Contents lists available at ScienceDirect

Chemie der Erde

journal homepage: www.elsevier.com/locate/chemer

Provenance, tectonic setting, source weathering and palaeoenvironmental implications of Middle-Upper Jurassic rocks of Ler dome, Kachchh, western India: Inferences from petrography and geochemistry



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GEOCHEMISTRY

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ARTICLE INFO

Handling Editor: Astrid Holzheid Keywords: Petrography Geochemistry Provenance Source-area weathering Palaeoenvironment Kachchh basin

ABSTRACT

The Middle-Upper Jurassic sandstones and shales of Ler dome (Chari and Katrol Formations), Kachchh, western India, have been analyzed for modal, bulk mineralogy and geochemistry to deduce their provenance, tectonic setting, source area weathering and palaeoenvironmental conditions. The detrital modes of Ler dome sandstones indicate that they were emanated from recycled orogen (uplifted shoulders of rift) and stable cratonic source in passive margin setting. Rapid deposition of sediments from a granitic source area can be predicted from feldspar abundance. A highly mature heavy mineral assemblage characterized in the form of high Zircon-Tourmaline-Rutile (ZTR) index also endorses these findings. The X-ray diffraction patterns (XRD) and scanning electron microscope (SEM) data show the presence of clay minerals depicting moderate to extensive chemical weathering in an oxidizing environment with periodic cycles of transgression and regression. The chemical index of moderate weathering conditions for sandstone and shales, respectively, that took place in low to moderate relief. We postulate that Ler dome sediments are derivative of the eroded and weathered parts of the Aravalli craton located on east and northeast of the basin and the Nagarparkar Massif placed to the north and northwest.

1. Introduction

The mineralogical and geochemical evidence of clastic sediments constitutes an archive to understand their provenance, tectonics and weathering state in their source region (e.g., Bhatia, 1983; Zou et al., 2016; Ramasamy et al., 2017a). The framework grains or modal signatures of sandstone commonly indicate the lithological characters of the source rocks and depositional settings (Dickinson, 1985; Garzanti et al., 2009; Fatima and Khan, 2012). Traditional petrographic analysis can reveal the provenance and tectonic setting (Quasim et al., 2017). The geochemical composition of clastic sedimentary rocks is the counterpart of the petrographic inferences that provides essential geochemical proxies for studying the provenance, tectonic setting as well as weathering conditions in their source region (McLennan et al., 1993). Noteworthy contributions have been made by several recent studies which mostly focused to understand the composition of the terrigenous clastic rocks (Quasim et al., 2017), to evaluate weathering processes (Absar et al., 2009) and to develop palaeogeographic reconstruction of source regions (Shi et al., 2016). As long as the bulk chemical composition is not totally modified the geochemical composition and/or

geochemical indices provide detailed information contained in clastic sedimentary rocks. For example, major element discrimination diagrams have been used to systematically categorize the tectonic settings of the sedimentary basins (Kundu et al., 2016; Pandey and Parcha, 2017). Occasionally, it may not be relevant to interpret the contributions from different sources due to weathering and sorting effect which in turn lead to erroneous quantitative estimations (Garzanti et al., 2009).

Further, trace elements and their ratios (e.g., Ni/Co, Th/Sc, Cr/Th, Sc/Nb, Rb/V) considered to be independent of weathering and hydraulic sorting effects are extensively used for discrimination of provenance and mixing between different sources as well as tectonic environments (Sawant et al., 2017 and references therein). Trace elements such as Sc, Y, Zr, Nb, Hf, Th and the rare earth elements (REE) are considered to be immobile during weathering, diagenesis and low to moderate levels of metamorphism, and the geochemical signatures are commonly preserved in the sedimentary rocks (Bhatia and Crook, 1986; McLennan et al., 1993). Trace elements also become more significant in evaluating tectonic environments (Bhatia and Crook, 1986; Condie,

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https://doi.org/10.1016/j.chemer.2018.06.002



Received 23 January 2018; Received in revised form 6 June 2018; Accepted 9 June 2018 0009-2819/ © 2018 Elsevier GmbH. All rights reserved.



Fig. 1. (a) Geological map of Kachchh basin (after Fürsich et al., 2001), Ler dome, western India (after Ghaznavi et al., 2015). (b) Lithostratigraphic section measured at the river.

1993). Therefore, during weathering and transportation trace elements are quantitatively transferred into clastic sediments which reflect the signatures of the parent material and palaeoenvironment conditions.

Furthermore, the clay mineralogy has been efficaciously used to interpret the palaeoclimate of the Jurassic and Cretaceous rocks (e.g., Khormali and Amini, 2015). Clay minerals are primarily a product of weathering except few that are produced during diagenesis as well as hydrothermal alterations (Velde, 1992). Since weathering is involved in the formation of these phyllosilicates, it is inevitable that these clay minerals depend on the rock type, climate, slope, and drainage characteristics of the area (Fürsich et al., 2005). Thus clay mineral assemblages seem to reflect the original composition derived from the source area and may be used for reconstruction of palaeoenvironment and palaeoclimatic condition of the source area (Fürsich et al., 2005).

The present work emphasise to provide a detailed account of source rock and the effect of tectonics, climate and weathering conditions of the sediments in Ler dome of Kachchh basin. The Kachchh basin has been studied for their biostratigraphy (Rai et al., 2015; Alberti et al., 2017), lithostratigraphy (Biswas, 1980; Fürsich and Pandey, 2003), microfossils (Gaur and Talib, 2009), basin analysis (Ahmad et al., 2014), petrofacies (Ahmad and Bhat, 2006; Ghaznavi et al., 2015), palaeoclimate (Fürsich et al., 2005; Alberti et al., 2017) and palaeogeography (Talib and Gaur, 2008). However, with the exception by Ghaznavi et al. (2015), very few attempts have been made to study the source rocks and the influence of climate and weathering with the aid of geochemical parameters (major and trace elements) and clay mineralogy of sandstones and shales in particular. For this approach, petrography as well as bulk mineralogy and geochemistry of both sandstones and shales have been used with the prime aim (a) to infer provenance, tectonic setting, and source weathering of Middle-Upper Jurassic rocks of Ler dome, Kachchh, western India, and (b) to interpret the palaeoenvironmental conditions that prevailed in the source regions by quantitative evaluation of clay minerals and geochemical data. This study will help to improve and expand models for the evolution of the Ler dome sediments of the Kachchh basin.

2. Study area - geological context

The Kachchh basin lies on the western most periphery of the Indian peninsula as a peri-continental, palaeo-rift basin (Biswas, 1987) between latitude 22°30' and 24°30'N and longitude 68° and 72°E (Fig. 1a). The basin is bordered by the Nagar Parkar fault in the north, Radhanpur-Barmer arch in the east and Kathiawar fault in the south (Biswas, 1982). Waagen (1875) by describing the rich ammonite faunas collected by Ferdinand Stoliczka, subdivided the Jurassic rocks into Patcham, Chari, Katrol and Umia groups in ascending order. The classification of Waagen (1875) was further modified by Biswas (1980) who introduced the Jhurio Formation below the Patcham Formation. Sequential rifting and repeated movements along Precambrian tectonic trends that took place in relation to the Indian Plate lead to the evolution of the basin (Ahmad et al., 2013). The Indian plate drifted northward after breakup of Gondwanaland in the Late Triassic and Early Jurassic (Biswas, 1987; Ahmad et al., 2013). Rifting along the Delhi trend was initiated in the Late Triassic as manifested by the continental Rhaetic sediments in the northern part of the basin (Koshal, 1984). The Kachchh basin was in fact formed by the subsidence of a block between Nagar Parkar Hills and the southwest extension of Aravalli Range during Jurassic times. In Bajocian, the basin was

Table 1

Percentages of detrital minerals of Dhosa Sandstones at Ler Dome, Kachchh, western India.

