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Recent Studies on the Singhbhum Craton

DHRUBA MUKHOPADHYAY* and ABDUL MATIN

Formerly of Department of Geology, University of Calcutta, India

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Introductory Remarks

Like other shield areas of the world, the Singhbhum Craton is made up of expansive granitoid terranes that are either flanked by or have internal screens of supracrustal belts in varying metamorphic states. The mafic-dominated supracrustal belts are thought to represent segments of the ancient oceanic crust and their isotopic characters provide information about the Hadean and Eoarchean mantle. Globally, evidence is accumulating about the existence of continental crust even in Hadean times (4.0-4.5 Ga) from the study of zircon grains in Archean sediments and xenocrystic zircon grains in granitic gneisses. In recent years, more refined isotopic studies of Pb, Sm, Nd, Lu, Hf, and particularly of short lived isotopes and oxygen isotopes have provided new information on the processes of crustal growth during the Hadean times and on the physical and chemical conditions existing on the surface of the early Earth. How much of the Archean felsic continental crust was juvenile and how much of it was due to reworking of older crusts is a matter of debate. There is also a raging controversy on whether the Archean geodynamic process was driven by subduction or by plume driven underplating or intraplating. The Singhbhum craton, with its supracrustal belts and the granitoid terrains of Paleoproterozoic to Neoproterozoic age is an ideal terrain to test the different models of Archean crustal growth. This review confines itself principally to the new information that has come during the last five years. The older work had earlier been summarized in Mukhopadhyay (2001) and Sarkar and Gupta (2012).

Singhbhum Craton

Singhbhum Craton, also designated as the Iron Ore

Craton (Saha, 1994), is one of the oldest cratonic blocks in Peninsular India, and encompasses an area of ~39,000 km² and forms a N-S elongate ovoid block between 84.5°E to 86.5°E and 21°N to 22.75°N in eastern India (Fig. 1). Deposition of different supracrustal sequences and emplacement of granitoid rocks built up the craton during the time span of 3.5-2.5 Ga. Evidence has recently emerged about the existence of continental crust during the Eoarchean-Hadean time.

The Singhbhum Craton (SC) (Fig. 1) lies south of the Central Indian Tectonic Zone (CITZ) and is bounded in the north by the arcuate, northerly convex Singhbhum Shear Zone (SSZ) (Fig. 1). Traditionally (Sarkar and Saha, 1962; Saha, 1994), the southern branch of the western boundary of SSZ is thought to veer towards SW, wrapping round the cratonic nucleus. But, later work (Mukhopadhyay *et al.*, 1991; Blackburn and Srivastava, 1994) suggests that the SSZ should pass south of Ongarbira Volcanics and Jojohatu Ultramafics.

The Barakot-Akul Fault Zone (Crowe *et al.*, 2003) and in its continuation the Sukinda Thrust (Fig. 1) demarcate the southern boundary between the SC and the Rengali Province (RP). The RP, principally made of high-grade Rengali granulite suite of rocks, is thrust over the SC (Bose *et al.*, 2015). The RP separates the SC from the main Eastern Ghats Granulite Belt (EGGB). Ghosh *et al.* (2016a) suggested that the RP shows imprints of a major 500 Ma transpressive deformation resulting from oblique collisional accretion of the Eastern Ghats belt to the Singhbhum Craton. They correlated this deformation with the Pan-African orogeny. Bose *et al.* (2015) and Mahapatro *et al.* (2012) have interpreted the granulite

*Author for Correspondence: E-mail: dhruba-38@yahoo.co.uk

Fig. 1: Simplified geological map of the Singhbhum craton (after Saha, 1994 and modified from other sources mentioned in the text). 1 - OMG, 2 - OMTG, 3 - Singhbhum Granite, 4 - Bonai Granite, 5 - Iron Ore Group, Malayagiri Supracrustals, 6 - BIF in western IOG, 7 - Kaptipada Granite, 8 - Pallahara Gneiss, 9 - Mayurbhanj Granite, 10 - Tamperkola Granite, 11 - Dhanjori Siliciclastics, 12 - Dhanjori-Simplipal-Jgannathpur-Malangtoli Volcanics, 13 - Darjing Group, 14 - Mahagiri-Keonjhar-Mankaharchua-Deograh Siliciclastics, 15 - Ongarbira Volcanics, 16 - Gabbro-anorthosite-ultramafics, 17 - Sukinda Ultramafics, 18 - Kolhan Group, 19 - Soda Granite, Arkasani Granite, Kuilapal Granite, Beldih Granite. 20 - Singhbhum Group, 21 - Dalma Volcanics, 22 - Chakradharpur Granite, 23 - Chhotanagpur Granite Gneiss, 24 - Rengali Province, 25 - Gondwanas, Younger Sediments and Alluvium 26 - hear Zone/Fault. A - Amjhor, B - Badampahar, Ba - Bahalda, Bh - Baharagora, Bi - Birtola, Bm - Birsampur, Bn - Brahmani, Bo - Bonai, Bp - Bangriposi, Br - Bhaunra, Bs - Besoi, Bu - Bhuasuni, C - Chaibasa, Ch - Chakradharpur, Cm - Chmpua, D - Daitari, Da - Darjing, Gr - Gorumahisani, H - Haludpukur, J - jaduguda, Ja - Jagannathpur, Jd - Joda, Jk - Jorapokhar, Jm - Jamda, Jo - Jojuhatu, Ka - Kaptipada, Ke - Keonjhar, Ko - Koira, Ku - Kumakela, L - Lotapahar, Ma - Mahagiri, Mg - Malangtoli, Mh - Manoharpur, Mk - Mankaharchua, Ms - Mosabani, N - Nilgiri, No - Noamundi, Np - Notopahar, Nu - Nuashahi, O - Ongarbira, On - Onlajori, P - Pallahara, R - Rairangpur, Ra - Rakha, Rk - Raurkela, Rp - Romapahari, Sa - Seraikela, Sk - Satkosia, So - Sonua, Su - Sukinda, T - Tatanagar, Th - Thakurmunda, Tm - Tamperkola, To - Tomka. The Singhbhum Group, the Dalma Volcanics and other associated rocks constitute the North Singhbhum Mobile Belt

facies lenses within the gneisses of the RP to represent thrust slices of the deeper crust of the Singhbhum Craton. The RP is interpreted by Bhattacharya *et al.* (2016) as an accretion zone, formed when the EGGB was welded to the Greater India landmass during the Grenvillian or Pan-African time.

The eastern boundary of the Singhbhum Craton is marked by the N-S to NNE-SSW trending, sinistral transcontinental deformation zone, named as the Eastern Indian Tectonic Zone (EITZ), developed during 876-784 Ma; EITZ continues northward cutting across the SSZ and the North Singhbhum Mobile Belt (Chatterjee *et al.*, 2010).

Four principal units make up the craton: (1) Older Metamorphic Group (OMG) and Older Metamorphic Tonalite Gneiss (OMTG), (2) large granitic massifs of Singhbhum Granite (SBG), Bonai Granite (BG), Kaptipada (Nilgiri) Granite (KG), Mayurbhanj Granite

(MBG) and Pallahara Gneiss (PLG), (3) supracrustal rocks of several Iron Ore Basins, and (4) belts of younger supracrustal rocks. In addition to these, there are the Malangtoli Volcanics, the Jagannathpur Volcanics and the Ongarbira Volcanics, the dyke swarms of Newer Dolerite, and intrusive bodies of ultramafics-gabbro-anorthosite suite. Insufficient geochronological data and rarity of contact relations in outcrops have hampered the understanding of temporal relation between these groups.

