Petrology and geochemistry of greywackes of the ~1.6 Ga Middle Aravalli Supergroup, northwest India: evidence for active margin processes

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Geochemical and petrological studies of the well-preserved greywacke horizon of the ‘Middle Aravalli Group’ were carried out to constrain the early evolution of the Aravalli basin. Petrological and geochemical attributes of Middle Aravalli greywackes (MAGs) such as very poor sorting, high angularity of framework grains, presence of fresh plagioclase and K-feldspars, variable Chemical Index of Alteration (CIA) index (46.7–74.5, avg. 61), and high Index of Compositional Variability (ICV) value (~1.05) suggest rapid physical erosion accompanying an active tectonic regime. The sediments record post-depositional K-metasomatism and extraneous addition of 0–25% (avg. ~10%) K is indicated. Assuming close system behaviour of immobile elements during sedimentation, various diagnostic element ratios such as Th/Sc, La/Sc, Zr/Sc, and Co/Th, Eu anomaly and rare earth element patterns of MAG suggest that the Archaean Banded Gneissic Complex (BGC) basement was not the major source of sediments. In conjunction with the dominant 1.8–1.6 Ga detrital zircon age peaks of Middle Aravalli clastic rocks, these data rather indicate that the sediments were derived from a young differentiated continental margin-type arc of andesite–dacite–rhyodacite composition. A highly fractionated mid-oceanic-ridge-basalt-normalized trace element pattern of MAGs, with characteristic enrichment of large-ion lithophile elements (LILEs), depletion of heavy rare earth elements, negative Nb-Ta, Ti and P anomalies, positive Pb anomaly, and distinctive Nb/Ta, Zr/Sm, Th/Yb, and Ta/Yb, Ce/Pb ratios envelop the composition of modern continental arc magmas (andesite–dacite) of the Andes, suggesting a subduction zone tectonic setting for precursor magma. High magnitude of LILE enrichment and high Th/Yb ratios in these sediments indicate that thick continental crust (~70 km) underlay the ‘Middle Aravalli’ continental arc, similar to the Central Volcanic Zone of the modern Andes. We propose that eastward subduction of Delwara oceanic crust beneath the BGC continent led to the formation of a continental volcanic arc, which supplied detritus to the forearc basin situated to the west. This model also explains the opening of linear enialic basins in the Bhilwara terrain, such as in Rajpura–Dariba and Rampura–Agucha in a classical back-arc extension regime, similar to the Andean continental margin of the Mesozoic. On the basis of the recent \textsuperscript{207}Pb/\textsuperscript{206}Pb detrital zircon age of Middle Aravalli sediment, a time frame between 1772 and 1586 Ma can be assigned for Middle Aravalli continental arc magmatism.

**Keywords:** greywacke; Aravalli Supergroup; trace element geochemistry; continental arc; provenance; weathering; K-metasomatism

**Introduction**

The clastic sedimentary archive is one of the main sources of information regarding past geological conditions that prevailed on the Earth’s surface. Such sediments preserve detritus from ancient orogens, which may get obscured later by tectonic overprinting or even eroded. In many cases, clastic rocks provide important clues to long-eroded or obscured source rocks (e.g. Saha et al. 2004; Chakrabarti et al. 2007; Guo et al. 2012; Wang et al. 2012). Clastic sedimentary rocks have also been used to understand orogenic progression, unroofing, tectonic setting of the provenance, and climatic conditions during the time of their deposition (Basu et al. 1990; Nie et al. 2012; references therein). The chemical record of clastic sedimentary rocks is considered to be influenced by source rock characteristics, chemical weathering, sorting processes during transport, sedimentation, and post-depositional diagenetic reactions (McLennan 1989; Nesbitt et al. 1996). However, certain relatively less mobile element pairs involving incompatible and compatible elements (such as Ti, Sc, Th, Zr, Hf, Cr, and Co), Rare Earth element (REE) patterns, and Nd isotope record of fine-grained clastic sedimentary rocks provide important clues regarding provenance and composition of upper continental crust (CC), and place important constraints on crustal extraction events and crustal growth models (e.g. Miller and O’Nions 1984; Taylor and McLennan 1985, 1995; Cullers 2000; Banner 2004; Saha et al. 2004; Condie 2005; Chakrabarti et al. 2007; Yang et al. 2012). Some aspects of the nature and intensity of chemical weathering, interaction between lithosphere–atmosphere–hydrosphere, can be estimated and inferred from the behaviour of different labile elements, such as alkalai and alkaline earth elements of clastic sedimentary rocks, by utilizing their analogous behaviour in modern weathering.

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profiles (e.g. Nesbitt et al. 1980; Nesbitt and Young 1982; Fedo et al. 1996; González-Álvarez and Kerrich 2012).

It is well known that much of the growth of CC takes place in the convergent continental margin (Condie 2005; Hawkesworth et al. 2010), and turbidites deposited along the continental margin provide important constraints on the evolution of the accretionary and juvenile magmatic episodes (e.g. Kemp et al. 2006; Cawood and Buchan 2007; Lamaskin et al. 2008; Condie et al. 2009, 2011; Lee 2009). The turbidites formed in such settings are transported down high gradient streams and commonly accumulated in forearc or intraarc basins. Because of high relief, significant recycling does not occur; as a result, these sediments retain the characteristics of their source rocks and are therefore ideal candidates for studying geodynamic evolution of sedimentary basins (Naqvi et al. 1988; Feng and Kerrich 1990; McLennan et al. 1990; Manikyamba et al. 1997; Guo et al. 2012). The Aravalli Mountain Range (AMR) of northwestern India (Figure 1, inset) represents one of the major Palaeo-Mesoproterozoic accretory orogens of the world. Consequently, understanding of its evolution has important bearing on Proterozoic crustal evolution vis-à-vis assembly and growth of the supercontinent Columbia. However, the early evolution of the orogen is still poorly understood. A thick greywacke unit in the Middle Aravalli Group shows typical features of turbidites (Roy and Jakhar 2002; Bhattacharya and Bull 2010) and offers an opportunity to unravel the early evolution of the Aravalli orogen. We have conducted geochemical and petrological studies on these well-preserved greywacke horizons to glean information on the weathering, diagenesis, and recycling history of the source terrain, as well as the provenance and tectonic setting of the deposition of the greywacke unit. Finally, with the aid of high-quality high-field-strength element (HFSE)–REE data, an attempt has been made to unravel the early geodynamic evolution of the Aravalli basin.

Geological setting

The AMR of northwest India extends over 700 km in length, with a general trend of NE–SW (Figure 1). It is composed of a Palaeo- to Neoarchaean cratonic nucleus (~3.3–2.5 Ga; Gopalan et al. 1990; Wiedenbeck and Goswami 1994; Roy and Kröner 1996; Wiedenbeck et al. 1996) and two major Proterozoic orogenic belts, i.e. the Aravalli fold belt and the younger Delhi fold belt (Figures 1 and 2). The cratonic nucleus is made up of tonalite-trondhjemite-ganodiorite (TTG) gneisses, intrusive granitoids with minor metasedimentary and metavolcanic units, collectively designated the Banded Gneissic Complex (BGC) or Mewar Gneiss Complex (Roy and Jakhar 2002). The supracrustal sedimentary sequences of the Aravalli Supergroup (~2.1–1.9 Ga; Deb and Thorpe 2004) and Delhi Supergroup (~1.85–1.70 Ga and ~1.0–0.85 Ga; cf., Kaur et al. 2011) constitute the orogenic belt, and there is a general consensus that these were deposited during episodes of continent rifting (e.g. Singh 1988; Sinha-Roy 1988; Bhattacharya and Bull 2010).

Rocks belonging to the Aravalli Supergroup cover a wide region in the eastern and southeastern belts of the Aravalli Mountains. The entire region can be divided from north to south into three sectors: (1) the Bhilwara sector; (2) Udaipur sector; and (3) Lunavada sector (Figure 1). Out of these three sectors, the Udaipur sector is known as the type area of the Aravalli rocks, where they show evidence of complete development of the stratigraphic succession and structural evolution (Figure 2). The Aravalli rocks of the Udaipur sector characteristically show two contrasting sedimentary facies associations which are distributed along two roughly N–S-trending belts. The eastern belt represents a shelf sequence, wherein carbonate rocks are the dominant component. The western belt, by contrast, is characterized by a carbonate-free, pelite-dominated sequence with thin intercalated arenites. The two sedimentary associations correspond to a near-shore shelf facies and a deep-sea or distal facies, respectively (Poddar 1966; Roy and Paliwal 1981; Roy et al. 1988, 1993; Roy 1990). Roy (2000) proposed a three-fold classification for the Aravalli Supergroup – Lower, Middle, and Upper Aravalli groups (Figure 3) – on the basis of unconformity-bounded litho-assemblages.

