MgO content ranges from 7.03 % to 11.02 % except one sample possessing low MgO (J13, 2.06 %). K_2O abundance is <I % in almost all samples. Large variation in MgO, Ni, and Cr contents coupled with low Mg numbers (Mg # avg. 54) are some of the features which indicate the evolved nature of the Bayana volcanic suite. Despite differences in $\sum$REE contents, the REE patterns of all the lavas are identical in shape and show parallel relationship. REE patterns of BYV in general, are moderately fractionated showing (La/Yb)n ratio between 5.32 14.27 without any significant Eu anomaly (Fig. 3). The ALV also show a narrow range of SiO_2 and (48.24 %–50.37 %) and Al_2O_3 (12.13 %–12.81 %). MgO abundance is between 6.17 % and 9.61 %) The evolved character of these basaltic rocks, as compared to BYV is indicated by low concentrations of compatible elements like Ni (19–61 ppm) and Cr (92–227 ppm). Unlike the Bayana and Khetri volcanics of NDFB, the ALV show nearly flat REE patterns (Fig. 3) with (La/Yb)n ratio (0.93–1.46). Few samples of these mafic rocks display minor deflection at Eu. The REE enrichment is about 11–18x chondrite for LREE and about 12–13x chondrite for HREE.

In KHV SiO_2 contents vary from 42.60 % to 50.60 %. Their MgO content is generally high (8.33 %–12.77 %, average 10.19 %). The FeO content shows considerably large range from 11.44 % to 18.58 %. Samples from Kolihan and Madhan–Kudhan section show some marked differences in their chemical characteristics, particularly in the concentration of TiO_2, Zr, Nb, Th and ferromagnesian elements which are relatively higher in the former. The mafic volcanics of Khetri basin are LREE enriched with (La/Yb)n ratio from 10.35 to 35.90. REE distribution of all the samples is almost similar except for the nature of Eu anomalies (Fig. 3). They all show strong enrichment with 60–136x chondrite for LREE and 4–7x chondrite for HREE in Kolihan samples and for about 78–107x chondrite for

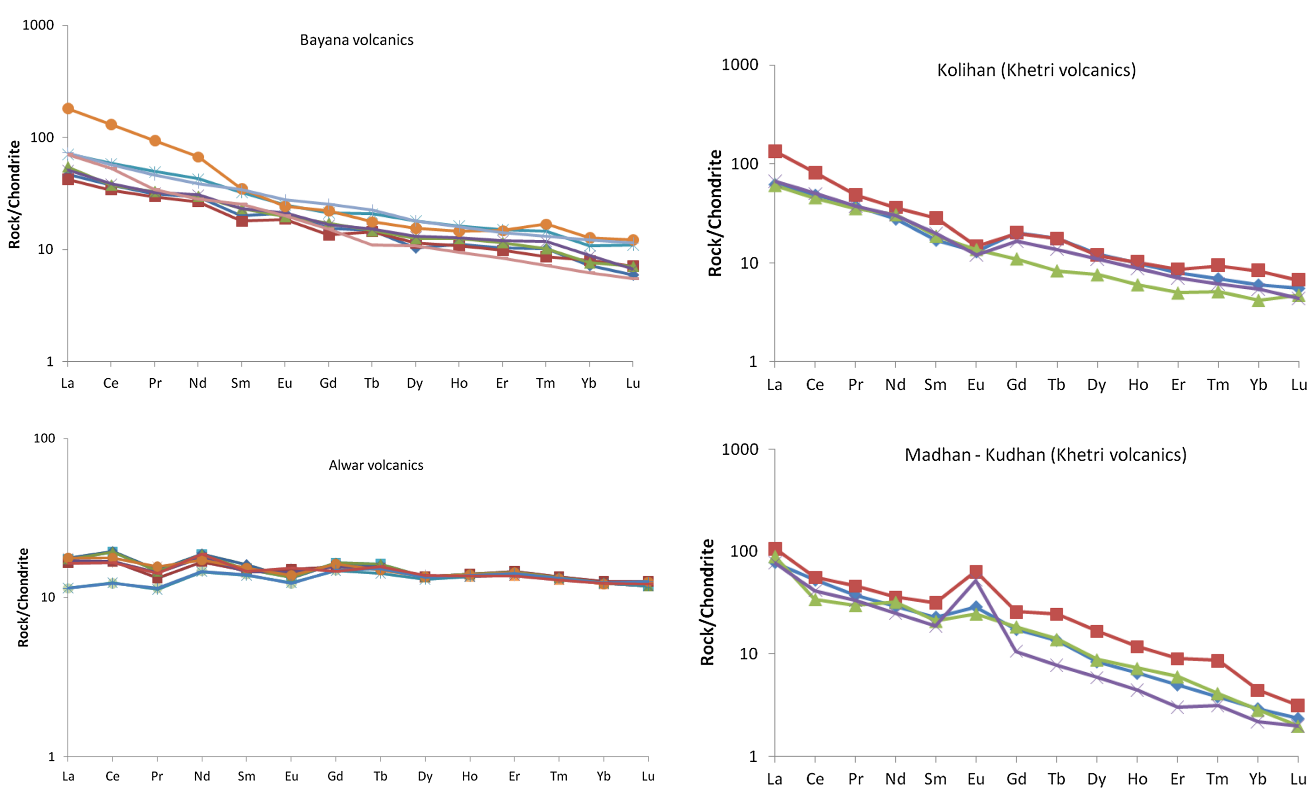

Fig. 3 Chondrite normalized REE diagram of the volcanic rocks of North Delhi fold belt. (Normalizing values from Sun and McDononough 1989). Some undetected values are interpolated for *smoothening curves*

LREE and about 2–3x chondrite for HREE in Madhan–Kudhan samples. While the former shows no or little negative Eu anomalies and low La/Yb_N ratio (10.35–16.12), later is characterized by strong positive Eu anomalies coupled with highly fractionated REE ($La/Yb_N = 24.37–35.91$). In addition to their parallel and generally similar REE patterns, narrow range of variation in the ratios such as La/Ce_N (1.28–2.65), La/Nd_N (1.98–3.71) and Ce/Nd_N (1.06–2.24) also suggest that REE abundances are primary and thus variation in Eu anomalies appears to be a source characteristic.

The volcanic rocks under consideration have undergone metamorphism ranging in grade from lower greenschist facies in Bayana and Alwar basins to lower amphibolite facies in Khetri basin, it is important to verify the effect of the post magmatic processes on the chemistry of NDF belt volcanism. The large ion lithophile elements (LILE) are considered quite susceptible to modifications during post magmatic time (Winchester and Floyd 1977; Humphris and Thompson 1978). Most of the LILE of NDFB except Th show large variation in their abundances that may be due to their mobilization during secondary processes (Lefleche et al. 1992).The effect of alteration on LILE concentration can be assessed using Rb/Sr ratios. It has been observed that Rb/Sr ratios of basaltic rocks vary from very low (0.007) in least altered basaltic rocks to very high (8) in highly altered basaltic rocks (Lefleche et al. 1992). In the studied volcanic rocks Rb/Sr ratios are low, showing a range of variation from 0.05 to 0.40 in BYV, 0.008 to 0.05 in ALV and 0.03 to 2.17 in KHV. Significantly low range of variation of Rb/Sr ratios in these rocks does not indicate any major effect of alteration on primary concentration of LILE. Th is considered as immobile even during high degrees of alteration (Lefleche et al. 1992). Th abundances of BYV (1.37–2.63 ppm), ALV (0.4–1.9 ppm) and KHV (1.6–9.88 ppm) negate their mobilisation during secondary processes. Moreover, similarity and parallelism of REE is maintained over a large range of composition and for the samples collected from different locations. This smoothness and parallelism of REE patterns and multi-element diagrams (Figs. 3, 6) are other evidences which suggest that incompatible elements including REE were not affected by secondary processes.