Granitoids in the Singhbhum Craton

Older Metamorphic Tonalite Gneiss (OMTG)

The OMTG rocks occupy a total area of about 900 km² east and south of Champua; smaller outcrops are scattered throughout the Singhbhum Granite body (Fig. 1). Prabhakar and Bhattacharya (2013) restrict the term OMTG only to the smaller enclaves of migmatitic

tonalite gneisses within the SBG. The bulk of what is shown as OMTG in the map of Saha (1994) (Fig. 1) belongs, according to them, to the TTG suite of Phase I of the Singhbhum Granite. This point is to be settled by careful field mapping coupled with more petrographical, geochemical and geochronological data. The OMTG rocks are gneissic in nature, typically tonalite-trondhjemite-granodiorite (TTG) in composition (Fig. 2A), and are comparable to the Archean TTG rocks of the world. Field relations in outcrops indicate intrusion of successive phases of tonalite to granite magma. Quartzo-feldspathic products of partial melting are present in boudin necks and in small scale shear zones (Hofmann and Mazumder, 2015). Upadhyay *et al.* (2014) and Pandey *et al.* (2019) have included some rocks compositionally falling in the granite field within the OMTG. However, as described by them, the TTG and the granites have different zircon dates (see

Fig. 2: A and B. Normative Ab-An-Or plots of granitoid rocks in the Singhbhum craton. Data source: Saha *et al.* (1984), Sengupta *et al.* (1991), Saha (1994), Tait *et al.* (2011), Nelson *et al.* (2014), Upadhyay *et al.* (2014), Dey *et al.* (2017), Pandey *et al.* (2019)

below), different REE patterns and different geochemical characteristics. No field relations between the two are described. We are of the opinion that the granite samples should not be grouped with the TTG of the OMTG.

Nelson *et al.* (2014) described gently sloping REE patterns with weak negative Eu anomaly and low HREE abundances for OMTG (Fig. 3A); this is consistent with a model of derivation by fluid absent partial melting of a lower crustal source with retained restitic feldspar following melt extraction. A tonalite enclave within the Singhbhum Granite near Keonjhar also shows a sloping LREE but flat HREE patterns and -ve Eu anomaly (Fig. 3A) (Dey *et al.*, 2017). A petrogenetic model of dehydration melting of a mafic source in absence of garnet with plagioclase as residual or fractionating phase has been suggested. The minimum temperature of the tonalite magma corresponding to the enclave has been estimated from zircon saturation temperature to be $\sim 750^\circ\text{C}$, and the pressure at emplacement depth is determined from Al in hornblende to be 3.5 ± 0.1 kb (~ 12 km depth) (Dey *et al.*, 2017). Pandey *et al.* (2019) reported that the OMTG trondhjemitic rocks have enriched LREE, no Eu anomaly and gently sloping to flat HREE, while the OMTG granite has strong Eu anomaly and flat HREE pattern (Fig. 3A). For the tonalitic-trondhjemitic rocks a petrogenetic model of dehydration melting in absence of garnet with plagioclase as residual or fractionating phase has been suggested. As already mentioned, it is debatable whether the granite samples should be included within the OMTG.

The radiometric dates obtained from OMTG samples by different workers cover a wide range (Table 1, Fig. 4). Geochronological data suggest that the TTG suite in OMTG was emplaced in different pulses at 3.52, 3.47-3.43 and ~ 3.40 -3.33 Ga; xenocrysts of 3.47-3.41 Ga zircon from the older pulse are present in the younger TTGs. The granitic components are distinctly younger at 3.31-3.28 Ga. The major metamorphic pulses were at ~ 3.3 -3.2, 3.14-3.11, 2.96-2.95 and 2.45 Ga. The ages of inherited zircons (xenocrystic cores) from older rocks range from ~ 4.2 to 3.6 Ga. It is interesting to note that the 3380 $\pm$ 11 Ma old granitoids on the Deo Nala display rafts of BIF and are clearly intrusive into the latter. The mean CHIME date of monazite in OMTG tonalite gneiss is 3306 $\pm$ 22 Ma (Prabhakar and Bhattacharya,

2013). This probably represents a metamorphic growth event.

Chaudhuri *et al.* (2018) reported that the Eoarchean (~ 3.7 Ga) cores of zircons in the OMTG and the Paleoarchean (~ 3.6 to ~ 3.3) zircon grains in the OMTG have negative $\epsilon_{\text{Hf(t)}}$ values of -4.7 ± 1.4 to -6.6 ± 1.1 and -0.4 ± 1.2 to -3.7 ± 1.8 respectively (Fig. 5). They interpret that the Hadean enriched reservoir persisted at least till Paleoarchean from which the felsic melts were derived, but these were variably mixed with depleted mantle-derived juvenile magma during Eoarchean to Paleoarchean. On the contrary, Dey *et al.* (2017) reported suprachondritic Hf isotopic signatures with positive values of $\epsilon_{\text{Hf(t)}}$ ranging from $+2.9$ to $+2.2$ and concluded that the 3.47 Ga old tonalite was generated by shallow melting of a juvenile mafic crust. Pandey *et al.* (2019) reported weak subchondritic to suprachondritic $\epsilon_{\text{Hf(t)}}$ values of -1.2 to $+4.5$ for the Paleoarchean OMTG zircon grains (Fig. 5). This implies the existence of a separate depleted reservoir during Paleoarchean which was probably complementary to the ancient enriched reservoir inferred by Chaudhuri *et al.* (2018). Therefore, different sources contributed to different pulses of the OMTG.

Singhbhum Granite

The large oval-shaped body of the Singhbhum Granite constitutes more than one-third of the Singhbhum craton (Fig. 1). It is a suite of granitoid rocks ranging in composition from tonalite to granite and these were emplaced in different pulses. Professor A. K. Saha and his co-workers demarcated the complex granite body into 12 separate units on the basis of mutual intrusive relations and distinctive primary foliation pattern of different units (Saha and Ray, 1984; Saha, 1994). On the basis of field relations, petrography and geochemistry, the 12 domal magmatic bodies were grouped into three closely related phases which were emplaced successively. The earliest phase (Phase I) is relatively K-poor, granodiorite-trondhjemitic in composition, while the later phases (Phase II and III) are granodiorite that grade to monzogranite and granite (Saha, 1994) (Fig. 2A). Phase III covers the largest area of the Singhbhum Granitic complex. Later work has brought into light greater range in composition of the different units of the SBG (Fig. 2A). The OMTG rocks are exclusively TTGs (tonalite-trondhjemitic-

Table 1: Selective geochronological data from the Singhbhum Craton

Material	Age (Ma)/Ga	Method	Reference	Remarks
OMTG				
Tonalite, Champua	3288+35 3280+130	WR Pb-Pb isochron, TIMS WR Rb-Sr isochron	Moorbath <i>et al.</i> , 1986	
Zircon from tonalite gneiss	~3.4 Ga ~3.2 Ga	207Pb/206Pb Small ion Microprobe	Misra, 1999	Formation of OMTG Metamorphism age
Zircon from Haludpukur tonalite	3448+19 3527+17	LA-ICP-MS	Acharyya <i>et al.</i> , 2010a	Upper intercept age
Monazite in tonalite Metamorphism age	3306+22	CHIME, EPMA	Prabhakar and Bhattacharya, 2013	
Zircon in tonalite intrusive into BIF	3380+11	SHRIMP	Nelson <i>et al.</i> , 2014	Concordia age
Zircon in tonalite, Champua	3331+9			Concordant age
Zircon in tonalite-trondhjemite suite, Champua	3437+23 3293+12 3194+24 2965+86	LA-ICP-MS	Upadhyay <i>et al.</i> , 2014	Crystallization age Main metamorphism Fluid activity
OMTG granite, Champua	3611+11 3415+9 3318+3 2953+41 ~2450 ~1100	LA-ICP-MS	Upadhyay <i>et al.</i> , 2014	Xenocryst core Crystallization age Later reequilibration events
Zircon in tonalite enclave	3471+24	LA-ICP-MS	Dey <i>et al.</i> , 2017	Upper intercept age, concordant
Zircon Xenocryst	4.24-4.03 Ga	SHRIMP	Chaudhuri <i>et al.</i> , 2018	Concordant age
Oscillatory zoned rim of xenocryst	3.9-3.8 Ga			Discordant age
Xenocryst	3.67-3.59 Ga			Concordant age
Oscillatory zoning, euhedral	3393+9, 3399+6			Upper intercept, Crystallization age of OMTG
Zircon in trondhjemite	3352+13, 3363+12 3440+37	LA-ICP-MS	Pandey <i>et al.</i> , 2019	Concordant crystallization age Concordant xenocryst age
Zircon in granite	3286+11 3472+57			Concordant crystallization age Concordant, xenocryst
Singhbhum Granite Phase I				
Zircon in granite	3350+20 3262+12 3165+17, 3015+16, 2790+27, 2521+44	LA-ICP-MS	Upadhyay <i>et al.</i> , 2014	Crystallization age Relatively high grade metamorphism Low grade fluid-induced metamorphism
Zircon	3397+9, 3267+6, 3289+10	SHRIMP	Chaudhury <i>et al.</i> , 2018	Upper Intercept, concordant