Sample No.	Monocrystalline	Pollycrystalline Qu	Feldspar	Mica		Chert	Rock	Heavies			
	Common Quartz	Recrystallized Metamorphic Quartz	Stretched Metamorphic Quartz	Microcline	e Plagioclase Orthoclase		Muscovite Biotite			iraginents	
D1(A)	87.78	4.49	1.31	4.77	1.17	0.00	0.00	0.00	0.14	0.35	0.00
D1(B)	85.35	7.82	0.57	3.41	1.00	0.00	0.00	0.43	0.43	1.00	0.00
D2(A)	91.74	0.91	2.67	3.04	0.85	0.00	0.12	0.00	0.00	0.24	0.43
D2(B)	86.19	2.99	4.58	2.66	1.00	0.00	1.59	0.00	0.13	0.33	0.53
D3(A)	90.36	1.50	1.71	2.79	1.43	0.00	1.07	0.00	0.29	0.29	0.57
D3(B)	87.21	2.06	1.91	5.74	0.59	0.00	0.44	0.00	0.88	0.00	1.18
D4(A)	89.56	0.88	0.75	3.90	2.64	0.00	0.00	0.00	0.25	1.01	1.01
D4(B)	84.86	2.46	0.91	6.99	1.03	0.00	1.03	0.00	0.52	1.29	0.91
D5(B)	82.94	2.38	0.00	5.83	3.89	0.00	0.43	0.00	0.65	2.59	1.30
D6	84.35	4.35	1.61	4.47	2.48	0.00	1.74	0.00	0.62	0.00	0.37
D7	91.90	0.98	0.00	3.56	2.45	0.00	0.86	0.00	0.00	0.25	0.00
D8	87.07	1.39	2.54	5.08	1.39	0.00	0.00	0.00	1.15	0.69	0.69
D9	81.63	5.56	1.94	4.01	3.49	0.00	0.78	0.26	0.52	0.91	0.91
D10	85.37	0.95	0.79	5.56	2.38	0.79	2.07	0.00	0.48	0.95	0.64
D11	76.62	7.19	1.80	6.29	2.16	0.00	2.52	0.00	0.36	1.80	1.26
D12	81.12	5.79	1.50	4.30	2.24	0.00	2.99	0.00	0.93	0.00	1.12
D13	84.35	3.00	1.43	4.30	2.09	0.00	2.87	0.13	0.52	0.65	0.65
D14	84.29	5.29	1.03	2.94	2.20	0.00	1.62	0.00	0.88	0.59	1.17
D15	80.93	2.62	2.99	4.86	2.06	0.75	2.24	0.00	0.93	0.56	2.06
L1	81.07	2.91	0.00	5.58	2.91	0.49	3.40	0.00	0.00	1.46	2.18
L2	83.23	3.16	0.00	4.75	3.48	0.95	1.27	0.00	0.63	1.58	0.95
L8	83.68	3.14	1.26	4.18	2.72	1.26	0.63	0.00	0.84	0.63	1.67
L10	88.37	1.94	1.16	3.68	2.33	0.00	0.39	0.19	0.97	0.39	0.58
L12	80.79	4.37	1.53	5.02	3.93	0.00	0.87	0.00	1.53	1.31	0.66
SII-3	84.06	2.36	0.72	4.53	2.36	1.09	3.08	0.00	0.91	0.54	0.36
SII-6	85.23	1.61	0.00	5.30	3.21	0.00	1.12	0.00	1.28	0.32	1.93
SII-7	87.24	1.87	0.68	3.91	1.70	0.00	2.38	0.00	1.19	1.02	0.00
SII-8	85.23	2.06	0.00	5.42	2.62	0.93	2.06	0.00	1.12	0.56	0.00
SII-9	83.54	1.86	0.00	5.28	2.80	0.00	4.97	0.00	0.93	0.62	0.00
SII-12	84.74	2.89	0.00	5.53	2.37	0.00	1.05	0.00	1.58	1.84	0.00
SII-15	88.56	2.13	0.00	4.52	1.60	0.00	0.53	0.00	1.86	0.80	0.00
LII-2	85.23	1.99	0.00	5.40	3.13	0.00	0.85	0.00	2.27	1.14	0.00
LII-5	87.30	4.99	0.00	3.14	1.85	0.00	0.71	0.00	0.86	0.86	0.29
Average	85.21	3.03	1.07	4.57	2.23	0.19	1.38	0.03	0.78	0.80	0.71
Std. Dev.	3.33	1.79	1.08	1.07	0.88	0.39	1.17	0.09	0.54	0.59	0.64

inundated by the sea that is evidenced by the first occurrence of marine sediments in that time. This marine sedimentation took place after a phase of terrestrial sedimentation in the Late Triassic and Early Jurassic (Fürsich et al., 2001). The basin was filled up and the sea began to recede in early Cretaceous time (Bardhan and Datta, 1987). In the late Cretaceous period, extensive regional uplift took place in the western part of India (Biswas, 1987).

2.1. Stratigraphy of Kachchh basin

Mesozoic rocks in Kachchh basin are exposed extensively in three areas- Kachchh Mainland, Wagad Uplift and Island belt (Pachham, Khadir, Bela and Chorar) – resting unconformably on the Precambrian crystalline basement (Bardhan and Datta, 1987). The thickness of the Jurassic strata varies from 700 to > 1000 m depending upon the locality (Fürsich et al., 2013). The three regions of the Kachchh basin have their own depositional history and differ in their sedimentary successions (Table S1). Most of the well-known sections are exposed in the Kachchh Mainland which occupies the largest part of the basin. It comprises of quaquaversal units referred to as domes. From north of Bhuj, these domes extend from Jara situated in the west to Habo dome lying in the east with intervening Jumara, Nara (Kaiya), Keera and Jhurio domes. Our study area lies at the Ler dome (Fig. 1a), located south of Bhuj, comprises well-exposed outcrops of Jurassic rocks.

In the Kachchh Mainland the Callovian to Oxfordian strata belong to Chari Formation. The Gypsiferous Shale Member (GSM) (Upper Callovian) is composed of bioturbated argillaceous silt with several levels of small concretions and abundant secondary gypsum. The absence of current-induced sedimentary structures and fine-grained sediments shows that they were deposited below storm wave base (Alberti et al., 2013a). It measures around 0.7 m in the study area (Fig. 1b). GSM is followed by the Dhosa Sandstone Member (DSM) (Upper Callovian-Lower Oxfordian). It has a thickness of 24.1 m in the study area. DSM is mainly composed of argillaceous coarse silt to finegrained sandstones (Ramkumar et al., 2013). The gradational boundary between the DSM and the GSM is diachronous (Alberti et al., 2013a). The DSM is overlain by Dhosa Oolite Member (DOM). The varying abundance of bioclasts and ferruginous, ooid-bearing rocks (Kulkarni and Borkar, 2000) as well as ferruginous concretions are embedded at different levels in DOM. DOM has a highly complex top unit, i.e., the Dhosa Conglomerate Bed (DCB), which is a characteristic marker horizon in most areas of the Kachchh basin (Singh, 1989). It is highly fossiliferous with reworked ammonites of early to middle Oxfordian age (Pandey et al., 2012). Upper Oxfordian and lower Kimmeridgian rocks are missing (hiatus). The youngest Jurassic strata in the study area belong to the siliciclastic Katrol Formation (Alberti et al., 2017).

3. Methodology

A total of fifty samples were collected from the Dhosa Sandstone Member and the Gypsiferous Shale Member at intervals of 0.5 m. Thirty-six samples were chosen for different analysis (Table S2). Out of these thirty three samples were studied petrographically. In each thin section 250–300 grains were counted according to the Gazzi–Dickinson method (Ingersoll et al., 1984). The nomenclature advocated by Dickinson et al. (1983) and Dickinson (1985) was used to determine the

Table 2

Modal composition based on Folk (1980) and detrital petrofacies modes based on Dickinson (1985) of the Dhosa Sandstones at Ler dome, Kachchh, western India. According to Folk (1980); Q: total quartz, F: total feldspar, R: total rock fragments. According to Dickinson (1985); Qt: total quartzose grain, F: total feldspar, L: lithic, Qm: monocrystalline quartz, Lt: total lithics, Qp: polycrystalline quartz, Lv: volcanic lithics, Ls: meta-sedimentary lithics, P: plagioclase feldspar, K: potassium feldspar.