4.3 Magma type

Low Nb/Y ratio of BYV (0.27–0.64, avg. 0.47), ALV (0.35–0.91, avg. 0.71) and KHV (1.28–0.14, avg. 0.44) classify them as sub alkaline basalts (Pearce and Gale 1977). In YTC ($Y = Y + Zr$, $T = TiO_2 \times 100$, $C = Cr$) diagram of Davies et al. (1979), which is an analogue of AFM but employs relatively immobile elements, BYV and ALV display tholeiitic nature whereas KHV show transitional nature between tholeiite and calc- alkaline basalts (Fig. 4). In order to evaluate extensional/subduction signatures in the genesis of NDFB volcanics, their data is plotted in La/Nb versus Y diagram (Fig. 5) of Floyd et al. (1991). In this diagram BYV and ALV display continental flood basalt affinity whereas the samples of KHV occupy the field of island arc basalts.

5 Discrimination of the magma source lithology

Determining the source lithology of a suite of magmatic samples is not only important for the overall geotectonic interpretation of the magma genesis but it is also crucial to petrological modeling based on parameterization of experimental data and observations made on mantle

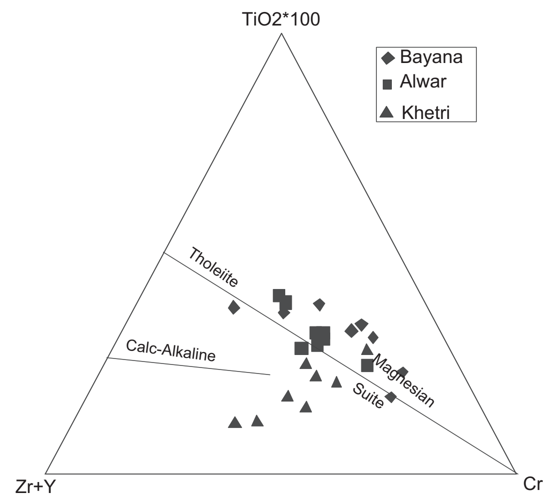

Fig. 4 Y (Zr + Y)−T (Ti × 100)−Cr ternary diagram (after Davies et al. 1979) of volcanic rocks of North Delhi fold belt indicating their tholeiitic to calc alkaline affinity

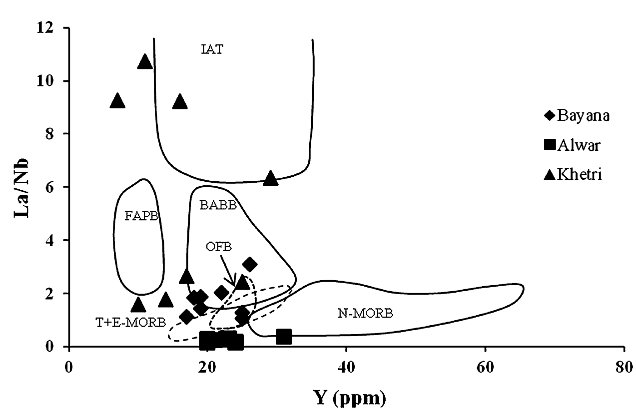

Fig. 5 La/Nb versus Yb diagram for the volcanic rocks of North Delhi fold belt (after Winchester et al. 1987) indicating their tectonic settings

peridotite sources. In recent years, the development of new petrological methods based on major element geochemistry (e.g., Sobolev et al. 2005; Dasgupta et al. 2007; Herzberg and Asimow 2008) provided new insights for the composition of magma source lithology. From the suite of lavas collected for this study there is a population (Khetri volcanics) of low-silica (<45 % SiO_2) samples characterized by high CaO contents, similar to experimental melts produced from a CO_2-metasomatized peridotite (Dasgupta et al. 2007) (Fig. 6a). The BYV and ALV which contain lower CaO and higher SiO_2 contents have modeled liquid lines of descent (Fig. 6a) that suggest that these samples are the result of normal fractional crystallization of a peridotite source primary magma. However, there is slight inclination of BYV towards pyroxenite component. This suggests that either their source had pyroxenite component or they have generated by large amounts of high pressure pyroxene fractionation in the mantle (Albarède et al. 1997; Herzberg and Asimow 2008). More detailed work (e.g. analyses of olivine compositions) is required to determine if there is a pyroxenitic component in the source of these lavas. Considering the presence of a possible pyroxenitic component in the form of veins that result from the reaction of silica-rich melts with a mantle peridotite (Feigenson and Carr 1993; Sobolev et al. 2005; Herzberg 2006) is a good option.