Singhbhum Granite Phase II

Granitoid rocks	3292+51	WR Pb-Pb isochron, TIMS	Moorbath and Taylor, 1988	
Zircon from Keonjhar-Bhaunra pluton	3280+7	207Pb/206Pb Small ion Microprobe	Misra, 1999	
Zircon from Keonjhar-Bhaunra pluton	3291+9 3496+5	SHRIMP	Tait <i>et al.</i> , 2011	Concordant age Inherited zircon (subhedral) age
Zircon in monzogranite	3326+5	SHRIMP	Nelson <i>et al.</i> , 2014	Near Rairangpur, Concordant age
Zircon in granite and trondhjemite	3336+4		LA-ICP-MS	Upadhyay <i>et al.</i> , 2014 Crystallization age
	3463+8 3281+4 3163+9			Xenocrystic core Metamorphism Low grade metamorphism
Zircon in porphyritic granite of Keonjhar-Bhaunra pluton	3308+31	LA-MC-ICPMS	Dey <i>et al.</i> , 2017	
Zircon in trondhjemite	3334+15	LA-ICP-MS	Pandey <i>et al.</i> , 2019	Concordant, Crystallization ages
Zircon in granite	3330+9			
Singhbhum Granite Phase III				
Zircon in granodiorite	3332+5 3368+7	SHRIMP	Nelson <i>et al.</i> , 2014	West of Gorumahishani Xenocryst from same sample
Zircon in granodiorite	3299+7			East of Keonjhar
Zircon in pegmatitic granodiorite	3285+7			North of Seraikela, Concordant age
Zircon in granite	3445+4 3347+8 3128+42, 2962±33	LA-ICP-MS	Upadhyay <i>et al.</i> , 2014	Crystallization age Metamorphism Fluid mediated alteration
Zircon in nonporphyritic granite	3308+31 3304+25	LA-MC-ICPMS	Dey <i>et al.</i> , 2017	Upper intercept Crystallization age Sub-concordant zircon
Older Core	3377+11, 3367+7	SHRIMP	Chaudhuri <i>et al.</i> , 2018	Concordant age
Euhedral	3286+6			Upper intercept
Zircon in trondhjemite	3470+10, 3292+19	LA-ICP-MS	Pandey <i>et al.</i> , 2019	Concordia age, Crystallization ages
Zircon in granite	3351+8			
Zircon in granodiorite	3300+7	SHRIMP	Olierook <i>et al.</i> , 2019	
Bonai Granite				
Zircon in tonalite xenolith	3380+64	U-Pb, TIMS	Sengupta <i>et al.</i> , 1996	Discordant, Upper intercept age
Xenocryst	3448			Discordant, Upper intercept
Tonalite xenolith	3369+57	WR Pb-Pb isochron		
Host granitoid	3163+126			Errorchron

Zircon in granite	3370+10 3033+38, 2809+29, 2531+ 22	LA-ICP-MS	Chakraborty <i>et al.</i> , 2019	Emplacement age Metamorphic/ hydrothermal ages
Tamperkola Granite				
Zircon in granite	2809+12 3391+8, 3251+8	207Pb/206PbSmall ion Microprobe	Bandyopadhyay <i>et al.</i> , 2001	Crystallization age Xenocrystic cores
Zircon in acid volcanics	2822+67			Volcanics related to granite
Mayurbhanj Grnite				
Zircon in granophyre	~2.8 Ga	LA-ICP-MS	Acharyya <i>et al.</i> , 2010a	Discordant grains, upper intercept of discordia curve
Zircon in aplogranite	3080+8, 3092+5	207Pb/206PbSmall ion Microprobe	Misra <i>et al.</i> , 1999	
Zircon in gneissic granite	3089+6	La-ICP-MS	Chakraborti <i>et al.</i> , 2019	Concordant age
Pallahara Gneiss				
Zircon in granite gneiss	2776+7 2721+10	LA-SF-ICP-MS	Chattopadhyay <i>et al.</i> , 2015	Concordant age, mantle & core Only mantle
Zircon in gneiss	2792+75 2788+59	LA-ICP-MS	Topno <i>et al.</i> , 2018	Upper intercept age Weighted average of most concordant 207Pb/ 206Pb ages
OMG				
Amphibole in amphibolite	3314± 4 3286±6	⁴⁰ Ar/ ³⁹ Ar	Bakshi <i>et al.</i> , 1987	Cooling age through 500°C
Amphibolite	3305+60	WR Sm-Nd isochron	Sharma <i>et al.</i> , 1994	Formation age
Amhibolite & Tonalite	3288+35			Combined OMG- OMTG isochron
Detrital zircon in quartzite and quartz-mica schist	3551+13 3583+25 3504+30 3628+38 3.43-3.38 Ga	207Pb/206PbSmall ion microprobe	Goswami <i>et al.</i> , 1995	Mean core age Detrital grain Detrital grain Detrital grain Detrital grains
Zircon overgrowth	~3.2 Ga			Metamorphism age
Detrital zircon	3504+30 3628+38 3.43-3.38 Ga	²⁰⁷ Pb/ ²⁰⁶ PbSmall ion microprobe	Misra, 1999	Round detrital Round detrital ? Metamorphic
Zircon in schist	3352+26	TIMS	Basu <i>et al.</i> , 1996	Pb-loss event
Euhedral zircon	3.35 Ga	SHRIMP	Saha <i>et al.</i> , 2012	Detrital/? Metamorphic grain
Monazite in mica schist	3377+45 3256+46	CHIME, EPMA	Prabhakar and Bhattacharya, 2013	Metamorphism ages
Zircon in quartzite	3375+3	SHRIMP	Nelson <i>et al.</i> , 2014	Concordant age Detrital grain
Rutile in mica schist	3317+8	LA-ICP-MS	Upadhyay <i>et al.</i> , 2014	Concordia age of metamorphism
Iron Ore Group				
Zircon in basal tuff	3583+25 3392+29	LA-ICP-MS	Basu <i>et al.</i> , 2008	Concordant age, Western BIF