Sample No.	Q%	F%	R%	Qt %	F%	L%	Qm%	F%	Lt%	Qp%	Lv%	Ls%	Qm%	P%	K%
D1(A)	93.00	5.90	1.10	93.08	5.84	1.09	87.23	5.90	6.86	86.21	0.00	13.79	93.66	1.25	5.08
D1(B)	95.51	4.49	0.00	95.51	4.49	0.00	86.96	4.49	8.55	100.00	0.00	0.00	95.09	1.11	3.80
D2(A)	96.08	3.92	0.00	96.08	3.92	0.00	92.47	3.92	3.61	100.00	0.00	0.00	95.93	0.89	3.18
D2(B)	96.25	3.75	0.00	96.25	3.75	0.00	88.48	3.75	7.77	100.00	0.00	0.00	95.93	1.11	2.96
D3(A)	93.57	4.21	2.21	93.71	4.12	2.17	90.36	4.21	5.43	71.03	0.00	28.97	95.54	1.51	2.95
D3(B)	91.45	6.34	2.21	91.63	6.20	2.16	87.46	6.34	6.19	73.68	0.00	26.32	93.24	0.63	6.13
D4(A)	92.71	6.65	0.64	92.76	6.61	0.64	91.05	6.65	2.30	78.26	0.00	21.74	93.19	2.75	4.06
D4(B)	90.21	8.20	1.59	90.36	8.07	1.56	86.77	8.20	5.03	76.00	0.00	24.00	91.36	1.11	7.52
D5(B)	87.97	10.02	2.00	88.21	9.83	1.97	85.52	10.02	4.45	68.97	0.00	31.03	89.51	4.20	6.29
D6	90.99	7.01	2.00	91.17	6.87	1.96	84.98	7.01	8.01	80.00	0.00	20.00	92.38	2.72	4.90
D7	93.92	6.08	0.00	93.92	6.08	0.00	92.93	6.08	0.99	100.00	0.00	0.00	93.86	2.51	3.63
D8	90.57	6.44	2.99	90.85	6.25	2.90	86.67	6.44	6.90	69.77	0.00	30.23	93.09	1.48	5.43
D9	90.07	7.58	2.35	90.29	7.41	2.30	82.48	7.58	9.93	80.85	0.00	19.15	91.58	3.92	4.50
D10	88.67	8.90	2.43	88.94	8.69	2.37	86.89	8.90	4.21	63.41	0.00	36.59	90.71	2.53	6.76
D11	89.98	8.88	1.13	90.09	8.79	1.12	80.53	8.88	10.59	90.32	0.00	9.68	90.06	2.54	7.40
D12	91.31	6.76	1.93	91.48	6.63	1.89	83.78	6.76	9.46	83.05	0.00	16.95	92.54	2.56	4.90
D13	88.44	6.36	5.19	89.01	6.05	4.94	84.03	6.36	9.61	64.91	0.00	35.09	92.96	2.30	4.74
D14	92.37	5.24	2.40	92.54	5.12	2.34	85.93	5.24	8.83	78.67	0.00	21.33	94.25	2.46	3.28
D15	87.19	7.72	5.08	87.81	7.35	4.84	81.54	7.72	10.73	67.86	0.00	32.14	91.35	2.32	6.33
L1	90.34	9.66	0.00	90.34	9.66	0.00	87.21	9.66	3.13	100.00	0.00	0.00	90.03	3.23	6.74
L2	86.94	9.24	3.82	87.42	8.90	3.68	83.76	9.24	7.01	64.71	0.00	35.29	90.07	3.77	6.16
L8	86.27	7.99	5.74	87.02	7.56	5.43	81.97	7.99	10.04	63.64	0.00	36.36	91.12	2.96	5.92
L10	89.06	5.85	5.09	89.59	5.57	4.85	86.04	5.85	8.11	61.43	0.00	38.57	93.63	2.46	3.90
L12	85.19	8.80	6.01	86.03	8.30	5.67	79.40	8.80	11.80	66.27	0.00	33.73	90.02	4.38	5.60
SII-3	89.07	8.15	2.78	89.37	7.93	2.70	85.93	8.15	5.93	68.09	0.00	31.91	91.34	2.56	6.10
SII-6	88.25	8.65	3.10	88.61	8.39	3.01	86.62	8.65	4.73	60.42	0.00	39.58	90.92	3.42	5.65
SII-7	89.49	5.59	4.92	89.98	5.33	4.68	86.95	5.59	7.46	60.27	0.00	39.73	93.96	1.83	4.21
SII-8	87.95	9.04	3.01	88.30	8.78	2.93	85.88	9.04	5.08	62.79	0.00	37.21	90.48	2.78	6.75
SII-9	89.58	8.47	1.95	89.78	8.31	1.92	87.62	8.47	3.91	66.67	0.00	33.33	91.19	3.05	5.76
SII-12	86.95	7.83	5.22	87.59	7.44	4.96	84.07	7.83	8.09	60.78	0.00	39.22	91.48	2.56	5.97
SII-15	89.97	6.07	3.96	90.36	5.84	3.81	87.86	6.07	6.07	60.53	0.00	39.47	93.54	1.69	4.78
LII-2	88.99	8.70	2.32	89.24	8.50	2.27	86.96	8.70	4.35	65.22	0.00	34.78	90.91	3.33	5.76
LII-5	88.39	4.78	6.83	89.13	4.48	6.39	83.61	4.78	11.61	62.96	0.00	37.04	94.59	2.01	3.40
Average	90.20	7.07	2.73	90.50	6.88	2.62	86.06	7.07	6.87	74.45	0.00	25.55	92.41	2.42	5.17
Std. Dev.	2.77	1.75	1.93	2.58	1.71	1.82	3.08	1.75	2.77	13.50	0.00	13.50	1.86	0.96	1.31

framework constituents. For the heavy mineral analysis, samples were first crushed then rinsed with water to remove any clay fraction. After that samples were sieved on mesh size $125-250 \,\mu$ m. Bromoform (specific gravity 2.89) was used for the separation of heavy minerals by using gravity method. The minerals that settled at the bottom of the separating funnel were washed with acetone and stannous chloride and then mounted on a glass slide with Canada balsam. The minerals were identified under the microscope by their diagnostic optical character. Field count method following the procedure outlined by Lindholm (1987) was used to determine their numbers for each horizon. Zircon-Tourmaline-Rutile (ZTR) values were calculated following Hubert (1962).

Scanning electron microscope (SEM) and X-ray diffraction (XRD) analysis were used to ascertain bulk mineralogy and abundance of 10 powdered samples. The XRD analyses were done using an X'Pert³ Powder X-ray diffractometer at Advanced Centre for Material Sciences (ACMS), Indian Institute of Technology, Kanpur (IITK). The measuring conditions such as Cu- K α ($\lambda = 1.54$ Å) source, 40 kV potential difference and 35 mA current at a scan speed of 1°/min, step size of 0.05°, and 2 θ values in the range 3° to 70° were opted and maintained for all XRD analysis.

Eleven representative samples (8 sandstones and 3 shales) have been selected for their geochemical analysis. Major oxides and selected trace element concentrations were determined on pressed powder pellets analyzed using a Bruker S8 Tiger wave-length dispersive X-ray fluorescence spectrometer (WD-XRF) at Wadia Institute of Himalayan Geology (WIHG), Dehradun. Calibration curves were obtained from regressions of more than 10 international rock standards from United States Geological Survey, Centre de Recherches Pétrographique et Géochimiques, Nancy, and Canadian Certified Reference Materials Project (CCRMP). Elements determined as weight % oxide in geological matrices are SiO₂, TiO₂, Al₂O₃, Fe₂O₃, MnO, MgO, CaO, Na₂O, K₂O, and P₂O₅. Before using them in various diagrams, they were recalculated to an anhydrous (LOI free) basis and adjusted to 100%. Trace elements analyzed in our samples include Ba, Sc, V, Cr, Co, Ni, Cu, Zn, Ga, Pb, Th, Rb, U, Sr, Y, Zr and Nb.

4. Results

4.1. Petrography

The Dhosa Sandstone is medium- to coarse-grained, moderately to well-sorted. Range and average of petrographic composition and detrital petrofacies (based on Dickinson, 1985) of the Dhosa Sandstone Member are listed in Tables 1 and 2, respectively (Ghaznavi et al., 2015). Quartz, feldspar and lithic fragment are the main framework components. The cement present is in form of carbonate (calcite), iron, infiltered clays, authigenic clays and silica. The individual mineralogical composition can be described as follows-

4.1.1. Quartz

Quartz occurs in all the samples with monocrystalline quartz being the dominant variety ranging from 76.62% to 91.90% (avg. 85.21). Few of the quartz grains have inclusions of minute mica, kaolinite and rutile needles (Fig. 2a). Straight to slightly undulose extinction is visible in these grains. The polycrystalline variety of quartz has recrystallized



Fig. 2. Representative microphotograph showing (a) presence of granulite facies rocks at the source area represented by rounded quartz with rutile inclusion; (b) Monocrystalline (common quartz) and polycrystalline (recrystallized and stretched quartz); (c) the most abundant feldspar variety, microcline, and calcite cement; (d) microcline and zircon grain, alteration of feldspar to sericitic mica; (e) bent muscovite grain between the rigid detrital grain of quartz; (f) sedimentary lithic fragment (chert); (g) detrital grain of tourmaline and monocrystalline quartz; (h) & (i) showing heavy minerals.

Table 3		
Percentage of heavy mineral	content in Dhosa sandstones	, Kachchh, western India.