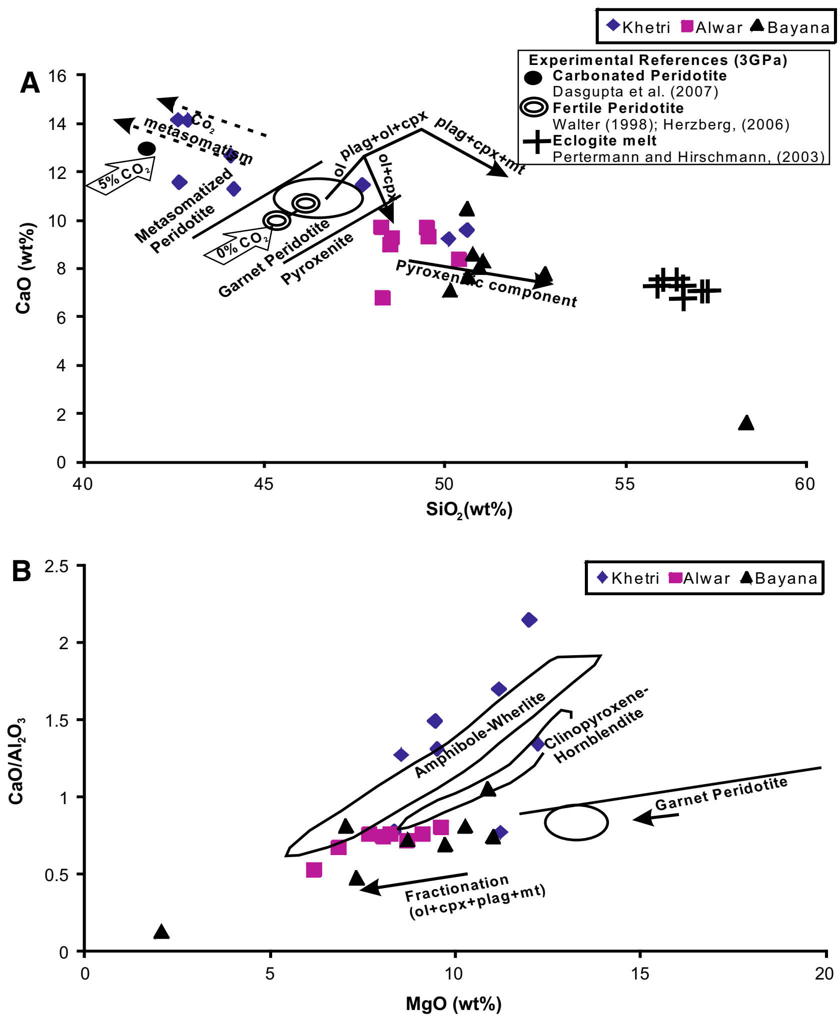

Fig. 6 Possible source lithologies and the effect of fractional crystallization for the samples collected in this study. **a** Petrological discrimination of magma sources for primary magmas produced in the garnet stability field (~3 GPa). The two diagonal lines separate magmas melted from metasomatized (carbonated) peridotite, garnet peridotite and pyroxenite. The low-silica (<45 % SiO_2) samples are characterized by high CaO contents, similar to experimental compositions of magmas produced by melting a carbonated peridotite. The population with lower CaO and higher SiO_2 (>47 %) shows similar compositions to magmas produced by melting second stage pyroxenite or derivative liquids of high-pressure pyroxene fractionation.The NDFB are compared to two low pressure (<1 GPa) experimental melting trends, amphibole-bearing wherlite and clinopyroxene bearing hornblendite and to primary garnet lherzolite. [Experimental values, fields and liquid lines of descent (LLD) from Gazel et al. 2011]

The low silica ($SiO_2 < 60$ %, Fig. 6a) type rocks are interpreted to have formed by the magmas produced from a mantle wedge metasomatized by silicic melts from the subducting slab (Martin et al. 2005). The KHV overlap with the low pressure experiments of melts produced by an amphibole bearing source (Fig. 6b). On the other hand BYV and ALV are consistent with an amphibole-free peridotite or pyroxenite source (Fig. 6a).

6 Geochemical variations between intraplate and arc rocks

The geochemical compositions and magma type classification of mafic rocks of NDFB suggest the presence of tholeiitic volcanism in Bayana and Alwar basins. The basaltic rocks of these two basins show close affinity with low-Ti continental flood basalts. On the other hand, KHV display transitional nature between tholeiite and calc-alkaline basalts characteristic of subduction magmas. To further confirm these affinities, it is appropriate to constrain their petrogenetic history using incompatible elements and element ratios.

In PM-normalized spiderdiagrams the samples of BYV are characterized by a general enrichment from less incompatible (Ti, Zr, Y) to more incompatible (P, Th, Ta, Nb). These are also enriched in LILE and LREE compared to PM (Fig. 7). These trace element characteristics of BYV are similar to most of the Proterozoic continental flood basalts and Proterozoic dyke swarms (Thompson et al. 1983; Tarney 1992). The enriched nature of BYV may be a consequence either of shallow level crustal contamination of the parent melt derived from a depleted mantle (asthenospheric) source (Arndt and Jenner 1986; Arndt and Christensen 1992) or the melts were generated in an enriched lithospheric source.

To assess the role of shallow level crustal contamination as a likely mechanism for enriched nature of BYV, incompatible element abundances and their ratios are used. Because their abundances in general and their ratios in particular, are not affected during the processes of partial melting and fractional crystallization (Condie 1990) and thus act as fingerprints of the source. In PM-normalized diagram, BYV display LREE enrichment with slight Nb, P, Zr and Hf anomalies suggesting the involvement of sialic crustal material (Fig. 7). The low Nb/La (0.65) and Ce/Nb (3.38) ratios of BYV with respect to upper continental crust (Nb/La = 0.89, Ce/Nb = 2.56) also suggest their enrichment most likely due to addition of crustal contaminants into their melts (Hawkesworth et al. 1995; Li et al. 2006; Sandeman et al. 2006).

ALV are chemically different from BYV as evident from PM-normalized spidergram (Fig. 7). Unlike BYV, the patterns of ALV are flat with positive anomalies at Ta and Nb. Higher Nb/Ce ratio (Avg. 1.63) with flat REE patterns of ALV suggest their derivation from an asthenospheric

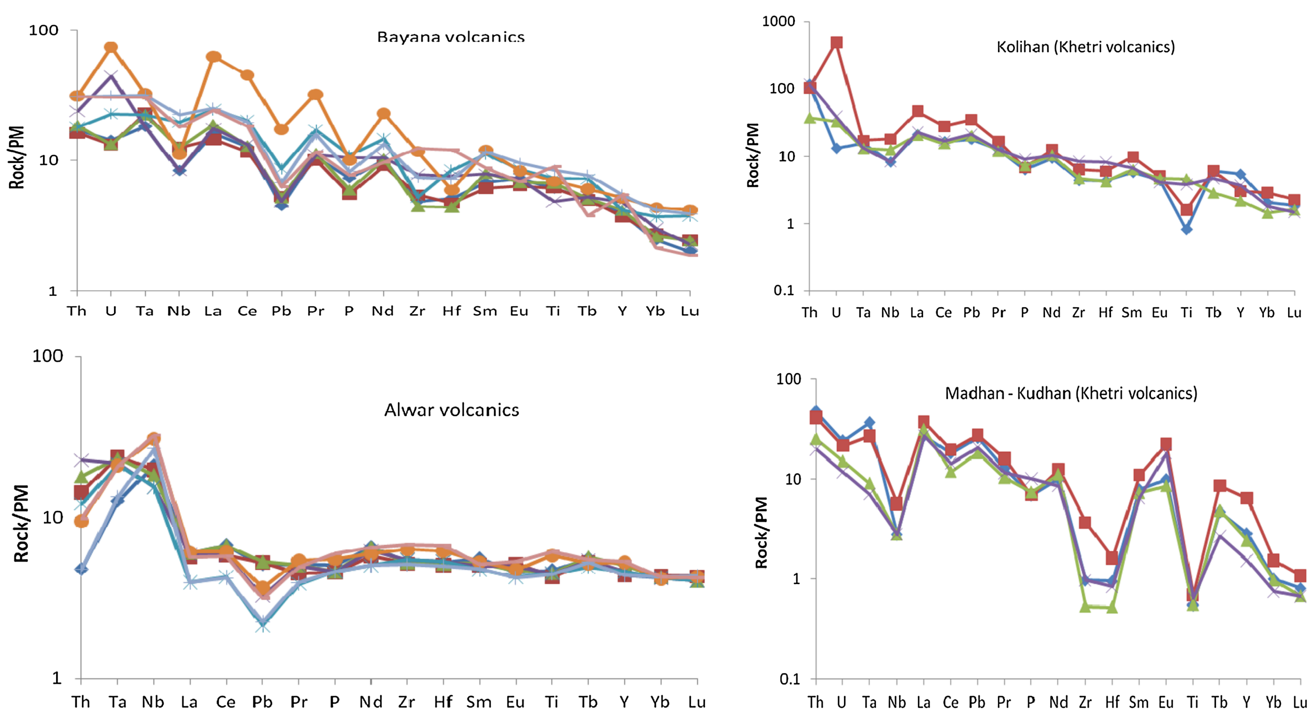

Fig. 7 PM normalized multielement diagrams of the volcanic rocks of different basins of North Delhi fold belt. (Primordial Mantle normalizing values from Sun and McDononough 1989). Some undetected values are interpolated for *smoothening curves*