In the Udaipur sector, the interface of the basement BGC and Aravalli Supergroup is characterized by the presence of alumina-rich and iron-poor palaeosols (Sreenivas et al. 2001a), indicating extreme weathering conditions and a large time gap before initiation of Aravalli sedimentation (Roy et al. 1988). The Lower Aravalli Group was deposited in an active rift basin during a marine transgression (Roy and Paliwal 1981). The basal Debari Formation is made up of an intercalated sequence of metabasaltic rocks and well-sorted feldspathic quartzites that were probably deposited in a shallow marine rift basin. Shallow-marine dolomites and carbonaceous phyllites belonging to the Jhamarkotra Formation occur stratigraphically above the Debari Formation. The sedimentary facies indicate stabilization of the rift basin and formation of stable passive-margin-like condition (Roy and Paliwal 1981). The Middle Aravalli Group commences with deep water turbidite sequences of the Udaipur Formation comprising metagreywacke and phyllite, indicating deepening of the carbonate platform and active tectonism. The overlying dolomite, quartzite, and phyllite–dolomite–quartzite interbands belonging, respectively, to the Mochia (Zawar), Bowa, and Tidi formations suggest shallow marine conditions and re-establishment of passive margin conditions. Debari Formation quartzites, Kabita Dolomite, and Lakahawali phyllites of the Upper Aravalli Group represent shallow water sedimentation in the eastern shelf, whereas mica schists and ultramafic rocks of the Jharol Formation represent deep-sea sedimentation in the western open marine basin (Roy and Jakhar 2002).
Figure 1. Generalized geological map of the AMR, northwestern India (modified after Roy et al. 1988), showing major Precambrian lithotectonic units. The location of the AMR is shown in the inset map as the solid grey shade. The inset map also depicts the framework of the Indian shield and major crustal discontinuities. The Central Indian Tectonic Zone separates the South Indian cratonic ensemble, consisting of the Dharwar, Bastar, and Singhbhum cratons, from the North Indian craton, comprising the Bundelkand–Aravalli craton. Geochronological information is obtained from compilations of Roy and Jakhar (2002), Deb and Thorpe (2004), and McKenzie et al. (2013). Abbreviations: Bnd, Bundelkhand craton; Bst, Bastar craton; Dh, Dharwar craton; Sbh, Singhbhum craton.
Geochronology of the Aravalli Supergroup

The Aravalli Supergroup is traditionally considered to have been deposited during 2.1–1.9 Ga (see review by Deb and Thorpe 2004). However, recent $^{207}\text{Pb}/^{206}\text{Pb}$ geochronological data of detrital zircons (McKenzie et al. 2013) strongly refute an early Palaeoproterozoic age of the Aravalli Supergroup and rather suggest a late Palaeoproterozoic to Mesoproterozoic age (1772–1586 Ma) for these strata. Zircons of lower Aravalli Delwara quartzites ($n = 96$) dominantly show 3.5–2.5 Ga age peaks with a single grain showing a 1709 ± 8 Ma age, precluding the possibility of precise depositional age constraint (McKenzie et al. 2013).

Figure 2. Detailed geological map of the Aravalli Supergroup around the Udaipur area (modified after Roy et al. 1988). Sample locations are shown for reference. Age information after McKenzie et al. (2013).
However, clastic rocks of the overlying Jhamarkotra Formation yielded a prominent age peak at 1772 Ma with subordinate peaks at the late Archaean (2.5–2.6 Ga) and early Palaeoproterozoic (2.2 Ga) (Figure 4). The youngest peak of 1772 Ma is interpreted to represent the maximum depositional age of the Jhamarkotra Formation. Middle Aravalli clastic rocks show distinct peaks at 1772, 1646, and 1586 Ma along with a prominent Neoarchean peak (2.6 Ga) (Figure 4). On the basis of these data, a maximum depositional age of 1586 Ma is assigned for the Middle Aravalli Group. On the basis of detrital zircon U–Pb ages, Kaur et al. (2011, 2013) suggested an ~1.7 Ga depositional age for the North Delhi Belt (Figure 1); whereas McKenzie et al. (2013) reported a 1.2–1.0 Ga depositional age for the South Delhi Belt (Figure 1). These recent works demonstrate that the Aravalli Supergroup and so-called North Delhi Belt rocks are coeval and represent contiguous sedimentation in an active margin basin, whereas the South Delhi Belt represents a younger phase of sedimentation. McKenzie et al. (2013) further postulated contiguous sediment sources for both the Aravalli and Vindhyan supergroups and suggested that tectonically deformed Aravalli–Delhi Orogenic Belt strata represent the distal margin equivalents of the Vindhyan successions.

These new age data on the Aravalli Supergroup call for reassessment of earlier geochronological data, which were
either rejected or interpreted as the metamorphic/reset age. Sreenivas (1999) obtained an Sm–Nd errorchron age of $1.83 \pm 0.51$ Ga ($\text{Nd}_i = 0.50989$, $\varepsilon_{\text{Nd}} = -7.44$, mean square weighted deviation = 2.35) for basal Aravalli mafic volcanics of the Delwara Formation of the Negaria area. Sreenivas (1999) also reported unpublished Sm–Nd data obtained by Gopalan and Macdougall of Delwara mafic volcanics, which yielded a similar age correlation, albeit with large error ($1.843 \pm 1.16$ Ga). These authors interpreted this age to be the reset age that corresponds to the closure of the Aravalli orogeny. However, with the new detrital zircon age data, we interpret an $\sim 1.83$ Ga age for emplacement of the basal Aravalli volcanic rocks.

**Sampling and analytical methods**

Fresh homogeneous representative samples of Middle Aravalli metagreywackes (MAGs) were collected from quarries, road/railway cuttings, and natural outcrops (Figure 2). Weathered and jointed surfaces with quartz or calcite veins were avoided. Chips of fresh rock samples were cleaned by ultrasonication in distilled water and air-dried. The sample chips were reduced to sand size by crushing in a steel mortar, and were then powdered to $\sim 200$ mesh in a agate ball mill and agate mortar to avoid contamination (cf. Sreenivas et al. 1995) and maintain the homogeneity of the sample. Various mineral phases were identified by the standard petrographic technique and powder X-ray diffraction (XRD) method at the Department of Earth Sciences, Pondicherry University.

We selected 16 samples for major element analysis and representative 7 samples were analysed for trace elements. Major element concentrations were estimated by the standard X-ray fluorescence (XRF) technique on fused glass beads and pressed pellets at the National Geophysical Research Institute, Hyderabad, and at the Central Instrumentation Facility (CIF) of Pondicherry University, Pondicherry, India. The XRF had been calibrated with international standards procured from the US Geological Survey. Accuracy of the XRF analyses is estimated to be better than 2% for major oxides. The details of the analytical procedures are given in Govil (1985). Determination of trace elements and REEs was carried out using inductively coupled plasma mass spectrometry (ICP-MS) techniques at the National Geophysical Research Institute, Hyderabad, India, following the procedure of Balaram et al. (1996). Precision and accuracy of ICP-MS analysis were better than 5% (Balaram et al. 1996, 1999).
are characterized by bimodal texture, with some large clasts (>1000 µm) of feldspar and relatively large (≈500 µm) framework grains, which are set in fine- to very-fine-grained (≈100 µm) groundmass. Some samples preserve the first-order volcaniclastic signature and show typical intergranular texture with thin plagioclase laths and opaque minerals (Figure 5C). All the studied samples contain 30–70% matrix, which is composed mainly of biotite–chlorite and fine quartz that are interspersed with fine Fe–Ti opaque minerals (Figure 5D). Some samples show development of foliation. Because of the relatively high proportion of matrix and partly metamorphic reconstitution, we did not carry out grain counting of framework grains. Among the framework grains, quartz is relatively abundant and is generally subangular to angular in shape, mostly monocrystalline, and lacks inclusions and undulatory extinction. Some quartz grains show splintery shapes, whereas others exhibit perfect idiomorphic shapes (rhomb shape), possibly indicating volcanic parentage for these. Some of the quartz grains show embayed boundaries, indicating partial dissolution of quartz during diagenesis. Angular feldspar is the dominant mineral fragment, which is generally fresh, and plagioclase is far more common than K-feldspar. Fresh plagioclase feldspars show both Carlsbad twinning and albite twinning. Lithic fragments are relatively rare and include fragments of polycrystalline quartz, quartzite, schistose, and volcanic rocks (cf. Banerjee and Bhattacharya 1994). A few grains of pyroxene and dark brown coloured altered Fe–Mg minerals were noticed in all the samples. One sample with well-preserved sedimentary texture indicates Q_35F_60L_5 composition of detrital modes. Current results rather suggest that these rocks differ from true greywacke and can be classified as feldspathic wacke.