Sample Id.	Zircon	Tourmaline	Rutile	Epidote	Biotite	Muscovite	Chlorite	Sillimanite	Hornblende	Staurolite	ZTR	Opaque	Transparent
D1(A)	35.0	32.7	18.0	5.6	3.3	5.6	0.0	0.0	0.0	0.0	85.6	24.8	75.2
D2(A)	31.8	22.1	14.0	18.5	0.0	7.5	3.2	2.9	0.0	0.0	67.9	22.0	78.0
D3(A)	43.2	28.9	13.6	0.0	9.2	5.1	0.0	0.0	0.0	0.0	85.7	21.4	78.6
D4(A)	39.6	20.4	30.4	0.0	0.0	9.6	0.0	0.0	0.0	0.0	90.4	33.3	66.7
D5(B)	36.8	23.9	12.3	11.3	5.8	0.6	4.3	2.5	0.6	1.8	73.0	26.8	73.2
D6	34.7	24.7	19.1	12.2	0.0	3.5	0.0	2.8	3.1	0.0	78.5	34.0	66.0
D7	39.2	22.7	16.9	7.7	3.8	2.3	0.4	2.3	4.6	0.0	78.8	23.2	76.8
D8	21.1	41.8	10.2	2.3	2.6	2.3	11.2	2.6	5.9	0.0	73.0	20.2	79.8
D9	35.2	22.5	15.4	12.3	1.6	5.9	2.8	2.0	1.2	1.2	73.1	28.6	71.4
D10	42.3	22.2	12.5	10.4	9.0	0.0	0.0	0.0	3.6	0.0	77.1	24.2	75.8
D11	34.7	15.3	5.9	15.3	6.6	3.5	9.7	3.5	2.8	2.8	55.9	26.3	73.7
D12	38.5	22.7	15.4	8.4	4.8	2.6	0.0	2.9	4.8	0.0	76.6	17.5	82.5
D13	25.0	33.3	16.7	8.3	0.0	0.0	0.0	8.3	8.3	0.0	75.0	30.8	69.2
D14	31.0	24.1	10.3	10.3	3.4	3.4	6.9	3.4	3.4	3.4	65.5	26.1	73.9
D15	42.9	25.0	13.6	11.4	7.1	0.0	0.0	0.0	0.0	0.0	81.5	27.1	72.9
L1	33.6	23.6	19.5	12.3	0.7	2.7	1.0	4.1	2.4	0.0	76.7	23.8	76.2
L2	38.6	24.8	12.4	11.1	4.7	1.0	4.0	2.0	1.0	0.3	75.8	22.7	77.3
L8	33.2	22.5	16.3	13.8	1.0	6.9	3.1	2.4	0.7	0.0	72.0	25.6	74.4
L10	40.4	21.4	10.4	9.6	7.1	1.1	3.2	4.3	1.4	1.1	72.1	24.0	76.0
L12	28.5	34.9	13.4	6.7	0.7	1.4	2.1	6.7	5.6	0.0	76.8	20.4	79.6
SII-3	29.2	36.0	13.3	5.7	1.1	1.5	1.5	5.3	4.9	1.5	78.4	22.4	77.6
SII-6	41.2	21.3	12.0	12.0	7.5	0.7	1.5	0.7	3.0	0.0	74.5	22.4	77.6
SII-9	44.4	21.4	11.5	8.5	7.7	1.3	1.3	0.9	1.7	1.3	77.4	23.6	76.4
SII-15	39.6	23.1	12.0	11.7	7.9	0.9	1.9	2.2	0.6	0.0	74.7	23.8	76.2
LII-5	29.6	34.8	16.6	7.9	0.8	4.3	0.8	2.0	1.6	1.6	81.0	24.9	75.1
Average	35.6	25.8	14.5	9.3	3.9	3.0	2.4	2.6	2.5	0.6	75.9	24.8	75.2



Fig. 3. Classification of Dhosa Sandstones based on classification of Folk, 1980 (after Ghaznavi et al., 2015). Here, Q = Quartz; F = Feldspar; R = Rock fragment.

(Fig. 2b) as well as stretched forms. Here, recrystallized ones with an average of 3.03% dominate over stretched variety constituting an average 1.07% (Table 1). The lensoid shaped sub individuals of recrystallized grains are sub-rounded to well-rounded. Stretched metamorphic quartz occurs in form of elongated micro quartz. However, they are completely absent in some sandstones.

4.1.2. Feldspar

Feldspar (avg. 6.99%) occurs both-as K-feldspar and plagioclase variety with the former being prominent. K-feldspar occurs mainly as microcline and minor as orthoclase. Microcline is represented by prismatic to rounded broken grains (Fig. 2c, d). Plagioclase forms an average of 2.2% of the total feldspars. Microcline and plagioclase occur in the form of sub-equant with subangular to subrounded outline. Alteration and leaching of these feldspars can be seen along cleavage plains and grain boundaries on account of their hydrothermal and weathering conditions (Fig. 2d). These alteration products occur in the form of sericitic mica and clay minerals (Fig. 2d).

4.1.3. Mica

Mica occurs in form of muscovite (avg. 1.38%) and biotite (avg. 0.03%). Tiny to long elongated flakes with frayed ends (Fig. 2e) are visible in the Dhosa Sandstone. The neighboring quartz grains exert mechanical compaction in these grains making their ends curved.

4.1.4. Lithic fragments

The lithic fragments constituting metamorphic as well as sedimentary rock fragments average to 0.80% in the Dhosa Sandstone. The metamorphic rock fragments include quartz schist and quartzite. The sedimentary rock fragments include siltstone, shale, and chert (Fig. 2f). Shale fragments on account of their soft nature have been squeezed occasionally in between the surrounding grains. Siltstones show subangular to sub-rounded silt-size quartz that is embedded in the matrix. They are sub-rounded to well-rounded and their size is similar to that of surrounding quartz grains.

4.1.5. Heavy minerals

The average heavy mineral concentration in the Dhosa Sandstone Member is 0.71%. The heavy mineral assemblages consist of transparent and opaque minerals. The transparent minerals comprise zircon, tourmaline, rutile, epidote, biotite, muscovite, chlorite, sillimanite, hornblende, and staurolite (Fig. 2d, g) (Table 3). Zircon (Fig. 2h), tourmaline and rutile are the most prominent heavy minerals resulting in high ZTR values (76.05%). Epidote, biotite, muscovite, hornblende, chlorite (Fig. 2i), sillimanite and staurolite occur in minor amounts (Table 3). Opaque heavy minerals include oxides and hydroxides of iron, primarily magnetite, hematite and limonite.

4.1.6. Cements

Cements occur in form of carbonates, iron, silica and clays in decreasing order of their abundance. The calcite cement occurs in form of well-developed rhombs and blocky crystals. These cements lodge in the pore spaces and cause porosity loss (Fig. 2c). Some of the calcite cement also shows patchy occurrence due to dissolution. The iron cement too occurs as pore filling as well as patchy material. Few corroded quartz grains and calcite cement occur with iron cement around them which suggests replacement of calcite cement by iron cement. Silica cement occurs as overgrowth whereas clays in form of kaolinite, chlorite, illite and smectite occur in various forms as booklet stacks, vermicules, sheets, rosette-shaped clusters and filamentous growth.

In order to classify the studied sandstones according to Folk's (1980) scheme, all the above essential constituents were recalculated to 100% ignoring the clay matrix, chemically precipitated cements, heavy minerals, and mica. The average composition of framework grains is as follows: 90.20% quartz, 7.07% feldspar, and 2.73% rock fragments. The studied sandstones mainly fall in the subarkose field. However, some of the samples fall in the quartzarenite field followed by a few of them in the sublitharenite field (Fig. 3) (Ghaznavi et al., 2015).

4.2. Bulk mineralogy

The mineralogical contents were also verified through XRD analysis of both sandstones as well as shale samples, based on semi-quantitative analysis described by Moore and Reynolds (1997) and Środoń (2006). Mineral peaks that were identified are represented in Fig. 4 and their palaeoenvironment and/or palaeoclimatic significance are listed in supplementary Table S3. The XRD spectra from bulk powder samples taken from all units reveal mineralogical assemblages are dominated by non-clay minerals. Quartz is the most abundant non-clay mineral (~92%) and is essentially present in all the samples (Fig. 4). This is followed by mica (~2%), illite (~1%), kaolinite (~1%), and zeolite (~1%). The clays are found to be distributed on the surface of framework grains and in pores.

4.3. Major oxides

Major element concentrations are given in Table 4 which shows the major oxide composition of shale and sandstones. For sandstones, SiO₂ is the most abundant oxide (avg. 66.5%), followed by CaO (avg. 10.8%), Al₂O₃ (avg. 6.3%), Fe₂O₃ (avg. 2.3%), K₂O (avg. 1.7%), TiO₂ (avg. 0.8%), Na₂O (avg. 0.2%), MgO (avg. 0.8%), P₂O₅ (avg. 0.04%), MnO (avg. 0.04%). For shales too, the most abundant oxide is SiO₂ (avg. 78.6%). It is followed by Al₂O₃ (avg. 6.1%), CaO (avg. 4%), Fe₂O₃ (avg. 0.4%), M₂O (avg. 1.6%), TiO₂ (avg. 1.1%), Na₂O (avg. 2.2%), MgO (avg. 0.4%), P₂O₅ (avg. 0.03%), MnO (avg.0.02%). Geochemical classification of Herron (1988) and Pettijohn et al. (1987) show that studied sandstones belong to quartzarenite, sublitharenite to Fe rich sand in composition (see supplementary Fig. SF1).

4.4. Trace elements

Trace element data are given in Table 4. Dhosa Sandstone Member samples have a high Zr concentration (avg. 367.2 ppm). In the Gypsiferous Shale Member, Zr and Ba constitute a very high average percentage of 673.00 ppm and 326.00 ppm respectively. Averages of most of the trace elements except Sc, Rb and Sr, are comparatively greater in the Gypsiferous Shale Member than in the Dhosa Sandstone Member.



Fig. 4. Representative X-ray diffraction patterns (sandstone and shale) from Chari Formation. I = Illite, K = Kaolinite, Q = Quartz, M = Magnetite, M' = Muscovite, P = Phengitic, Z = Zeolite.