source (Saunders et al. 1992). It is widely accepted that the magmatic production in a volcanic arc results from the partial melting of the mantle wedge triggered by fluids from the subducted slab. PM-normalized patterns of KHV (Fig. 7) are characterized by enriched LILE including Th and LREE, depleted HFSE with distinct negative anomalies of Nb, P and Ti and steep REE patterns ($La/Yb_n > 10$) (Fig. 7). These features, along with tholeiite to calc-alkaline nature of these volcanics are typical of basaltic rocks that are generated by subduction related magmatic processes wherein mantle derived magmas are metasomatised by slab derived-fluids and/or melts (Saunders et al. 1980; Pearce 1983). Prominent negative anomalies at P and Ti suggest the fractionation of plagioclase, apatite and Ti-bearing phase. Geochemical attributes like high LILE/HFSE ratio, relatively low Nb abundances and behavior of Ti as compatible element during differentiation of the magma are considered characteristic features of subduction generated lavas (Pearce 1982; Holm 1985; Smith and Holm 1987). Furthermore, higher La/Yb indicates a lower degree of partial melting or derivation from a more enriched source. The average La/Yb_n ratio of Kolihan and Madhan–Kudhan volcanics (13.36 and 29.73 respectively) is the testimony of differentially enriched sources for the two suites of rocks. These geochemical variations along the arc reflect regional changes in the extent and type of metasomatic processes caused by the subducting input and magma source composition alongwith the degree of partial melting.

Fe/Mn ratio has been found useful in characterizing the basalts from various tectonic settings (Li et al. 2006). Plume derived basalts typically exhibit Fe/Mn ratio of 65–71 (Humayun et al. 2004) whereas island arc basalts usually have this ratio 54–59 (Wilson 1989). The BYV and ALV possess high Fe/Mn ratios (avg. 64 and 69 respectively) comparable with those of Hawaiian basalts indicating their affinity to plume derived magma. However, Fe/Mn ratio of KHV is low (avg. 52) and compares well with those of island arc basalts. Fractionation of HFSE (e.g. Nb, Ta, Zr, Ti, Hf) relative to the large ion lithophile and light rare earth elements is well known for convergent margin igneous rocks (Fig. 7). This fractionation is related to subduction processes in which a residual mineral phase in the subducting slab (rutile) or mantle wedge (amphibole) holds the HFSE (e.g., Ringwood 1990; Foley et al. 2000, 2002). The Nb/Nb* denotes the variations in the typical arc depletion of Nb relative to U or Th and La, when normalized to the McDonough and Sun (1995) primitive mantle reference. Magmas produced in a subduction zone will have $Nb/Nb^* < 1$, whereas magmas produced by mantle upwelling (decompression melting) with no subduction signature (intraplate settings) are characterized by $Nb/Nb^* > 1$. The KHV have significant subduction signature with $Nb/Nb^* < 0.5$ (Fig. 7, KV avg. 0.37, MKV avg. 0.15, Table 1). Conversely, the ALV have high positive Nb anomalies ($Nb/Nb^* >$ avg. 3.24) and BYV nearly transitional $Nb/Nb^* > 0.6$) (Fig. 7). This suggests that these lavas were produced by mantle upwelling, but with some influence of components. HFSE depletions ($Nb/Nb^* < 0.5$) are restricted to geographic locations directly above the subducting slab. In contrast, samples with $Nb/Nb^* > 0.5$ are located in areas with no evidence of a subducting slab or well-behind it, and the correlations are consistent with the HFSE depletions in the arc magmatism being controlled by the presence of a HFSE retaining phase (e.g. residual rutile) in the subducting slab. Magmas produced by arc volcanism are also characterized by having enrichments in fluid mobile elements (e.g., Ba, K, Pb, Sr) relative to other trace elements. In contrast, the behavior of fluid mobile elements is the opposite in within plate magmas. In summary, the correlation between La/Yb, Nb/Nb*, LILE and HFSE for the lavas of different basins suggests that these geochemical variations, are controlled by the composition of the magma sources (mantle source and subducting oceanic crust components). The geochemical variations of the magmas across the arc depend on the absence of, or distance from, the subducting slab. The Bayana basalts are located in the areas that have no evidence of a subducting slab and have geochemical signatures controlled by decompression melting (mantle upwelling). The Alwar volcanics are located close to the edge of the subducting slab (Khetri belt) and thus posses signatures of mantle upwelling with little component of slab. While Kolihan and Madhan Kudhan lavas (Khetri volcanics) are situated above the slab in the down slope direction and thus posses important slab signature. In view of the preceding discussion the emplacement of mafic volcanic rocks of Bayana and Alwar basins may be related to the rifting of a pre- Delhi Archaean basement (BGC) triggered by upwelling plume whereas lavas of KHV were derived from the melting of the mantle segment overlying a Proterozoic subduction zone. This interaction between a mantle plume and the subduction system caused geochemical and geodynamic deviations from “normal” arc magmatism because the arc is close enough to the influence of a mantle plume. The interaction between the mantle plume (Bayana and Alwar volcanism) and an arc system produced arc lavas with calc alkaline to tholeiitic geochemical signature (Khetri volcanics) in a subduction setting (Wendt et al. 1997; Turner and Hawkesworth 1998; Smith et al. 2001).

7 Petrogenetic history

To evaluate the melting conditions, hypothetical REE compositions of the sources have been calculated. Such sources are commonly tailored so that they provide a perfect fit to the analytical data, and the hypotheses founded upon such models are extremely difficult to rebut (Thompson et al. 1986). This approach provides a method of demonstrating how several compositionally different suites may be derived from the same source by petrogenetically reasonable models. Though it does not provide a unique solution but the results are much better constrained than those based on rock suites modeled individually. The calculated source compositions may be the result of a variety of contributions made to the melts, possibly at different stages in its evolution. The models used in these calculations are designed to be as simple as possible and are constrained by mineralogical and chemical considerations. Hypothetical sources have been calculated from the chemical data in terms of simple batch melting and fractional crystallization (Hanson 1980; Thompson et al. 1984; Smith and Holm 1987, 1990). In modeling the tholeiitic suites extensive partial melting has been considered, however, some workers argued for lower degree of melting (Thompson et al. 1984). To constrain true melting conditions, hypothetical REE compositions have been calculated for the most primitive sample of individual group in each volcanic suite. A primary magma is a silicate liquid that initially separates from a mantle source. In most cases primary magmas are modified by crystallization during transport to the surface and eruption (e.g., O'Hara 1968). Mg# is a sensitive indicator about the extent of magmatic evolution the rock has undergone. Accordingly high Mg# number samples have been chosen as the primitive members because none of the sample of any rock suite contains accumulated olivine. The source calculation is based on the La, Ba, Ce and Nd contents of the sample using the equation $Cl/Co = 1/F$, when $D = 0$ (Hanson, 1980); $D =$ bulk distribution coefficient, $Cl =$ concentration of the element in the melt, $Co =$ initial concentration of the element in the source, $F =$ fraction of the melt.

7.1 Bayana volcanics

The volcanism in Bayana basin took place in three phases as evident from the occurrence of three groups of lava flows at different stratigraphic levels. In order to precisely constrain the melt generation for BYV, one primitive sample of each group i.e. K-6 (lower group), Kr-5 (middle group) and By-15 (upper group), is selected for source modeling. About 9 %, 7 % and 5 % partial melting have been calculated for K-6, Kr-5 and By-15 samples respectively. The REE patterns for 1 %, 2 %, 4 %, 6 %, 8 % and 10 % melting of the modeled sources of each group are calculated using the batch melting equation and kd values of Hanson (1980) and assuming lherzolite mantle (olivine + orthopyroxene + clinopyroxene) originally consisting of 55, 25 and 20 % respectively of each mineral mode. The assumed melting proportion of olivine (Olv), orthopyroxene (Opx) and clinopyroxene (Cpx) used for calculation are 20 %, 25 % and 55 % respectively (Hanson 1980).