Since these greywackes contain >50% matrix, we have carried out XRD studies on these samples. These samples typically contain quartz, plagioclase (mostly albite), microcline, biotite, chlorite (both chamosite and clinochlore), muscovite, titanomagnetite, and magnetite.

**Geochemistry**

Major element oxide data, various oxide ratios, and computed geochemical parameters such as the Chemical Index of Alteration (CIA), Plagioclase Index of Alteration (PIA), and Index of Compositional Variability (ICV) of 16 greywacke samples are presented in Table 2. Compositionally, MAGs are characterized by a restricted range in SiO_2/Al_2O_3–Fe_2O_3/K_2O space (Supplementary Figure 1a, after Herron 1988, see http://dx.doi.org/10.1080/00206814.2014.999355 for supplemental figures) and can be classified as wacke. In comparison to normal greywacke, the MAG samples of the present study are chemically distinct. They show distinct enrichment of K over Na, significant enrichment in Fe and Mg, and can be classified as ferroan potassic sandstone in a Fe_2O_3 + MgO–Na_2O–K_2O ternary diagram (Supplementary Figure 1b, after Blatt et al. 1980). This is in contrast to the earlier reported geochemical data of MAG (by

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**Figure 5.** Photomicrographs (cross-polarized light) of MAG showing: (A) general texture with poor sorting and presence of angular clasts of plagioclase, K-feldspar, and embayed monocrystalline quartz within a biotite/chlorite matrix; (B) presence of opaque minerals (magnetite–titanomagnetite) in groundmass; (C) typical intergranular texture with thin plagioclase laths and opaque minerals, indicating a first-order volcaniclastic nature; (D) development of foliation and presence of lithic clast. Abbreviations: p, plagioclase; kf, K-feldspar; mq, monocrystalline quartz; pq, polycrystalline quartz; ls, schistose lithic fragment; o, opaque minerals.
Banerjee and Bhattacharyya 1994), which document diverse composition ranging from shale to subarkose (Supplementary Figure 1a) and have much higher Na/K ratios compared with the present data. A similar diverse compositional range was also observed for Hindoli greywacke (Saxena and Pandit 2012). One sample (sample no N8), however, is arkosic in nature with higher SiO2 content (93.76%) and high Si/Al ratio (25.13) (Supplementary Figure 1a, Table 1). This sample appears to be an anomalous outlier and indicates fictitious correlation among other samples in the set, and is, therefore, excluded from further statistical computation. Harker-type variation diagrams show a wide scattering of data points with SiO2 exhibiting no significant correlation with other major oxides \( r = -0.29 \) (Al2O3), \(-0.41\) (TiO2), \(-0.45\) (Fe2O3), \(-0.29\) (MnO), \(-0.6\) (MgO), \(-0.2\) (CaO), \(0.34\) (Na2O), \(-0.52\) (K2O), \(-0.59\) (P2O5)]. Similarly, Al2O3 does not show any correlation with other oxides except K2O \( r = 0.85\). The absence of strong negative correlation between Al and Si \( r = -0.29\) suggests that sedimentary sorting and fractionation of framework silicate and phyllosilicate minerals between bedload and suspended load did not take place in the catchment area (e.g. Fralick and Kronberg 1997). The high order correlation between K and Al suggests that K is hosted in K-Feldspars and muscovites. In comparison to upper CC (UCC, Rudnick and Gao 2003), MAGs are characterized by slight enrichment of Fe and Mg, minor depletion of Ti, moderate depletion in Na and Ca, and high enrichment in K (Supplementary Figure 2). These indicate that the source terrain was somewhat more mafic in composition compared with UCC and perhaps underwent mild weathering and significant post-depositional K-metasomatism. In comparison to predominantly volcanogenic early Proterozoic greywacke (Condie 1993), MAG samples are slightly depleted in Fe and Ti contents but enriched in both Mg and K contents (Supplementary Figure 2).

The trace and REE data are presented in Table 2. The transition trace elements such as Sc, V, and Cr are strongly correlated \( r > 0.7\) with Fe and Ti but are only weakly correlated with Mg, suggesting that these elements are hosted in Fe–Ti oxides. Ni and Co are decoupled from other transition elements, the reason for which is not clear. The large-ion lithophile elements (LILeS) and HFSEs such as Rb, Cs, Nb, and Ta are moderately \( r > 0.56\) correlated with Si, preserving their original magmatic signature. The trace and REE do not show any significant correlation with Al, precluding weathering-related clay mineral control and suggest that the elements are hosted in primary silicate, oxide, and phosphate minerals. ΣREEs, light rare earth elements (LREEs), and Th exhibit strong negative \( r = -0.7\) correlation with Mg and Mn, indicating that these elements are excluded from mafic minerals and concentrated in felsic minerals. The heavy rare earth elements (HREEs) and Y are moderately correlated with P, Ti, and Fe \( r = 0.54\), indicating that they are hosted in accessory minerals such as xenotime, allanite, and titanite. The absence of correlation between Zr and HREEs, and, between Hf and HREEs rules out control of zircon on HREE chemistry. Similarly, the lack of correlation of ΣREE with Ca \( r = -0.2\), Sr \( r = 0.16\), and P \( r = 0.17\) indicates that REEs are not hosted in apatite. However, strong positive correlation between Eu and Sr \( r = 0.83\) suggests that Eu is probably hosted in plagioclase feldspars. In comparison to present day UCC (Rudnick and Gao 2003, Supplementary Figure 3), MAGs are slightly enriched (1.5–2 X UCC) with LILEs such as K, Rb, Cs, and Th. The ferromagnesian transition elements (Sc, V, Co, Cr) show variable concentration with about half of the samples enriched (1.5–2 X UCC) and the rest of the samples depleted (0.6–0.7 X UCC). The average Aravalli greywackes are marginally enriched (3–11%) in transition elements compared with UCC. Some of the samples exhibit negative anomalies for Na and Sc, and positive anomaly for Ba. These features can be explained by more intense weathering conditions, more felsic provenance compared with UCC, and Ba addition during K-metasomatism. Large variations of labile elements of Ca, Sr, and Na (0.1–1.0 X UCC) in these probably indicate a non-steady state of weathering.

As far as REE concentration is concerned, the MAGs are moderately enriched in LREEs (20%), slightly enriched in middle rare earth elements (MREEs) (3%), and depleted in HREEs (4.5%) in comparison to UCC (Supplementary Figure 3). In comparison to UCC, MAGs are characterized by contrasting geochemical features such as (a) more fractionated REE patterns and (b) positive Eu anomalies, indicating adakite–sanukitoid-type source rocks. The MAG exhibits coherent chondrite normalized REE patterns with high LREEs \( La = 95–236\) X chondrite, avg. 169), moderate MREE \( Sm = 16–47\) X chondrite, avg. 33), and depleted HREEs \( Yb = 6–18\), ~10 X chondrite, avg. 11) contents. All the samples are characterized by high total REE fractionation \( La/YbN = 11.68–22.18\), avg. 15.69), with both LREE \( La/SmN = 4.48–5.56\), avg. 5.22) and HREE \( Gd/YbN = 1.33–2.28\), avg. 1.81) fractionation and moderate to no Eu anomalies \( Eu/Eu* = 0.52–1.0\), avg. 0.8).