4.5. Bivariant relationship between major oxides and trace elements

Correlation coefficient is a numerical measure which establishes a statistical relationship between two variables. For correlation coefficient, values range from -1 to +1, where -1 represents strongest possible disagreement and +1 represent strongest possible agreement, are assumed (Table S4). As Al_2O_3 is regarded to be immobile during the processes of weathering, diagenesis and metamorphism (Bauluz et al., 2000), major and trace element abundances are plotted against Al_2O_3 in

order to verify the primitive geochemical character. Fig. 5 shows that for sandstones, Al_2O_3 shows a weak to moderately positive correlation with Na_2O (r = 0.2), MgO (r = 0.3), CaO (r = 0.2), TiO₂ (r = 0.2), MnO (r = 0.3) and SiO₂ (r = 0.5) and strong positive correlation with P_2O_5 (r = 0.8), K_2O (r = 0.8), Fe_2O_3 (r = 0.9), Nb (r = 0.8), Zr (r = 0.8), V (r = 1.0), Th (r = 0.8), and Y (r = 0.9). In shale samples, Al_2O_3 shows a moderately positive relation with Nb (r = 0.4) and strong positive correlation with Na_2O (r = 0.9), P_2O_5 (r = 0.9), K_2O (r = 1.0), CaO (r = 0.9), MnO (r = 1.0), Fe₂O₃ (r = 0.9), Zr (r = 0.7),

Table 4

Major and trace element compositions of Chari Formation rocks Ler dome, Kachchh, western India.

S. id	D-1	D-2	D-12	D-5	D-7	D-10	D-11	D-15	CLSII-1	CLSII-4	CLSII-3
Major elemen	ts (wt%)										
Na ₂ O	0.22	0.15	0.20	0.22	0.29	0.23	0.22	0.27	5.98	0.20	0.54
MgO	0.20	0.11	2.78	0.28	0.61	0.60	1.91	0.63	0.39	0.40	0.41
Al ₂ O ₃	3.42	4.56	6.25	8.28	4.41	5.90	8.61	10.09	9.26	6.07	3.14
SiO ₂	56.13	69.14	63.09	81.72	66.98	64.46	62.67	71.71	73.68	78.75	81.51
P_2O_5	0.02	0.02	0.04	0.04	0.02	0.02	0.04	0.11	0.04	0.03	0.03
K ₂ O	1.46	1.68	1.56	2.06	1.77	1.70	1.81	1.93	1.82	1.66	1.36
CaO	20.87	12.67	8.67	1.25	15.01	13.66	9.97	5.02	1.16	5.98	5.62
TiO ₂	0.87	0.87	0.92	0.71	0.87	0.68	0.95	0.99	1.22	0.91	1.38
MnO	0.04	0.02	0.08	0.03	0.02	0.05	0.08	0.03	0.02	0.02	0.02
Fe ₂ O ₃	0.87	0.76	3.79	2.62	0.78	1.28	4.52	4.48	1.99	0.84	0.77
LOI	15.91	10.03	12.61	2.78	9.22	11.42	9.22	4.74	4.44	5.14	5.20
Total	100.01	100.01	99.99	99.99	99.98	100.00	100.00	100.00	100.00	100.00	99.98
Geochemical	indices										
CIA	59.71	66.31	72.70	73.62	60.43	69.38	76.14	77.29	29.96	71.21	49.02
ICV	1.33	0.92	2.10	0.76	1.38	1.00	1.37	0.87	2.75	0.86	2.10
CIW	82.53	90.27	90.53	91.92	81.98	88.58	92.10	92.04	32.00	90.28	63.73
PIA	71.72	84.77	87.45	89.24	71.99	84.21	90.01	90.16	27.04	86.72	48.18
Trace elemen	ts (ppm)										
Sc	14	8	10	6	11	11	10	8	8	8	10
Со	24	21	25	47	20	26	21	25	42	38	40
Ni	10	9	12	25	11	13	18	17	17	18	13
Cu	2	3	9	10	3	5	7	10	12	5	4
Zn	BDL	4	16	38	2	10	17	27	37	9	BDL
Ga	7	7	8	10	7	7	9	10	11	8	7
Pb	20	20	21	22	21	21	21	21	22	21	21
Th	10	14	19	17	11	9	18	23	40	17	22
Rb	22	28	38	55	28	33	45	51	48	34	26
U	1	1	BDL	1	1	2	2	2	3	BDL	1
Sr	100	121	107	81	126	136	152	91	117	108	107
Y	8	10	14	15	9	8	15	19	27	13	15
Zr	274	351	445	417	272	247	402	530	961	449	609
Nb	10	11	13	15	10	8	13	16	34	15	26
Ва	208	278	276	431	260	268	282	336	386	338	254
Cr	11	25	52	81	28	39	52	116	111	75	91
v	30	36	62	71	36	39	65	88	95	55	62

Th (r = 0.8), Y (r = 0.8) and V (r = 0.8). However, the correlation is negative with MgO (r=-1.0), SiO₂ (r=-1.0),TiO₂ (r=-0.3) and Sc (r=-0.8). All these relationships suggest that geochemical abundances of these rocks are near primary. But MgO, SiO₂, TiO₂ and Sc have undergone redistribution process.

5. Discussion

5.1. Tectonic setting and provenance

Petrofacies were calculated (Appendix 1) (Table 2) and plotted on the standard ternary diagrams viz. Qt-F-L, Qm-F-Lt, Qp-Lv-Ls and Qm-P-K as specified by Dickinson (1985). In the Qt-F-L plot, samples lie in continental block provenance (Fig. 6a). Some of the samples also fall in recycled orogenic provenance. The high percentage of Qt (90.50%) and moderate feldspars (6.8%) (see Table 2) corroborated with these findings suggest moderate relief and proximity of source and depositional basin. In Qm-F-Lt petrofacies, the data points are in between F and Lt near the apex quartzose recycled orogen provenance (Fig. 6b). The small extent of an overall feldspar percentage shows they can possibly be a derivative of recycled orogeny source. In the Qp-Lv-Ls triangular diagram, the data plots completely lie on the Qp-Ls leg which suggests rifted-continental margin type of tectono--provenance (Fig. 6c). In the Qm-P-K petrofacies, data points are inclined more toward the Qm-K leg. It suggests maturity of continental block provenance (Fig. 6d). However, since the overall provenance seems closely to be of mixed type, the provenance model should not be over interpreted and can further be improved by including quantified knowledge about the processes that govern the generation of these sediments (Weltje, 2006).

Furthermore, the heavy mineral data were plotted in the Nechaev and Isphording (1993) diagram to distinguish three groups, i.e. MF, GM and MT. The MF component has mafic minerals that include olivine, iddingsite, pyroxenes and green-brown hornblende. GM group holds zircon, tourmaline, staurolite, sillimanite, andalusite, monazite and kyanite. The MT component has basic metamorphic minerals such as pale colored and blue-green amphiboles, garnet and epidote. Our samples plot in the GM-MT field, i.e., a mature passive continental margin (see supplementary Fig. SF2). The presence of mature heavy mineral assemblages (ZTR \sim 76%) and high quartz content in the samples are in sync with these findings. The subsidence curves are relatively smooth for the Jurassic reflecting mature passive-margin sedimentation (Sciunnach and Garzanti, 2012).

The geochemical composition of clastic rocks has been successfully utilized to assess the initial composition of source rocks (Raza et al., 2010). However, geochemical components of the basin sediments are prone to alteration by weathering, hydraulic sorting and diagenesis (e.g. Roser and Korsch, 1988) but chemical signatures to the original source terrain still prevail. In order to further support the results of major element geochemical data, it is vital to corroborate them with petrographical data so that suitable tectonic setting can be ascertained.

In order to confirm the tectonic setting, we used Verma and Armstrong-Altrin (2013) diagrams that is based on the \log_{e} ratio transformation of major-elements for high-silica and low-silica clastic sediments. These diagrams are of great importance since they are capable of differentiating arc sediments from the collision and continental rift sediments unlike the traditional plots that discriminate amongst ocean island arc, continental island arc, active continental margin and passive margin (Bhatia, 1983; Roser and Korsch, 1986) but does not clarify that in what way continental (island) arcs are different



Fig. 5. Variation in major oxides and trace elements abundances versus Al₂O₃ (wt%) content of Chari Formation rocks.

from active continental margins. The effects of sedimentary provenance (source parameters like lithology, climate, relief) were minimized by only working on siliciclastic sediments so that effect of the tectonic settings (in terms of plate tectonics) are likely to dominate the chemical composition of each type of sediment (Verma and Armstrong-Altrin, 2013). For this reason, the samples were divided into high-silica ((SiO₂)_{adj} = 63%–95%) and low silica ((SiO₂)_{adj} = 35%–63%) types. These discriminant-function-based diagrams were plotted from nine

normally distributed natural logarithm of major-element ratios with $(SiO_2)_{adj}$ as the common denominator. The discriminant function diagram for high silica samples shows that they belong to continental rift and collision setting with Dhosa Sandstone Member having an equal contribution from both the setting (Fig.7a). However, two of the Gypsiferous Shale Member belongs to rift setting. For the low silica samples, DF1 vs DF2 plot shows that both the sandstone samples belong to collision field (Fig.7b). The results are in consonance with the geology



Fig. 6. Classification of the Dhosa Member Sandstone (according to Dickinson, 1985) (after Ghaznavi et al., 2015) The provenance field in (a) and (b) are (I) Continental Block: I-A: Craton Interior, I-B: Transitional Continent, I-C: Continent uplift; (II) Recycled Orogen: II-A Quartzose, II-B: Transitional, II-C: Lithic; (III) Magmatic Arc: III-A: Dissected, III-B: Transitional III-C: Undissected and Mixed (IV); in (c) Rifted continental margin: I, Subduction complex: II, Collision suture and fold thrust belt: III, Arc Orogen: IV; and in (d) Circum pacific volcano plutonic suites: I. The arrow indicates maturity and stability from continental block provenance.

of the Kachchh basin where rifting was prominent.

