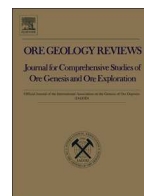




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Uraniferous paleoplacers of the Mesoarchean Mahagiri Quartzite, Singhbhum craton, India: Depositional controls, nature and source of >3.0 Ga detrital uraninites

Joydip Mukhopadhyay^{a,d,*}, Brundaban Mishra^b, Kalyan Chakrabarti^c, Saheli De^a, Gautam Ghosh^a

^a UGC-Centre for Advanced Studies, Department of Geology, Presidency University, 86/1 College Street, Kolkata 700073, India

^b Atomic Minerals Directorate for Exploration and Research, Shillong, India

^c Atomic Minerals Directorate for Exploration and Research, Bengaluru, India

^d PPM Research Group, Department of Geology, University of Johannesburg, Auckland Park, Johannesburg, South Africa

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ABSTRACT

The quartz-pebble conglomerate (QPC)-hosted detrital uranium mineralization is unique in character in terms of their restricted distribution before 2.2 Ga atmosphere during pre-Great Oxidation Event (pre-GOE). Such QPC paleoplacer deposits over the world are good targets for moderate to high tonnage and low grade uranium deposits and more importantly for their gold content. The Mahagiri Quartzite, dated c. 3.02 Ga for their youngest detrital zircon population, is developed unconformably over the Mesoarchean Singhbhum Granite (3.44 Ga to 3.1 Ga). The Mahagiri Quartzite includes a conglomerate-pebbly sandstone dominated subaerial alluvial fan to coastal braided plain sequence in the lower parts and shallow marine mature quartz arenite in the upper parts. The alluvial fan-braided plain deposits in the lower parts host a number of pyritiferous and uraniferous conglomerate and pebbly sandstone beds. The uraninite grains are rounded to subrounded in outline suggesting mechanical transport and detrital origin. Together with detrital pyrite and uraninite constitute the example of >3.0 Ga paleoplacer closely comparable to the Witwatersrand Au–U deposits. EPMA and SEM-EDS studies suggest that the uraninite grains are rich in Th (>4 wt.%), S and REE–Y. Chemical formula calculations from EPMA analyses suggest uraninite grains belong to two populations with different oxidation states as revealed from Y/REE and cation $U^{4+}:U^{6+}$ [apfu] ratios. The U contents of the detrital uraninite grains from Mahagiri are significantly lower than that of the ideal stoichiometric composition of UO_2 . This is mainly due to higher amount of heterovalent cationic substitution by Th, REE, Y, Pb, and Ca in Mahagiri QPC uraninite structures, and partial alteration and metamictization of uraninites. Alteration due to metamictization resulted in elevated concentration of Si, Al, P, and Ca in more altered and metamict uraninite grains. The REE pattern is typically flat with comparable LREE–HREE concentration. The high Th content flat REE-pattern suggests that the uraninites presents high temperature phases (>350 °C) and are magmatic in origin. The Mahagiri detrital uraninite grains suggest existence of highly felsic and K-rich (richer than TTG) granodiorite–granite–monzogranite suites (GGM) of rocks older than 3.1 Ga in the Singhbhum craton.

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1. Introduction

Sediment-hosted uranium deposits rank among the world's most important uranium mineralization mined so far (Tillsley, 1988). Uranium mineralization in sedimentary rocks is commonly classified into several major genetic types such as the quartz-pebble conglomerate mineralization (QPC type), unconformity-related uranium mineralization, sandstone-hosted uranium mineralization, phosphorite-uranium mineralization, and black shale-hosted uranium mineralization (Dahlkamp, 2009; Cuney, 2010a, 2010b, 2014). Among these the QPC-

hosted uranium mineralization is unique in character in terms of their restricted distribution in geologic time and depositional controls. The Archean–Paleoproterozoic QPC deposits over the world are good targets for moderate to high tonnage and low grade (0.03–0.1% of U_3O_8) deposits because they contain detrital uraninite and gold primarily as a product of physical liberation and mechanical transportation from granite-greenstone terrane of Archean age (Frimmel et al., 2005). The deposition of uraninite as detrital sedimentary component was possible because of the low atmospheric oxygen concentration before 2.2 Ga Great Oxidation Event (GOE) (Grandstaff, 1980; Frimmel, 2005). The distribution of the detrital uranium–thorium-bearing minerals in the sedimentary systems is primarily the functions of the depositional environment, hydrodynamic conditions, provenance, and tectonic setting of sedimentation (Plant et al., 1999). There are only a few occurrences of

* Corresponding author at: UGC-Centre for Advanced Studies, Department of Geology, Presidency University, 86/1 College Street, Kolkata 700073, India.
E-mail address: joydip17@gmail.com (J. Mukhopadhyay).

the uraniferous QPCs that are relatively well documented in terms of depositional conditions e.g., Elliot Lake, Canada (Bergen and Fayek, 2012) and Witwatersrand, South Africa (Minter, 1976, 2006; Frimmel, 2005, 2014). Although alternative views of hydrothermal U-mineralization (e.g., Phillips and Powell, 2011) as well as microbially mediated syngenetic U-mineralization model (Horscroft et al., 2011) exist in the case of the Witwatersrand deposits, opinions so far weigh in favour of detrital (placer) origin of the Wits uraninite albeit modified during subsequent alterations in these QPCs (Frimmel et al., 2014).

The QPC-type uraninite mineralization in India known over half a century (Dunn and Dey, 1942; Rama Rao, 1974), is reported from the three Archean cratons, viz, the Dharwar, Bastar, and the Singhbhum cratons (Fig. 1A). Several uraniferous QPC prospects have been reported from the Neoproterozoic Bababudan and Chitradurga Groups and their equivalents in the Western Ghats in the Dharwar craton (Rama Rao, 1974; Srinivasan and Ojakangas, 1986; Fareeduddin et al., 1988; Vasudeva Rao et al., 1988; Viswanath et al., 1988; Mahadevan, 1988, 1995; Radhakrishna, 1988; Roy, 1996). Uraniferous QPCs are described from Neoproterozoic to Paleoproterozoic supracrustals of the Dhanjori Group and its equivalents bordering along the northern periphery of the core of the Singhbhum craton (Vasudeva Rao et al., 1988), and its equivalents in the Mahagiri–Keonjhar–Mankachhura siliciclastics along the southern periphery of the Singhbhum craton (Fig. 1B) (Mahadevan, 1988; Viswanath et al., 1988; Mishra et al., 2008; Chakrabarti et al., 2013; Mukhopadhyay et al., 2014). We have studied the uraniferous QPCs of the Mahagiri Quartzite in the Mahagiri Hill ranges (Figs. 1B and 2) of the Jajpur District in Odisha from the southern parts of the Singhbhum craton in recent times.

Uraniferous QPC deposits are predominantly believed to be paleoplacer deposits that record detrital uraninite (U_3O_8) accumulation in some successions from Archean–Paleoproterozoic times when the atmospheric oxygen concentration remained insufficient for the surface oxidation of the extremely redox-sensitive uraninite. There has been considerable research on the depositional controls and nature of mineralization on the uraniferous paleoplacers of the Witwatersrand and Elliot Lake deposits. In general these are primarily pre-GOE (the Great Oxygen Event) fluvio-deltaic deposits (see for reviews Minter, 1976; Bergen and Fayek, 2012; Frimmel, 2014). Detailed mineralogical and sedimentological studies from these deposits reveal not only the nature of source rock, but also suggest controls of the distribution of economic grade ores in sedimentary successions and signatures of post-depositional remobilization (e.g., Minter, 1976; Frimmel, 2014). In this paper we report the depositional environment and controls on the deposition of the uraniferous siliciclastics in the Mahagiri Quartzite (Fig. 1B). The Mahagiri Quartzite forms a part of the Mesoarchean siliciclastic succession (with youngest zircon age of 3.02 Ga, Mukhopadhyay et al., 2014) that is comparable to or even slightly older than the lower part of the Au–U-bearing siliciclastics of the Witwatersrand succession (U–Pb detrital zircon maximum age of 2985 ± 14 Ma for West Rand Group sedimentation by Kositsin and Krapez, 2004). The Mahagiri U-QPC deposits are thus not only the oldest uraniferous deposit in India but rank among oldest such deposits in the world comparable to the Witwatersrand paleoplacers that record the uraninite older than 3 Ga and provide here a rare opportunity to identify the nature and source of sediment-hosted Mesoarchean uraninite deposit.

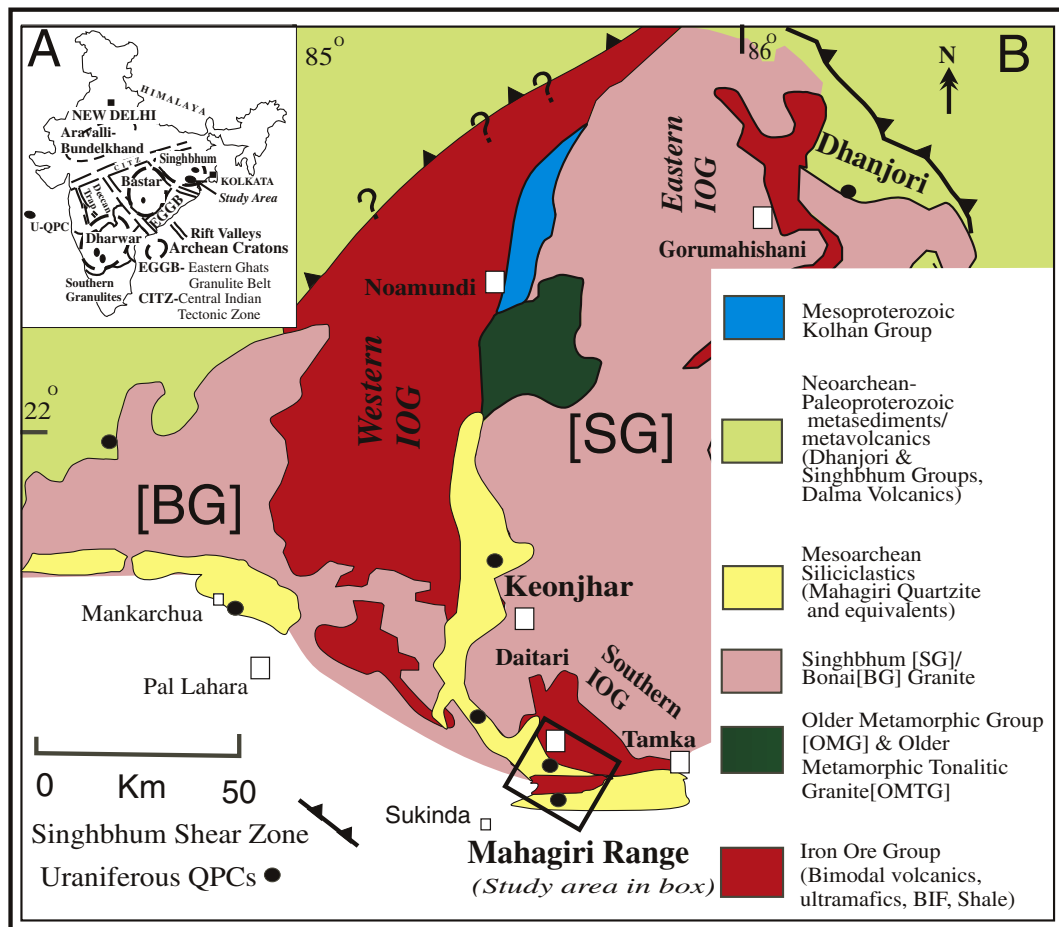


Fig. 1. A) Outline map of India with distribution of five major Archean cratonic nuclei that constitute the Indian shield. Note the occurrences of uraniferous paleoplacers in Dharwar, Bastar, and Singhbhum cratons. B) Generalized geological map of the Singhbhum craton (modified after Saha, 1994).

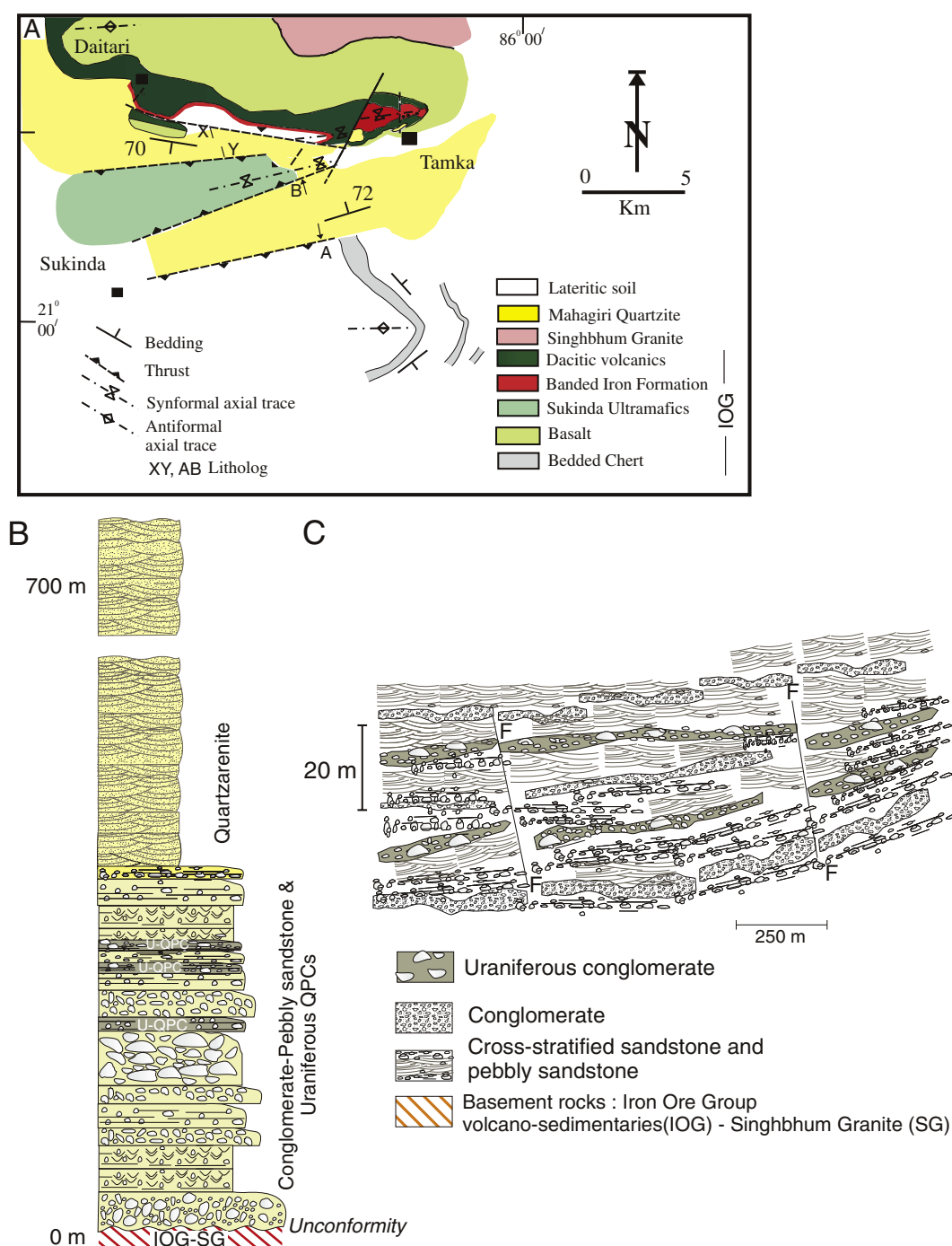


Fig. 2. A) Generalized geological map of the Daitari-Tamka area (after Mukhopadhyay et al., 2012), B) a generalized stratigraphic log of the Mahagiri Quartzite from the southern limb of the Mahagiri synform (along AB in Fig. 2A), C) schematic panel diagram showing mode of occurrence of the uranium-bearing QPCs in the lower part of the Mahagiri Quartzite.