Discussion

Weathering and K-metasomatism

The intensity of chemical weathering of the source terrain can be quantitatively evaluated by computing CIA of derived clastic rocks (Nesbitt and Young 1982, 1984) and is defined as: CIA = \([Al2O3/(Al2O3 + CaO* + Na2O + K2O)] × 100\). In the expression of CIA, all elements are in molecular proportion and CaO* represents Ca in silicate fraction. Selective removal of labile cations (e.g. Ca2+, Na+, K+) over stable residual constituents (Al3+, Ti4+) during
Table 1. Major element compositions of greywackes of the Middle Aravalli Supergroup, along with some important geochemical indices and ratios.

| Sample No. | Udaipur Formation | Zawar Formation |
|------------|-------------------|-----------------|
|            | N1    | N2    | N3    | N4    | N5    | N6    | S1    | S2    | S3    | S4    | S5    | S6    | S7    | N7    | N8    | N9    | Average |
| SiO₂       | 60.33 | 67.78 | 73.59 | 68.39 | 64.79 | 67.88 | 66.35 | 59.47 | 67.50 | 62.80 | 75.43 | 68.44 | 66.61 | 93.76 | 64.32 | 66.78 |
| Al₂O₃      | 17.66 | 17.45 | 13.63 | 14.08 | 18.64 | 14.24 | 17.12 | 14.09 | 11.70 | 16.03 | 15.65 | 15.89 | 18.62 | 15.75 | 3.73  | 20.64 |
| TiO₂       | 0.56  | 0.39  | 0.66  | 0.63  | 0.58  | 0.62  | 1.01  | 0.66  | 0.46  | 0.38  | 0.49  | 0.72  | 0.83  | 0.05  | 0.61  | 0.61  |
| Fe₂O₃      | 6.87  | 4.35  | 5.61  | 6.03  | 6.14  | 4.49  | 6.43  | 10.91 | 9.56  | 4.77  | 2.76  | 2.60  | 3.79  | 6.71  | 0.33  | 2.49  | 5.57  |
| MnO        | 0.06  | 0.08  | 0.01  | 0.06  | 0.04  | 0.09  | 0.12  | 0.06  | 0.25  | 0.08  | 0.13  | 0.03  | 0.12  | 0.10  | 0.03  | 0.08  |
| MgO        | 4.16  | 1.83  | 2.38  | 3.29  | 3.35  | 2.27  | 2.01  | 3.44  | 3.35  | 4.44  | 1.08  | 2.66  | 3.07  | 3.71  | 0.09  | 1.62  | 2.84  |
| CaO        | 0.64  | 2.81  | 0.26  | 2.71  | 0.42  | 4.36  | 3.21  | 2.56  | 4.25  | 4.25  | 0.29  | 2.21  | 0.75  | 0.35  | 0.05  | 0.28  | 1.96  |
| Na₂O       | 1.06  | 0.45  | 1.44  | 1.39  | 0.61  | 1.09  | 0.46  | 3.27  | 2.37  | 4.44  | 1.08  | 2.66  | 3.07  | 1.60  | 0.14  | 1.08  | 1.42  |
| K₂O        | 8.23  | 4.69  | 2.11  | 3.13  | 5.08  | 3.34  | 3.92  | 3.39  | 2.62  | 5.25  | 3.87  | 4.16  | 5.25  | 2.51  | 1.69  | 8.53  | 4.41  |
| P₂O₅       | 0.19  | 0.09  | 0.17  | 0.13  | 0.12  | 0.13  | 0.19  | 0.31  | 0.22  | 0.15  | 0.06  | 0.13  | 0.12  | 0.21  | 0.04  | 0.20  | 0.16  |
| Fe₂O₃/K₂O  | 0.83  | 0.93  | 2.66  | 1.92  | 2.12  | 1.42  | 1.64  | 3.21  | 3.65  | 0.91  | 0.71  | 0.63  | 0.72  | 2.67  | 0.19  | 0.29  | 1.56  |
| SiO₂/Al₂O₃ | 3.42  | 3.88  | 5.40  | 4.86  | 3.48  | 4.77  | 3.87  | 4.22  | 5.77  | 3.92  | 4.82  | 4.31  | 3.58  | 4.32  | 25.13 | 3.12  | 4.25  |
| K₂O/Na₂O   | 1.20  | 1.04  | 1.46  | 2.25  | 3.38  | 1.33  | 42.33 | 0.72  | 31.50 | 3.33  | 9.95  | 1.26  | 5.02  | 1.57  | 12.10 | 7.90  | 9.01  |
| Al₂O₃/TiO₂ | 31.31 | 45.21 | 20.78 | 22.93 | 29.45 | 24.38 | 27.72 | 13.95 | 17.61 | 34.61 | 32.34 | 25.94 | 19.02 | 76.14 | 33.62 | 26.16 |
| Fe₂O₃ + MgO | 11.03 | 6.18  | 7.99  | 9.32  | 9.50  | 6.76  | 8.43  | 14.35 | 12.91 | 9.21  | 3.85  | 5.26  | 6.86  | 10.42 | 0.41  | 4.11  | 8.41  |
| ICV        | 1.22  | 0.84  | 0.91  | 1.22  | 0.87  | 1.24  | 0.95  | 1.85  | 1.76  | 1.31  | 0.57  | 0.98  | 0.79  | 1.00  | 0.63  | 0.71  | 1.08  |
| CIA        | 72.49 | 61.49 | 57.07 | 63.34 | 59.93 | 72.00 | 62.58 | 46.76 | 52.26 | 50.03 | 74.50 | 53.28 | 68.01 | 47.60 | 72.66 | 64.17 | 60.37 |

Note: Major elements in oxide wt%, CIA molecular ratio of \([\text{Al}_2\text{O}_3/(\text{Al}_2\text{O}_3 + \text{CaO}^* + \text{Na}_2\text{O} + \text{K}_2\text{O}) \times 100]\), where \(\text{CaO}^*\) represents Ca in silicate fraction (Nesbitt and Young 1982), ICV = \([\text{Fe}_2\text{O}_3 + \text{K}_2\text{O} + \text{Na}_2\text{O} + \text{CaO} + \text{MgO} + \text{MnO} + \text{TiO}_2]/\text{Al}_2\text{O}_3\) (Cox et al. 1995).
weathering in a warm and humid climate results in high CIA values (Nesbitt and Young 1982; Fedo et al. 1995). Near absence of chemical alteration results in low CIA values, which may reflect cool and/or arid conditions or alternatively rapid physical weathering and erosion under an active tectonic setting (Fedo et al. 1995; Nesbitt et al. 1997; Singh 2009, 2010). Fresh igneous rocks and minerals have CIA values of 50 or less. CIA values of intensely weathered residual material, such as gibbsite or kaolinite, approaches 100, whereas natural clastic sedimentary rocks have intermediate values depending on the intensity of weathering. The CIA values of MAG range between 46.7 and 74.5 with an average value of 61, indicating a low to moderate degree of weathering. Similarly, the ICV (Cox et al. 1995) is potentially useful to evaluate the degree of chemical weathering, which depicts formation of aluminous clay minerals over framework silicate minerals and is quantified as: ICV = [(Fe₂O₃ + K₂O + Na₂O + CaO + MgO + MnO + TiO₂)/Al₂O₃]. The ICV values of the MAG are quite high and vary between 0.56 and 1.56 with an average of 1.05, indicating the presence of primary silicate minerals and incipient nature of weathering. The weathering relationship of MAG is portrayed in the Al₂O₃–(CaO + Na₂O)–K₂O compositional diagram (A–CN–K, Figure 6). The
Following the procedure of Figure 7A, fine-grained fresh framework silicate minerals are characterized by a narrow range of SiO$_2$ values (3.41–59x10$^{-1}$) and plotted close to the feldspar join. Out of the remaining 10 samples, 6 samples are characterized by moderate CIA values (61.5–68), and 4 samples show moderately high CIA values (72–74.5). High variability of CIA values suggests a non-steady state of weathering and rapid physical erosion under an active tectonic regime and tapping of the complete zone of weathering profile (cf. Nesbitt et al. 1997). The UCC normalized multi-element diagram of MAG (Supplementary Figure 3) shows only minor depletion of smaller cations of alkaline earth elements (Ca, Na, Sr) relative to large alkali cations (K, Rb, Cs), possibly indicating a lesser degree of weathering (Nesbitt et al. 1980; Wronkiewicz and Condie 1987). This is noteworthy in the context of a worldwide moist and warm climate, high surface temperatures, and plume outbreaks during the Palaeoproterozoic (Condie et al. 2001; Eriksson et al. 2004; González-Álvarez and Kerrich 2012), which favoured extreme chemical weathering. Therefore, the relatively unweathered nature of MAGs suggests that these were originated and deposited under an active tectonic setting.