The REE patterns for 1 %, 2 %, 4 %, 6 %, 8 % and 10 % melting of the modeled sources of each suite are shown in Fig. 7. It is evident from this figure that the calculated patterns in shape match very well with those of the samples of each group. However, to verify the cogenetic nature of all flows of BYV, the source composition for most primitive sample K-6 amongst all flows is calculated. It is evident from the diagram (Fig. 9) that all the samples of BYV could be generated from similar source(s) by varying degree of melting (1 %–10 %) and subsequent fractional crystallization.

7.2 Alwar volcanics

In Alwar basin the samples of mafic volcanics were collected from two different localities (Al-2, Al-3, Al-4 & Al-5 from one location and Al-6, Al-7, Al-9 & Al-10 from the other). In order to determine the melting condition of ALV one most primitive sample from each location (i.e. samples Al-3 & Al-6) has been selected and the respective sources are modeled. About 14 % for Al-3 and 18 % for Al-6 partial melting have been calculated (Hanson 1980). The REE patterns for 1 %, 2 %, 4 %, 6 %, 8 %, 10 %, 15 %, 20 % and 22 % melting of modeled sources of Al-3 and Al-6 are shown in Fig. 7. It is clear from the diagram that these patterns mimicked with the samples of ALV of their respective locality. It is observed from the diagram (Fig. 8) that the samples Al-2, Al-3, Al-4 and Al-5 were generated by 12 %–15 % partial melting whereas the range of partial melting for the samples Al-6, Al-7, Al-9 and Al-10 is between 12 % and 20 %. Like BLV the possibility of cogenetic origin of ALV is also explored by choosing Al-6 as the most primitive sample. It is inferred from the diagram that the ALV were generated from same source by 12 %–20 % melting followed by fractional crystallization (Fig. 9).

7.3 Khetri volcanics

Since KHV are well exposed in two distinct belts i.e. Kolihan (KL) and Madhan–Kudhan (MK), and accordingly one primitive sample of each belt (KL-7 for Kolihan and KH-13 for Madhan–Kudhan) are selected for source modeling. Partial melting of about 6 % and 15 % have been calculated for KL-7 and KH-13 respectively. The REE patterns for 1 %, 2 %, 4 %, 6 %, 8 %, 10 %, 12 % and 15 % melting of modeled source of KL-7 and 1 %, 2 %, 4 %, 6 %, 8 %, 10 %, 15 %, 20 %, 25 % and 30 % melting of modeled source of KH-13 are shown in the

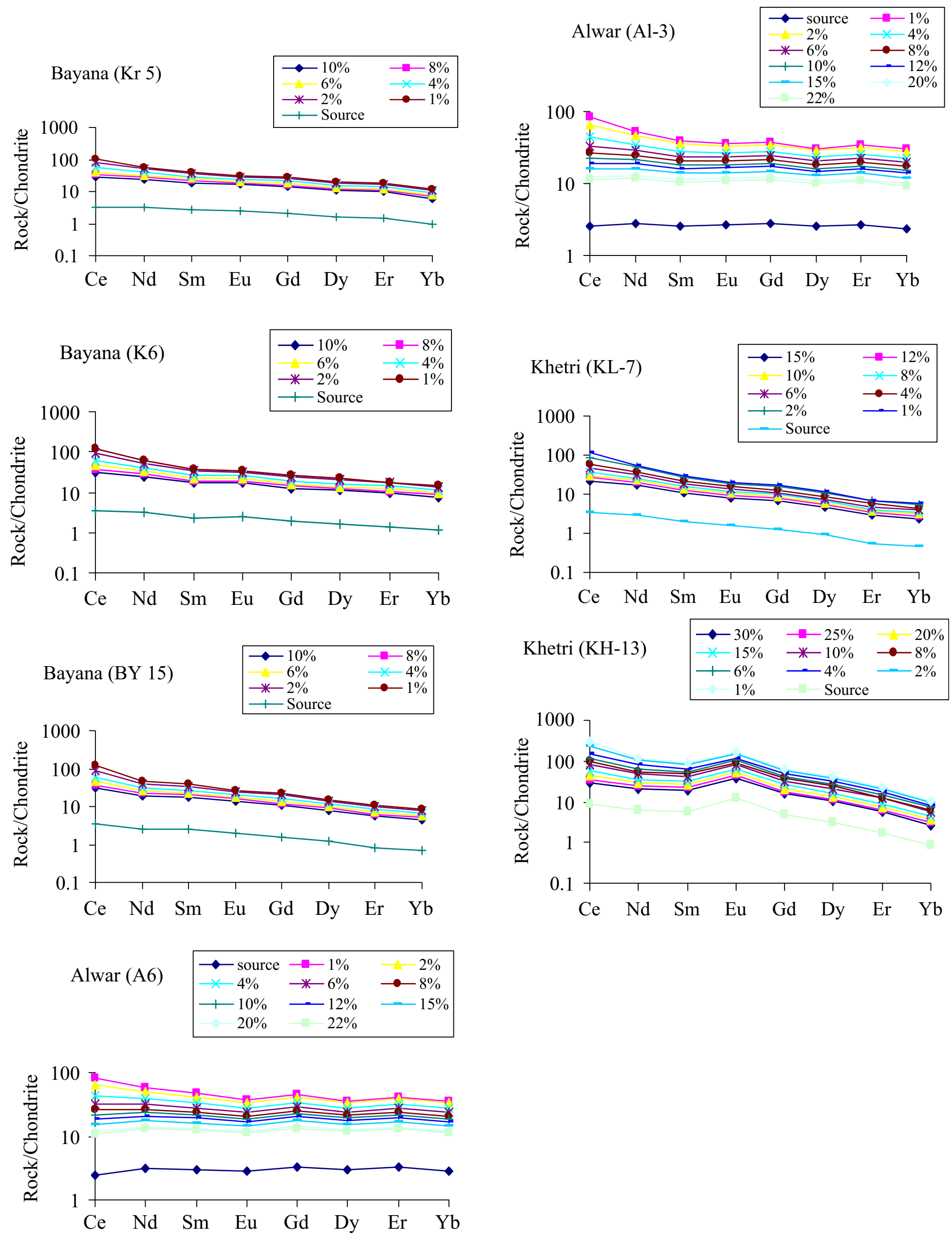

Fig. 8 Showing calculated source from the most primitive sample (indicated with the name of the basin) of each volcanic flow of the three basins and various percent of melting of the source

diagram (Fig. 8). The calculated REE patterns for different degrees of melting of each sample match very well with respective representative samples of each belt. These results profess 1 %–6 % melting for Kolihan and 15 %–30 % melting for Madhan–Kudhan belts followed by fractional crystallization (Fig. 9).