2. Overview of the Singhbhum craton

The Singhbhum craton in eastern India is the one among the five major Archean cratonic nuclei constituting the Indian shield (Fig. 1A). Like most of the ancient cratonic blocks the core of Singhbhum craton includes major TTG (tonalite–trondhjemite–granodiorite) batholiths with enclaves of greenstone belts (Acharyya, 1993) and metasediments, and supracrustal cover sequences. The oldest stratigraphic unit so far dated is reported from the southern margin of the craton of a Paleoproterozoic BIF-bearing greenstone succession, the southern Iron Ore Group (S-IOG), (dated from U–Pb SHRIMP zircon 3506 ± 2 Ma; Mukhopadhyay et al., 2008) that occurs as enclave within the Paleo-

Mesoarchean TTG plutons known as the Older Metamorphic Tonalite Gneiss (OMTG) and Singhbhum Granite (SBG) with three phases of intrusion at 3.44 Ga, 3.32 Ga, and 3.1 Ga. The earliest phase of SBG (SBG-1) is considered as equivalent to the earliest OMTG component (reviews Saha, 1994; Mishra et al., 1999; Misra, 2006; Acharyya et al., 2010a; Nelson et al., 2014; Upadhyay et al., 2014). The TTG cratonic core also includes metasedimentary and metavolcanic enclaves of various dimensions which are collectively termed as Older Metamorphic Group (OMG) (Saha, 1994). The stratigraphic status of the OMG is debatable. Based on higher grade of metamorphism and higher intensity of deformation and intrusive relationship of tonalitic gneisses with them, the OMGs are considered by many workers as the oldest unit in the

Singhbhum craton (Saha, 1994; Mukhopadhyay, 2001; Mazumdar, 1996; Mazumder et al., 2012a, 2012b). The absence of any granitoids or tonalitic gneisses older than 3.44 Ga and the 3.51 Ga precision U–Pb zircon date from the southern IOG tuff beds led some workers to speculate that some of the OMGs are likely to be equivalent to the IOGs of the southern Singhbhum (southern IOG). However, recent detrital zircon U–Pb ages of ~3.3 Ga from some of the metasedimentary enclaves of the OMG (e.g., Nelson et al., 2014) suggest that some OMG enclaves are younger than the first phase of Singhbhum Granite/OMTG emplacement (see also Mukhopadhyay et al., 2008; Acharyya et al., 2010a). Two more major greenstone enclaves similar to the southern IOG occur along the eastern (eastern IOG) and western (western IOG) periphery of the granitoid core of the craton. U–Pb LA-ICPMS age determined from the western IOG acid volcanics suggests an age of 3.29 Ga (Basu et al., 2008). There is no direct age data from the eastern IOG, however, from the intrusive relationship with granitoids within the eastern IOG a minimum age of 3.1 Ga has been suggested (Saha, 1994). The siliciclastic successions unconformably exposed along the southern, western, and northeastern periphery of the Singhbhum Granite in the Mahagiri Range, Keonjhar–Pallahara area and in the basal part of the Dhanjori basin record Mesoarchean sedimentation during <3.1 Ga to 2.8 Ga (Acharyya et al., 2010b; Mukhopadhyay et al., 2014). These siliciclastics have been intruded by granitoids emplaced during ~2.8 Ga (e.g., the Tamperkola and Mayurbhanj Granites, Bandyopadhyay et al., 2001; Acharyya et al., 2010b).

The core of the Singhbhum craton is bordered in the north by a metasedimentary and metavolcanic mobile belt successions collectively known as the North Singhbhum Mobile Belt (NSMB). The southern boundary of the NSMB with the Singhbhum Granite of the craton core is a tectonic contact represented by a northerly dipping crustal scale north-side up shear zone known as the Singhbhum Shear Zone (SSZ) (Saha, 1994). The NSMB from south to north includes stratigraphic units such as the Dhanjori Group, the Singhbhum Group, and the Dalma Volcanics (Gupta et al., 1980). The Chhotanagpur Granite Gneissic Complex (CGGC) is upthrust on the NSMB succession by another major southerly dipping shear zone known as the North Purulia Shear Zone. The age of the successions within the NSMB is yet to be well constrained. The ages of metamorphism, mineralization and granite emplacement within NSMB suggest sedimentation event between 2.7 and 2.1 Ga, deformation and mineralization events from 1.9 Ga to 1.0 Ga with a peak of deformation and mineralization at ~1.6 Ga (Mukhopadhyay, 2001; Gupta and Basu, 2000; Mahato et al., 2008; Mazumder et al., 2012a, 2012b; Pal et al., 2011; Pal and Rhede, 2013; Prabhakar and Bhattacharya, 2013; De et al., 2014).

The youngest stratigraphic unit in the Singhbhum craton is the Kolhan Group. The Kolhan Group is a siliciclastic shelf to deep-water carbonate platform succession (Mukhopadhyay et al., 2006). Mukhopadhyay et al. (2006) considered the Kolhan Group to be equivalent to terminal Mesoproterozoic carbonate platform succession developed in other parts of peninsular India.

3. Geological setting of the Mahagiri uraniferous siliciclastics

The Mahagiri Quartzite covers an extensive area mainly as weathering resistant linear ridges forming the Mahagiri Hill along the southern and western periphery of the metavolcanic-BIF-bearing greenstones of the southern Iron Ore Group (Figs. 1B and 2). The Mahagiri Quartzite is predominantly composed of low-grade metamorphosed arenitic sandstones that unconformably overlie the Paleoproterozoic greenstone succession of the southern Iron Ore Group (S-IOG) and the Paleo-Mesoarchean TTG (tonalite–trondhjemite–granodiorite) of the Singhbhum Granite (Fig. 2). The siliciclastics extend to the farther North as detached outcrop belts beyond Keonjhar Town (Prasada Rao et al., 1964; Acharya, 2002) (Fig. 1B). In the west, the siliciclastics can be traced to the north of the Pallahara Town in the Mankarchua basin (Chakrabarti et al., 2011) as well as at the western margin of Bonai Granite pluton (Kumar et al.,

2012) (Fig. 1B). The siliciclastics unconformably overlie either southern or western IOG greenstones or the Paleo-Mesoarchean Singhbhum Granites (3.4 to 3.1 Ga) and their equivalents with locally well preserved Mesoarchean paleosol (the >3.02 Ga Keonjhar Paleosol, Mukhopadhyay et al., 2014), along the unconformity. Based on LA-ICPMS U–Pb ages of youngest population of detrital zircons from these siliciclastic rocks, Mukhopadhyay et al. (2014) suggested 3.02 Ga maximum age of deposition of these siliciclastics. The minimum age of the siliciclastics has been constrained from the ~2.8 Ga granites that have intruded the siliciclastics in the southern and western outcrop belts (Mukhopadhyay et al., 2014).

The Mahagiri Quartzite, a weakly metamorphosed pack of sediments, is deformed and the outcrop pattern defines broad east–west trending synforms and antiforms with shallow but variable plunge directions in the Daitari–Tamka–Sukinda area. The metamorphic grade in general remained within low greenschist grade as revealed by the presence of sericite/muscovite and chlorite in matrix of the quartzite. The Mahagiri Quartzite is deformed primarily at the vicinity of the Sukinda Thrust (Chakraborty et al., 1980; Ghosh et al., 2010). The megascopic fold pattern includes a synformal closure in the south (Mahagiri synform) and an antiformal closure in the north. The high-grade granulite terranes of the Rengali-Eastern Ghats Province (Ghosh et al., 2010; Bose et al., 2014) is upthrust on the Mahagiri quartzite along the EW to WNW–ESE trending Sukinda Thrust with an upward movement towards north against the northerly dipping southern limb of the Mahagiri synform (Ghosh et al., 2010). The dip of the southern limb of the Mahagiri synforms ranges from 50° to 60° towards north whereas the northern limb is gentler with an average 40° dip towards south. The core of the synform is occupied by the Sukinda Ultramafics of the southern Iron Ore Group, in tectonic contacts with both the limbs (Mukhopadhyay et al., 2012) (Fig. 2).

4. Sedimentary facies in the Mahagiri Quartzite

A combined stratigraphic succession of both the limbs of the Mahagiri synform suggests variation in the composition of about 1000 m thick Mahagiri Quartzite from bottom to top. The lower part of the succession is primarily a pebbly sandstone-conglomerate association of about 400 m thickness (Fig. 2B). The sandstones are light greenish in colour with arenaceous to micaceous matrix showing poor to moderate sorting. The upper part of the Mahagiri Quartzite is essentially clean, mature, and well sorted quartz arenite.

Radioactive mineralization has been recorded in variable extent in many of the conglomerate-pebbly sandstone beds from the lower part of the succession with very poor lateral continuity in most cases (Fig. 2C). However, the uraniferous conglomerate and pebbly sandstone could be traced as laterally discontinuous beds for over a km along the strike with several cm to about a metre-thick uraniferous beds occurring as tabular stratabound intervals with lateral persistence of tens to hundred metres.

To understand the depositional controls of the uraniferous conglomerate-pebbly sandstone beds, the sedimentary facies characters of the entire Mahagiri succession were studied in detail. Here we report in brief the major facies variants in the Mahagiri Quartzite in order to interpret the depositional setting. Based on parameters such as grain size, composition, matrix-content, grain-to-grain contact relation, primary sedimentary structures, bed thickness, and bed geometry, we describe here the primary depositional characters of the Mahagiri Quartzite.

4.1. Matrix-supported conglomerate

Conglomerate beds occur in the lower part of the succession in both the southern and northern limbs of the Mahagiri synform. Conglomerate beds consist of boulder-to-cobble-size clasts in sandy to clayey matrix and are matrix-supported (Fig. 3A). Clasts do not have preferred orientation and are essentially constituted by the clasts of bedded chert, black chert, and vein quartz with a few clasts of altered BIF. Clasts

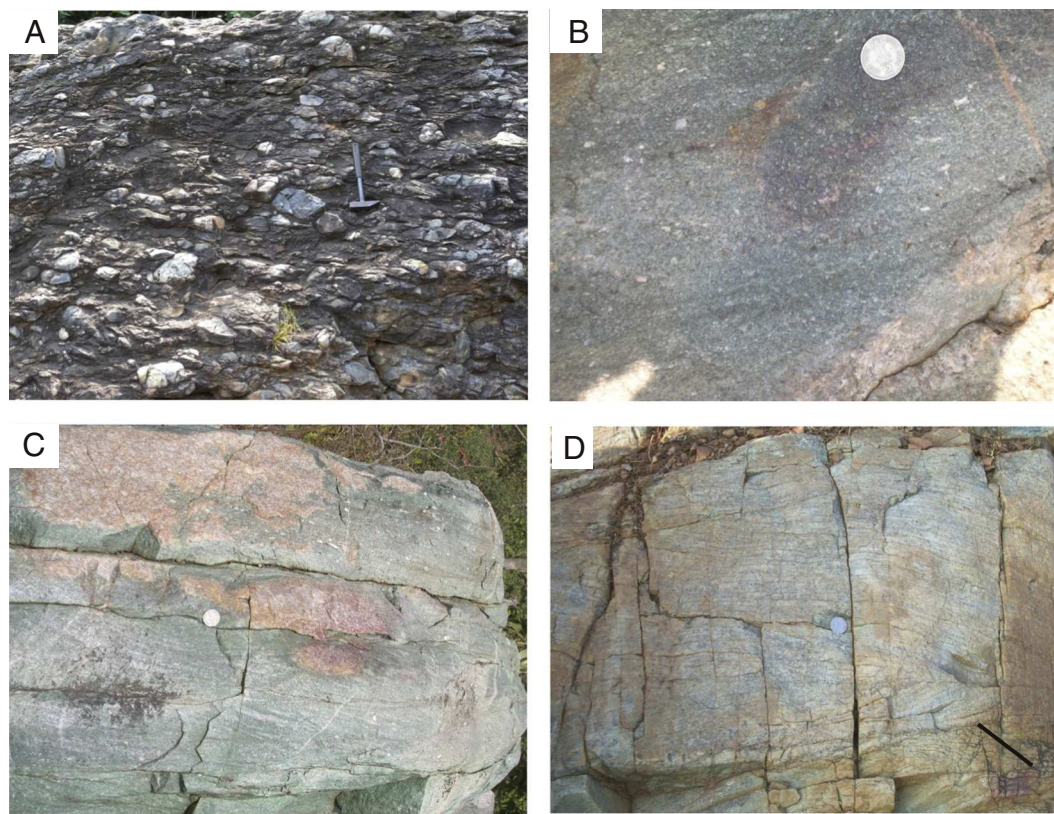


Fig. 3. Sedimentary depositional characters of uraniumiferous and non-uraniferous deposits in the Mahagiri Quartzite: A) matrix-supported boulder-to-cobbly uraniumiferous conglomerate of debris flow origin representing proximal subaerial fan setting, B) uraniumiferous pebbly conglomerate and sandstone beds with locally erosive lower bounding surface, C) cross-stratified uraniumiferous pebbly sandstones. Note tabular sheet-like geometry, D) supermature medium-grained cross-stratified arenite. Cross-strata occur in sets and cosets defining compound strata that locally show reversal of paleocurrent direction.

are subangular in outline and do not show any imbrication. The bed-thickness varies from 30 cm to 8 m. The conglomerate beds are shallow channelized to tabular sheet like bodies with or without limited basal erosion (Fig. 3A). However, the upper bounding surface varies from slightly convex up that locally renders a lateral thickness variation as the beds taper down along the margins. Beds are internally massive and are non-graded.

4.2. Massive-to-weakly stratified pebbly sandstone

Pebbly sandstone beds are coarse-to-medium-grained with thinly dispersed floating pebbles (Fig. 3B). The individual sandstone bed-thickness varies from 2 cm to 15 cm. The size of the pebbles varies from 2.5 cm to 3 cm. The sandstone is non-graded with traces of weak planar stratification and locally developed isolated planar cross-stratification. The sandstone beds are parallel-sided, sheet-like with non-erosive bounding surfaces.