Petrographic evidence, for example the presence of abundant angular fresh plagioclase and K-feldspars grains along with volcanic and sedimentary rock fragments, indeed suggests a very low degree of chemical weathering, rapid physical erosion, and deposition under a turbiditic regime. Petrographic study also revealed distinct bimodality, with coarse-grained fresh framework silicate minerals (30–70%) and a fine-grained biotite–chlorite–clay-mineral-dominated matrix component (70–30%), which possibly reflect a dual nature of weathering. We argue that framework grains are the product of rapid physical weathering and uplift of plutonic rocks under active tectonism, while the matrix component represents the incipient to moderately weathered concomitant volcanic detritus, which is usually highly susceptible to chemical weathering. These together explain the variability of CIA values of MAG. The relatively unweathered nature of the sediments is also indicated by lack of correlation between Al and Ti ($r = -0.19$, Figure 7A) (cf. Sreenivas and Srinivasan 1994; Young and Nesbitt 1998). Another important feature is that the samples of the present study plot along a linear trend (trend-B, Figure 6) with a distinctly lower slope compared with the predicted weathering path (trend-A, Figure 6), indicating post-depositional K-metasomatism (cf. Fedo et al. 1996, 1997; Dey et al. 2008). This is also reflected by a high degree of correlation between K and Al ($r = 0.85$). Following the procedure of Fedo et al. (1996), we have computed the extent of K addition. About 50% of the population (eight samples) is affected by K-metasomatism and extraneous K-addition in these range between 6% (two samples) and 25% (three samples), with an intermediate value of 10% (three samples).

### Sedimentary recycling

MAG samples are both texturally and mineralogically immature and characterized by poor sorting, presence of angular grain, and fresh plagioclase feldspars, indicating the first-cycle nature of the sediments. Furthermore, MAGs are characterized by a narrow range of SiO$_2$/Al$_2$O$_3$ values (3.41–5.77, avg. 4.25) and SiO$_2$ does not antipathetically co-vary with the other major oxides. The absence of strong negative correlation between Al and Si ($r = -0.29$) suggests that sedimentary sorting and fractionation of tectosilicate and phyllosilicate minerals between the bedload and suspended load did not take place in the catchment area during transport (e.g. Fralick and Kronberg 1997). These features are consistent with the first-order primitive nature of the sediment debris and minimum partitioning of elements in to different mineralogical and grain size fractions during surface sedimentary processes, indicating rapid deposition under a turbiditic regime.

Some trace element ratios, such as Th/Sc, Zr/Sc, Th/U, and Rb/Sr, are useful for identifying the degree of sediment recycling and heavy mineral sorting (McLennan et al. 1993; Roser 2000). A positive linear correlation between the Th/Sc and Zr/Sc ratios expresses an igneous differentiation trend (McLennan et al. 1993). Addition of zircon during recycling causes a fast rise in Zr/Sc ratios compared to Th/Sc ratios, resulting in the change in slope of the correlation line. The Th/Sc–Zr/Sc diagram (Figure 8C) shows that the slope of the correlation line for Aravalli greywacke samples is similar to that of the igneous differentiation trend, ruling out sedimentary recycling or zircon addition. Coherence to the primary source
trend also suggests a first-cycle volcanioclastic nature of MAG sediments (McLennan et al. 1993; Roser et al. 2002). UCC and the mid-oceanic-ridge-basalt (MORB) normalized multi-element diagram (Supplementary Figure 3, Figure 9) show that HFSEs (Zr, Hf) are not decoupled from MREEs, further ruling out heavy mineral addition. Th/U ratios in MAG range between 4.5 and 9.2 (avg. 6.0), which is slightly higher than the upper crust (3.8), indicating that significant oxidative weathering and loss of hexavalent U did not take place (McLennan and Taylor 1980; McLennan et al. 1993). Rb/Sr ratios of MAG are also low (0.18–1.5, avg. 1.15) and only marginally higher than the Rb/Sr ratio of the upper crust (0.26, Rudnick and Gao 2003), indicating a simple recycling history (e.g. McLennan et al. 1993).

**Provenance**

Numerous factors, including chemical composition and weathering conditions of the source area, hydraulic sorting during transport and deposition, and diagenesis, affect the overall composition of clastic sediments. To characterize the provenance of terrigenous sediments, it is necessary to rely on the elements that are least mobile during these processes. Taylor and McLennan (1985) and McLennan and Taylor (1991) suggested that the REEs, Th, Sc, Co, Ti, and HFSEs (Zr, Hf) are especially useful for monitoring source area composition. These elements have very short residence times in sea water and are transferred almost quantitatively into the sedimentary record. Additionally, this array includes both incompatible (Th, REEs, HFSEs) and compatible elements (Sc, Co, Ti), ratios of which are useful in differentiating felsic from mafic source components. However, during recycling and transport, HFSEs are strongly partitioned in sand size grains and can be decoupled from other element groups because of heavy mineral fractionation (Taylor and McLennan 1985; McLennan 1989). But discussions from preceding sections apparently suggest that MAG samples suffered only mild weathering and are relatively unaffected by hydraulic sorting, and therefore we assume that the immobile elements behaved as a closed system during sedimentary processes and can be utilized to place constraints on the provenance composition.

Petrographic parameters such as the presence of abundant fresh plagioclase and K-feldspar, splintery and...
Figure 8. Trace element-based provenance discrimination diagrams for MAG. (A) Th/Sc–Sc diagram (BGC end members and mixing curve after Bhat and Ghosh 2001), (B) Co/Th–La/Sc diagram, (C) Th/Sc–Zr/Sc diagram, (D) La–Th–Sc ternary diagram indicating felsic to intermediate source rocks. Various source end members are also plotted for reference (HAB, high alkali basalt; LSA, low silica andesite; AND, andesite; DAC, dacite; RHY, rhyolite; T, TTG; G, granite; A, andesite; B, basalt, after Condie 1993; Roser 2000). La–Th–Sc diagram (after Bhatia and Crook 1986) also indicating deposition in a predominantly island arc tectonic setting. (Abbreviations are similar to Figure 7).

Figure 9. Normal-mid-oceanic-ridge-basalt (N-MORB) normalized multi-element diagram of MAG showing enrichment of LILEs, depletion of HREEs, negative anomalies of Nb–Ta, Ti, and P, and conspicuous positive anomaly of Pb. These features overlap the patterns of modern continental arc magmas of the Andes and suggest a similar tectonic setting with thick (~70 km) subarc CC.
idiomorphic monocrystalline quartz grains, minor lithic clasts of schists, volcanic rocks, and polycrystalline quartz grains suggest predominantly first-cycle igneous provenance with only a minor contribution from recycled orogen. The enrichment of Mg, Fe, and K in the UCC normalized diagram (Supplementary Figure 2) suggests a somewhat more mafic source, such as high Mg andesite or sanukitoid-type source rocks that are also enriched with LILEs (Kelemen et al. 2003; Martin et al. 2005). The concentration of mafic mineral compatible elements such as Ti, Fe, and Mg is strongly dependent on the composition of the source rocks (cf. Lamaskin et al. 2008). Although relative enrichment of Ti and Fe can take place in a weathering profile because of mass transfer of labile elements and the resultant volume loss (Tripathi and Rajaman 2007), this is not significant in case of mildly weathered sediments. The Ti vs. Fe + Mg diagram suggests that the source area was significantly mafic compared with granite and TTG end members of BGC basement and contribution from an andesitic to mafic source component appears plausible (Figure 7B). Al, Ti, and Zr remain immobile in sedimentary processes and their ratios can be utilised to infer source rock composition (Hayashi et al. 1997). Al$_2$O$_3$/TiO$_2$ ratios of igneous rocks are primarily controlled by igneous fractionation processes. Since Al resides mostly in feldspars and the Ti in mafic minerals (e.g. olivine, pyroxene, hornblende, biotite, ilmenite), the Al/Ti ratios of igneous rocks generally increase with increasing SiO$_2$ contents (Hayashi et al. 1997). The Al–Ti ratio diagram (Figure 7A) suggests felsic to intermediate igneous source rocks for MAG; similar inference can be drawn from Ti/Zr ratios (Figure 7C after Hayashi et al. 1997). Initial composition of the source rock can also be obtained from the weathering trend line at the feldspar join in the A–CN–K molecular proportion diagram (Figure 6). MAG samples fall along the correlation line that includes granodiorite, indicating felsic-intermediate source rock composition (Figure 6). Similarly, the CIA vs. ICV diagram (Figure 7D after, Potter et al. 2005; Lamaskin et al. 2008) suggests derivation of MAG sediments from dominantly andesitic source rocks. However, some samples fall along the basalt and granite mixing line, indicating mixed igneous provenance. Relatively weak and non-steady state weathering in the source areas are indicated by the large variability of CIA values and variable concentration of labile elements such as Ca, Na, and Sr (cf. Nesbitt et al. 1997). In this context, we infer that the sediments were supplied from a rapidly uplifted source area.