Zircon in basal tuff	3507+2	LA-ICP-MS	Mukhopadhyay <i>et al.</i> , 2008	Concordia age, Southern BIF
	3505+5	LA-ICP-MS	Sreenivas <i>et al.</i> , 2019	
Dhanjori Group				
Metavolcanics	2072±106	WR Sm-Nd TIMS	Roy <i>et al.</i> , 2002	Samples close to shear zone, isotopes likely to be disturbed
Metavolcanics	2787+270 2794+210	WR Sm-Nd WR Pb-Pb	Misra and Johnson, 2005	
Zircon in basal acid tuff	3330+58 and 3358+110	LA-ICP-MS	Acharyya <i>et al.</i> , 2010b	Discordant grains. Upper intercept of discordia curve
Detrital zircon in basal conglomerate	~3.04 to ~3.09	U-Th-Pb chemical age	Sunilkumar <i>et al.</i> , 1996	Age of provenance
Younger Siliciclastic Rocks				
Detrital zircon, Keonjhar Quartzite, Mahagiri Quartzite	3584+7.1 to 3024+13	LA-ICP-MS	Mukhopadhyay <i>et al.</i> , 2014	Concordant ages
Detrital zircon in Mahagiri Quartzite	3950 - 2909 3510 - 3176 3438 - 3220Peak ~3.35 Ga	SHRIMP and LA-ICP-MS	Sreenivas <i>et al.</i> , 2019	Entire Population Concordant ages 86% data Dominant age Population
Malayagiri Belt				
Zircon in dacite, Malayagiri belt	2806+6	Shrimp	Nelson <i>et al.</i> , 2014	
Detrital Zircon in schist, Malayagiri	2788+25	LA-ICP-MS	Chattopadhyay <i>et al.</i> , 2015	Youngest age of detrital zircon
Ultramafic suite				
Gabbroic suite, Baula, Nuasahi	3123+7, 3119+6, 3122+5	SHRIMP	Auge <i>et al.</i> , 2003	Concordant ages
Newer Dolerite				
Ultramafic dyke NNE'ly trend	2613+177	WR Rb-Sr isochron	Roy <i>et al.</i> , 2004	
Dolerite dyke WNW'ly trend	1766+1 and 1765+1	TIMS	Shankar <i>et al.</i> , 2014	207Pb/206Pb age from baddeleyite Weighted Mean age
Dolerite dyke NNEly trend	2762+2 and 2800+0.7	TIMS	Kumar <i>et al.</i> , 2017	
Dolerite dyke NE'ly to ENE'ly trend	2256+6	TIMS	Srivastava <i>et al.</i> , 2019	Upper intercept age from baddeleyite
River Sands				
Detrital zircon	4015+9 3.62-3.55 Ga 3.50-3.22 Ga 3.10-3.06	LA-ICP-MS	Miller <i>et al.</i> , 2018	1 grain 5% data 87% data 6% data

granodiorite), whereas the SBG rocks are granite, monzogranite, granodiorite and trondhjemite (Fig. 2A).

Prabhakar and Bhattacharya (2013) contested

the view of Saha and his co-workers and opined that no magmatic fabric is observable in the Singhbhum granitoids. However, they have described the presence

Fig. 3: A-E. REE patterns in different plutonic units in the Singhbhum Craton from different sources. Selected curves from different sources are shown. Note the variability within each unit. See text for discussion. Data source: Saha and Ray (1984), Saha (1994), Sharma *et al.* (1994), Tait *et al.* (2011), Nelson *et al.* (2014), Dey *et al.* (2017), Pandey *et al.* (2019)

of magmatic flow texture defined by imbricate euhedral plagioclase grains. They recognize a single set tectonic foliation in the SBG granitoids and different

sets of locally developed shear zones. There are two large NNE-SSW striking shear zones, the western Champua Shear Zone (CSZ) along the boundary

Fig. 4: Radiometric dates from the different units in the Singhbhum craton from all sources, (Data from Table 1). Mostly concordant zircon ages are shown. Rb-Sr ages are shown as metamorphic ages. D – Dhanjori Group, YS – Younger siliciclastic rocks like Keonjhar Quartzite etc. Other abbreviations are as in the text

between the OMG and the SBG, and the eastern Rairangpur Shear Zone (RSZ) within the SBG, both showing sinistral-normal (west-down) shear sense. They hypothesized that a period of crustal shortening responsible for the folding of the western Iron Ore Group belt and its accretion to the OMG-OMTG basement, was followed by extensional tectonism marked by the ascent and emplacement of the granitoids of the SBG pluton in pulses along steep-dipping sinistral-normal ductile shear zones. However, the evidence is inadequate to confirm this hypothesis. Prabhakar and Bhattacharya (2013) subdivided the SBG into three types on the basis of deformation textures: Type I with low temperature deformation microstructures, common throughout the pluton; Type II with high temperature deformation microstructures, found exclusively along the western margin of the pluton; and Type III, which is sheared pegmatoidal granitoids associated with the shear zones. According to them the deformation textures developed in response to D_4 shearing that is recorded also in the OMG and the OMTG. The bulk of the pluton cooled appreciably below the solidus prior to deformation,

while along the western margin of the pluton the granitoids were emplaced later and were deformed while still hot.

According to Saha (1994) and Pandey *et al.* (2019) the SBG-Phase I is characterized by smoothly sloping REE pattern with fractionated LREE and moderately fractionated HREE and with or without Eu-anomaly (Fig. 3B). Some samples of SBG-Phase II show sloping REE patterns similar to that of OMTG, while other samples show strong Eu anomaly (Fig. 3C) (Nelson *et al.*, 2014; Dey *et al.*, 2017; Pandey *et al.*, 2019). In SBG-Phase III, REE patterns show fractionated LREE, flat HREE or slightly fractionated HREE and negative Eu-anomaly, but some granodiorite samples show no Eu anomaly and a monzogranite even shows positive Eu-anomaly (Fig. 3D). Therefore, petrographically and in REE pattern different units of SBG show more variability than originally supposed by Saha (1994) (Figs. 2 and 3).

Most of the Singhbhum granitoid samples fall in low to medium pressure category (Pandey *et al.*, 2019). The minimum value of magma temperature as

estimated from zircon saturation temperature is $800^{\circ}\pm 30^{\circ}\text{C}$ for a porphyritic granite/trondhjemite from a Phase II pluton and $754^{\circ}\pm 13^{\circ}\text{C}$ for a non-porphyritic granite from a Phase III pluton. Pandey *et al.* (2019) have pointed out that all the granitoids in the SC have broadly similar geochemical characters and REE patterns which hint towards the derivation of these granitoids from chemically and mineralogically similar source(s). The negative or positive Eu anomalies observed in some granitoids indicate plagioclase fractionation or accumulation respectively during differentiation.

Recent geochronological data (Fig. 4) throw doubt on the three-phase chronology of the SBG as proposed by Saha and his co-workers. It is clear that there is no systematic age difference between the OMTG, the SBG Phase I, the SBG Phase II and the SBG Phase III, as these are defined now. However, there are different pulses of intrusion of granitic magma, and enclaves of older intrusions occur in the younger phases, which is reflected in the ages of zircon xenocrysts. The total period of granitic intrusion lasted from 3.52 to 3.28 Ga and distinct pulses are dated to be at 3.52 Ga (only in OMTG), 3.43-3.47 Ga (noted in OMTG and SBG III), 3.35-3.40 Ga (noted in OMTG only), 3.32-3.35 Ga (noted in OMTG, SBG I, SBG II and SBG III), and 3.30-3.28 Ga (noted in OMTG, SBG II and SBG III). It is also to be noted that younger 3.28-3.30 ages in the OMTG come from samples of OMTG granites and not tonalites. We are of the view that samples of granite are not to be placed within OMTG. Similarly, we recommend that the 3470 Ma trondhjemite (Pandey *et al.*, 2019) and 3445 Ma granite (Upadhyay *et al.*, 2014) be taken out of Phase III and be placed within the OMTG. Olierook *et al.* (2019) have recommended to place the boundary between the OMTG and the SBG at ca. 3380-3365 Ma, which marks the cessation of trondhjemite magmatism and the onset of monzogranite emplacement. However, more mineralogical, geochemical and geochronological data are necessary to fix a firm boundary between magmatic pulses having distinct compositional characteristics. It is interesting to note that the ages of detrital zircon grains in the river sand reported by Miller *et al.* (2018) show a prominent peak at ~ 3.35 Ga and smaller peaks at ~ 3.5 Ga and 3.1 Ga.