4.3. Cross-stratified pebbly sandstone

The sandstone beds are predominantly medium-grained and are moderately sorted. Matrix constitutes about 10 vol.% of the rock and is represented by sericite or fuchsitic mica. The sandstone beds are commonly sheet-like with wavy upper surface and locally erosive lower bounding surface (Fig. 3C).

4.4. Cross-stratified mature sandstone

This type of sandstone constitutes the upper half of the Mahagiri Quartzite. The sandstones are supermature white medium-grained arenite. Internally the sandstones show tabular as well trough cross-

stratification (Fig. 3D). Cross-strata occur in sets and cosets defining compound strata that locally show reversal of paleocurrent direction (Fig. 3D). The sandbodies show wavy geometry and mega-rippled tops.

5. Mineralogy and chemistry of Mahagiri uraninite

5.1. Samples and methods

Systematic sampling along and across the strike length of radioactive QPC bands/lenses were carried out from the least-weathered bedrock exposures and the drill core samples, wherever available, at selected intervals of the radioactive zones. The radioactivity of the samples was checked by the gamma-ray detecting hand-held NaI (TI) scintillation counter. Based on the intensity of the gamma radiation detected in hand specimen, a few samples were selected for preparation of polished thin section (30 μ m) and investigated under optical microscope using Nikon (model EL-600) microscope attached with an image analyser using BIOVIS v. 1.5 and Mutech software. 100 μ thick transparent cellulose Nitrate films (CN-film; CN-85, Kodak make) were used for autoradiographic studies of the thin sections for locating radioactive mineral grains in the section on the basis of alpha track registered on the film.

Selected thin-sections of uraninite-bearing QPC matrix were studied by analytical electron microscopy. The back-scattered electron (BSE) images were obtained at scanning electron microscopy (make: TESCAN VEGA LSU) equipped with Energy Dispersive X-ray Spectroscopy (EDS – OXFORD) facility at the UGC-CAS Laboratory of Presidency University, Kolkata. Samples were run under high vacuum with the accelerating voltage of 15–20 kV. The semi-quantitative analytical data supplied by the EDS spectrometers helped in identification of mineral phases. The chemical composition of uraninite, its alteration products and other radioactive phases were determined by electron microprobe

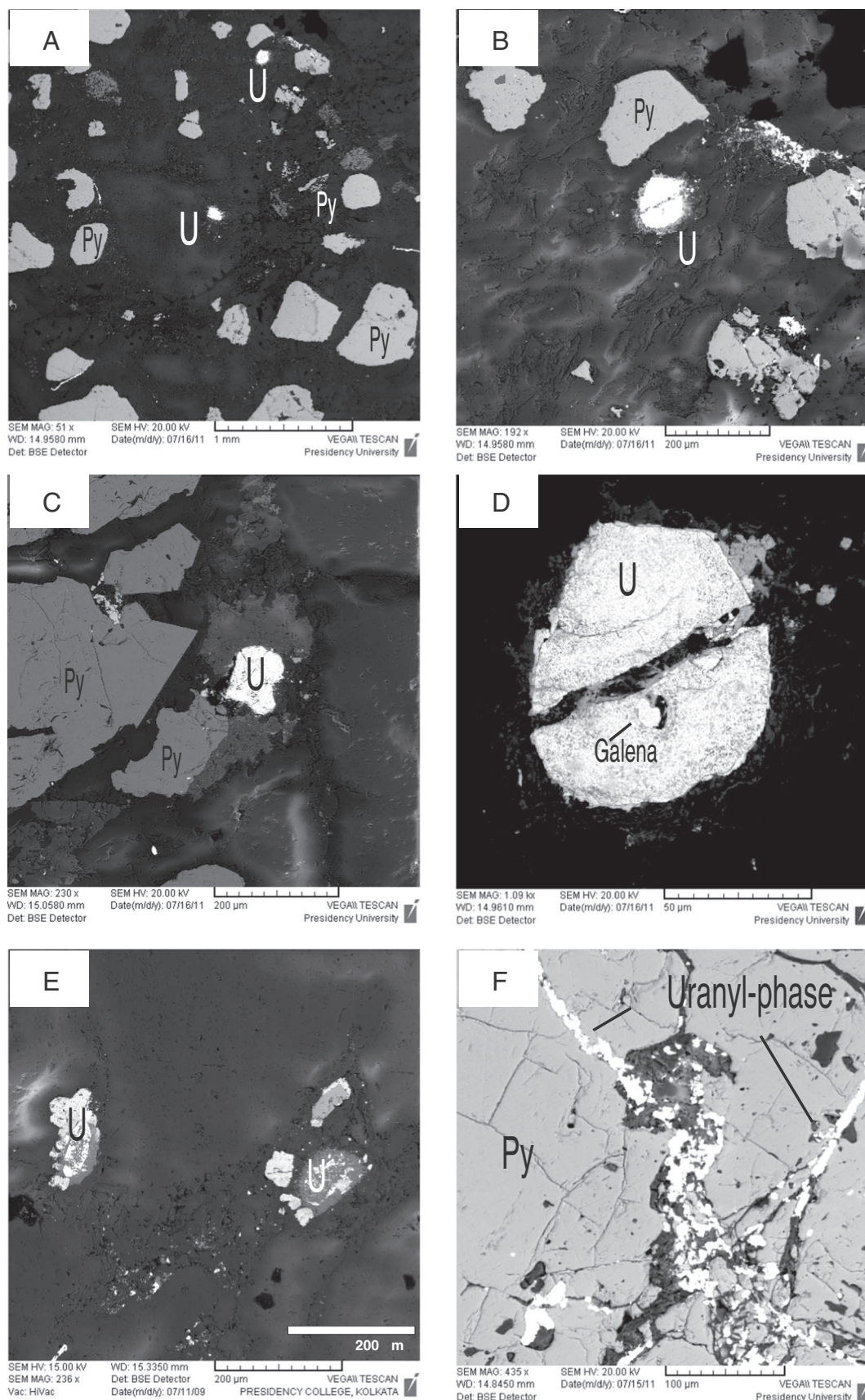


Fig. 4. Detrital uraninites and pyrites under SEM EDS: (A to D) detrital uraninites (brighter) and detrital pyrite grains as framework in Mahagiri Quartzite. Note rounded outline of pyrites and uraninites. The dark background is framework quartz. Note patches of galena within uraninite (D) occurring as product of sulphidisation of radioactive Pb. E) variably altered detrital uraninite (U) grains. Pseudomorphous alteration to sulphides and coffinite of detrital uraninites. Note relict uraninite preserved in brighter shades within altered grains. F) Pyrite grain with secondary veins of uranyl minerals (brighter).

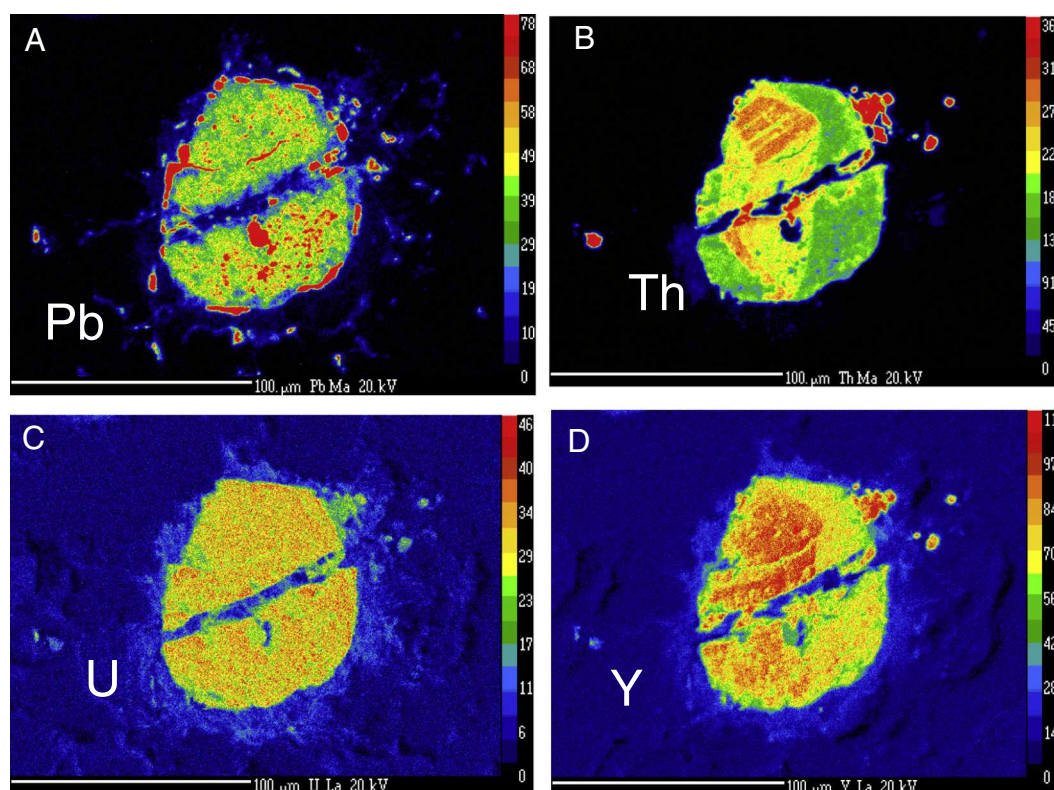


Fig. 5. EPMA elemental map of a uraninite grain. A) distribution of Pb. Note patches of Pb-alteration (sulphidization), B) distribution of Th. Note Th-enrichment in the core of the grain maintaining the euhedral habit of the early phase uraninite, suggesting magmatic origin, C) homogeneous distribution of uranium throughout the grain, D) Y-distribution with enrichment closely comparable to Th suggesting HREE-Y incorporation during early phase of crystallization.

analysis (EMPA, a Cameca SX-100) at the EPMA laboratory of Central Analytical Facility of Department of Geology and Geophysics, Indian Institute of Technology, Kharagpur. Analysis was completed using an accelerating voltage of 20 kV and beam current of 50 nA and a beam spot of $\sim 1 \mu\text{m}$ in diameter. The characteristic X-rays wavelengths were diffracted by crystals, detected and counted by five vertical Wavelength Dispersive Spectrometers (WDS) with a counting time of 120 s to determine the sample composition. In addition to the common major elements the trace elements, REE, Y, Zr, Ba, Sr, U, Th, and Pb of the uranium- and thorium-bearing mineral phases were determined by EPMA analyses. In order to avoid contamination from associated sulphides and adjacent other minerals the spot analyses of the fractures or near grain boundaries were excluded ($< 3 \mu\text{m}$ area). The representative chemical analyses presented here have been selectively taken from a number of point analyses of uraninite of uniform composition. Their compositional uniformity was also ensured from EDS composition and BSE images with smooth uniform brightness. This helped in nullifying any compositional heterogeneity that may arise due to inclusions of other mineral phases within a uraninite grain.

For analytical purpose, the specific X-ray spectral line counted, diffraction crystals, and standards used for each element and their respective detection limits (in ppm) are tabulated in Appendix A. At 20 kV accelerating voltage, 50 nA beam current stability remained well within $\pm 0.05\%$ error (modal value is $\pm 0.025\%$ for the present analyses). Among the REEs Europium, Terbium, Thulium, and Lutetium concentrations could not be measured because both the $L_{\alpha 1}$ and $L_{\beta 1}$ lines of these REEs suffered strong interferences with lines of other REEs. Eu could not be measured because of the serious interference of Eu- $L_{\alpha 1}$ line by other REEs, particularly due to high concentration of Pr and Nd in the samples and low concentration of Eu ($< 0.1 \text{ wt.}\%$). Eu- $L_{\beta 1}$ line was also strongly interfered by Dy and HREE concentrations which are subequal in concentration to LREE in Mahagiri samples. In

the REE-plots they were interpolated from chondrite-normalized abundances of the neighbouring elements.

5.2. Results

5.2.1. Petrography and mineralogy of the Mahagiri Quartzite

The radioactive conglomerates are typically matrix-supported, poorly sorted cobble-pebble conglomerate with wide range of roundness in the framework grains set in a sand-mud matrix. Petrographic studies revealed that the framework grains are constituted by both monocrystalline as well as polycrystalline grains of predominantly well-rounded to subangular vein quartz, subangular to angular clasts of chert of various colours. The fine sericite-muscovite flakes in the matrix of sandstone represent altered feldspars derived from granitic/acidic sources and fuchsite and chlorite are likely to be contributed from basic/ultrabasic parentage.

The effect of deformation is mild in the uraniferous conglomerate and the associated coarse sandstone of Mahagiri Quartzite. They are manifested by undulose extinction and crude orientation in a few thin sections. Pressure-solved boundaries of the quartz grains in coarser clastics are suggestive of post-depositional diagenetic alteration. However, some of the pebbles of vein quartz and cherts show effect of strong deformation representing structurally deformed source rocks. The mineral assemblages of the sandstone matrix have chlorite as the main metamorphic mineral and are characterized by the absence of any other metamorphic minerals of higher metamorphic grade. This fact suggests a low greenschist grade of metamorphism for the Mahagiri Quartzites.

Heavy minerals contain variable amount of rounded to sub-rounded pyrite, uraninite, zircon, monazite, chromite, rutile, Th-silicates and U-phosphatic phases with epigenetic galena, minor chalcopyrite, and possibly authigenic U-Ti phases. Carbonaceous matter locally observed in the matrix.

Table 1

EPMA composition of detrital uraninite from Mahagiri Quartzite (detailed methodology and standards used in Appendix A).