In the UCC normalized multi-element diagram (Supplementary Figure 3), MAG samples are only marginally enriched in LILEs such as Th, La, Rb, Ba, and K, indicating a source terrain of felsic composition. However, compatible transition elements such as Sc, V, Co, Cr, Ti, and Fe show variable concentrations. Whereas, some samples are considerably enriched, others are depleted, indicating a mafic-intermediate component in the source.

Relative contributions from felsic vs. mafic sources can be deduced from ratios of various immobile incompatible to compatible element pairs, such as Th/Sc, Zr/Sc, and La/Sc, La/Co (e.g. Taylor and McLennan 1985; Tran et al. 2003; Raza et al. 2010b; Fatima and Khan 2012; Wang et al. 2012; many others). The data of MAGs are plotted in various diagrams involving these element pairs, along with various possible basement end members and idealized source components for comparison (Figure 8). Th/Sc vs. Sc systematics is considered to be more robust (Bhat and Ghosh 2001; Manikyamba and Kerrich 2006; Manikyamba et al. 2008; Absar et al. 2009); in this diagram (Figure 8A), MAG defines a smooth mixing curve between TTG and the mafic end member of BGC, indicating mixing of felsic and mafic end members. Mass balance calculation suggests that the majority of the samples are consistent with the mixture of 60% mafic and 40% felsic end members, while some samples show a 90% felsic end member component. It is interesting to note that one sample (Sample no S2) indicates a 90% contribution from a mafic source, and these figures are likely to increase if the andesite end member is considered. A similar situation is observed in Co/Th/La/Sc (Figure 8B), Th/Sc–Zr/Sc (Figure 8C), and La–Th–Sc ternary diagrams (Figure 8D), which indicate variable mixtures between andesite, dacite, and rhyolite (or their plutonic equivalents) source rocks.

Based on the characteristic of REE patterns, the MAG samples can be grouped into three types. Type-I (Figure 10A) is characterized (samples – S-4, S-6, S-7) by a highly fractionated REE pattern (La/Yb$_N$ = 15.28–22.19, avg. 18.69), high HREE fractionation (Gd/Yb$_N$ = 1.78–2.29, avg. 2.10), and mild to no Eu anomaly (0.74–1). Type-II (Figure 10B, samples-S-1, S-2, S-3) shows much lower total REE fractionation (La/Yb$_N$ = 11.68–13.04, avg. 12.34) compared with Type-I, chiefly because of low HREE fractionation (Gd/Yb$_N$ = 1.34–1.54, avg. 1.43), and mild to no Eu anomaly (0.72–1). Type-III (Figure 10C, sample-S-5) is characterized by a fractionated REE pattern (La/Yb$_N$ = 16.69, Gd/Yb$_N$ = 2.06) with large negative Eu anomaly (Eu/Eu* ~ 0.5). These fractionated REE patterns suggest evolved felsic source rocks. Based on geochemistry and Nd isotopic systematics, McLennan et al. (1990, 1993) suggested four distinctive provenance components for modern deep-sea turbidites: (1) old upper CC; (2) young undifferentiated arc; (3) young differentiated (intracrustal) arc; and (4) MORB. Although the REE patterns of MAG are subparallel to TTG (BGC, ~3.3 Ga) and granitoid (Berach Granite, ~2.5 Ga) end members of local BGC basement, contribution from these old upper crustal sources can be ruled out since the MAG samples lack prominent negative Eu anomalies, unlike these basement end members (Eu/Eu* ~ 0.5, Gopalan et al. 1990; Ahmad and Tarney 1994; Raza et al. 2010a; Rahaman and Mondal 2014). This is further evidenced by the presence of prominent 1772–
1586 Ma detrital zircon age peaks in Middle Aravalli clastic rocks (Figure 4). The deep negative Eu anomaly of BGC basement is not an artefact of sampling bias; since sedimentary rocks locally derived from these, for example, feldspathic arenites of basal Aravalli riftogenic facies of Delwara quartzite \( (n = 8) \), basement Naharmagra quartzite \( (\sim 2.8 \text{ Ga}, n = 6) \), and Upper Aravalli passive margin Debari quartzites \( (n = 7) \) are indeed characterized by large negative Eu anomalies \( (\text{Figure 10D, data from Sreenivas et al. 1999}) \). The dominant BGC source components of Delwara quartzites are also evident from detrital zircon ages, which show prominent Mesoarchaean to late Neoarchaean peaks (McKenzie et al. 2013). Therefore, the possibility of a detritus contribution from a more juvenile source needs to be evaluated. We have compiled REE and trace element data of andesite, dacite, and rhyodacite end members of the Andean mountain belt from the large database of GEOROCK (http://georoc.mpch-mainz.gwdg.de), and these are plotted in REE diagrams for reference. The Type-I REE pattern (Figure 10A) closely matches with dacite and andesite members and their intrusive equivalent of the Andean mountain belt, indicating similar sources for MAG. Old Andean-type continental arc magmas, such as Portar Bay Complex andesites and the Wathaman granitoid batholith of the \( \sim 1.92–1.85 \text{ Ga} \) Trans-Hudson Orogen (Maxeiner and the Rayner 2011), also showed remarkably similar patterns (Figure 10D), indicating that source rocks of MAG were very similar to these in composition and tectonic association. The high HREE fractionation \( (\text{Gd/Yb}_\text{N} > 2) \) in both Types I and III may imply (1) a deep-seated mantle source for precursor igneous melts where garnet and amphibole were the dominant residual phase or (2) a component of slab melt was involved (e.g. Stern and Killian 1996; Rapp et al. 1999). The Type-II pattern is characterized by HREE enrichment compared with Type-I and is consistent with a higher contribution from andesite and/or medium K series arc basalts (Hollings...
and Kerrich 2006) along with dacites and their intrusive equivalents. This is also corroborated from their corresponding incompatible to compatible element ratios, and Type-II samples are characterized with much lower Th/Sc ratios (avg. 0.89) compared with Type-I (avg. 2.05) and Type-III (2.4) samples. The Type-III pattern is characterized by a similar pattern to that of Type-I but with large negative Eu anomaly. These are consistent with a rhyodacite source (Figure 10C) and indicate intracrustal differentiation of the precursor magma in a shallow magma chamber. However, a Type-III REE pattern can also represent an old upper crustal source, since it exhibits patterns parallel to that of basement TTG and granites. But because of their intimate association with Type I and II samples, we interpret that these too were derived from new magmatic sources that underwent intracrustal differentiation. The recent detrital zircon ages of mature quartzose sandstones of the Bowa Formation in the Machla Magra area (McKenzie et al. 2013) showed prominent peaks at 1772, 1646, and 1586 Ma along with a major 2.6 Ga peak (Figure 4), confirming detritus shed off from a young arc source. In comparison to these, the texturally and mineralogically immature greywackes of the Udaipur Formation appear to be derived entirely from a new magmatic arc source. However, more data on the Nd isotope of sediments and Hf isotope data on detrital zircons are needed to firmly establish the fact. Therefore, from the above discussions, it is clear that the detritus of MAG was dominantly derived from a young differentiated arc of intermediate to felsic composition, with possibly a minor component from old UCC.

**Tectonic setting of deposition**

The major and trace element compositions of sediments have been widely used to infer the plate tectonic setting of ancient sedimentary basins (Bhatia and Crook 1986; Roser and Korsch 1986; Mader and Neubauer 2004; Verma and Armstrong-Altrin 2013). But such an approach has been contested by some researchers (e.g. Armstrong-Altrin and Verma 2005; Ryan and Williams 2007), who, on the basis of the large compositional database of sediments of modern basins of known tectonic setting, have demonstrated the ineffectiveness of such tectonic setting discriminant diagrams. Therefore, before drawing such conclusions, associated geological, sedimentological, and stratigraphic parameters need to be taken into consideration (e.g. Manikyamba et al. 2008; Absar et al. 2009).