Pandey *et al.* (2019) classified the Singhbhum

granitoid samples (OMTG, SBG I, SBG II, SBG III) into older and younger granitoids, and described that their Sm-Nd isotopic compositions define linear arrays in the standard Sm-Nd isochron diagram. The apparent ages calculated from these arrays are 3344 ± 160 Ma and 3142 ± 390 Ma for the older and younger granitoids respectively. The $\epsilon_{\text{Nd}(t)}$ values of the samples have been calculated using their respective U-Pb zircon crystallization ages. Most of the samples have near-chondritic $\epsilon_{\text{Nd}(t)}$ values. The $\epsilon_{\text{Nd}(t)}$ values for the older granitoids (3440 Ma) range from +0.1 to +2.2. The younger granitoids (3263 Ma – 3363 Ma) have $\epsilon_{\text{Nd}(t)}$ values between ~ 0.2 and +1.6. A few samples have higher values of +2.7 to +5.2.

The major relatively high grade metamorphic episode is recorded at 3.34 - 3.26 Ga. This is coeval with magmatic episodes in the SBG and the OMTG. Thus magmatism and metamorphism went hand in hand in the Singhbhum Craton. Younger metamorphic resettings are recorded in OMTG and SBG at 3.16-3.11 Ga and 3.0-2.8 Ga. This appears to be broadly synchronous with the emplacement of younger felsic plutons such as the Mayurbhanj Granite (3.1 Ga; Misra *et al.*, 2002). Even younger dates of 2.45-2.50 Ga are sporadically recorded by zircon in the granitoids and gneisses (Upadhyay *et al.*, 2014). Prabhakar and Bhattacharya (2013) determined CHIME ages of monazite in the SBG granitoids. The range of monazite spot ages in the three textural types recognized by them are as follows. Type I: 2973 ± 101 to 3255 ± 110 Ma (granodioritic); 3137 ± 132 to 3450 ± 238 Ma (granitic); Type II: 2920 ± 138 to 3339 ± 166 Ma; and Type III: 2878 ± 109 to 3534 ± 129 Ma. They have shown that the weighted mean ages of monazites are identical for all granitoids: 3.28 ± 0.04 Ga for Type-I, 3.28 ± 0.07 Ga for Type-II, and 3.27 ± 0.02 Ga for Type-III. This age is interpreted to correspond to the major deformation and metamorphic growth event. The slightly older mean monazite age of OMTG enclaves in SBG (3306 ± 22 Ma) provides an upper limit for the emplacement of SBG.

Dey *et al.* (2017) reported that the mean $\hat{a}_{\text{Hf}(t)}$ values for the 3.47, 3.35, and 3.30 Ga granitoids from the SBG are $+2.1\pm 2.8$ to $+4.8\pm 2.8$, $+1.8\pm 3.0$ to $+4.0\pm 3.6$, and $+0.8\pm 2.8$ to $+3.7\pm 2.8$ respectively (Fig. 5), suggesting derivation from ancient crustal rocks. Because of the positive values they suggested melting of a tonalite dominated source and minor mafic source.

Fig. 5: $\epsilon_{\text{Hf}(t)}$ vs. age plots of magmatic and detrital zircon grains from the Singhbhum Craton (modified from Sreenivas *et al.*, 2019; data source: Chaudhuri *et al.*, 2018, Miller *et al.*, 2018, Sreenivas *et al.*, 2019, Pandey *et al.*, 2019, Olierook *et al.*, 2019). DM- Depleted mantle curve for mantle extraction at 4.5 Ga. Mafic crust (enriched reservoir) curve for $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ fits the plots for the Hadean and Eoarchean zircon grains. The isotopic trajectory of the upper continental crust that formed at 4.5 Ga is drawn for the $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.005. The Hadean-Eoarchean grains (>3.7Ga) lie within a band around a TTG trajectory with $^{176}\text{Lu}/^{177}\text{Hf} = 0.008$ that corresponds to a TTG like source that formed at ~4.3-4.2 Ga. The younger Eoarchean grains lie on a trajectory that corresponds to a TTG-like source that formed at ~3.95 Ga. Note that the excursion of the $\epsilon_{\text{Hf}(t)}$ towards positive values starts at ~3.5 Ga. The three distinct periods of crust formation in Singhbhum are indicated by three arrows at the bottom

Pandey *et al.* (2019) report similar, but in general lower, mean $\epsilon_{\text{Hf}(t)}$ values for the Singhbhum Granite samples (~0.3+0.9 to +2.0+0.6). The latter values are near-chondritic and they argue against the recycling of large amounts of much older evolved crust as the source of granitoids of the SC and opine that the data are more consistent with a primitive near-chondritic mantle source. Pandey *et al.* (2019) have also reported that, samples showing high $\epsilon_{\text{Nd}(t)}$ values of +1.6 to +5.2 (suggesting a strongly depleted source) have near-chondritic $\epsilon_{\text{Hf}(t)}$ values (suggesting primitive mantle source). They have commented that $\epsilon_{\text{Nd}(t)}$ is determined from whole rock compositions and is sensitive to later modifications affecting the rocks. The Hf isotopic data from zircon are more robust to later modifications of the whole rock, and hence Hf isotopes in zircon yield more reliable measures than whole rock Nd isotopes.

Other Granite Bodies

The Bonai Granite (BG) is dominantly porphyritic and also equigranular trondhjemite-granodiorite with subordinate granite (Fig. 2B) exposed west of the main SBG body. Xenoliths of trondhjemite-tonalite, banded migmatite, and older supracrustals, such as, quartzite, BIF and mafic-ultramafic rocks are common in both varieties of the BG. Chakraborty *et al.* (2019) reported zircon concordia age of 3370+10 Ma as emplacement age of the Bonai Granite, and 3033+38 Ma, 2809+29 Ma, 2531+ 22 Ma as metamorphic/hydrothermal ages.

The Mayurbhanj Granite (MBG) forms a batholithic body, covering ~1000 km² in the southeastern side of the Singhbhum craton adjacent to the Singhbhum Shear Zone and is intrusive into the SBG and the IOG (Fig. 1). The MBG is a K₂O-rich suite dominantly of granite (Fig. 2B) having slightly enriched LREE with conspicuous Eu-anomaly and flat HREE pattern (Saha, 1994; Nelson *et al.*, 2014) (Fig. 3E); one sample shows no Eu anomaly and high HREE abundance (Fig. 3E). Zircon grains from a porphyritic granophyric granite have yielded a concordia age of 3087+3 Ma (Nelson *et al.*, 2014) which confirms the age (3092+5 Ma) determined by Misra *et al.* (2002). Chakraborti *et al.* (2019) determined an age of 3089+6 for zircon from gneissic granite belonging to the MBG suite. Therefore, the MBG is a Mesoarchean unit involved in Proterozoic deformation and metamorphism because of its proximity to the Proterozoic fold belt of North Singhbhum. More detailed work is necessary to resolve the problem of relation between the MBG and the SBG. The crystallization age of the MBG marks the time of stabilization of the craton.

A body of A-type alkali-feldspar bearing granitic gneisses within host mica schists was named as Mayurbhanj Granite by Saha (1994); it has recently been named as Bangripasi Granite by Chakraborti *et al.* (2019). The mafics include iron rich biotite and hastingsite. The foliation in the gneisses is mylonitic. Their REE pattern is similar to that of the MBG (Fig. 3E) with LREE enrichment, prominent Eu anomaly and flat HREE. The concordia ages of oscillatory zoned zircon grains are 3079+7 Ma and 3115+10 Ma, similar to the age of the MBG. The T_{DM} is 3.4 Ga and the $\epsilon_{\text{Nd}(t)}$ varies between 0.5 and 1.4.

The *Pallahara Gneiss (PG)* occurs in the southwestern part of the craton as an undifferentiated gneissic terrain (Fig. 1). Having an overall compositional range from granite to quartz monzonite (Fig. 2B), it is characterized by distinct gneissic banding with alternate dark hornblende-magnetite-rich bands and light quartzo-feldspathic bands. Chemically the rocks are calc-alkaline, ferroan, weakly metaluminous granites, depleted in Al, Ca and Mg and rich in LILE and HFSE and are classified as A-type granites with high Y/Nb ratios. In the Y-Nb discrimination diagram their plots fall in the field of within-plate granite. Zircon dating of Pallahara Gneiss fixes its time of crystallization at ~2.8 Ga (Chattopadhyay *et al.*, 2015; Topno *et al.*, 2018). The intrusion of Pallahara granitoids is suggested to be coeval with prominent mafic magmatism in the Singhbhum Craton (e.g., the Dhanjori mafic volcanic rocks and mafic dyke swarm) and with the Tamperkola Granite (2809±12 Ma).