To determine the extent of partial melting for petrogenetic evaluation of mafic rocks of Bayana, Alwar and Khetri (Kolihan and Madhan–Kudhan separately) basins, the average REE concentrations (chondrite normalized) of each suite are compared with the melts generated through different extents of melting of assumed lherzolite mantle source originally consisting of 55 % olv, 25 % opx and 20 % cpx (Fig. 10). It is evident from the diagram that models obtained through various percents of partial melting of assumed sources are similar and almost identical with the average composition of respective volcanic suites. The BYV show their derivation by 4 % partial melting of the modeled source leaving a residual mineralogy of 56 % olv, 25 % opx and 19 % cpx. Whereas the ALV describes a residual mineralogy of 61 % olv, 25 % opx and 15 % cpx at 14 % partial melting of lherzolite source.

In case of Khetri basin the samples collected from Kolihan and Madhan–Kudhan are dealt separately. The mafic rocks from Kolihan show 4 % of partial melting of the calculated source with a residual mineralogy of 56 %

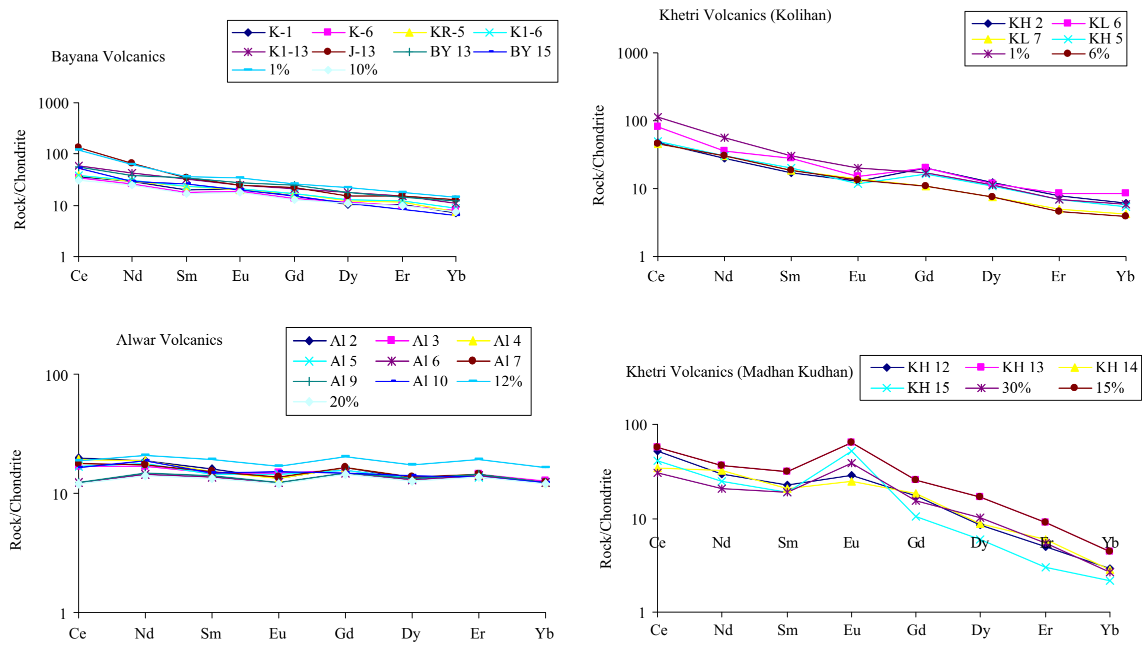

Fig. 9 Showing range of melting to accommodate all samples of each volcanic basin

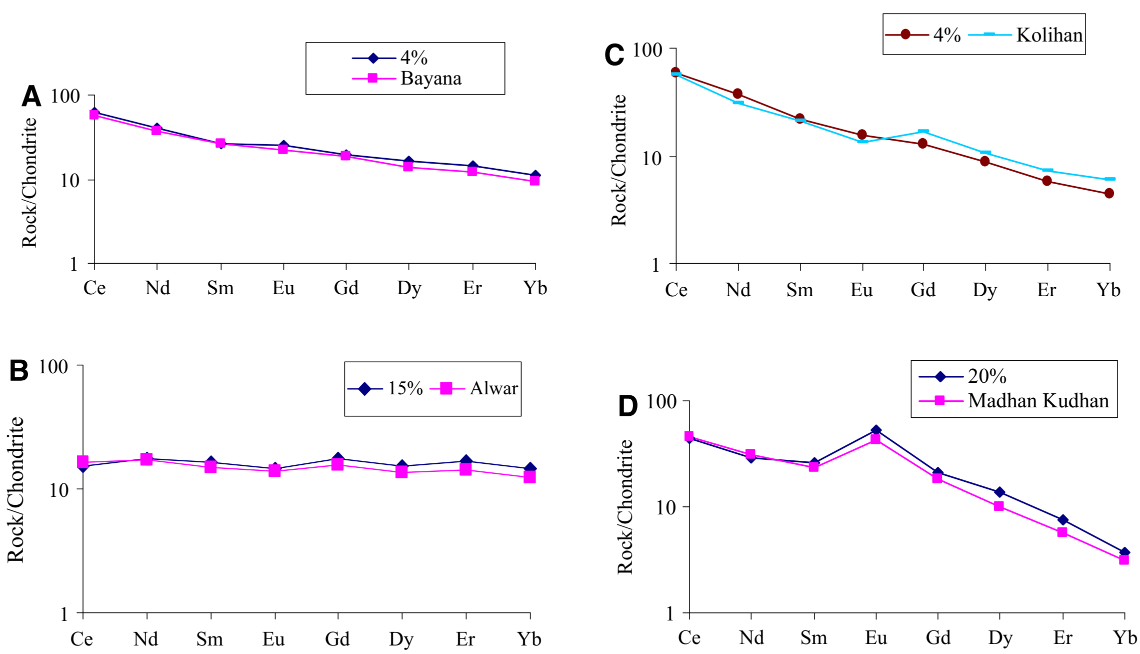

Fig. 10 Showing extent of melting compared to average composition of volcanic suites of the three basins

olv, 25 % opx and 19 % cpx while the samples from Madhan–Kudhan gives the residual mineralogy of 64 % olv, 25 % opx and 11 % cpx at 20 % partial melting.

The geochemical compositions and magma type classification of mafic rocks of NDFB have suggested the presence of tholeiitic volcanism in Bayana and Alwar basins. The basaltic rocks of these two basins show close affinity with low-Ti continental flood basalts (CFB). On the other hand, the mafic volcanics of Khetri basin are distinguished as transitional basalts between tholeiite and calc-alkaline basalt (Raza et al. 2007). To confirm these affinities, source characteristics have been assessed using incompatible element contents and ratios. To assess the source characteristics of basaltic rocks the most effective method is the use of multi-element spidergrams and incompatible element ratios such as Nb/La, Zr/Nb, Ta/Th (Condie 1987), La/Sm, Sm/Yb, Zr/Y (Floyd et al. 1991), Nb/Y, Th/Nb, Zr/Ti (Kuzmichev et al. 2005; Teklay 2006) etc. The CFB affinity of BYV and ALV is well established from their PM-normalized patterns (Fig. 7) and other chemical signatures like low values of Ti, Zr, Zr/Y, and Ti/Y.