The Ti vs. Fe + Mg diagram (Figure 7B, after Bhatia 1983) suggests an active continental margin to island arc tectonic setting for MAG. Low-to-moderate CIA values of MAG are also indicative of unsteady weathering under an active tectonic setting. In the La–Th–Sc ternary diagram (Figure 8D, Bhatia and Crook 1986), most of the MAG samples plot within the continental island arc field and in Zr-based ternary discriminant plots (Th–Sc–Zr/10, Th–Co–Zr/10, diagram not shown), the MAG samples plot outside the field because these are depleted in Zr. Furthermore, the Th/Sc–Zr/Sc diagram (Figure 8C) reveals that MAG samples exhibit lower Zr/Sc ratios compared with their Th/Sc ratios, and plot above the normal igneous fractionation trend. This indicates (a) lack of heavy mineral (zircon) sorting and (b) that the HFSE-depleted nature of the igneous precursor of these sediments possibly originated in a subduction-related tectonic setting (cf. Foley 2008). The tectonic scenario is further evaluated by employing a N-MORB normalized multi-element diagram (normalization value after Sun and McDonough 1989) and MREE–HFSE systematics. The trace element data of MAG are plotted in a N-MORB normalized multi-element diagram (Figure 9), where the elements are arranged according to progressively decreasing incompatibility from the left to the right. Figure 9 shows a highly fractionated pattern with extreme enrichment of Cs, Rb, Ba, and Th (>100 X), moderate-to-high enrichment of K, La, and Ce (10–60 X), depletion of HREEs (<0.3–0.9 X), characteristic negative anomalies of Nb-Ta, Ti, and P, and conspicuous positive anomaly of Pb. These are characteristic features of magmas that are generated in subduction zone tectonic settings (Tatsumi and Eggins 1995; Stern 2002; Kelemen et al. 2003; Tatsumi 2005). For reference, we have also plotted the average andesite, dacite, and rhyodacite of the modern Andean continental margin that are obtained from the large database (over 1000 analyses) of GEOROCK (http://georoc.mpch-mainz.gwdg.de). We also compiled the data for ancient magmatic rocks that originated in Andean-type continental margins such as Portar Bay Complex andesites and the Wathamath batholith of the ~1.92–1.85 Ga Trans-Hudson Orogen (Maxeiner and Rayner 2011). It can be seen (Figure 9) that MAG samples show a trace element concentration and pattern exactly similar to those defined by continental margin magmas, indicating an Andean-type continental margin setting of the precursor magma. Distinctive partitioning behaviour of Nb, Ta, Zr, and Sm during partial melting of low magnesium amphibolites or hydrous basalts of eclogite facies in a subduction zone setting has recently been considered for generation of tonalitic magma (Foley et al. 2002, 2003; Rapp et al. 2003; Foley 2008) and has been widely implicated for the growth of Precambrian CC. To glean information in this regard, MAG data are plotted in a Nb/Ta vs. Zr/Sm diagram (Figure 11). Also shown in the diagram are the fields of modern MORB, OIB, and island arc basalts (IAB), CC, and Archaean TTG. The intersection of the two lines in Figure 11 shows the primitive mantle (PM) ratios for Nb/Ta and Zr/Sm as a reference point. MAG sediments are characterized by sub-PM Nb/Ta ratios (7.7–13.2) and sub- to super-PM Zr/Sm ratios (16.4–47.8), which are characteristics of volcanic arc magma and modern adakites (Foley et al. 2002). Therefore, the data clearly
suggest slab-induced melt generation in a subduction zone tectonic setting. It may be noted that these samples have much higher Nb/Ta and lower Zr/Sm compared with Archaean TTG (Figure 11), the feature which also affirms the notion that the TTG of BGC basement was not involved as a source. Low Ce/Pb ratios (1.34–12.03, avg. 4.24) and a conspicuous positive spike of Pb in the MORB normalized multi-element diagram suggest extreme enrichment of Pb in the protolith. The Ce/Pb vs. Ce systematics (Figure 12) attest the compositional signature of global arcs, and the efficient non-magmatic transfer of mantle-derived Pb into the source of convergent margin magmas has been suggested as cause of such Pb enrichment (Miller et al. 1994). In the Th/Yb vs. Ta/Yb diagram (Figure 13, Pearce and Peate 1995), the samples distinctly plot above the MORB compositional array and suggest enrichment of the water-mobile element Th over Ta, which is a tell-tale signature of continental arc. The high Th/Yb ratios (4.8–10.0) indicate the alkaline–shoshonitic nature of the precursor melt (Pearce and Peate 1995) that might have traversed over considerably thickened CC and therefore suggest a mature nature of the continental arc. The Central Volcanic Zone of the Andes is known to have the highest crustal thickness (~70 km) among continental arcs (e.g. Rogers and Hawkesworth 1989). The MORB normalized multi-element diagram (Figure 9) of MAG is virtually similar to that of the Central Volcanic Zone of the Andes with similar degrees of element enrichment or depletion, indicating a similarly thick nature of the crust for the Aravalli continental arc (cf. McMillan et al. 1993). As stated earlier, the relative depletion of HREEs suggests that melting took place at considerable depth, i.e. in the stability field of garnet and amphibole; and possibly the subducted slab, in addition to the metasomatized mantle wedge, contributed to the melt. Furthermore, the presence of relatively young (1772–1586 Ma) zircon grains (Figure 4) in the Middle Aravalli sediments (McKenzie et al. 2013) provides evidence for syntectonic deposition of first-cycle immature sedimentary materials along the active convergent margin (e.g. Cawood et al. 2012).

**Geodynamic implication**

The Aravalli Mountain Belt of northwestern India represents one of the major Proterozoic accretionary orogens of the world, and its geodynamic evolution has been considered

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**Figure 11.** Nb/Ta vs. Zr/Sm diagram (after Foley et al. 2002) of MAG suggesting a subduction zone tectonic setting, analogous to present day IAB, and adakites. Note compositions of MAG are significantly different from Archaean TTG and the solitary Bundelkhand granite sample (data source: Chakrabarti et al. 2007), which is represented by an open square.

**Figure 12.** Ce/Pb vs. Ce diagram (after Miller et al. 1994) of MAG indicating Pb enrichment and overlapping with the field of global arcs. Fields for MORB, OIB, PM, and CC are also shown.

**Figure 13.** Th/Yb vs. Ta/Yb diagram (after Pearce and Peate 1995; Lamaskin et al. 2008) of MAG showing enrichment of a water-mobile element, Th over Ta, indicating a subduction zone tectonic setting.
critical for better understanding of Proterozoic crustal growth and assembly of the Supercontinent Columbia (e.g. Zhao et al. 2004; Kaur et al. 2009, 2011, 2013; Rogers and Santosh 2009; Bhowmik et al. 2010). Based on lithostratigraphic records, structural analyses, and regional geophysical data, various tectonic evolutionary models for the Aravalli Mountain Belt have been proposed. These include (a) two separate Wilson cycles represented by the Palaeoproterozoic Aravalli Supergroup and Palaeo-Mesoproterozoic Delhi Supergroup (Sinha-Roy 1988, 2004; Deb and Sarkar 1990; Vijaya Rao et al. 2000), (b) one Wilson cycle (Sugden et al. 1990), (c) ensialic orogenesis (Roy et al. 1988; Roy 1990; Sharma 1995), and (d) inversion tectonics (Verma and Greiling 1995). Although two Wilson cycles involving opening and closing of the Palaeoproterozoic Aravalli ocean and Mesoproterozoic Delhi ocean were recognized, these models are still tentative, and the finer details of the evolution of the orogen are still poorly understood. On the basis of geochemical study, Banerjee and Bhattacharya (1994) suggested that the Aravalli Supergroup represents sedimentation on a passive to active continental margin. They envisaged closure of the Aravalli ocean due to westward subduction of Aravalli oceanic crust and collision of the Aravalli continental margin with the arc-microcontinent along the Rakhabdev lineament. However, these models are fraught with serious shortcomings such as (1) these do not explain the formation of linear Aravalli sub-basins at the Bhilwara terrain during the tectonic cycle, (2) tectonic episodes are wrongly construed because of improper stratigraphic correlation of various litho-units of the Aravalli Supergroup, wherein the Upper Aravalli Debari Formation is being considered as the basal horizon (e.g. Roy et al. 1993; Roy and Jakhar 2002), and (3) these do not provide a unified tectonic model for basin evolution and formation of mineral deposits. Therefore, in the light of the present geochemical dataset, more refined stratigraphic record of the Aravalli Supergroup (Roy and Jakhar 2002), and recent detrital zircon U–Pb ages (McKenzie et al. 2013) of Aravalli clastic rocks, we have attempted to reconstruct the tectonic evolution of the basin, especially during the early–middle stage of its evolution.