Older Supracrustal Belts

Older Metamorphic Group (OMG)

OMG occurs as enclaves of upper greenschist to amphibolite facies pelitic schists (biotite + muscovite + quartz + garnet + sillimanite), quartz-muscovite schist, quartz-magnetite-cummingtonite schist, fuchsite quartzite, banded calc-gneiss, and layered (?para-) and massive to schistose (ortho-) amphibolites spread all over the Singhbhum Granite outcrop (Fig. 1). The type area of the OMG is Champua where the OMG outcrops cover an area of ~200 km² (Fig. 1). On the basis of major element and trace element composition of the OMG biotite-muscovite schists Hofmann and Mazumder (2015) concluded that the protolith was siltstone or greywacke and the composition was similar to that of Archean greywackes and shales such as those of the Fig Tree Chert of the Barberton greenstone belt. The schists contain detrital zircon grains suggesting granitic or granite-derived sedimentary provenance of the metasediments. The REE pattern of the OMG quartzite shows high LREE enrichment with a small negative Eu anomaly and low HREE abundances, similar to that of the associated tonalite gneisses (Nelson *et al.*, 2014). This is consistent with their derivation from principally granitoid rocks. Relatively high Cr, Ni and Mg content in some schists and fuchsite quartzite suggests presence of mafic/ultramafic material also in the

source area (Hofmann and Mazumder, 2015). No komatiite has been reported so far from the OMG, but from Jagannathpur area (Fig. 1) Manikyamba *et al.* (2015) have reported that the geochemical signatures from some samples of metamorphosed mafic rocks indicate boninitic affinity; these are associated with samples having composition of island arc tholeiites. The authors invoke magma generation in an intra-oceanic subduction zone setting and infer a depleted, refractory mantle source that has been subsequently enriched by subduction-related slab-dehydration process. The rock association and the geochemical signatures suggest that the OMG outcrops represent fragments of the Archean oceanic crust, probably in an ancient (Paleoarchean) subduction system similar to, but younger than, those recorded from the Nuvvuagittuq Greenstone Belt of Canada (O'Neil *et al.*, 2012) and Isua Greenstone Belt of Greenland (Polat *et al.*, 2002, 2011).

The OMG was considered to be the oldest stratigraphic unit of the craton (Saha, 1994). Older Sm-Nd isochron dating fixes the age of formation of the OMG amphibolite as 3305 ± 60 Ma (Sharma *et al.*, 1994) (Table 1). ⁴⁰Ar/³⁹Ar ages of amphibole in amphibolite tallies with this age (Bakshi *et al.*, 1987). Therefore this age is likely to represent a metamorphic age. Ages of detrital zircon grains in sedimentary rocks help in fixing the upper limit of the age of deposition. The available age data from the OMG indicate that the deposition of the supracrustal rocks may be bracketed within ~3.37~3.35 Ga (Table 1, Fig. 4). Saha *et al.* (2012) opined that the unimodal distribution of detrital zircon age indicates either a volcanoclastic origin of the rock or derivation by erosion of a single igneous source. U-Pb concordia age from metamorphic rutile in OMG mica schist fixes the metamorphic age as 3317±8 Ma (Upadhyay *et al.*, 2014). Hence not much time elapsed between the deposition and the earliest metamorphism.

The OMG supracrustal assemblage was probably formed as a greenstone-belt-type succession in a supra-subduction setting at ~3.37 Ga, and it has been suggested that they represent higher grade equivalents of the Iron Ore Group of rocks (Hofmann and Mazumder, 2015).

Iron Ore Group (IOG)

The Archean volcano-sedimentary greenschist facies

supracrustal rocks of the SC conventionally go by the name Iron Ore Group (IOG). These occur either as peripheral basins around the Singhbhum Granite batholithic complex or as keels within it (Fig. 1). The three principal volcano-sedimentary supracrustal belts, also known as the IOG basins, are: the western IOG belt (Noamundi-Jamda-Koira belt), the eastern IOG belt (Gorumahisani-Badampahar belt) and the southern IOG belt (Tomka-Daitari belt) (Fig. 1). The stratigraphic relation between the SBG and the IOG has long been a point of debate (Mukhopadhyay, 1976, 2001; Sarkar and Saha, 1977; Saha, 1994). The main reason for assigning a younger age to the IOG is the interpretation of a sandstone-conglomerate unit unconformably overlying the SBG as the basal unit of the IOG (Murty and Acharya, 1975; Mukhopadhyay, 1976). However, recent work by Ghosh *et al.* (2015), Ghosh *et al.* (2016b) and Mukhopadhyay *et al.* (2007, 2013, 2014, 2016) have shown that the siliciclastic rocks along the eastern margin of the western IOG belt and along the southern and southeastern margin of the southern IOG belt have an unconformable or tectonic contact with the IOG and do not constitute the basal unit of the IOG.

Recent geochronological data have confirmed the antiquity of the IOG belts. The $\sim 3380 \pm 11$ Ma age of the granite intruding the IOG at the Deo River (Nelson *et al.*, 2014) stipulates that the BIF of the western belt cannot be younger than ~ 3.38 Ga. Zircon grains from a tuff layer less than 1 m below the BIF of the western belt yield concordant age of 3392 ± 29 Ma (Basu *et al.*, 2008). The age of the monzogranite near Rairangpur (3326 ± 5 Ma, Nelson *et al.*, 2014) with enclaves of BIF also fixes a lower limit of the age ($> \sim 3.3$ Ga) of the eastern IOG. Zircon grains from a dacitic lava immediately underlying the BIF of the southern belt (Tomka-Daitari) conformably yield a concordia age of 3507 ± 2 Ma and 3505 ± 5 (Mukhopadhyay *et al.*, 2008, 2012; Sreenivas *et al.*, 2019). Therefore, the deposition of the IOG supracrustals is as old as ~ 3.5 - 3.4 Ga and is of the same order as or older than the age of the OMG supracrustals. The ages of all the three IOG belts fall in this bracket.

Banded Iron Formation

It is probable that the IOG was deposited in three separate basins, which had somewhat different

depositional settings, and small differences in depositional age. There are some differences in the mineralogy, geochemistry and lithological association of the BIFs in different belts, particularly between the western and the eastern belts. BIFs are more magnetite-rich in the eastern and the southern basins than in the western one. Sarkar and Gupta (2012) have summarized the differences between the three IOG basins.

In the western IOG belt, basal massive or pillowed metabasalt is followed upwards successively by chert breccias, lower phyllitic ferruginous shale/tuff (with local thin carbonates and manganiferous units in the lower part), BIF-iron ore, and upper phyllitic shale that is ferruginous or white (Ghosh and Mukhopadhyay, 2007; Beukes *et al.*, 2008). The chert breccia overlying the mafic lavas indicates falling sea level and aerial exposure. Structures resembling stromatolites in some of the chert clasts suggest that the breccias could have formed through solution collapse and silicification of an earlier carbonate buildup during exposure. This solution collapse breccia is overlain by a transgressive shale unit with manganiferous shale being deposited in the early stages of transgression or during low strand, which is overlain by ferruginous shale. Iron formation was deposited on top of it during maximum flooding when the basin became starved of siliciclastic input.

Trace and REE patterns with strong Eu anomalies of the iron oxide phases of the BIFs suggest exhalative hydrothermal fluids with variable, but overall high, temperatures as source for the major part of the Fe; other trace elements and REEs were contributed to the bottom seawater by diagenetic fluids and fluvial discharge to the ocean. Variable Eu anomalies in the samples indicate mixing of hydrothermal fluids of different temperatures with the bottom seawater. Insignificant Ce anomaly is attributed to anoxic nature of the Archean bottom sea water (Bhattacharya *et al.*, 2007).