In order to discriminate active and passive margin settings from the isometric log-ratio transformation of major and trace elements, two new discriminant function-based multidimensional diagrams (Verma and Armstrong-Altrin, 2016) were made. Both the diagrams (Fig. 8a, b) show that the all the samples of sandstones and shales exclusively plot in passive margin setting.

For determining provenance characteristics, polycrystalline quartz versus non-undulatory and undulatory monocrystalline quartz was plotted in a double-triangular diagram given by Basu et al. (1975) which was later modified by Tortosa et al. (1991) and presented in Fig. 9. Conjoin study of undulosity in monocrystalline quartz with polycrystalline quartz and their crystals per unit grain helps to discern sands of plutonic, low rank and high-rank metamorphic parentage (Basu et al., 1975). The majority of samples fall in plutonic source and few of them lie in the medium- to the high-grade metamorphic field. They have a similar distribution in Qp2–3 and Qp > 3 fields and data tend to approach the Qm non (non-undulatory monocrystalline quartz) vertex.

TiO₂ versus Zr diagram was plotted in order to have an insight into the source rock characteristics (Hayashi et al., 1997). According to this diagram most of the sandstones and all of the shale samples are derived from felsic igneous source lithologies (see supplementary Fig. SF3). Few of the sandstones also depict an intermediate igneous rock parentage. But during transport and deposition of clastic sediments, fractionation of zircon and titanium-bearing minerals may be significant and thus the TiO₂/Zr ratio of clastic sediments is not always a good indicator of the nature of their source rocks (Hayashi et al., 1997).

The NE-SW trending Aravalli mountain ranges are the main structure of the Aravalli craton of the north-west Indian shield. The belt constitutes rocks of Aravalli and Delhi Supergroup that have undergone deformation and metamorphism (Meert et al., 2010). The rocks of the Aravalli Supergroup are composed of low-grade meta-arenites, carbonates, greywackes, interbedded fine-grained silty arenites and metapelites (Banerjee and Bhattacharya, 1994) producing more sub-rounded and polycyclic sediments (Dubey and Chatterjee, 1997). They overly on the Archaean basement complex of gneisses (Banded Gneissic Complex i.e., BGC), schists and high-grade metasediments (Banerjee and Bhattacharya, 1994). Moreover, Nagarparkar Complex is a part of Precambrian fragment of the Western Indian Shield and the granites of the complex are considered to be the extension of the post-Aravalli magmatism in the Late Proterozoic (Muslim et al., 1997). The sediments deposited by them are sub-angular and monocyclic with an abundance of heavy minerals like zircon, tourmaline and rutile, monazite, etc. and polycrystalline quartz grains (Dubey and Chatterjee, 1997). The samples in the present study show similarity with the above said sources in terms of presence of metamorphic rock fragments (schist, phyllite), polycrystalline quartz (RMQ, SMQ and chert), heavy mineral assemblages (staurolite, sillimanite, epidote, garnet indicating metamorphic source; biotite, pink tourmaline, muscovite, euhedral zircon and rutile confirming igneous sources) and large carbonate content (observed as both cement in thin section and major oxide geochemistry). Marked presence of epidote in the studied samples also strengthens our interpretation of source since it is commonly found in the joints and veins of Nagarparkar Massif (Kazmi and Khan, 1973). These findings are in consonance with the earlier studies (Dubey and Chatterjeee, 1997; Ghaznavi et al., 2015) where they have concluded that detritus was derived from a mix provenance constituting granitegneiss with some meta sedimentary supracrustals-possibly a derivative of the Aravallis and Nagarparkar Massif. The findings of above discriminant diagram and bivariants conclude that these sediments were derived from a passive margin tectonic setting with an intermediate to felsic igneous provenance.



Fig. 7. New discriminant-function multi-dimensional diagram proposed by Verma and Armstrong-Altrin (2013) for high-silica clastic sediments from three tectonic settings (arc, continental rift, and collision). The subscript $_{m1}$ and $_{m2}$ in DF1 and DF2 represents the high-silica and low silica diagram respectively, based on $\log_{e^{-1}}$ ratios of major-elements. The discriminant function equations are: For High-silica: (SiO₂)_{adi} = > 63%- \leq 95%

$$\begin{split} DF1_{(Arc-Rift-Col)m1} &= (-0.263 \times In(TiO_2/SiO_2)_{adj}) + (0.604 \times In(Al_2O_3/SiO_2)_{adj}) \\ &+ (-1.725 \times In(Fe_2O_3^t/SiO_2)_{adj}) + (0.660 \times In(MnO/SiO_2)_{adj}) \\)+ (2.191 \times In(MgO/SiO_2)_{adj}) + (0.144 \times In(CaO/SiO_2)_{adj}) + (-1.304 \times In(Na_2O/SiO_2)_{adj}) \\ \end{split}$$

$$\begin{split} & (1.504\times10(1.502)_{adj}) + (0.144\times10(CdO/SiO_2)_{adj}) + (-1.504\times10(Xa_2O/SiO_2)_{adj}) + (0.054\times10(Xa_2O/SiO_2)_{adj}) + (-0.330\times10(P_2O_5/SiO_2)_{adj}) + 1.588. \\ & DF2_{(Arc-Rift-Col)m1} = (-1.196\times10(TiO_2/SiO_2)_{adj}) + (1.604\times10(Al_2O_3/SiO_2)_{adj}) + (0.303\times10(Fe_2O_3^t/SiO_2)_{adj}) + (0.436\times10(MnO/SiO_2)_{adj}) \\ & + (0.303\times10(Fe_2O_3^t/SiO_2)_{adj}) + (0.436\times10(MnO/SiO_2)_{adj}) \\ \end{split}$$

$$\begin{split}) + & (0.838 \times In(MgO/SiO_2)_{adj}) + (-0.407 \times In(CaO/SiO_2)_{adj}) + (1.021 \times In(Na_2O/SiO_2)_{adj}) + (-1.706 \times In(K_2O/SiO_2)_{adj}) + (-0.126 \times In(P_2O_5/SiO_2)_{adj}) - 1.068. \\ For low-silica: (SiO_2)_{adj} = > 35\% - \le 63\% \end{split}$$

$$\begin{split} DF1_{(Arc-Rift-Col)m2} &= (-0.608 \times In(TiO_2/SiO_2)_{adj}) + (-1.854 \times In(Al_2O_3/SiO_2)_{adj}) \\ &+ (-.299 \times In(Fe_2O_3^{t}/SiO_2)_{adj}) + (-0.550 \times In(MnO/SiO_2)_{adj}) \\ &+ (.120 \times In(MgO/SiO_2)_{adj}) + (0.194 \times In(CaO/SiO_2)_{adj}) + (-1.510 \times In(Na_2O/SiO_2)_{adj}) \\ &+ (.120 \times In(MgO/SiO_2)_{adj}) + (0.194 \times In(CaO/SiO_2)_{adj}) + (-1.510 \times In(Na_2O/SiO_2)_{adj}) \\ &+ (.120 \times In(MgO/SiO_2)_{adj}) + (0.194 \times In(CaO/SiO_2)_{adj}) + (-1.510 \times In(Na_2O/SiO_2)_{adj}) \\ &+ (.120 \times In(MgO/SiO_2)_{adj}) + (0.554 \times In(TiO_2/SiO_2)_{adj}) + (-0.003 \times In(P_2O_5/SiO_2)_{adj}) - 0.294. \\ &+ DF2_{(Arc-Rift-Col)m1} = (-0.554 \times In(TiO_2/SiO_2)_{adj}) + (-0.003 \times In(P_2O_5/SiO_2)_{adj}) + (-1.391 \times In(MnO/SiO_2)_{adj}) + (1.765 \times In(Fe_2O_3^{t}/SiO_2)_{adj}) + (0.225 \times In(CaO/SiO_2)_{adj}) + (0.713 \times In(Na_2O/SiO_2)_{adj}) + (0.330 \times In(K_2O/SiO_2)_{adj}) \\ &+ (0.637 \times In(P_2O_5/SiO_2)_{adj}) - 3.631. \end{split}$$