The mafic volcanic rocks of Bayana and Alwar basins are interpreted as having erupted in a rifted basin, formed on attenuated continental crust (Raza et al. 2007). This is based on their geochemical characteristics and lithologic associations (association of sedimentary rocks). Therefore, consideration may be given to the possibilities that the source(s) have been modified by deep mantle magmatic and metasomatic processes which induce mineralogical and chemical heterogeneities (Basaltic Volcanism Study Project 1981; Vidal et al. 1984; Allegre and Turcotte 1985; Duncan et al. 1986; Gerlach et al. 1986; Menzies and Hawkesworth 1987) like crustal contamination (Thompson et al. 1984; Carlson et al. 1981) or components available in subduction zone (Pearce 1983; Hawkesworth et al. 1984; Thompson et al. 1984; Arculus and Powell 1986; Hickey et al. 1986; Carlson et al. 1981).

In order to assess the role of garnet, if any, in the genesis of BYV the ratios Ti/Y and Rb/Ba may be of considerable help. Ti/Y ratio is less sensitive to the process of partial melting until garnet is present as a residual phase (Hergt et al. 1991) and Rb/Ba ratio is also insensitive to the degree of melting even in the presence of residual phlogopite (Hawkesworth et al. 1986). Thus melting in the absence of residual garnet would not fractionate Ti/Y and Rb/Ba ratios. The low range of variation of Ti/Y (288.85–501.69) and Rb/Ba (0.07–0.17) ratios of BYV are not in favour of residual garnet. Furthermore, the combination of low Ti/Y (average 421.43) with low Zr/Y (average 0.108) as found in BYV is not consistent with mantle processes involving garnet (Hergt et al. 1991). In view of these arguments, garnet can not be considered as a residual phase during the genesis of Bayana melts. The derivation of low-Ti continental basalts from a source containing residual garnet is not favoured even by those who argue their derivation from an asthenospheric source region (e.g. Arndt et al. 1993).

The asthenospheric source for ALV is well constrained on the basis of positive Nb–Ta anomaly with flat PM patterns (Fig. 7). Nb/Ce ratio (average 1.63) and flat REE patterns (Fig. 3) also suggest their derivation from an asthenospheric source (Saunders et al. 1992). Therefore, it may be interpreted that the melts for ALV were produced from an asthenospheric source which ascended to a shallower depth where it was subjected to a higher degree of melting and produced Nb-enriched basaltic melts with low Ce/YbN ratios (avg. 1.33).

The generation of melt of Alwar tholeiites has been suggested by large degree of partial melting at shallow depth relative to BYV as they are characterized by low La/SmN (mean 1.64) and Sm/YbN (mean 1.08) ratios than those of BYV (La/SmN = mean 4.19 and Sm/YbN = mean 2.58). Therefore, it is possible that the two suites of rocks were derived from the same source(s). Furthermore, the two suites of ALV are closely associated in space and show similar geochemical signatures with few exceptions, it is more appropriate to model them as resulting from same source materials in two separate episodes by variable degrees of partial melting followed by fractional crystallization. And an explanation for their geochemical differences may be adhered to the mechanism of their emplacement or tectonic set up.

The lack of strong fractionation in REE patterns of the volcanics of Bayana and Alwar basins and other evidences as discussed in preceding sections, indicate that there was no garnet in the refractory residue of the source materials. This, together with their average Cr contents (BYV—238 ppm, ALV—139 ppm), imply that both the volcanics were derived from spinel lherzolites. Spinel is not used in modeling because distribution coefficients are only available for REE and not for Zr, Ti, Y and Nb of this mineral. Partial calculations using REE show that approximately 5 % spinel may replace olivine without greatly affecting models(Smith and Holm 1987). The implied absence of garnet in the lherzolite sources and MgO/FeO ratios of BYV (avg. 0.16) and ALV (avg. 0.58) suggest that partial melting took place at pressure of <20 Kb and at temperature between 1,300 and 1,400 C (Smith and Holm 1987).

The sources of both BYV and ALV have their own characteristic trace element signatures indicating heterogeneity in the mantle though both volcanic suites have been derived from spinel lherzolite. Such heterogeneity in the mantle has commonly been reported and is interpreted as resulting from multiple depletion and enrichment of lithophile elements by the processess of partial melting and fluid metasomatism (Basaltic Volcanism Study Project 1981).

According to present understanding about the basalt petrogenesis, the enriched characteristics of low-Ti continental tholeiites could either be inherited from the metasomatised subcontinental lithosphere or reflect addition of felsic materials from continental crust. A lithospheric mantle origin is not favoured because it needs substantial heat and a lot of water to initiate melting (Arndt et al. 1993). On the hand asthenospheric source has plenty of heat and does not require water to initiate the melting. Therefore, it may be inferred that basaltic suites of Bayana and Alwar basins of NDFB have been derived from a common asthenospheric source which suffered variable degrees of melting producing one magma type i.e. low-Ti continental basalt with different chemical signatures.

The samples of KHV both from Kolihan and Madhan–Kudhan areas maintained similarity and parallelism in their REE patterns and PM-normalized spiderplots irrespective of their different locations though some compositional variations are observed between them (Table 1). The REE distribution patterns of all the samples are almost similar and show strong LREE enrichment with respect to HREE despite the fact that MKV display strong positive Eu anomaly whereas samples of KLV are characterized by little or no negative Eu anomaly. Parallelism and general similarity in REE patterns (Fig. 3) along with very narrow range of variation in ratios of REE such as La/Ce_n (1.28–2.65), La/Nd_n (1.98–3.71) and Ce/Nd_n(1.06–2.24) suggest that the REE concentrations are inherited from source.

For the positive Eu anomaly as shown by the MKV, it is possible that the source for MKV magma was a plagioclase cumulate assemblage. Melting of such a material would result in liquids possibly characterized by a positive Eu anomaly of a magnitude smaller than that in the parent rock. But this possibility appears remote remote, because plagioclase is not a stable phase above pressure of 8.6–11 Kb, depending on its exact composition (Herzberg 1972). In any event, melting at shallow depths (<35 km) within the plagioclase peridotite stability field would require an exceptionally high geothermal gradient and would also preclude direct origin of such magma from deep mantle plume (Leeman 1976). Furthermore, the assumed modeling of MKV suggested their derivation from higher extent of melting followed by fractional crystallization. It is possible that the fractionation has greatly participated in the magma evolution causing plagioclase accumulation. It is therefore, important to assess the effects of this process for Eu enrichment.

Low pressure crystallization of moderate amount of olivine and plagioclase will tend to enrich the magma in all of the REEs and to cause a relative depletion in Eu. The net effect is to decrease the Eu anomaly of the melt. In case of MK magma, the observed magnitude of positive Eu anomaly would be minimal, and original melt would be characterized by a larger Eu anomaly. High pressure fractionation of a phase such as clinopyroxene, which has a slight negative Eu anomaly in its partition coefficient pattern, would tend to enrich Eu in the melt relative to the other REEs. However, more than 70 % of pure clinopyroxene is required to produce an anomaly of the magnitude observed in MKV under normal terrestrial redox conditions (Grutzeck et al. 1974). Such extensive fractionation seems unlikely, and if it occurred, it would possibly deplete the residual liquid in Cr relative to Ni. Such depletions are not observed in these rocks (Cr = avg. 30.75; Ni = avg. 15). And thus it seems more reasonable to interpret that the positive Eu anomalies of Madhan–Kudhan samples were inherited during the partial melting (Leeman 1976).