The overall structural trend in the different IOG supracrustal belts change from NNW-SSE in Gorumahisani-Badampahar to NE-SW near Nuasahi, ENE-WSW in Tomka-Daitari, NW-SE near Pallahara-Malayagiri and NNE-SSW in Noamundi-Jamda-Koira (Sarkar and Saha, 1977). Thus the trends curve round the Singhbhum Granite body, suggesting

that the major deformation in the supracrustal belts was moulded round the granitic block (Mukhopadhyay, 1976; Sarkar and Gupta, 2012). The eastern belt probably represents a subvertical synformal keel within a granitic terrane. In the western IOG, Ghosh and Mukhopadhyay (2007) concluded that two phases of deformation have given rise to the outcrop pattern in the 'horse-shoe' structure (Fig. 6A), which is structurally a synclinorium. Its western limb is overturned and steep westerly dipping and the eastern limb is gently westerly dipping. On the eastern limb a second order asymmetrical fold pair congruous to the major synform is developed; on the subhorizontal short limb of this asymmetric fold pair (their eastern asymmetric anticline) third order symmetrical folds with subvertical axial planes are developed (Fig. 6B). In the southern IOG belt the first and second generation folds are co-axial and have E-W trending axial planes; the third generation folds are N-S trending (Ghosh *et al.*, 2010a; Bose *et al.*, 2015). The D₂ structures are thought to be related to the top-to-the-north motion along the Sukinda Thrust.

IOG Metavolcanic Rocks

The basal parts of the supracrustal sequences in all the three IOG basins are mafic-ultramafic volcanics, often with well preserved pillow structures (Sengupta *et al.*, 1997; Bose, 2009). The IOG volcanics often have high MgO (up to 13%) with high SiO₂ (up to 52%) and some may be characterized as siliceous high magnesia basalt (SHMB). The mafic rocks in all the three IOG belts are mostly tholeiite, but basaltic andesite and andesite variants are also found. Rare ultramafic rocks with cumulate texture are present in all the three belts.

Komatiitic types with spinifex texture have been reported from several localities in the eastern belt (Chaudhuri *et al.*, 2015; Ghosh *et al.*, 2019). Both Al-depleted (ADK) and Al-undepleted komatiites (AUK) are present. The basal cumulate part of the komatiite is chemically similar to ADK which is characterized by subchondritic Al₂O₃/TiO₂ (9.38-10.5) ratio and depleted HREE, elucidating presence of majorite in residuum. The upper spinifex-textured lava typically displays suprachondritic Al₂O₃/TiO₂ (~22-28.5) ratio and enriched HREE indicating majorite-free residuum. A primitive mantle-like source is envisaged which melted at greater depth (>300 km)

to produce the parental liquid of ADK (basal cumulate) and at shallower depth (~150 km) to produce the AUK melt (spinifex-textured lava) (Chaudhuri *et al.*, 2015; Ghosh *et al.*, 2019). Sm-Nd isotopic studies suggest a mantle source with long term depletion ($\epsilon_{Nd} = \sim +2$ to $+4$). Intrusion of 3.3 Ga granitoid belonging to the Singhbhum Granite suite into the eastern IOG belt (Nelson *et al.*, 2014) suggests that emplacement of these komatiites is older than 3.3 Ga. The similarity between the Singhbhum komatiite and the Barberton komatiite has been pointed out and a plume-arc geodynamic setting has been suggested.

The southerly dipping sequence in the southern IOG belt has massive or pillowed metabasalt and

Fig. 6: (A) Generalized geological map of the western IOG. (B) Schematic cross section along A-B (in Fig. 6A). The cross section shows a regional syncline with a congruous dextral large scale fold on the eastern limb

interbanded chert in the basal part which is overlain by a thick unit of dacitic lava flows and pyroclastic rocks followed by BIF (Mukhopadhyay *et al.*, 2008, 2012). The dacitic volcanic unit (coherent dacitic flow and ash-poor, non-vesiculated, welded pyroclastics) is interpreted to have formed in a deep marine setting and erupted under a steam cupola (Mukhopadhyay *et al.*, 2012). In the upper part of the volcanics there is evidence of transition from a deep water low-height volcanic chain to a shallower subaqueous eruption in an aggradational volcanic chain. This was followed by deposition of BIF. The bimodal volcanics association and the trace elements and REE characteristics of the volcanic units in the southern IOG are similar to the rocks in Phanerozoic suprasubduction arc-fore arc settings (Mukhopadhyay *et al.*, 2012).

The 3.5 Ga zircons from the dacitic lava in the southern IOG belt yield a $\varepsilon_{\text{Hf}(t)}$ value of +0.5 to +7.8 (suprachondritic $\varepsilon_{\text{Hf}(t)}$ and radiogenic Hf isotopic composition) (Sreenivas *et al.*, 2019) and mostly plot close to the depleted mantle (DM) curve (Fig. 5), thereby indicating that by ~3.5 Ga the depleted mantle was established under the Singhbhum Craton and was available for magma generation.

Malangtoli and Jagannathpur Volcanics

Contrasted views exist about the stratigraphic position of the Jagannathpur Volcanics (or Dungoaposi Volcanics) north of Noamundi and the Malangtoli Volcanics south of the horse-shoe closure of BIF (Fig. 1). Some workers (Saha, 1994; Singh *et al.*, 2016; Singh *et al.*, 2017) regard these to be younger than the IOG and lying unconformably over the latter, while others regard them to be equivalent to the basal IOG mafic volcanics (Dunn, 1940; Acharyya, 1993; Sengupta *et al.*, 1997). The Malangtoli Volcanics with subhorizontal primary layering underlies the gently folded BIF with an overall flat attitude (Sengupta *et al.*, 1997), and hence the field relations confirm an infra-BIF position of the Malangtoli Volcanics. There are no direct age data on the Malangtoli Volcanics. Singh *et al.* (2016, 2017) made a comparative study of the IOG Volcanics near Noamundi and the Malangtoli Volcanics and pointed out that the IOG and the Malangtoli basalts exhibit similar mineralogical composition, but the IOG volcanics samples are andesitic with a calc-alkaline affinity, while the Malangtoli Volcanics samples are basalt to andesite

with tholeiitic to calc-alkaline trend. The Malangtoli Volcanics have both low-MgO and high-MgO varieties; some are boninitic in composition. The IOG basalts show enriched LREE and flat HREE and negative Eu-anomaly; some samples of the Malangtoli Volcanics show a similar REE pattern while the others have nearly flat LREE and HREE (Singh *et al.*, 2016, 2017). The IOG basalts show low to moderate PGE contents (26.23-68.35 ppb of SPGE), whereas the Malangtoli basalts show a moderate to high concentration of PGE (SPGE = 43.01-190.43 ppb). The geochemical attributes point to a subduction-related origin for both. The IOG volcanics were generated in an intra-oceanic arc setting, while the Malangtoli Volcanics originated in an arc to rift-controlled back-arc setting. Arc basalts and boninites in the Malangtoli Volcanics are interpreted to be the products of juvenile subduction processes involving shallow level partial melting of mantle wedge under hydrous conditions triggered by slab-dehydrated fluid flux. High magnesian andesites and the Nb-enriched basalts represent the intermediate and mature stages of subduction respectively (Singh *et al.*, 2017) and slab-melting and mantle wedge hybridization processes were involved in their melt generation. The IOG volcanic rocks are modelled to have been derived by <1-1% partial melting of garnet lherzolite whereas the Malangtoli basalts were generated by 3-<10% partial melting from spinel to garnet lherzolite. Banerjee *et al.* (2016), on the basis of tectonic discrimination diagrams, supported the idea that the Malangtoli Volcanics have island arc tholeiitic affinity; in other words they were erupted in a supra-subduction setting.