Sample	DT-391-11-11	DT-391-11-11	DT-391-11-11	DT-391-11	DT-391-11	DT-391-11	DT-391-11-11	DT-391-8/1	DT-1-79-2-11-11	DT-1-79-2-11-11	DT-1-79 4/1	DT-1-79-2-11-11	DT 1-79 6/1	DT-1-79
Phase	Uraninite	Uraninite	Uraninite	Uraninite	Uraninite	Uraninite	Uraninite (Oxd?) Fourmarierite	Uraninite	Uraninite (Oxd?)	Uraninite	Uraninite	Uraninite	Uraninite (Metamict?)	Uraninite (Metamict?)
Analyses	1/1	2/1	3/1	4/1	5/1	6/1	7/1	8/1	1/1	3/1	4/1	5/1	6/1	7/1
UO ₂	63.615	72.431	66.329	61.85	67.38	68.655	67.862	75.451	67.878	69.194	63.371	69.66	61.459	57.264
ThO ₂	9.080	6.701	6.931	11.708	7.292	7.372	4.401	5.267	7.65	4.24	8.115	7.59	4.107	7.801
SiO ₂	2.096	1.366	1.321	1.026	0.724	0.522	0.807	0.73	0.419	0.26	0.514	0.70	0.158	0.468
TiO ₂	0.033	0.048	0.03	0.017	0.026	0.025	0.042	0.094	0.282	0.237	0.384	0.20	0.196	0.377
ZrO ₂	0.096	0.000	0.023	0.061	0.028	0.017	0.083	0.078	0.198	0.358	0.369	0.22	0.323	0.536
PbO	12.045	8.889	14.908	15.066	11.804	14.016	12.577	9.224	6.795	9.151	7.617	10.02	10.02	8.039
MnO	0.081	0.102	0.039	0.014	0	0.004	0.009	0	0.193	0	0	0.00	0	0
FeO	0.316	0.397	0.372	0.187	0.131	0.22	0.211	0.341	0.547	0.306	0.451	0.19	0.513	0.138
MgO	0.032	0.016	0.034	0.035	0	0.018	0.027	0.013	0.128	0.052	0.073	0.12	0	0.077
CaO	1.200	1.230	1.13	0.982	1.32	1.055	1.216	0.991	1.323	1.012	2.106	1.47	2.013	3.014
SrO	0.000	0.000	0	0	0	0	0	0	0.132	0.093	0.382	0.00	0.32	0.619
Na ₂ O	0.043	0.049	0.007	0.169	0.054	0.06	0.031	0.031	0.17	0.208	0.227	0.26	0.165	0.273
K ₂ O	0.180	0.181	0.18	0.253	0.163	0.155	0.209	0.202	0.212	0.199	0.21	0.22	0.182	0.198
Al ₂ O ₃	0.116	0.109	0.157	0.185	0.031	0.065	0.07	0.076	2.966	1.132	3.276	0.06	5.013	7.16
P ₂ O ₅	0.874	0.339	0.141	0.138	0.017	0.014	0.083	0	1.892	2.743	5.04	2.21	5.012	6.019
SO ₃	2.164	1.372	2.073	2.092	4.126	2.167	1.477	1.026	1.13	0.215	0.268	0.40	0.418	0.409
Y ₂ O ₃	1.116	1.225	0.758	0.996	0.845	0.832	1.493	1.289	1.586	3.27	1.637	1.18	2.894	1.654
La ₂ O ₃	0.000	0.000	0	0	0	0	0	0	0	0.024	0.002	0.00	0.011	0.027
Ce ₂ O ₃	0.173	0.184	0.049	0.07	0.038	0.034	0.143	0.148	0.172	0.302	0.219	0.13	0.305	0.386
Pr ₂ O ₃	0.408	0.300	0.387	0.464	0.37	0.337	0.319	0.036	0.41	0.384	0.382	0.36	0.26	0.34
Nd ₂ O ₃	0.242	0.398	0.108	0.1	0.088	0.133	0.207	0.228	0.277	0.309	0.181	0.28	0.421	0.156
Sm ₂ O ₃	0.187	0.199	0.13	0.052	0	0.101	0.117	0.12	0.142	0.168	0.128	0.00	0.217	0.115
Eu ₂ O ₃	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Gd ₂ O ₃	0.533	0.054	0.342	0.357	0.207	0.281	0.661	0.206	0.27	0.437	0.41	0.37	0.593	0.327
Tb ₂ O ₃	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Dy ₂ O ₃	0.29	0.36	0.26	0.229	0.126	0.175	0.389	0.36	0.42	0.70	0.416	0.35	0.667	0.277
Ho ₂ O ₃	0.00	0.20	0.11	0.007	0	0	0.021	0.00	0.02	0.20	0	0.00	0.131	0
Er ₂ O ₃	0.21	0.21	0.06	0.143	0.181	0.135	0.252	0.19	0.24	0.52	0.246	0.21	0.436	0.284
Tm ₂ O ₃	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Yb ₂ O ₃	0.13	0.16	0.04	0.022	0.125	0.155	0.06	0.06	0.17	0.37	0.2	0.18	0.4	0.162
Lu ₂ O ₃	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd

Total	95.25	96.52	95.92	96.22	95.08	96.55	92.77	96.17	95.62	96.09	96.22	96.37	96.23	96.12
U (total)	0.634	0.703	0.713	0.675	0.688	0.746	0.739	0.771	0.568	0.632	0.496	0.672	0.478	0.395
<i>Number of cations based on 1 cation per formula unit</i>														
Th	0.075	0.058	0.060	0.101	0.063	0.066	0.040	0.048	0.058	0.034	0.057	0.064	0.027	0.047
Si	0.076	0.052	0.050	0.039	0.027	0.020	0.032	0.029	0.014	0.009	0.016	0.026	0.005	0.012
Ti	0.001	0.001	0.001	0.000	0.001	0.001	0.001	0.003	0.007	0.006	0.009	0.006	0.004	0.008
Zr	0.002	0.000	0.000	0.001	0.001	0.000	0.002	0.002	0.003	0.006	0.006	0.004	0.005	0.007
Mn	0.003	0.003	0.001	0.000	0.000	0.000	0.000	0.000	0.005	0.000	0.000	0.000	0.000	0.000
Fe	0.010	0.013	0.012	0.006	0.004	0.007	0.007	0.011	0.015	0.009	0.012	0.006	0.013	0.003
Mg	0.002	0.001	0.002	0.002	0.000	0.001	0.002	0.001	0.006	0.003	0.003	0.007	0.000	0.003
Ca	0.047	0.050	0.046	0.040	0.054	0.044	0.052	0.042	0.048	0.038	0.069	0.058	0.063	0.086
Sr	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.003	0.002	0.007	0.000	0.005	0.010
Na	0.003	0.004	0.001	0.012	0.004	0.005	0.002	0.002	0.011	0.014	0.014	0.019	0.009	0.014
K	0.008	0.009	0.009	0.012	0.008	0.008	0.011	0.010	0.009	0.009	0.008	0.010	0.007	0.007
Al	0.005	0.005	0.007	0.008	0.001	0.003	0.003	0.004	0.117	0.047	0.118	0.003	0.173	0.224
P	0.027	0.011	0.005	0.004	0.001	0.000	0.003	0.000	0.054	0.082	0.131	0.069	0.124	0.135
S	0.059	0.039	0.059	0.059	0.117	0.064	0.044	0.031	0.028	0.006	0.006	0.011	0.009	0.008
Y	0.022	0.025	0.015	0.020	0.017	0.017	0.032	0.027	0.028	0.062	0.027	0.023	0.045	0.023
La	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ce	0.002	0.003	0.001	0.001	0.001	0.000	0.002	0.002	0.002	0.004	0.002	0.002	0.003	0.004
Pr	0.005	0.004	0.005	0.006	0.005	0.005	0.005	0.001	0.005	0.005	0.004	0.005	0.003	0.003
Nd	0.003	0.005	0.001	0.001	0.001	0.002	0.003	0.003	0.003	0.004	0.002	0.004	0.004	0.001
Sm	0.002	0.003	0.002	0.001	0.000	0.001	0.002	0.002	0.002	0.002	0.001	0.000	0.002	0.001
Eu	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Gd	0.006	0.001	0.004	0.004	0.003	0.004	0.009	0.003	0.003	0.005	0.004	0.005	0.006	0.003
Tb	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Dy	0.003	0.004	0.003	0.003	0.002	0.002	0.005	0.005	0.005	0.008	0.004	0.004	0.006	0.002
Ho	0.000	0.002	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.002	0.000	0.000	0.001	0.000
Er	0.002	0.002	0.001	0.002	0.002	0.002	0.003	0.002	0.003	0.006	0.002	0.002	0.004	0.002
Tm	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Yb	0.001	0.002	0.000	0.000	0.001	0.002	0.001	0.001	0.002	0.004	0.002	0.002	0.004	0.001
Lu	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Total	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000
<i>Calculated U^{4+}, U^{6+}, and number of oxygen in formula unit (based on charge balance method, after Janeczek and Ewing, 1992)</i>														
U^{4+}	0.330	0.473	0.577	0.587	0.340	0.670	0.049	0.486	0.196	0.367	0.245	0.460	0.296	0.177
U^{6+}	0.304	0.230	0.137	0.088	0.348	0.075	0.690	0.285	0.372	0.265	0.251	0.211	0.182	0.218
ΣO	2.272	2.161	2.102	2.041	2.374	2.049	2.623	2.217	2.235	2.150	2.115	2.118	2.023	2.029

nd: not detected.

Uraninite is the most important primary U-bearing phase in the Mahagiri QPC. Other U-bearing phases present in the conglomerate include altered U–Th and U–Ti phases in subordinate proportions. Uraninite occurs as dispersed grains oval to rounded in outline (Fig. 4A, B and C). Subrounded uraninites show broken outline and abraded corners and edges providing testimony of mechanical transport (Fig. 4D). The inner part of the grains often shows pitted appearance and is partially dissolved and replaced by sulphides (e.g. pyrite and galena) (Fig. 4E). Minor amount of coffinite is also present as microcrystalline, relatively heterogeneous phase in such altered uraninite grains and in the matrix (Fig. 4B). Late oxidative alteration resulted goethitisation of the pyrite grains and are often mildly radioactive due to adsorption of uranium liberated from the altered uraninite grains and other U-bearing phases (Fig. 4F). Textural evidence suggests that the galena is mostly secondary in origin and assumed authigenic blebs, stringers or drop-like shapes within as well as along the periphery of the rounded uraninite grains (Fig. 4D and 5A). They were most likely to have been formed by sulphidisation of radiogenic lead. Pyrite grains also occur as sub-angular to well-rounded grains with abraded corners (Fig. 4A) suggesting their mechanical abrasion during transport. However, locally euhedral overgrowth of pyrite surrounding the relict rounded outline of the original detrital pyrite (Fig. 4C) bears the testimony of post-depositional sulphidisation.

5.2.2. Chemical composition of Mahagiri uraninite

A number of uranium bearing mineral phases have been identified from observations using both optical microscope and SEM–BSE analyses. EPMA studies carried out on Mahagiri uraninite provided a comprehensive mineral chemical composition (Table 1). The total elemental oxide content varied from 92.77 to 96.55 wt.%. Analytical data falling short of 100% may have reasons such as 1) deviation from ideal UO_2 formula in common uraninite, displaying heterovalent cationic substitution (Janeczek and Ewing, 1992), 2) higher concentration of some other elements which were not been analysed and 3) alteration and metamictization of the uraninite grains resulting in hydration of uraninite, where H_2O cannot be analysed in EPMA.

5.2.3. Major elemental distribution of Mahagiri uraninite

Electron microprobe analyses of fourteen grains of uraninites (about 32 spot analyses) from six selected samples are presented here. They show UO_2 -content varying from 57.26 to 75.45 wt.% (average 66.60 wt.%), ThO_2 -content varying from 4.10 to 11.71 wt.% (av. 7.02 wt.%) and PbO varying from 6.80 to 15.07 wt.% (av. 10.73 wt.%) (Table 1). Other elements include SiO_2 (0.16–2.09%, av. 0.79), Al_2O_3 (range: 0.03–7.16 wt.%; av. 1.46 wt.%), CaO (range: 0.98–3.01 wt.%, av. 1.43 wt.%), P_2O_5 (Traces to 6.02 wt.%, av. 1.75), MgO (traces from 0.13 wt.%, av. 0.04), and K_2O (0.16–0.25 wt.%; av. 0.19). The other important trace elements in the uraninite include Zr (traces to 0.54 wt.%; av. 0.17 wt.%) and Sr (traces to 0.62 wt.%; av. 0.13 wt.%).

EPMA elemental mapping carried out in one of the uraninite grains (Fig. 5) shows even distribution of U throughout the grain (Fig. 5C) and also an overgrowth on an originally euhedral U–Th-bearing zoned mineral phase at the core (Fig. 5B). The Y map also simulates the outline of the Th-bearing original nucleus suggesting that the Th-phase was also rich in Y (Fig. 5D) whereas concentration of Pb along metamict alterations (Fig. 5A).

As described above the Mahagiri uraninite characteristically contains highly elevated Th. This gives here an opportunity to test variation of a minor or trace element (thorium) in a matrix of a major, relatively invariant element (uranium). UO_2/ThO_2 ratio versus ThO_2 plot (Fig. 6) shows UO_2/ThO_2 ratio as a function of ThO_2 concentration and depicts a hyperbolic trend (Fig. 6). Such hyperbolic relationship in the variation of a minor or trace element in a matrix of a major and relatively invariant element is likely to arise during crystallization from a melt and does not support U-compositions from minute inclusions of galena or silicate within the uraninite grains. In such cases, those data would have more

likely to be plotted in a mixing line with a constant UO_2/ThO_2 value equal to that of the original uraninite grain making a high angle to the expected hyperbolic line (cf. Grandstaff, 1981). Similar hyperbolic trends were also reported in detrital uraninite samples from Witwatersrand as well as Blind River areas (Grandstaff, 1980).

5.2.4. REE and Y composition of Mahagiri uraninite

Mahagiri uraninites show REE-oxide concentration ranging between 0.98 and 2.98 wt.% (av. 1.75 wt.%) and Y-oxide content varies from 0.96 % to 2.98% (Table 1). Chondrite-normalized (Anders and Grevesse, 1989) REE pattern of Mahagiri uraninite shows broadly flat REE pattern with slight LREE depletion and prominent negative La-anomaly and Pr-enrichment (Fig. 7A). As the Eu content of the uraninite could not be estimated due to interference of spectral lines with those of Pr and Nd, a possible moderate negative Eu-anomaly is expected. The uraninite grains show significant uniformity in composition in the total REE vs. LREE/HREE plot (Fig. 7B) and total REE vs. U/Th field (Fig. 7C).

5.2.5. Chemical formula of Mahagiri uraninite

Naturally occurring uraninite can be expressed as $\text{UO}_2 + x$, where x is the amount of U^{4+} converted to U^{6+} due to auto-oxidation. Thus the EPMA data of the Mahagiri uraninite was recalculated for U^{4+} and U^{6+} (Table 1, detailed calculation principle in Appendix A after Janeczek and Ewing, 1992). Impurities in these uraninite grains include Si, Al, Zr, Th, Y, REE, and radiogenic Pb. Thus, more elaborate structural formula for natural uraninite can be expressed as $[\text{U}^{4+}_{1-x-y-z-u} \text{U}^{6+}_x (\text{Th}^{4+})_u \text{REE}^{3+}_y \text{M}^{2+}_z \text{O}_2 + x - (0.5y) - z]$, displaying heterovalent cationic substitution, where M represents bivalent cations, mainly constituting Pb^{2+} and Ca^{2+} , REE^{3+} incorporates Al^{3+} and Th^{4+} incorporates Zr^{4+} . The presence of H_2O and other volatiles in associated uranyl phases due to uraninite alteration complicates the formula calculation (cf. Zhao and Ewing, 2000).