Due to close spatial relationship between the samples of Kolihan and Madhan–Kudhan, the large variation in their degrees of partial melting (avg. 4 % and 20 % respectively) seems unlikely. However, it is in accordance with the observations of Sun and Nesbitt (1978) based on TiO_2 contents and CaO/TiO_2 and Al_2O_3/TiO_2 ratios to separate different magma types. They noted that these ratios in primitive MORB magmas increase with the degree of partial melting up to 20 and 17 respectively at approximate 0.8 % TiO2. Low Ti (<0.6 % TiO2) basalts from ophiolite complexes, island arcs and inter-arc basins have much higher ratios (up to 60). The MKV samples are characterized by very high CaO/TiO2 (93.86) and Al2O3/TiO2 (63.72) ratios compared to those of primitive MORB. These high ratios can not be produced by increasing degrees of melting from any known mantle mineralogy and the Ti depletion must be inherited from the source. One possible origin suggested for these high CaO/TiO2 and Al2O3/TiO2 ratio low-Ti basalts is remelting of a source depleted by an earlier magmatic episode (Gaskarth and Parslow 1987). In the light of these arguments, the petrogenetic processes which played a part in the origin of the KHV, is the melting of a descending plate. This mechanism involved two stages of magma generation i.e. production of magma due to low degree of partial melting ($\sim$4 %) at shallow depths producing Kolihan volcanics followed by a second episode of magma generation at deeper level through remelting of the source (already depleted by the magmatic episode of Kolihan) due to progressive subduction causing high degree of partial melting ($\sim$20 %) that supplied lavas for Madhan–Kudhan volcanics.

Khetri basin is in linear alignment with South Delhi belt lying $\sim$200 km in south and hosts Phulad ophiolite suite (Gupta et al. 1997). Geochemical studies on the magmatic rocks of this ophiolitic complex and associated sediments have established that Phulad ophiolite developed in a fore arc setting and the deposition of South Delhi sediments in back arc setting (Khan et al. 2005; Fatima and Khan 2012). Thus it can be interpreted that the process of generation of oceanic crust which got initiated in the Khetri basin was completed in the South Delhi belt probably due to the change in subduction direction/angle. Examples for this type of crustal evolution are found in central and southern Chile (Aberg et al. 1984; Barthlomew and Tarney 1984) and Grenville Province, Canada (Smith and Holm 1987). It has been proposed that the Andean continental margin has been subjected to alternating periods of extension and compression resulting in complex migration of the loci magmatism and changes in the character of magmatism. Calc alkaline activity is associated with compressional periods silicic mafic volcanism occurred in the basins developed by extension.

8 Conclusions

The geologic record shows that Bayana, Alwar and Khetri lavas outcrop in NDFB at ~1,832 Ma (Guerot 1993). Major element compositions suggest that most Bayana and Alwar lavas are derived from a mantle peridotite source traversed by pyroxenite veins. The Khetri lavas, on the other hand, are consistent with magmas from the mantle wedge metasomatized by melts from subducting slab. The petrogenetic modeling shows that the BYV suite was formed from the melts derived from 1 % to 10 % (avg. 4 %) of the partial melting of a spinel lherzolite source giving a residual mineralogy of 56 % olv, 25 % opx and 19 % cpx. Whereas the ALV suite appeared to be evolved through 12 %–20 % (avg. 15 %) partial melting of the same source with a residual mineralogy 61 % Olv, 25 % Opx and 14 % Cpx. This source modeling authenticates the earlier interpretation (Raza et al. 2007), that the melts of Bayana and Alwar tholeiites were generated by partial melting of a common source within the spinel stability field under the influence of mantle plume. The geochemical attributes e.g. similar magma type but different degrees of enrichment in LREE, Nb Ta and HFSE and HREE, of Bayana and Alwar volcanics in conjunction with their associated shallow water sediments suggest their eruption in a continental rift. Thus these volcanic suites in NW Indian shield may be considered to have been evolved in an extensional environment produced by stretching and attenuation of the old subcontinental lithosphere in response to upwelling plume. Consequently linear rifts were developed in the Achaean basement along the weak zones in the lithosphere which paved the routes for upwelling melts. In Bayana area the asthenosphere suffered ~4 % partial melting to produce tholeiitic basalt with flood basalt signatures. In response to the same extension, further attenuation of the crust caused more shallowing of the asthenosphere which resulted in its further melting (higher degree of partial melting ~15 %). These melts were erupted in Alwar basin along with sedimentary processes to produce tholeiite with OIB type signatures. During their ascent to the surface the Bayana tholeiites underwent moderate contamination/modification as they traversed a relatively thick continental crust and thus acquired enriched character. Whereas, due to more attenuated nature of the crust in the Alwar basin, tholeiitic magma ascended unhindered causing little variation in its chemistry.

The volcanic suites of Khetri basin show subduction signatures and all the features which are consistent with their development in an ensialic back arc basin. Modeling of KLV suite suggests two tier magma production in Khetri region. First suite of magma was derived from 1 %–6 % (avg. 4 %) partial melting with residual mineralogy 56 % Olv, 25 % Opx and 19 % Cpx whereas the second batch of magma was generated by 15 %–30 % partial melting of the same source leaving behind 64 % Olv, 25 % Opx and 11 % Cpx as residual mineralogy. The transitional nature of KHV between tholeiite and calc-alkaline basalt along with distinctive subduction signatures may be explained by operation of three petrogenetic processes, (i) mixing of mantle-derived basaltic magma with sialic crustal material and (ii) melting of a descending plate, (iii) opening of a slab "window" caused by detachment of the subducting slab. First two possibilities can not explain the coexistence of such two magmas of KLV and MKV. The third option i.e. the detachment of the subducting slab will produce a slab-free area in the form of a large window caused by the subduction of an active spreading center. Structural weakness in the subducting slab can trigger a tear and lead to the rapid sinking of the detached slab into the mantle (e.g., Davies and von Blanckenburg 1995; Wortel and Spakman 2000). This will produce an area free of subducting lithosphere that propagates laterally as the oceanic plate continues to tear along the strike of the subducting slab. The detached slab is replaced by upwelling asthenosphere (Levin et al. 2002; Ferrari 2004; Pallares et al. 2007). It may be suggested that the collision of the Bayana–Alwar tracks with the Khetri arc clogged or slowed the subduction processes and triggered the detachment of a segment of the subducting slab. The detached slab was replaced by hot and buoyant asthenosphere that produced tholeiitic basalts. Nevertheless, the most plausible model for origin of Khetri mafic volcanic rocks despite being complex, offers a reasonable solution. A mantle wedge resting upon a subducted oceanic plate partially melted (supra subduction zone setting) and later ward subducted plate itself got melted. Source modeling of Kolihan volcanics suggests that they were derived from the melts generated at low degree of partial melting (~4 %) by subduction at shallow depth. While a significantly high melting (~20 %) is accounted for Madhan–Kudhan volcanics which is possible at deeper level in the subduction zone. This implies progressive subduction resulted in the melting of the subducted plate itself. The geochemical features of Khetri volcanics in conjunction with their source modeling advocate for the evolution of Khetri basin probably on a continental crust in the back of an arc system. It further suggests that the initiation of the formation of oceanic crust in this part/region of Indian shield during Proterozoic which was completed in further south as evident from the emplacement of Phulad ophiolite in South Delhi belt. Thus, full Wilson cycle operated that built up the Precambrian crust of the NW Indian shield at about 1,832 Ma, the age of mafic volcanics of Khetri basin; Guerot 1993). Moreover, the spatial existence of three suites of mafic volcanics of diverse chemical signatures is best example of subduction–plume interaction.

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