5.2. Source area weathering

Palaeoweathering is one of the most significant processes in the source area that causes alteration of rocks by depleting them in alkalis and alkaline earth elements and preferential enrichment of Al₂O₃ (Cingolani et al., 2003). This weathering operation affects the mineralogical and geochemical composition of sedimentary rocks (Zou et al., 2016). The quantitative assessment of source area weathering effect can be evaluated by calculating the molecular percentage of the oxide components. For this purpose chemical index of alteration (CIA= $[Al_2O_3/(Al_2O_3 + CaO^* + Na_2O + K_2O] \times 100)$ proposed by Nesbitt and Young (1982) and chemical index of weathering (CIW = $[Al_2O_3/$ $(Al_2O_3 + CaO^* + Na_2O] \times 100)$ proposed by Harnois (1988) are widely used. Here, the value of each component is expressed as a molar proportion. CaO^{*} obtained by the method of McLennan et al. (1993) which is restricted to the ones that are derived from silicate minerals (excluding calcite, dolomite, and apatite). However, the CaO content was very high in the samples so apatite correction was made (mol $CaO_{(corrected)} = mol CaO - [10/3 mol P_2O_5]$). McLennan et al. (1993)

suggested that if the CaO value is greater than Na₂O, then the CaO^{*} concentration is set to be equal to that of Na₂O. In all the studied samples, the calculated CaO^{*} was greater than that of Na₂O, therefore, $CaO^* = Na_2O$ was taken. The CIA value in the unweathered upper crust and unaltered plagioclase and K-feldspar is ~50 indicating first cycle sediments. It increases with increasing degree of chemical weathering (Nesbitt and Young, 1982). The average in shale is ~70-75, which increases up to 100 in case of highly weathered residual soils. The CIW values of studied samples range from 82.5 to 92.1 (avg. 88.7) in sandstones and 32 to 90.28 (avg. 62) in shales. The CIA in sandstones ranges from 59.7 to 77.2 (avg. 69.4), and 29.9 to 71.2 (avg. 50.0) in shales (Table 4). Thus proposes moderate to high and low to moderate weathering conditions for sandstones and shales of the study area, respectively (see Nesbitt and Young, 1982; Harnois, 1988). For the analyzed samples, the value of CIW index is higher than CIA values due to exclusion of K₂O from the index. ACNK ternary diagram (Nesbitt and Young, 1984) was also used to evaluate the mobility of the elements (Fig. 10). Here, $A = Al_2O_3$ (mol %), $CN = CaO^* + Na_2O$ (mol%) and $K = K_2O$ (mol%). Increased weathering leads destruction of plagioclase that lead to removal of CaO and Na2O and the samples plots close to A-K boundary (Fig. 10). Further, K is removed with respect to Al from Kfeldspar due to which the trend is near the Al₂O₃ apex which is attributed to post-depositional K-metasomatism modification (Fedo et al., 1995). This metasomatism can take place either by conversion of aluminous clay minerals (such as kaolinite) to illite or by conversion of plagioclase to K-feldspar. Chemical weathering can be inferred by plagioclase weathering through calculating Plagioclase Index of Al-(PIA); following teration the formula $(Al_2O_3-K_2O)/$ $(Al_2O_3 + CaO^* + Na_2O-K_2O) \times 100$ proposed by Fedo et al. (1995). This may be used an alternative of CIW because plagioclase is significantly present in silicate rocks and dissolves relatively fast. For unweathered plagioclase, PIA values around 50. PIA is close to 100 in case complete conversion of plagioclase into secondary aluminous clay minerals such as kaolinite, illite and gibbsite (Fedo et al., 1995) takes place. The PIA values in sandstone samples range from 71.72 to 90.16 (avg. 83.6) but shales show a wide range from 27.0 to 86.7 (avg. 53.9) which indicate moderate to high and low to moderate weathering, respectively (Table 4).

Climate is an important factor which determines weathering intensity and composition of detritus (Suttner and Dutta, 1986). For interpreting the palaeoclimate of Ler dome sandstones, bivariant log/log plot was used. Here, the ratio of log values of polycrystalline quartz to feldspar plus rock fragments was plotted against the ratio of total quartz to feldspar plus rock fragments (Suttner and Dutta, 1986). The plot of samples shows a humid to sub-humid palaeoclimate (Ghaznavi et al., 2015) (Fig. 11a). The findings were further corroborated by geochemical data. The plot between SiO₂ and Al₂O₃+ K₂O + Na₂O depict that majority of the samples belong to humid to semi-humid environment attaining a high chemical maturity (Fig. 11b).

The petrographical data was also plotted in Weltje et al. (1998) diagram which shows that all the plots fall in field number 1 (Ghaznavi et al., 2015) (Fig. 12). This depicts that the detritus was deposited in a temperate to sub-humid climate in moderate relief. India as part of Gondwanaland experienced a humid to tropical climate (Chandler et al., 1992). The point counts for the sandstones were also plotted on the ternary plot by Cox et al. (1995) (and modified by Alam, 2002) which shows a metamorphic source, possibly deposited under humid climate (see supplementary Fig. SF4).

Jurassic Period witnessed a large amount of carbonate precipitation due to the warm and humid climatic conditions similar to that in the tropics. It was a favorable time for accumulation of carbonate (Budyko et al., 1987). The increase in carbon is related to increased productivity and/ or/ low oxygen/anoxic bottom water and/ or high burial rates of organic carbon favoring preservation of organic carbon (Tyson, 1987). Along with this, intensified seafloor spreading and volcanic activity increased flux of volcanic CO_2 to the atmosphere (Berner, 1994).



Fig. 8. New multi-dimensional discriminant function diagram for the discrimination of active (A) and passive (P) margin settings based on major element (M) and combined major and trace element (MT) proposed by Verma and Armstrong-Altrin (2016). The function $(DF_{(A-P)M})$ and $(DF_{(A-P)MT})$ is calculated from Eq. 1 and Eq. 2, respectively. (A-P)_{boundary} is the boundary dividing the two fields.

(a) Eq. 1. $DF_{(A-P)M} = (3.0005^{*}ilr1_{Tim}) + (2.8243^{*}ilr2_{Alm}) + (-1.596^{*}ilr3_{FeM}) + (-0.7056^{*}ilr4_{MnM}) + (-0.3044^{*}ilr5_{MgM}) + (0.6277^{*}ilr6_{CaM})$

 $+(-1.1838*ilr7_{NaM})+(1.5915*ilr8_{KM})+(0.1526*ilr9_{PM})-5.9948;$

(b) Eq. $2.DF_{(A-P)MT} = 3.2683*ilr1_{TiMT}) + (5.3873*ilr2_{AlMT}) + (1.5546*ilr3_{FeMT}) + (3.2166*ilr4_{MnMT}) + (4.7542*ilr5_{MeMT}) + (2.0390*ilr6_{CaMT})$

+ $(4.0490*ilr7_{NaMT})$ + $(4.7542 ms_{MgMT})$ + $(2.3688*ilr9_{PMT})$ + $(2.8354*ilr10_{CrMT})$

 $+ (0.9011^{\circ}ilr11_{NbMT}) + 1.9128^{\circ}ilr12_{NiMT}) + (2.9094^{\circ}ilr13_{VMT}) + (4.1507^{\circ}ilr13_{VMT}) + (4.1507^{\circ}il$

 $r14_{YMT}$) + (3.4871*ilr15_{ZrMT})-3.2088.

Presence of high CaO content can be corroborated with the above factors. Water cycles accelerated and coincided with greenhouse episodes (Del Genio et al., 1991) which have certainly left imprints on the accumulation history of these clastics (Weissert and Mohr, 1996). Moreover, high rates of cementation took place after considerable burial by groundwater saturated with calcium carbonate that moved through the pores (Tandon and Friend, 1989). This is evidenced in form of carbonate cement confirmed by petrography and geochemistry.

5.3. Hydraulic sorting

Hydraulic sorting aid in sorting detrital minerals according to their grain properties. This type of sorting significantly influences the chemical composition of bulk sediments. The Index of Compositional Variability (Cox et al., 1995) was calculated in order to evaluate geochemical variability due to hydraulic sorting. Here, $ICV = (Fe_2O_3 + K_2O + Na_2O + CaO + MgO + MnO + TiO_2)/Al_2O_3)$

According to Cox et al. (1995), rock-forming minerals like plagioclase, K-feldspars, amphiboles and pyroxenes show ICV values > 1. ICV values < 1 for alteration products such as kaolinite, illite and muscovite. The ICV values of sandstones range from 0.7 to 2.1 (avg. 1.2) and shale from 0.8 to 2.7 (avg. 1.9) (Table 4). This suggests that the samples are enriched primarily in rock-forming minerals. Both, sandstones as well as shales can be considered as first cycle sediments and are related to areas of most uplift that are not associated with extensive chemical weathering (Hajalilou et al., 2016). Textural maturity can also be defined by SiO₂/Al₂O₃ ratios. High values of these ratios are representatives of matured sediments (e.g., Ahmad and Chandra, 2013). In the Dhosa Sandstone Member, they range from 7.1 to 16.4 (avg. 11.5) and in Gypsiferous Shales, they range from 7.9 to 25.9 (avg. 15.6). Therefore, the Dhosa Sandstone Member belonging to recycled orogeny may be considered a derivative of quartzose recycled rocks having medium to high maturity.