The U contents (50.48–65.63 wt.% U) of the uraninite grains from Mahagiri are much lower than the ideal composition of UO_2 (i.e., 88.15% U). Ca content is elevated (up to 3.01 wt.%) with some amount of Al (average 1.46 wt.% Al_2O_3), Mg (traces from 0.13 wt.% MgO) and K (0.16–0.25 wt.% K_2O).

The data reveal that uraninite grains are rich in Th, Y, and REE with a very narrow range of Y/REE [apfu] ratio (0.60–1.55) averaging around 1.00 ($n = 14$). Two groups of uraninite, one with Y/REE ratios averaging <1 (range: 0.60–0.97 with av. 0.60) and another with ratio ≥ 1 (range 1.00–1.55 with av. 1.23) could be distinguished. When the $\text{U}^{4+}:\text{U}^{6+}$ [apfu] ratios are plotted against the Y/ ΣREE [apfu] ratio, two distinct fields emerge (Fig. 8A).

The calculated amount of oxygen [apfu] vs. Ca [apfu] for Mahagiri uraninites are plotted with stability fields of $\text{UO}_2 + x$, at $x = 0, 0.25, 0.33$, and 0.66 (Fig. 8B). Majority of these data plot in the stability field of cubic uraninite ranging from 2.023 to 2.217 of oxygen [apfu]; only two points deviate significantly from cubic field (2.272, and 2.623 oxygen [apfu]).

6. Discussion

6.1. Depositional setting and control on uranium mineralisation

The lower half of the Mahagiri Quartzite that hosts a number of uraniferous conglomerate and coarse-grained sandstone/pebbly sandstone beds includes texturally and compositionally less mature, matrix-rich sandstones interbedded with conglomerates. The rounded-to-subrounded and angular outline of the uraninite grains together with their dispersed distribution in the conglomerate and sandstone (Fig. 4A, B and C) strongly suggest their detrital nature. Subrounded pyrite grains with abraded corners and edges also provide testimony of mechanical transport. Detrital uraninites distinguish the Mahagiri uraniferous QPCs as a typical example of paleoplacer deposit.

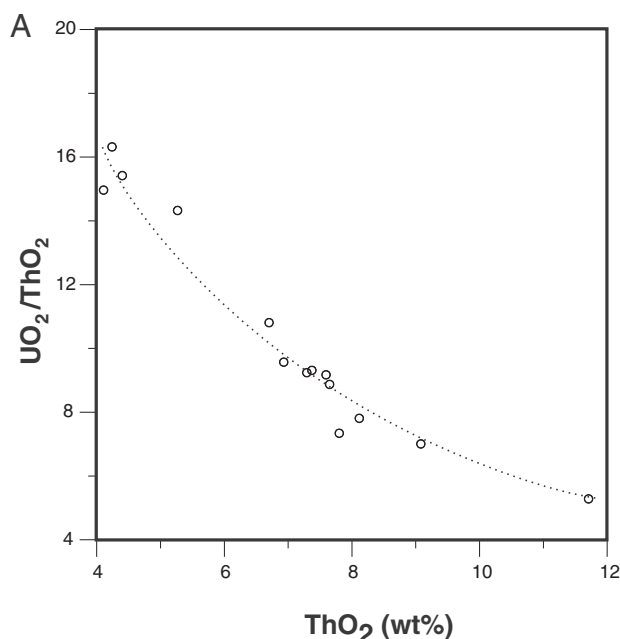


Fig. 6. UO_2/ThO_2 vs. UO_2 variation from EPMA composition of uraninite. The hyperbolic relationship shown by the majority of the grains in the variation of a minor or trace element (Th) in a matrix of a major and relatively invariant element (U) is likely to suggest compositional variation during crystallization from melt and primary magmatic nature of the uraninite.

Matrix-supported conglomerate and matrix rich pebbly sandstone in the lower part of the Mahagiri Quartzite with non-graded character and lack of internal structure suggest that these conglomerates are products of cohesive debris-flows (Nemec and Steel, 1988). The sandstones with floating pebbles are likely to be deposited from sheet floods. The matrix-rich debris flow as well as sheet flood deposits mainly develop in high-gradient subaerial fans (alluvial fan) and associated braided fluvial systems. Cross-stratified mature sandstone that constitutes the upper half of the Mahagiri Quartzite is distinguished by their extremely homogeneous composition and strikingly monotonous nature. Cross-stratified sandstone interbedded with the debris-flow conglomerate and sheet-flood deposits are the fluid-gravity flow bedload counterparts in the braid plains that develop downstream of an alluvial fan system (Blair and McPherson, 1994). Wavy geometry of these sandbodies, megaripled top, and supermature nature of the arenites in the upper half of the succession, on the other hand, suggests that they have been deposited in a shallow-marine shelf with strong wave-current reworking. Local reversal of paleocurrent direction of cross-strata further supports deposition in shallow shelf with tidal currents. The low-amplitude high-wave length sandbody geometry indicates that these sands accumulated as aggrading to prograding sandwave complexes in shoreface to foreshore setting.

A composite stratigraphic log of the Mahagiri Quartzite from both the limbs of the Mahagiri synform suggests a subaerial alluvial fan to shallow marine shelf through a transitional braided channel-delta depositional system (Fig. 9A, B, C). The transition from subaerial fan/braid plain depositional settings to shallow marine shelf suggests an overall condition of base level rise (Catuneanu, 2002). The uraniferous lithologies that are confined to the lower part of the succession occur within the alluvial fan-braid delta depositional setting (Fig. 9A, B). The coarsest of all the sedimentary lithologies, the matrix-supported cobbly/bouldery conglomerate of cohesive debris flow origin represent the proximal alluvial fan deposits. The more sheet-like uraniferous pebbly sandstone and uraniferous trough-cross-stratified sandstone are likely to represent subaerial braided-plains farther downstream of the fan. Evidences of wave-reworking on top of such beds point to a coastal

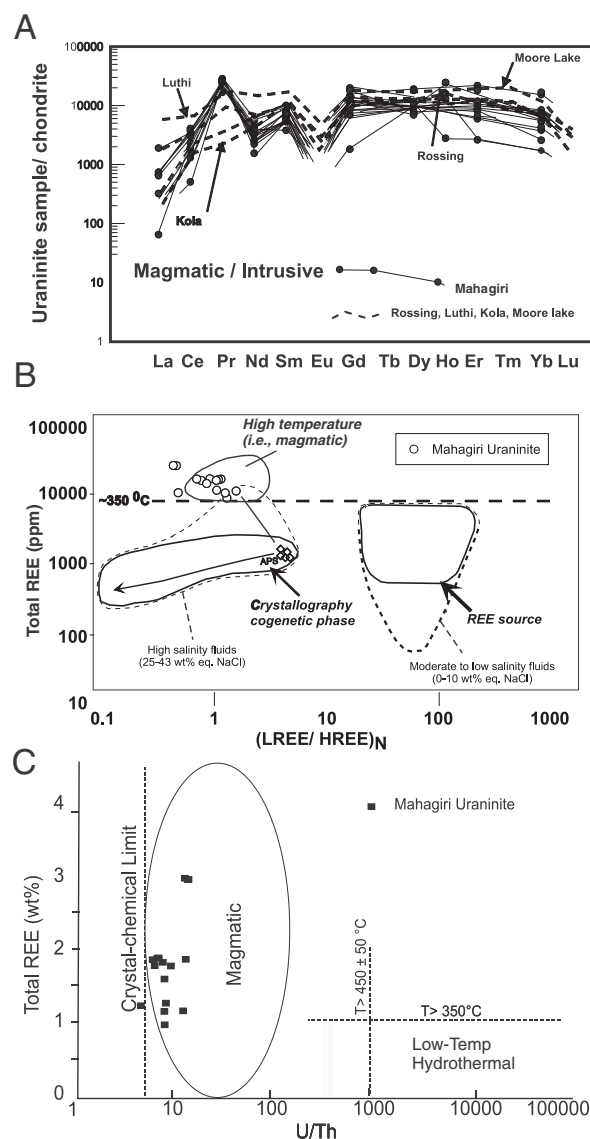


Fig. 7. REE plots of uraninites from EPMA analyses: A) chondrite normalized REE pattern of Mahagiri uraninites. Note overall flat pattern with possible Eu anomaly. Plots for other deposits are adopted from Mercadier et al. (2011). B) Total REE vs. $(\text{LREE}/\text{HREE})_{\text{N}}$ plot. Note the Mahagiri uraninite compositions are clustered with the high-temperature magmatic uraninite field. C) Total REE vs. U/Th plot with Mahagiri uraninite clustering within the magmatic uraninite field (fields after Frimmel et al., 2014).

braid plain setting (Fig. 9C). The trough cross-stratified mature Quartz arenites that constitute the upper part of the Mahagiri Quartzite were deposited with the base level rise in a wave-/tide-dominated high-energy shelf setting. The development of number of centres of such alluvial fan/braid plain associations along strike indicates that the uraniferous conglomerates and sandstones are developed as coastal alluvial cones at the base of a major transgressive sequence (Fig. 9B).

6.2. Source of detrital uraninite

6.2.1. Pb-content in Mahagiri uraninite

High amount of Pb content in the Mahagiri uraninite points to their old age (Archean) of crystallization. Wide variability in Pb-content in different uraninite grains indicates their derivation from source rocks of different ages. However, some amount of variability together with weak negative correlation between U and Pb may attribute to a component of Pb-loss of the Mahagiri uraninite. Mahagiri uraninite shows S-

rich character similar to the detrital uraninite grains (up to 2.49 wt.% S) reported from the classical Mesoarchean Witwatersrand QPC deposits (Depiné et al., 2013). Inclusions and exsolution of radiogenic Pb and subsequent formation of galena in the presence of high S-rich environment have been reported by many authors from Precambrian QPC occurrences of the world (Grandstaff, 1980; Depiné et al., 2013).

6.2.2. Th-content in Mahagiri uraninite

Th-content is the function of temperature of crystallization of uraninite. The insoluble nature of Th in low- to medium-temperature hydrothermal processes is the reason for its absence or extremely low (≤ 0.01 wt.%) concentration in epigenetic uranium mineralization (Grandstaff, 1981; Oberthür and Saager, 1986; Roy and Roy, 2008; Mercadier et al., 2011; Depiné et al., 2013; Cuney, 2010b; Cuney, 2013; Frimmel et al., 2014). Thus, high (>4 wt.%) and the widely

variable Th-content of Mahagiri uraninite suggest their derivation from a high temperature sources and multiple source rocks with different conditions of uraninite crystallization (Fig. 7). Similar variable and elevated Th content in detrital uraninite of Witwatersrand QPC ($\text{ThO}_2 > 4.0$ wt.%) is attributed to the magmatic origin, i.e., granitic/pegmatitic provenance (Grandstaff, 1981; Oberthür and Saager, 1986; Frimmel et al., 2014). With this analogy, the very fact of the presence of high temperature uraninite in a low grade metamorphic siliciclastic rock strongly places Mahagiri uraninite as detrital and further, precludes a hydrothermal or syngenetic uraninite mineralization in the Mahagiri quartzite. Th/U ratio of uraninite is another important parameter that helps in distinguishing low-temperature epithermal uraninite from high temperature mesothermal or pegmatitic uraninite and from main stage magmatic uraninite (Frimmel et al., 2013, 2014). All low-temperature hydrothermal uraninite typically display Th/U ratio < 0.001 , whereas those formed at a higher temperature, i.e. above 450 ± 50 °C, exhibit Th/U ratios much greater than 0.001, when derived from a granite-hosted magmatic uraninite. Th/U ratio less than 0.001 is typically absent in Mahagiri uraninite. They display Th/U ratios comparable with those reported from detrital uraninite of Witwatersrand [Th/U = 0.02–0.019 (Frimmel, 2014) and 0.1 to 0.5 (Reimer, 1987; Rantzsche et al., 2011; Depiné et al., 2013)]. The range of Th-content of Mahagiri uraninite reflects their crystallization from a possible peraluminous melt that generally crystallizes uraninite with 1–12 wt.% ThO_2 (Cuney and Friedrich, 1987). Interestingly, Th/U ratio of Mahagiri uraninite grains largely overlaps the range of those of pre-2.2 Ga granite and granitic pegmatoid (e.g., Witwatersrand Granite with Th/U ratio of 0.15–3.7) (Cuney, 2010b). Similar high Th/U ratio of Mahagiri uraninite grains in conglomerate matrix suggests their derivation from high temperature hydrothermal, pegmatitic or granitic sources.

6.2.3. REE-Y signature in Mahagiri uraninite

The fractionation of REEs in uranium oxide phases is less sensitive to redox changes. The ionic radii of REEs (1.16–0.977 Å) are close to that of U^{4+} (1 Å) in eightfold coordination in uraninite structure (Shannon, 1976), which allows them to easily substitute for U with only minor modification by post-crystallization episodes (Janeczek and Ewing, 1992). Pioneer work on REE contents in uranium oxides was carried out by Fryer and Taylor (1987) and Pagel et al. (1987), and more recently by Mercadier et al. (2011), Depiné et al. (2013), Frimmel (2014), Frimmel et al. (2014). Mahagiri uraninite shows wide range of REE content and comparable concentration of LREE and HREE (ΣLREE : 1580–15150 ppm; av. 8549 ppm and ΣHREE : 2640–24890 ppm; av. 11645 ppm) (Table 1). The flat REE pattern for uraninite (Fig. 7A) is believed to indicate crystallization at a temperature corresponding to their magmatic origin (Mercadier et al., 2011). At that high temperature (> 350 °C) all the REEs in sufficient quantity are likely to get incorporated from magmatic residual melt into uraninite structure without any preference and thereby attains $\Sigma\text{LREE}/\Sigma\text{HREE}$ ratio of around 1, giving rise to a flat chondrite-normalized REE pattern. A pronounced negative Eu-anomaly is also observed in REE pattern of magmatic uraninite because of early fractionation of plagioclase in the silicate melt for intrusive deposits (Cuney, 2010b). Below 350 °C, the total REE content decreases and a fractionation occurs between REEs due to the crystallographic control of the mineral structure, with preferential incorporation of REEs having ionic radii close to that of U^{4+} (Tb–Er), leading to an enrichment of MREE compared to LREE and HREE that results in a bell shaped REE pattern centred on Tb to Er. A possible negative Eu anomaly is visible for some deposit types (synmetamorphic and vein-type for example) because of possible reducing conditions prevalent during the precipitation such uranium oxides (Mercadier et al., 2011). However, processes other than pure crystallographic control were also responsible for hydrothermal U-deposits. In unconformity-related deposits, LREE addition into uraninite structure is reduced by the crystallization of cogenetic LREE-rich Al–Phosphate–Sulphate (APS) minerals which preferentially partition the LREE into the APS minerals