As discussed in preceding sections, the petrological and geochemical characteristics of texturally immature MAGs are consistent with their derivation from a continental volcanic arc terrain. On the basis of (1) the highly evolved, LILE- and REE-enriched alkaline nature of the protoliths (Figure 9), which would require traversing of melt through considerably thick CC, (2) the presence of lithic clasts, such as quartzite and schists, which are typical components of BGC basement (cf. Banerjee and Bhattacharya 1994), and (3) the absence of unequivocal evidence of occurrences of mature CC in the west of Rakhabdev suture zone, we infer that the sediments were derived from a volcanic arc that was situated to the east of depositional site. The plate tectonic evolutionary model during the early–middle Aravalli stage has been depicted in series of cartoons in Figure 14. We postulate that the Aravalli basin was initiated during continental rifting of the BGC protocontinent at ~1830 Ma (stage-A, Figure 14). The early rift stage is represented by interbedded mafic volcanics and feldspathic quartzites of the basal Delwara Formation. The rift basin gradually evolved into a passive margin during ca. 1772 Ma (stage-B, Figure 14) and extensive deposition of Jhamarkotra carbonates and phosphorites took place in a shallow water carbonate platform. Proliferation of phosphate-binding algae (stromatolites), conspicuous positive δ13 C anomaly (Sreenivas et al. 2001b) in Jhamarkotra carbonates, and presence of thick uranium-bearing carbonate shales in the deeper water of Umra basin (Roy and Jakhar 2002) indicate deep burial of organic matter and subsequent release of oxygen to the atmosphere (Sreenivas et al. 2001b). During this stage, the Aravalli basin possibly evolved into a full-fledged oceanic basin in the western margin of the present BGC continent with development of oceanic crust. Overlying turbidites of the Udaipur Formation and basal Zawar Formation indicate deepening of the basin due to active tectonics in the hinterland (cf. Eriksson et al. 2001). In order to explain the geochemical features of MAG, the stratigraphic trend and the presence of young detrital zircon grains in Middle Aravalli clastics, we propose eastward subduction of Delwara oceanic crust beneath the BGC continent in an Andean-type plate margin and deposition of volcanic arc debris in the forearc basin that was situated on the west of volcanic arc (Stage-C, Figure 14). On the basis of the U–Pb detrital zircon age (Figure 4, McKenzie et al. 2013) of clastic rocks of the Middle Aravalli Bowa Formation of the Machlamagra area, a time frame of 1646–1586 Ma can be assigned for the active ‘Middle Aravalli arc’. This continental arc model is consistent with the field setting, which suggests rapid drowning of the Jhamarkotra carbonate platform and deposition of Middle Aravalli turbidites in a compressive regime. This model also explains the opening of linear ensialic basins in the Bhilwara terrain, such as in Rajpura–Dariba (RD) and Rampura–Agucha (RA) in a classical back-arc extension regime similar to that of the Andean continental margin of the Mesozoic (Maloney et al. 2013; Rossel et al. 2013). Alternatively, these linear basins may have opened as strike-slip basins in the back-arc zone in a minor compressive to extensive regime, similar to that of the present day Tibetan plateau behind the Himalayan mountain chain (Yin and Harrison 2000). The absence of Lower Aravalli Group sediments such as Delwara volcanics and Jhamarkotra carbonates in these linear basins of the Bhilwara region also supports the notion of post-‘Lower Aravalli’ opening for these. Based on the stratigraphic trend, we further propose that this subduction-compressive regime terminated after the deposition of Udaipur Formation turbidites at ca 1500 Ma, possibly because of collision with an oceanic
Figure 14. Schematic cartoons demonstrating the tectonic setting and evolution of the Aravalli basin. Stage-A (~1830 Ma): formation of intracontinental rift basin and deposition of rift volcanics and quartz arenites of Delwara formation in syn-rift stage. Stage-B (1772 Ma): opening of Delwara ocean basin, formation of passive continental margin, and deposition of Jhamarkotra carbonates and phosphorites in platform areas and black carbonaceous shales in a deep shelf. Stage-C (1646–1586 Ma): eastward subduction of Delwara oceanic crust (spontaneous conversion of passive to active margin, cf. Stern 2004), formation of continental volcanic arc and deposition of a thick turbidite sequence in the forearc basin (Udaipur basin) and simultaneous opening of linear ensialic back-arc extensional basins at the Bhilwara terrain. Stage-D (~1500 Ma): termination of the subduction regime possibly because of collision with an oceanic arc/oceanic plateau and re-establishment of a passive margin. During this stage, deposition of carbonates in the shallow eastern shelf and carbonate-free sediments in the deep marine western Jharol oceanic basin took place, and these constitute the sedimentary piles of the Middle Aravalli and Upper Aravalli groups.
Conclusions

(1) Petrological and geochemical attributes of MAG, such as very poor sorting, high angularity of framework grains, presence of fresh plagioclase and K-feldspars, variable CIA index (46.7–74.5, avg. 61), high ICV value (~1.05), and relatively minor depletion of Ca, Na, Sr over K, Rb, Cs in the UCC normalized multi-element diagram, suggest a non-steady state of weathering and rapid physical erosion under an active tectonic regime.

(2) Very poor sorting of framework grains and lack of anti-correlation between Si with other major elements rule out grain-size-related partitioning of tectosilicate and phyllosilicate minerals in the bedload and suspended load, and therefore suggest deposition under a turbiditic regime. Low Zr/Sc ratios and coherent behaviour of MREEs and HFSEs suggest a first-order nature of the sediments and rule out heavy mineral sorting.

(3) The sediments witnessed post-depositional K-metasomatism with K enrichment from 0 to 25% (avg. 10%).

(4) Various diagnostic incompatible to compatible element ratios and REE patterns suggest variable contribution from new magmatic materials of andesite–dacite–rhyodacite composition that are significantly different from BGC basement components. The lack of significant negative Eu anomaly of MAG samples and young detrital zircon ages of Aravalli clastic rocks suggest that they were derived from a young differentiated continental margin type arc. The high HREE fractionation in both Types I and III supports a deep-seated mantle source for precursor igneous melts with garnet and amphibole as the dominant residual phase.

(5) The highly fractionated MORB normalized trace element pattern of MAG, with characteristic enrichment of LILEs, depletion of HREEs, negative Nb-Ta, Ti, and P anomalies, positive Pb anomaly, and distinctive trace element composition in Nb/Ta–Zr/Sm, Th/Yb-Ta/Yb, Ce/Pb–Ce systematics suggest a subduction zone tectonic setting for precursor magma. These patterns mimic the pattern of continental arc magmas such as modern Andean andesite the dacite and the Paleoproterozoic Portar Bay Complex and Wathamian batholith of the Trans-Hudson orogen, indicating a continental arc tectonic setting. Relatively high enrichment of LILEs in these also suggests considerably thick CC (~70 km) below the lower Aravalli continental arc, similar to that of the Central Volcanic Zone of modern Andes.

(6) We postulate eastward subduction of Delwara oceanic crust beneath the BGC continent and formation of a continental volcanic arc that supplied detritus to the forearc basin situated on the west of volcanic arc. This model also explains the opening of linear ensialic basins in the Bhilwara terrain, such as in Rajipura–Dariba and Rampura–Agucha in a classical back-arc extension regime, similar to that of the Andean continental margin of the Mesozoic. On the basis of the recent $^{207}$Pb/$^{206}$Pb detrital zircon age of Middle Aravalli sediment (McKenzie et al. 2013), a time frame of 1646–1586 Ma can be assigned for the ‘Middle Aravalli arc’.

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