5.4. Implications for palaeoenvironment

Palaeoenvironmental and palaeoclimatic implications vary for authigenic and diagenetic clays. Also, clays from different sources, such as primary and secondary rocks bear different insinuations on palaeoenvironmental and palaeoclimatic condition. These implications of the detrital clays have been summarized in supplementary Table S3. Weathering produces four important clay mineral groups-chlorite, illite, smectite and kaolinite (refer Fig. 4). Illite and chlorite are formed during the initial stages of weathering by alteration of micas and ferromagnesium minerals, respectively (Fürsich et al., 2005). Smectite and kaolinite groups are formed at the advanced stages of weathering. Therefore, the clay mineral assemblages that are produced during



Fig. 9. Diamond-shaped provenance-discrimination diagrams of (a) Basu et al. (1975); (b) Tortosa et al. (1991); Qm non = non-undulatory monocrystalline quartz; Qm un = undulatory monocrystalline quartz; Qp 2-3 = polycrystalline quartz with 2-3 crystals; Qp > 3 = polycrystalline quartz with more than 3 crystals.



Fig. 10. A-CN-K ternary diagram of molecular proportions of Al_2O_3 (CaO + Na₂O) and K₂O respectively for the Chari Formation rocks (after Nesbitt and Young, 1984).



Fig. 11. (a) Bivariant Log/Log plot for Dhosa Member Sandstone according to Suttner and Dutta, 1986 (Ghaznavi et al., 2015); **(b)** Bivariant plot for major oxides to determine the chemical maturity.

weathering are a function of chemistry of water, water-rock ratio, duration of weathering, climate, and slope conditions (Fürsich et al., 2005). Authigenic kaolinite occurs in two forms-as bridge growing in between the detrital grains that occlude the pores. The second variety is in form of thin subhedral plates (Fig. 13a,b). Usually, kaolinites are an alteration product of feldspars formed by intense chemical weathering in tropical to sub-tropical humid climatic condition (Hallam et al., 1991). It may also develop by the circulation of acid solutions during diagenesis (Ghandour et al., 2003). Acidic conditions lead to hydrothermal alteration or weathering of alumino-silicate minerals (feldspar) which facilitate leaching of Ca, Mg, Na and Fe ions (Sheldon and Tabor, 2009). Steep slopes, high water-rock ratio and good drainage seem to be present during the period. The presence of kaolinite in our samples can possibly be a product of continental weathering under warm and



		Physiography (relief)					
	Semi –quantitative	High	Moderate	Low			
	Weathering index	(Mountain)	(Hill)	(Plains)			
			0	1	2		
Climate (Precipitation)	(Semi) Arid and Mediterranian	0	0	0	0		
	Temperate subhumid	1	0	1	2		
	Tropical humid	2	0	2	4		

Fig. 12. Bivariant log ratio plot after Weltje et al. (1998). Q = Quartz, F = Feldspar, RF = Rock fragments (after Ghaznavi et al., 2015).

humid climate that lead to its erosion and transportation from rivers to the sea. Recycling, dissolution and kaolinization of feldspar is also an important contributing factor towards formation of kaolinite. Smectite too, was formed in warm climate with seasonal contrast in humidity (Singer, 1984). Chlorite occurs as a pore-lining and a replacement mineral of detrital grains like ferromagnesian minerals and rock fragments growing tangentially to the surface of detrital grains in the form of rosette-shaped clusters (Moore and Reynolds, 1997) (Fig. 13c, d). Illite is found in a minor amount (Fig. 13e) mostly within the matrix and sometimes in pore spaces. It was formed by neoformation in the pores as well as replacement of kaolinite and smectite precursors (Aynur and Ali, 2005). In our samples, they occur as filamentous growth that may play a vital role in determining reservoir quality on account of their tendency to migrate and clog pore throats (Seemann, 1979). Few iron oxide peaks, solely represented by magnetite, occur in almost all the samples. The presence of magnetites is possibly interlayered with quartz. They can be a result of the diagenetic modification or clastic accumulation of iron-bearing minerals that were initially formed in separated environments. A transgression or regression activity can also be a possible reason for the deposition of magnetite (Fralick and Pufahl, 2006). During transgression, haematite may grade to magnetite in the deeper part of the basin. In the course of regression, sideritic rock accumulated in near-shore positions grade seaward into magnetite bodies.

5.5. Remarks on palaeogeography

After the break-up of Gondwana in the Late Triassic/ Early Jurassic, a continuous seaway known as Transgondwana/ Indo-Malagasian/ South African seaway developed that connected the southern Pacific and Tethys along the East Africa coast (see supplementary Fig. SF5). The opening and spreading of Indian Ocean floor and further northward drift of India led to the final breakup of India from East Africa/ Madagascar and Antarctica (Westermann, 1988). The present shape and size of Indian plate (Ali and Aitchison, 2008) and the nature of the Indian Ocean basin (Sammartin, 2002) were determined by the Wallaby and Zenith plateaus and the Wallaby–Zenith fracture zone which is located offshore of western Australia. Situated on the eastern margin of the southerly extension of the Malagasy Gulf, the Kachchh basin laid on the palaeolatitude of 33 °S during the Middle Jurassic (Dercourt et al.,



Fig. 13. (a) SEM photograph of kaolinite in form of pseduhexagonal vermicular booklet stacks; (b) EDX analysis (qualitative) of (a); (c) SEM photograph showing smectite (S) and chlorite (Cl); (d) EDX analysis of (c); (e) SEM photograph of illite (I).

2000). Rifting is evident from the present studies which have also been described by the previous investigation (Biswas, 1982, 1987). The regional tectonic events or the fault-controlled subsidence kept the sedimentation rate in pace which was further coupled with eustatic sea level changes which lead to cycles of transgression and regression (Fürsich et al., 2001). The sediments deposited during transgression (Oxfordian) were characterized by rock fragments and heavy minerals of metamorphic origin along with polycrystalline and undulatory quartz grains (Dubey and Chatterjee, 1997). This is similar to our samples. Metamorphic rock fragments like schist and phyllite and polycrystalline quartz are found in addition to heavy minerals like staurolite, sillimanite and epidote (medium to high grade metamorphic source) and chlorite, biotite and pale yellow and brown tourmaline (low rank metamorphic source). The sediments deposited during regression were characterized by sandy lithology, and heavy minerals of

igneous origin. The present study depicts zircon (euhedral), tourmaline (pink) and rutile depicting an igneous source. Our interpretation is further strengthened by the geochemical data which has established that the nature of the rocks is nearly primary and the primeval geochemical character is preserved.

6. Conclusions

Following conclusions can be drawn from petrography and geochemistry of Middle-Upper Jurassic rocks of Ler dome, Kachchh (western India):

1 Texturally, the studied sandstones are medium to coarse-grained and moderately to well sorted of subarkosic and quartzarenitic composition. They also show variation in lithic and heavy mineral compositions, indicating different proximal source areas.

- 2 The sediments of Ler dome were derived from quartzose recycled orogen and continental block provenance primarily from the passive continental margin. The provenance settings represent the eroded and weathered parts of the present day Aravalli craton which is situated east and northeast of the basin and the Nagarparkar Massif situated to the north and northwest.
- 3 Weathering indices including CIW, CIA, PIA and ICV were calculated in order to estimate the weathering conditions. These values depicted low to moderate and moderate to high weathering conditions for shales and sandstones, respectively under humid to subhumid palaeoclimatic condition that leads to high chemical maturity of the detritus.
- 4 Geochemical data corroborated with heavy mineral analysis also confirm chemical and mineralogical maturity of these sediments.
- 5 The bulk mineralogy delineates the presence of abundant quartz and phyllosilicate clay minerals (primary products of chemical weathering) points towards warm, humid to sub-humid climate with seasonal rainfall that promoted moderate to extensive chemical weathering in the oxidizing environment under transgression-regression cycles.

Acknowledgements

The authors are grateful to the Chairperson, Department of Geology, AMU, Aligarh for providing the necessary facilities during the study. The authors are also thankful to Dr. Debajyoti Paul, Department of Earth Sciences, Indian Institute of Technology Kanpur and Director, Wadia Institute of Himalayan Geology, Dehradun for providing laboratory facilities. We are also thankful to Prof. M. Masroor Alam, Department of Civil Engineering, AMU for his valuable suggestions and constant encouragement and two anonymous reviewers who have appreciated our efforts and helped in improving the quality of the manuscript. AAG is also thankful to the University Grants Commission (UGC), New Delhi, India for the award of Senior Research Fellowship (No.F1-17.1/ 2015-16/MANF-2015-17-UTT-58619 /(SA-III/ Website).

Appendix A. Supplementary data

Supplementary material related to this article can be found, in the online version, at doi:https://doi.org/10.1016/j.chemer.2018.06.002.

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