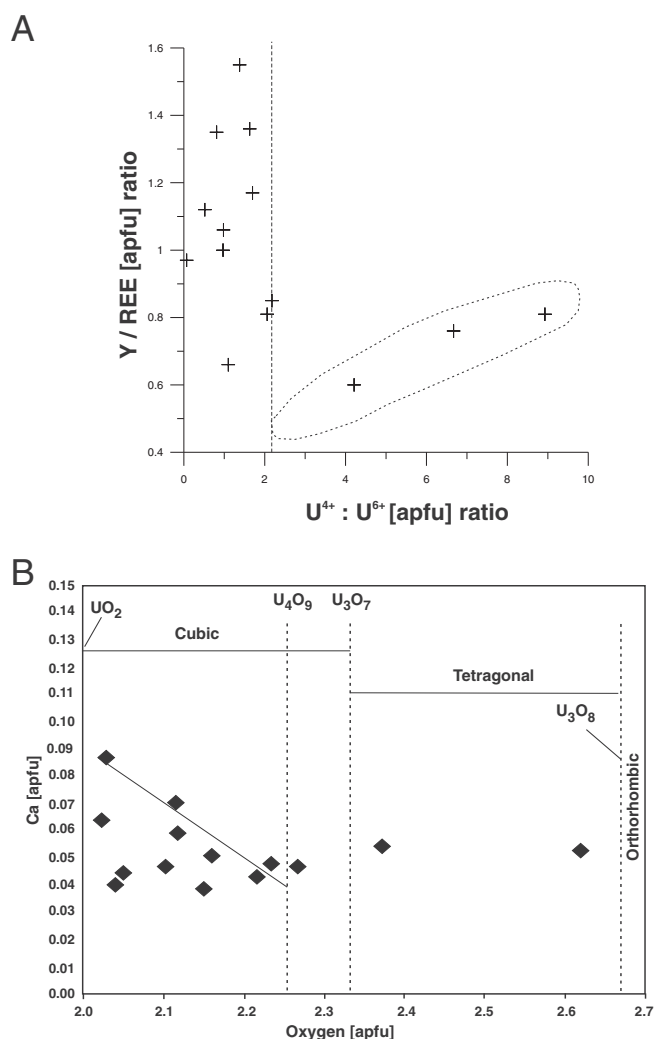


Fig. 8. Elemental composition plots based on chemical formula calculations (based on Janeczek and Ewing, 1992). A) $\text{U}^{4+}/\text{U}^{6+}$ [apfu] ratios vs. Y/REE (wt.%) plots showing two distinct fields, where the less oxidized (with $\text{U}^{4+}/\text{U}^{6+}$ [apfu] ratio $> \sim 2$) uraninite shows distinct positive correlation with the Y/ ΣREE [apfu] ratios, while the highly oxidized uraninite (with $\text{U}^{4+}/\text{U}^{6+}$ [apfu] ratio $< \sim 2$) displays very wide range of Y/ ΣREE [apfu] ratios (0.66–1.55). See text for explanation. B) Calculated amount of oxygen [apfu] vs. Ca [apfu] for Mahagiri uraninite are plotted with stability fields of $\text{UO}_2 + x$, at $x = 0, 0.25, 0.33$, and 0.66 (fields after Deditius et al., 2007). Majority of these data plot in the stability field of cubic uraninite ranging from 2.023 to 2.217 of oxygen [apfu]; only two points deviate significantly from cubic field (2.272 and 2.623 oxygen [apfu]).

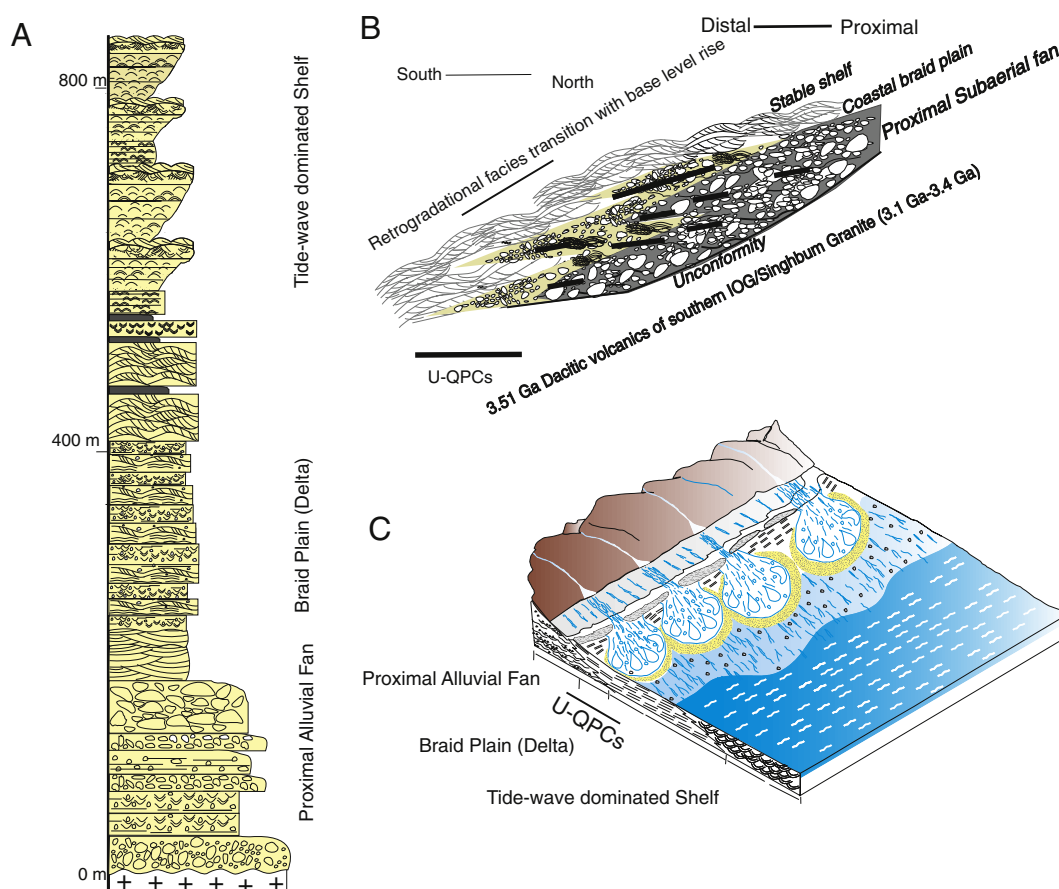


Fig. 9. Depositional setting of the Mahagiri Quartzite. A) Composite litholog (along XY in Fig. 2A) showing different depositional facies associations from proximal alluvial fan to wave-tide dominated shelf, B) schematic model for facies transition in response to base level rise in section from the southern limb of the Mahagiri synform, C) schematic model for the depositional setting of the Mahagiri Quartzite. The uraniferous paleoplacers occur mainly in coastal braid plain and proximal fan lobes.

resulting in a LREE-depleted asymmetric bell shaped REE pattern (Gaboreau et al., 2007; Mercadier et al., 2011). Uraninites from vein type and sandstone type deposits strongly simulate the REE patterns of the source rock despite their different geological and geochemical mechanism of formation of uranium deposits (Mercadier et al., 2011). Flat REE pattern of the detrital uraninites from Mahagiri area (Fig. 7A) and their clustering around (LREE/HREE)_N ratio of 1 (Fig. 7B) in the high temperature (>350 °C) field confirms their high temperature magmatogenic granitic source. La depletion in REE pattern of the Mahagiri uraninites possibly may be the result of two reasons. Firstly, the average ionic radius of U⁴⁺ (1.00 Å) is closer to those of HREE (Gd³⁺: 1.053 Å to Lu³⁺: 0.977 Å) than LREE with largest being La³⁺ (La³⁺: 1.16 Å) and, hence, even with minor lowering of temperature in magmatic uraninite Lanthanum will be the first among the REEs to be comparatively least incorporated in uraninite structure, and secondly, La has possibility to be preferentially admitted in the low temperature LREE-rich authigenic monazite and Ca–U–Th bearing phosphatic phases developed in QPC matrix in post-depositional stages (Cuney and Mathieu, 2000).

This approach of characterizing the different types of uranium deposits based on the typical REE patterns of uraninite have a direct bearing to resolve the genetic controversies that exist for QPC-hosted mineralization in oxygen deficient environment. Disagreement exists between detrital (sedimentary placer) and hydrothermal origin of metals (Maier et al., 2009). If the uranium-bearing minerals present in this sedimentary basin are the result of mechanical disintegration of

Archean magmatic source, then the uraninites must bear the signature of a flat REE pattern with negative Eu-anomaly, typical of magmatic high temperature uraninites (e.g., Mercadier et al., 2011; Depiné et al., 2013; Frimmel et al., 2014).

The total REE and Y contents in uraninite is believed to be the function of temperature of formation and low temperature hydrothermal uraninite has been suggested to contain <1 wt.% ΣREE (e.g., Fryer and Taylor, 1987; Pagel et al., 1987; Mercadier et al., 2011; Depiné et al., 2013; Frimmel et al., 2014). REE concentration together with the U/Th ratio is a better proxy for estimating temperature of uraninite formation (Frimmel et al., 2014). In ΣREE vs. U/Th binary diagram (Fig. 7C) Mahagiri uraninite typically plots within the magmatic uraninite field formed at >450 °C temperature and cluster distinctively away from the field of low-temperature hydrothermal uraninite (Fig. 7C) (after Mercadier et al., 2011 and fields after Frimmel et al., 2014). The plots further cluster within the limits of high temperature ultra-metamorphic and magmatic uraninites set to have U/Th ratio <10³ and REE more than 0.12 wt.% (Frimmel et al., 2014).

6.2.4. Significance of the chemical formula of Mahagiri uraninite

The U contents of the detrital uraninite grains from Mahagiri are significantly lower than the ideal composition of UO₂. This may be mainly due to higher amount of heterovalent cationic substitution by Th, REE, Y, Pb, and Ca in Mahagiri uraninite structure (cf. Janeczek and Ewing, 1992), and partial alteration and metamictization of uraninites. Alteration due to metamictization resulted in elevated

concentration of Si, Al, P, and Ca, as they are found to be higher in more altered and metamict grains compared to the less altered uraninite grains. In altered Mahagiri uraninite Si content is slightly elevated (up to 2.2 wt.%) likely due to coffinitisation. However, ionic radius of Si^{4+} (0.26 Å) is too small to be accommodated in the uraninite structure (Frimmel, 2014). Elevated Ca content partially contributes together with some amount of Al, Mg, and K to the low temperature post-depositional alteration of uraninite and elemental redistribution among the rock-forming and ore minerals. Mahagiri uraninites, however, accommodated some amount of Ca in uraninite (fluorite) structure presumably due to similarity in ionic radii of Ca^{2+} (1.06 Å) and U^{4+} (1 Å).

Despite very narrow range of Y/REE [apfu] ratio (0.60–1.55) averaging around 1.00, two groups of uraninites, one with Y/REE cation ratio averaging < 1 and another with ratios ≥ 1 suggest that they are derived from different uraninite populations (Fig. 8A) that in turn support their detrital nature (cf. Frimmel et al., 2013, 2014). However, their overall average around 1 possibly suggests their non-preferential incorporation in uraninite crystals at a higher temperature, preferably in the main stage magmatic (granitic) temperature.

The $\text{U}^{4+}:\text{U}^{6+}$ [apfu] ratios of Mahagiri detrital uraninite again define two groups: a) less oxidized uraninite and b) more oxidized uraninite. When the $\text{U}^{4+}:\text{U}^{6+}$ [apfu] ratios are plotted against the Y/ ΣREE [apfu] ratio, two distinct fields emerge; where the less oxidized uraninite grains show distinct positive correlation with the Y/ ΣREE [apfu] ratios within a limiting value of < 1 . On the other hand, the population of the more oxidized uraninite (with $\text{U}^{4+}:\text{U}^{6+}$ [apfu] ratio < 2) displays very wide range of Y/ ΣREE [apfu] ratios (0.66–1.55) (Fig. 8A). This relationship suggests that Y was preferentially incorporated at lower oxygen fugacity stage of uraninite crystallization which further indicates that Y- is incorporated at early stage of uraninite crystallization at higher temperature. The more oxidized uraninite compositions with lesser degree of Y-incorporation are likely to suggest that uraninite crystallization must have continued at a higher oxygen fugacity with later stage of magmatic crystallization. The elemental map of one Mahagiri uraninite (Fig. 5) clearly depicts such early Y fractionation that took place during early phase of Th-rich core formation from a high temperature melt.

Mahagiri uraninite accommodated some amount of Ca in uraninite (fluorite) structure as expected due to similarity in ionic radii of Ca^{2+} (1.06 Å) and U^{4+} (1 Å). The oxygen [apfu] vs. Ca [apfu] plots for Mahagiri uraninites (Fig. 8B) display major grouping in the stability field of cubic uraninite ranging from 2.023 to 2.217 of oxygen [apfu] excepting for two samples that deviate significantly from cubic field (2.272, and 2.623 oxygen [apfu]). Such deviation may arise because of alteration of primary uraninite into uranyl minerals such as Schoepite (cf. Deditius et al., 2007). Significant negative correlation ($r = -0.6$) exists between the amount of oxygen and Ca [apfu] for the first group of uraninite (Fig. 8B) in the cubic field. The negative correlation between Ca and oxygen displayed by less oxidized uraninite group is likely to suggest that Ca incorporation into the uraninite structure is again related to stage and temperature of crystallization of uraninite. The U^{4+} rich phases are early to crystallize and Ca preferentially enters into the early phase uraninite structure. Ca incorporation into the uraninite structure further created oxygen vacancies (Shannon, 1976) in order to maintain charge balance in the Ca-rich group of uraninite (cf. Janeczek et al., 1993).

6.3. Implications for provenance from the Mesoarchean detrital uraninite

The mineral chemical analyses and textural studies suggest that the uraninite grains in Mahagiri QPC matrix has been derived from a late magmatogenic fluid of granitic composition at a high temperature ($> 350^\circ\text{C}$). REE data together with the Th/U ratios and elevated Y-content further confirm their high temperature origin. However, variation in uraninite composition with regard to Th, Y, Ca, and REEs

further indicates significant differences in the granitoids/pegmatoids in the source for these detrital uraninites. Based on the present geo-chronological data, the maximum age of deposition of Mahagiri Quartzite has been suggested at 3.02 Ga (Mukhopadhyay et al., 2014). The youngest detrital zircon population from the Mahagiri Quartzite and its equivalent stratigraphic units in the southern Singhbhum craton has been dated at 3.02 Ga (U–Pb LA-ICPMS age by Mukhopadhyay et al., 2014). Though we have no direct age measurement from the Mahagiri uraninite, given the fact that the chemical composition of the Mahagiri uraninite unequivocally suggesting magmatic origin, the age of the youngest detrital zircon population may as well suggest a reasonable lower age limit for the detrital grain population in the Mahagiri Quartzite. The early event of crustal enrichment of uranium is believed to have started during Paleo–Mesoarchean transition with the uranium enrichment processes in the metaluminous komatiitic and basaltic melts from partial melting of the primitive mantle ($\text{U} = 20.5$ ppb, Javoy, 1998). The fractional crystallization of these melts led to formation of the Archean tonalite–trondhjemite–granodiorite (TTG) association (Martin, 1994; Moyen, 2011) with minor U enrichment (1–2 ppm U) (Cuney, 2014). In such pre-Mesoarchean felsic rocks U was mainly hosted in accessory mineral structures (e.g., monazite, apatite, allanite, zircon and titanite). However, due to the insufficient initial concentration of U in TTGs, such TTG suites are not likely to be the source rocks for detrital uraninites in the > 3 Ga old Mahagiri QPC. During Mesoarchean time (3.2–2.8 Ga), highly felsic, K-rich (richer than TTG) granodiorite–granite–monzogranite suites (GGM) of rocks, high-K metaluminous to slightly peraluminous granitic and pegmatitic complexes also came into existence (Cuney and Kyser, 2008; Clemens et al., 2010; Cuney, 2014). K-rich true granite has been reported with ages as old as ~ 3.6 Ga from Bastar craton, India (Rajesh et al., 2009). The detrital uraninite in the Mahagiri QPC thus possibly owes its origin to the K-rich GGM suites and suggests the existence of such K-rich felsic suites older than 3.0 Ga in the Singhbhum craton.

7. Conclusions

Uraninite paleoplacers of the Mahagiri Quartzite are deposited in a subaerial alluvial fan to coastal braided plain environment. The uraninite placers are deposited with quartz pebble conglomerate as debris flow deposits of proximal fan feeders and pebbly sandstone of braid plain to mid-fan complex as sheet-flood and high-density fluid flow deposits. The mineral composition of the Mahagiri uraninite with elevated Th content suggests their magmatic source. High total REE abundance is likely to have resulted from increased substitution into uranium oxide structure at high temperature compared to those crystallized at low temperature. Binary plots of ΣREE vs. U/Th clearly indicate the high temperature magmatic source of the Mahagiri uraninite. The chemical formula of the uraninite suggests multiple source population and variations in oxidation states. Mahagiri detrital uraninite points towards the existence of highly felsic and K-rich (richer than TTG) granodiorite–granite–monzogranite suites (GGM) of rocks older than 3.1 Ga in the Singhbhum craton.

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Appendix A. Standards used in EPMA analyses and steps in chemical formula calculations

Element	Spectral line	Crystal	Standard	Detection limit (ppm)
Na	K α	TAP	Jadeite_NaSp1_076	301
Si	K α	TAP	Orthoclase_SiSp1_062	217
Mg	K α	TAP	MgO_MgSp1_049	186
Al	K α	TAP	Al2O3_AlSp1_052	249
P	K α	LPET	Apatite_P_Sp3_011	284
S	K α	LPET	FeS2_S_Sp3_011	353
K	K α	PET	Orthoclase_KSp4_054	202
Ca	K α	PET	Apatite_CaSp4_015	209
Ti	K α	PET	TiO2_TiSp4_023	342
Fe	K α	LIF	Fe2O3_FeSp2_015	1183
Y	L α	TAP	YAG_Y_Sp1_034	232
Zr	L α	LPET	ZrSiO4_ZrSp3_008	802
Ba	L α	LPET	BaSO4_Ba_Sp3_034	412
La	L α	LPET	REE3_LaSp3_038	244
Ce	L α	LPET	REE2_Ce_Sp3_026	209
Sm	L α	LIF	REE3_Sm_Sp2_027	832
Gd	L β	LIF	REE3_Gd_Sp2_026	2535
Dy	L α	LIF	REE1_Dy_Sp2_016	1130
Nd	L β	LIF	REE4_Nd_Sp2_036	854
Yb	L α	LIF	REE3_Yb_Sp2_006	728
Pr	L β	LIF	REE1_Pr_Sp2_035	1350
Ho	L β	LIF	REE2_HoSp2_025	2306
Er	L α	LIF	REE1_Er_Sp2_006	715
Pb	M α	LPET	Pyro_Pb_Sp3_017	485
U	M β	PET	UO2_U_Sp4_036	1471
Mn	K α	LIF	Rhodinite_Mn_Sp2_036	854
Th	M α	PET	ThO2_Th_Sp4_038	2385
Sr	L α	PET	Celestine_Sr_Sp4_006	519

From the above table it is apparent that both the short (TAP, PET, LiF) as well as large (LPET) crystals were used as diffracting media for the secondary X-rays and both the natural minerals and synthetic materials, viz. Jadeite, orthoclase, alumina, apatite, rhodonite, celestine, uraninite, thorinite, and synthetic glasses like MgO glass, FeS₂, TiO₂, Fe₂O₃, YAG, synthetic zircon, synthetic barite, synthetic REE1, REE2, REE3 glasses, and pyro were used as standards for analyses of Na, Si, Al, P, Mn, Sr, U, Th, Mg, Fe, Ti, Fe, Y, Zr, Ba, La, Ce, Sm, Gd, Dy, Nd, Yb, Pr, Ho, Er, and Pb with detection limits of <0.1 wt.% for most of the elements, except a few HREE. The Eu was not measured because of the serious interference of Eu-L α 1 line from other REE, particularly due to higher concentration of Pr and Nd in the samples. The Eu-L β 1 line also suffered interference, mainly from Dy and HREEs concentrations which are comparably subequal to LREE. Moreover, its concentration in samples were lower than 0.1 wt.% (cf. Reed and Buckley 1998). Terbium, Thulium, and Lutetium concentrations could not be measured because both the L α 1 and L β 1 lines of these HREEs suffered strong interferences with lines of other REE. Thus, they were interpolated from chondrite-normalized abundances of the neighbouring elements.

While attempting calculation of chemical formula from EMPA data, following procedure was followed: 1) entire Pb content was assumed to be radiogenic and PbO is recalculated to UO₂; 2) after converting PbO to UO₂, if the EMPA total is less than 100 wt.%, U⁴⁺ was converted to U⁶⁺, adding one oxygen per U atom; 3) if all U⁴⁺ is converted to U⁶⁺ and the total is still less than 100 wt.%, then H₂O and CO₂ may exist in the mineral phase; 4) if all U⁴⁺ is converted to U⁶⁺ and the total is more than 100 wt.%, then only part of U⁴⁺ can be converted to U⁶⁺, thus suggesting a U⁴⁺–U⁶⁺ phase. The cationic fractions of all elemental analyses for each sample with U⁴⁺ and U⁶⁺ components and excess oxygen content in each formula unit of uraninite were also calculated on the basis of cationic charge balance method (after Janeczke and Ewing, 1992; Zhao and Ewing, 2000) (Table 1). The Mahagiri uraninite range in U⁶⁺ content from

0.0.075 to 0.690 [apfu] corresponding to the amount of excess oxygen from 2.049 to 2.623 [apfu] (Table 1).

References

- Acharya, S., 2002. The Daitari–Tomka Basin—its structural and stratigraphic evolution & genesis of associated iron ores. *Indian J. Geol.* 74, 49–81.
- Acharyya, S.K., 1993. Greenstones from Singhbhum Craton, their Archean character, oceanic crustal affinity and tectonics. *Proc. Natl. Acad. Sci., India, Sect. A* 63 (A), 211–222.
- Acharyya, S.K., Gupta, A., Orihashi, Y., 2010a. New U–Pb zircon ages from Paleo–Mesoproterozoic TTG gneisses of the Singhbhum Craton, Eastern India. *Geochim. J.* 44, 81–88.
- Acharyya, S.K., Gupta, A., Orihashi, Y., 2010b. Neoproterozoic–Paleoproterozoic stratigraphy of the Dhanjori basin, Singhbhum Craton, eastern India: and recording of a few U–Pb zircon dates from its basal part. *J. Asian Earth Sci.* 39, 527–536.
- Anders, E., Grevesse, N., 1989. Abundances of the elements: meteoritic and solar. *Geochim. Cosmochim. Acta* 53, 197–214.
- Bandyopadhyay, P.K., Chakrabarti, A.K., DeoMurari, M.P., Misra, S., 2001. 2.8 Ga old anorogenic granite–acid volcanics association from western margin of the Singhbhum–Orissa Craton, Eastern India. *Gondwana Res.* 4 (3), 465–475.
- Basu, A.R., Bandyopadhyay, P.K., Chakraborti, R., Zou, H., 2008. Large 3.4 Ga alga type BIF in the Eastern Indian Craton. *Goldschmidt Conference 2008, Vancouver (Canada)*, Abstract Volume. *Geochim. Cosmochim. Acta* 72, (125), p. A59.
- Bergen, L., Fayek, M., 2012. Petrography and geochronology of the Pele Mountain quartz-pebble conglomerate uranium deposit, Elliot Lake District, Canada. *Am. Mineral.* 97, 1274–1283.
- Blair, T.C., McPherson, J.G., 1994. Alluvial fans and their natural distinction from rivers based on morphology, hydraulic processes, sedimentary processes, and facies assemblages. *J. Sediment. Res.* 64, 450–489.
- Bose, S., Guha, S., Ghosh, G., Das, K., Mukhopadhyay, J., 2014. Tectonic juxtaposition of crust and continental growth during orogenesis: example from the Rengali Province, eastern India. *Geosci. Front.* <http://dx.doi.org/10.1016/j.gsf.2014.09.002>.
- Catuneanu, O., 2002. Sequence stratigraphy of clastic systems: concepts, merits, and pitfalls. *J. Afr. Earth Sci.* 35, 1–43.
- Chakrabarti, K., Ecka, N.R.R., Mishra, B., Ramesh Babu, P.V., Parihar, P.S., 2011. Paleoproterozoic quartz pebble conglomerate type uranium mineralisation in Mankarhachua area, Angul District, Orissa, Eastern India. *J. Geol. Soc. India, Bangalore* 77, 443–449.
- Chakrabarti, K., Ecka, N.R.R., Mishra, B., Mahendra Kumar, K., Katti, V.J., Umamaheswar, K., Parihar, P.S., Mukhopadhyay, J., Ghosh, G., 2013. Gold, silver and platinum group of elements mineralization in Precambrian uraniferous quartz-pebble conglomerates of Mankarhachua area, Angul District, Odisha. *Curr. Sci.* 105, 978–983.
- Chakraborty, K.L., Chakraborty, T.L., Majumder, T., 1980. Stratigraphy and structure of the Precambrian banded iron formations and chromite-bearing ultramafic rocks of the Sukinda Valley, Orissa. *J. Geol. Soc. India* 21, 398–404.
- Clemens, J.D., Belchery, R.W., Kisters, A.F.M., 2010. The Heerenveen Batholith, Barberton mountain land, South Africa: Mesoarchean, potassic, felsic magmas formed by melting of an ancient subduction complex. *J. Petrol.* 51, 1099–1120.
- Cuney, M., 2010a. The extreme diversity of uranium deposits. *Mineral. Deposita* 44, 3–9.
- Cuney, M., 2010b. Evolution of uranium fractionation processes through time: driving the secular variation of uranium deposit types. *Econ. Geol.* 105, 449–465.
- Cuney, M., 2013. Uranium and thorium resources and sustainability of nuclear energy. In: Burns, P. (Ed.), *Uranium: Cradle to Grave*, MAC Short Course Series 43. 15, pp. 417–438.
- Cuney, M., 2014. Felsic magmatism and uranium mineralization. *Bull. Geol. Soc. France* 185, 75–92.
- Cuney, M., Friedrich, M., 1987. Physicochemical and crystal-chemical controls on accessory mineral paragenesis in granitoids. Implications on uranium metallogenesis. *Bull. Mineral.* 110, 235–247.
- Cuney, M., Kyser, K., 2008. Recent and not-so-recent developments in uranium deposits and implications for exploration. *Mineral. Assoc. Canada Short Course Series* 39 (257 pp.).
- Cuney, M., Mathieu, R., 2000. Extreme light rare earth element mobilisation by diagenetic fluids in the geological environment of Oklo natural reactor zone, Franceville basin, Gabon. *Geology* 28, 743–746.
- Dahlkamp, F.J., 2009. *Uranium Deposits of the World*. Springer, Berlin.
- De, S., Mazumder, R., Ohta, T., Hegner, E., Yamada, Bhattacharya, T., Chiranzelli, J., Altermann, W., Arima, M., 2014. Geochemical and Sm–Nd isotopic characteristics of the Late Archean–Palaeoproterozoic Dhanjori and Chaibasa metasedimentary rocks, Singhbhum craton, E. India: Implications for provenance, and contemporary basin tectonics. *Precambrian Res.* 265, 62–78.
- Deditius, A.P., Utsunomiya, S., Ewing, R.C., 2007. Fate of trace elements during alteration of uraninite in a hydrothermal vein-type U-deposit from Marshall Pass, Colorado, USA. *Geochim. Cosmochim. Acta* 71, 4954–4973.
- Depinè, M., Frimmel, H.E., Emsbo, P., Koenig, A.E., Kern, M., 2013. Trace element distribution in uraninite from Mesoproterozoic Witwatersrand conglomerates (South Africa) supports placer model and magmatogenic source. *Mineral. Deposita* 48, 423–435.
- Dunn, J.A., Dey, A.K., 1942. The geology and petrology of eastern Singhbhum and surrounding areas. *Mem. Geol. Surv. India* 69 (2), 281–450.
- Fareeduddin, Janardhan, A.S., Basavalingu, B., 1988. Sedimentology, mineralogy and geochemistry of the Kalasapura conglomerate. *Mem. Geol. Soc. India* 65–82.
- Frimmel, H.E., 2005. Archean atmospheric evolution: evidence from the Witwatersrand gold fields, South Africa. *Earth Sci. Rev.* 70, 1–46.

- Frimmel, H.E., 2014. A giant Mesoarchean crustal old-enrichment episode: possible causes and consequences for exploration. *Soc. of Econ. Geol. Special Publication* 18, 209–234.
- Frimmel, H.E., Groves, D.J., Kirk, J., Ruiz, J., Chesley, J., Minter, W.E.L., 2005. The formation and preservation of the Witwatersrand goldfields, the largest gold province in the world. In: Hedenquist, J.W., Thompson, J.F.H., Goldfrab, R.J., Richards, J.P. (Eds.), *Society Econ. Geol., 100th Anniv.* Vol. 9, pp. 769–797.
- Frimmel, H.E., Depiné, M.P., Emsbo, A.E., Koenig, Kern, M., 2013. Reply to comments by T. Oberthür on “trace element distribution in uraninite from Mesoarchean Witwatersrand conglomerates (South Africa) supports placer model and magmatogenic source”. *Mineral. Deposita* (ISSN: 0026-4598) 48 (8), 1051–1053.
- Frimmel, H.E., Schedel, S., Bratz, H., 2014. Uraninite chemistry as forensic tool for provenance analysis. *Appl. Geochem.* 48, 104–121.
- Fryer, B.J., Taylor, R.P., 1987. Rare-earth element distributions in Uraninites implications for ore genesis. *Chem. Geol.* 63, 101–108.
- Gaboreau, S., Cuney, M., Quirt, D., Beaufort, D., Patrier, P., Mathieu, R., 2007. Significance of aluminium phosphate–sulfate minerals associated with U unconformity-type deposits: the Athabasca basin, Canada. *Am. Mineral.* 92.
- Ghosh, G., Kumari, S., Patil, S.K., Mukhopadhyay, J., Ray, A., 2010. Superposed deformation fabrics in the Precambrian metabasic rocks of the Iron Ore Group, Singhbhum craton, Eastern India: evidences from anisotropy of magnetic susceptibility studies. *J. Struct. Geol.* 32, 249–261.
- Grandstaff, D.E., 1980. Origin of uraniferous conglomerates at Elliot Lake, Canada and Witwatersrand, South Africa: implications for oxygen in the Precambrian atmosphere. *Precambrian Res.* 13, 1–26.
- Grandstaff, D.E., 1981. Microprobe analyses of uranium and thorium in uraninite from the Witwatersrand, South Africa, and Blind River, Ontario, Canada. In: Armstrong, F.C. (Ed.), *Genesis of Uranium- and Gold-Bearing Precambrian Quartz-Pebble Conglomerates*. U.S. Geol. Surv. Professional Paper, pp. J1–J5.
- Gupta, A., Basu, A., 2000. North Singhbhum proterozoic mobile belt Eastern India—a review. *Geol. Surv. India Spec. Publ.* 55, 195–226.
- Gupta, A., Basu, A., Ghosh, P.K., 1980. The proterozoic ultramafic and mafic lavas and tuffs of the Dalmia greenstone belt, Singhbhum, eastern India. *Can. J. Earth Sci.* 17, 210–231.
- Horscroft, F.D.M., Mossman, D.J., Reimer, T.O., Hennigh, Q., 2011. Witwatersrand metallogenesis: the case for (modified) syngeneses. *SEPM Spec. Publ.* 101, 75–95.
- Janeczek, J., Ewing, R.C., 1992. Structural formula of uraninite. *J. Nucl. Mater.* 190, 128–132.
- Janeczek, J., Ewing, R.C., Thomas, L.E., 1993. Oxidation of uraninite: does tetragonal U_3O_8 occur in nature? *J. Nucl. Mater.* 207, 177–191.
- Javoy, M., 1998. The birth of the Earth's atmosphere: the behaviour and fate of its major elements. *Chem. Geol.* 147, 11–25.
- Kositcin, N., Krapez, B., 2004. SHRIMP U–Pb detrital zircon geochronology of the Late Archaean Witwatersrand Basin of South Africa: relation between zircon provenance age spectra and basin evolution. *Precambrian Res.* 129, 141–168.
- Kumar, A., Venkatesh, A.S., Ramesh Babu, P.V., Nayak, S., 2012. Genetic implications of rare uraninite and pyrite in quartz-pebble conglomerate from Sundargarh district of Orissa, Eastern India. *J. Geol. Soc. India* 79 (3), 279–286.
- Mahadevan, T.M., 1988. Exploration for quartz pebble conglomerate type uranium deposits in India — problems and prospects. *Mem. Geol. Soc. India* 9, 19–28.
- Mahadevan, T.M., 1995. Identifying high-grade uranium deposits in the proterozoic basins of India — a challenge to exploration. *Explor. Res. At. Miner.* 8, 49–60.
- Mahato, S., Goon, S., Bhattacharya, A., Mishra, B., Bernhardt, H.J., 2008. Thermo-tectonic evolution of the North Singhbhum mobile belt: a view from the western part of the belt. *Precambrian Res.* 162, 102–107.
- Maier, W.D., Barnes, S.J., Campbell, I.H., Fiorentini, M.L., Peltonen, P., Barnes, S.J., Smithies, R.H., 2009. Progressive mixing of meteoritic veneer into the early Earth's deep mantle. *Nature* 460, 620–623.
- Martin, H., 1994. The Archaean grey gneisses and the genesis of the continental crust. In: Condie, K.C. (Ed.), *Archean Crustal Evolution*, 11. *Dev. Precamb. Geol.* Elsevier, Amsterdam, pp. 205–259.
- Mazumdar, S.K., 1996. Precambrian geology of Peninsular India. *Indian Miner.* 50, 139–174.
- Mazumdar, R., van Loon, A.J., Mallik, L., Reddy, S.M., Arima, M., Altermann, W., Eriksson, P.G., De, S., 2012a. Mesoarchaeoan–Palaeoproterozoic stratigraphic record of the Singhbhum crustal province, eastern India: a synthesis. In: Mazumdar, R., Saha, D. (Eds.), *Palaeoproterozoic of India*. *Geol. Soc. London Spl. Publ.* 365, pp. 29–47.
- Mazumdar, R., Eriksson, P.G., De, S., Bumby, A.J., Lenhardt, N., 2012b. In: Mazumdar, R., Saha, D. (Eds.), *Palaeoproterozoic sedimentation on the Singhbhum Craton: global context and comparison with Kaapvaal*. *Geol. Soc. London Spl. Publ.* Vol. 365, pp. 51–76.
- Mercadier, J., Cuney, M., Iach, P., Boiron, M.C., Bonhoure, J., Richard, A., Mathieu, L., Kister, P., 2011. Origin of uranium deposits revealed by their rare earth element signature. *Terra Nova* 23, 264–269.
- Minter, W.E.L., 1976. Detrital gold, uranium pyrite concentrations related to sedimentology in the Precambrian Vaal Reef placer, Witwatersrand, South Africa. *Econ. Geol.* 71, 157–176.
- Minter, W.E.L., 2006. The sedimentary setting of Witwatersrand placer mineral deposits in an Archaean atmosphere. *Evolution of Earth's Early Atmosphere, Hydrosphere, and Biosphere—Constraints from Ore Deposits*. *Geol. Soc. America Memoir* Vol. 198, pp. 105–119.
- Mishra, S., Deomurari, M.P., Wiedenbeck, M., Goswami, J.N., Ray, S., Saha, A.K., 1999. $^{207}Pb/^{206}Pb$ zircon ages and the evolution of the Singhbhum Craton, eastern India: an ion microprobe study. *Precambrian Res.* 93, 139–151.
- Mishra, B., Pande, D., Gogoi, J., Kumar, A., Ramesh Babu, P.V., Parihar, P.S., 2008. Quartz-pebble conglomerate type uranium mineralisation in Balia–Rankia Area of Daitari–Tomka basin, Jaipur district, Orissa. *Explor. Res. Atomic Miner., AMD, DAE, Hyderabad* 18, 1–14.
- Misra, S., 2006. Precambrian chronostratigraphic growth of Singhbhum–Orissa craton, eastern Indian shield: an alternative model. *J. Geol. Soc. India* 67, 356–378.
- Moyen, J.F., 2011. The composite Archaean grey gneisses: petrological significance, and evidence for a non-unique tectonic setting for Archaean crustal growth. *Lithos* 123, 21–36.
- Mukhopadhyay, D., 2001. The Archean nucleus of Singhbhum: the present state of knowledge. *Gondwana Res.* 4, 307–318.
- Mukhopadhyay, J., Ghosh, G., Nandi, A.K., Chaudhuri, A.K., 2006. Depositional setting of the Kolhan Group: its implications for the development of a Meso to Neoproterozoic deep-water basin on the South Indian craton. *South Afr. J. Geol.* 109, 183–192.
- Mukhopadhyay, J., Beukes, N.J., Armstrong, R.A., Zimmermann, U., Ghosh, G., Medda, R.A., 2008. Dating the oldest greenstone in India: a 3.51 Ga precise U–Pb SHRIMP Zircon Age for Dacitic Lava of the Southern Iron Ore Group, Singhbhum Craton. *J. Geol.* 116, 449–461.
- Mukhopadhyay, J., Ghosh, G., Zimmermann, U., Guha, S., Mukherjee, T., 2012. A 3.51 Ga bimodal volcanic BIF ultramafic succession from Singhbhum Craton: implications for Palaeoarchaeoan geodynamic processes from the oldest greenstone succession of the Indian subcontinent. *Geol. J.* 47, 284–311.
- Mukhopadhyay, J., Crowley, Q.C., Ghosh, G., Ghosh, S., Chakrabarti, K., Mishra, B., Bose, S., 2014. Oxygenation of the Archaean atmosphere: New paleosol constraints from eastern India. *Geology* 42, 923–926.
- Nelson, D.R., Bhattacharya, H.N., Thern, E.R., Altermann, W., 2014. Geochemical and ion-microprobe U–Pb zircon constraints on the Archaean evolution of Singhbhum Craton, eastern India. *Precambrian Res.* 255, 412–432.
- Nemec, W., Steel, R.J., 1988. What is a fan delta and how do we recognize it? In: Nemec, W., Steel, R.J. (Eds.), *Fan Deltas: Sedimentology and Tectonic Settings*. Blackie and son, pp. 1–13.
- Oberthür, T., Saager, R., 1986. Silver and mercury in gold particles from the Proterozoic Witwatersrand placer deposits of South Africa. *Metallogenic and geochemical implications*. *Econ. Geol.* 81, 20–31.
- Pagel, M., Pinte, G., Rotach-Toulhoat, N., 1987. The rare earth elements in natural uranium oxides. In: Friedrich, G., Gatzweiler, R., Vogt, J. (Eds.), *Uranium Symposium on the Occasion of the DMG-GDMB-SGA Joint Meeting; Aachen (Germany, F.R.)*. Monograph Series on Mineral Deposits Vol. 27, pp. 81–85.
- Pal, D.C., Rhede, D., 2013. Geochemistry and chemical dating of Uraninite in the Jaduguda uranium deposit, Singhbhum Shear Zone, India—implications for uranium mineralization and geochemical evolution of uraninite. *Econ. Geol.* 108, 1499–1515.
- Pal, D.C., Chaudhuri, T., McFarlane, C., Mukherjee, A., Sarangi, A.K., 2011. Mineral chemistry and in situ dating of allanite and geochemistry of its host rocks in the Bagjata uranium mine, Singhbhum shear zone, India — implications for chemical evolution of REE mineralization and mobilization. *Econ. Geol.* 106, 1155–1171.
- Phillips, G.N., Powell, R., 2011. Origin of Witwatersrand gold: a metamorphic devolatilisation-hydrothermal replacement model. *Appl. Earth Sci. (Trans. Inst. Min. Metall. B.)* 120, 112–129.
- Plant, J.A., Simpson, P.R., Smith, B., Windley, B.F., 1999. Uranium ore deposits—products of the radioactive earth. In: Burns, P.C., Finch, R. (Eds.), *Uranium: Mineralogy, Geochemistry and the Environment*. Reviews in Mineralogy Vol. 38. *Miner. Soc. of Amer.*, pp. 255–319.
- Prabhakar, N., Bhattacharya, A., 2013. Paleoproterozoic partial convective overturn in the Singhbhum Craton, Eastern India. *Precambrian Res.* 231, 106–121.
- Prasada Rao, G.H.S.V., Murthy, Y.G.K., Deekshitulu, M.N., 1964. Stratigraphic relations of Precambrian iron formations and associated sedimentary sequences in parts of Keonjhar, Cuttack, Dhenkanal and Sundargarh Districts, Orissa, India. *Proceed. 22nd Internat. Geol. Cong., New Delhi, India*, Section 10, pp. 72–87.
- Radhakrishna, B.P., 1988. Economic significance of quartz-pebble conglomerates. *Geol. Soc. India Mem* 9, 7–17.
- Rajesh, H.M., Mukhopadhyay, J., Beukes, N.J., Gutzmer, J., Belyanin, G.A., Armstrong, R.J., 2009. Evidence for an early Archaean granite from Bastar craton, India. *Jour. Geol. Soc. London* 166, 193–196.
- Rama Rao, B.V., 1974. Discovery of uraniferous pre-Cambrian conglomerates at Chikmagalur, Karnataka. *Curr. Sci.* 44, 174–175.
- Rantzeck, U., Gaurer, C.D.K., Van der Westhuizen, W.A., Duhamel, I., Cuney, M., Beukes, N.J., 2011. Mineral chemical study of U-bearing minerals from the Dominion Reefs, South Africa. *Mineral Deposits* 46, 187–196.
- Reimer, T.O., 1987. The Late Archaean dominion conglomerates (South Africa): new aspects of their derivation and their relationship with those of the Witwatersrand. *Neues Jb. Mineral. Abh.* 158, 13–46.
- Roy, M.K., 1996. Quartz-Pebble Conglomerate Type Uranium Deposits. *Exploration for Atomic Minerals, Part II, AMD Training Course Handbook*, v.3, pp. 38–51.
- Roy, M., Roy, M., 2008. Temperature dependent textural and chemical variations in uraninites from diverse geological environments of India. *J. Geol. Soc. India* 72, 155–167.
- Saha, A.K., 1994. Crustal evolution of Singhbhum–North Orissa, eastern India. *Mem. Geol. Soc. India* 27, 341.
- Shannon, R.D., 1976. Revised effective ionic radii and systematic studies of interatomic distances in halides and chalcogenides. *Acta Crystallogr. A* 32, 751–767.
- Srinivasan, R., Ojakangas, R.W., 1986. Sedimentology of quartz-pebble conglomerates and quartzites of the Archaean Bababudan group, Dharwar craton, South India: evidence for crustal stability. *J. Geol.* 94, 199–214.
- Tilsley, J.E., 1988. Genetic considerations relating to some uranium ore deposits. In: Roberts, R.G., Shaehnan, P.A. (Eds.), *Ore Deposit Models*, Geosciences Canada Reprint Series 3. *Geol. Assoc. Canada*, pp. 91–101.
- Upadhyay, D., Chatopadhyaya, S., Kooijman, E., Mezger, K., Berndt, J., 2014. Magmatic and metamorphic history of Paleoproterozoic tonalite–trondhjemite–granodiorite

- (TTG) suite from the Singhbhum craton, eastern India. *Precambrian Res.* 252, 180–190.
- Vasudeva Rao, M., Sinha, K.K., Mishra, B., Balachandran, K., Srinivasan, S., Rajasekharan, P., 1988. Quartz-pebble conglomerate from the Dhanjori – a new uranium horizon of Singhbhum uranium Province. *Mem. Geol. Soc. India* 9, 89–95.
- Viswanath, R.V., Roy, M.K., Pandit, S.A., Narayan Das, G.R., 1988. Uranium mineralisation in the quartz pebble conglomerates of Dharwar supergroup, Karnataka. *Mem. Geol. Soc. India* 9, 33–41.
- Zhao, D., Ewing, R.C., 2000. Alteration products of uraninite from the Colorado Plateau. *Radiochim. Acta* 88, 739–749.