The Jagannathpur Volcanics exposed near Noamundi along the eastern margin of the western IOG belt dips gently towards SW and show gentle domes and basins structure. Some of the volcanics resemble boninites with high MgO, low Ti/Zr and high Zr/Y. The Jagannathpur and Malangtoli Volcanics are apparently undeformed, but the AMS studies (Mukherji *et al.*, 2004; Ghosh *et al.*, 2010b) reveal that they share the same magnetic fabric as the IOG mafic rocks and on this ground the three are correlated.

Younger Supracrustal Belts

Dhanjori Basin

Dhanjori basin is situated in the north-eastern corner

of the Singhbhum Craton (Fig. 1). It is a NW-SE trending, elongate basin that is caught up in the deformation of the Mesoproterozoic orogeny represented by the North Singhbhum Mobile Belt. The Dhanjori succession, predominantly consisting of volcanics and volcanoclastic and siliciclastic sediments, unconformably overlies the granitoids of the SBG pluton. The Dhanjori Group has been subdivided into Upper and Lower Formations (Mazumder and Sarkar, 2004; De *et al.*, 2015). The Lower Dhanjori Formation consists of pebbly phyllite, quartzite and thin conglomerates with minor mafic-ultramafic intrusions and the Upper Dhanjori Formation consists of volcanics and volcanoclastic rocks along with quartzites and phyllites. The Phuljhari Formation disconformably underlies the polymictic conglomerate-quartzite of the Lower Dhanjoris. The basal Dhanjoris are profusely intruded by the ~3.1 Ga old Mayurbhanj Granite Acharyya *et al.* (2010). Hence, at least the lower part of the Dhanjori Group is older than ~3.1 Ga.

On the basis of sediment characteristics and Sm-Nd isotopic data De *et al.* (2015) suggested proximal sources and poorly mixed provenance with both mafic and felsic rocks. They further suggested a provenance in a juvenile felsic Andean-type magmatic arc situated on the Archean basement. Mazumder (2006) observed that the magmatic component of the Dhanjori Formation represents a Large Igneous Province (LIP), and Mazumder and Arima (2009), on the basis of major, trace, rare earth elements composition, and selected elemental ratios (Nb/Th, Zr/Nb, Nb/Y, and Zr/Y) from the Upper Dhanjori volcanics and volcanoclastics and the sedimentological and stratigraphical characteristics of the interbanded metasedimentary rocks, inferred an intracontinental rift setting for Dhanjori volcanism and sedimentation. Bose (2009) also inferred extensional tectonics for the Dhanjori basin.

Earlier whole rock Sm-Nd isochron dating led to contradictory results (Roy *et al.*, 2002; Misra and Johnson, 2005) (Table 1). Acharyya *et al.* (2010) carried out U-Pb dating of zircon grains from the acid tuff in the Phuljhari Formation. The U-Pb dates from the zircon grains were highly discordant and no precise age estimate can be made. The upper intercepts of the discordia lines with the concordia curve give ages of 3330±58 and 3358±110 Ma. Since the Mayurbhanj

Granite with probable age of ~3.1 Ga intrudes (see Table 1) the Phuljhari Formation the Dhanjori Group ought to be >3.1 Ga old.

The *Simlipal Basin* has a circular outcrop pattern and is surrounded by the SBG or undifferentiated granite (Fig. 1). The volcano-sedimentary sequence of the basin overlies the Singhbhum and Kaptipada Granites and IOG of the Badampahar belt. It starts with basal quartz pebble conglomerate, orthoquartzite and feldspathic quartzite; the basal units are overlain by a volcanic pile (spilitic lavas and tuffs), with three intercalated bands of cross bedded quartzites (Iyengar and Banerjee, 1964). The origin and age of the basin remains unresolved. Along the flanks a suite of gabbro-anorthosite and the Mayurbhanj Granite intrude into the volcanics (Saha 1994), testifying to its pre-MBG status. The Simlipal Volcanics is correlated with the Dhanjori Group mainly because of similar lithological associations in the two basins and their close proximity, but definite evidence about the correlation is lacking. Naqvi and Rogers (1987) floated the idea that the Simlipal basin may be an astrobleme or impact crater, but so far no evidence of any meteorite impact has been found.

Other Metasedimentary Sequences

Mahagiri Quartzite, Keonjhar Quartzite, Mankaharchua Quartzite, Malayagiri Supracrustals, Devgarh Quartzite

Extensive siliciclastic rocks are exposed along the western boundary of the Singhbhum Granite pluton separating the latter from the western IOG belt (Ghosh *et al.*, 2015) (Fig. 1). The ~1.5-km-thick siliciclastic succession, exposed between Joda and Keonjhar, has been named as Keonjhar Quartzite (Ghosh *et al.*, 2016b). Similar rocks extend from west of Keonjhar along the south-eastern margin of the SBG up to Sukinda where the rocks are named as Mahagiri Formation or Mahagiri Quartzite (Mukhopadhyay *et al.*, 2012, 2014) (Fig. 1). Siliciclastic rocks are also exposed north of Pallahara near Mankaharchua lying unconformably over the Malangtoli Volcanics (Fig. 1). U-Pb ages determined from detrital zircons from samples of sandstone coming from a large area extending from Keonjhar to Mahagiri and Mankaharchua cluster between 3950 to 2909 Ma (Mukhopadhyay *et al.*, 2014; Sreenivas *et al.*, 2019).

The age of the granitic basement on which the sandstone lies is 3.29 Ga, and this sandstone is intruded by ~2.8 Ga old granitoids at the Mahagiri Range (Mukhopadhyay *et al.*, 2014). Thus the time of deposition of the Mahagiri Quartzite must be between 2.91 and 2.8 Ga. Paleo- and Mesoarchean ages (3510–3176 Ma) are dominant among the detrital zircon population. Some Eoarchean grains have ages of 3950–3601 Ma.

Ghosh *et al.* (2016b) inferred a continental interior setting for the sandstones. Several geochemical proxies indicate their highly matured and recycled nature and they have a mixed source of predominantly granodioritic and primitive mafic rocks belonging to a differentiated felsic upper continental crust (UCC). Evidence also points to the presence of both unweathered and deeply weathered source rocks. The rare earth element patterns of the siliciclastics are similar to those of the post-Archean shales. Recycling and intracrustal fractionation processes imply craton stabilization and confirm the existence of a Mesoarchean continental crust comparable to the post-Archean UCC from other parts of the world (Ghosh *et al.*, 2016b). The detrital uraninite grains in Mahagiri and Keonjhar Quartzites suggest the existence of highly felsic and K-rich (richer than TTG) granodiorite-granite-monzogranite suites of rocks older than 2.9 Ga in the Singhbhum craton. Dey *et al.* (2017) have reported superchondritic $\epsilon_{\text{Hf}(t)}$ values (+1 to +8.75) from detrital zircons in the Mahagiri Quartzite, suggesting predominantly juvenile mantle source. The radiogenic Hf isotopic ratio in the depleted mantle source is interpreted to be typical of I-type granite that largely forms in island-arc type accretionary settings.

The siliciclastic sediments are underlain by pyrophyllite-bearing paleosol (Bandyopadhyay *et al.*, 2010) formed on the surface of the 3.29 Ga old (Tait *et al.*, 2011) granite. This is the oldest documented Archean paleosol in the Indian shield (Mukhopadhyay *et al.*, 2014). Soft-sediment deformation in the paleosol near the contact with the overlying sandstone indicates that the paleosol was wet at the onset of deposition of the overlying siliciclastic sediments. The soil type suggests alternation of very wet and very dry climate (Bandyopadhyay *et al.*, 2010). Geochemical evidence indicates that oxidative weathering occurred in the original upper soil profile,

and the paleosol was formed before the deposition of sediments. On the other hand, presence of detrital uraninite and pyrite at the base of the siliciclastic sequence points to low surface oxygen condition during the time of deposition. Hence, Mukhopadhyay *et al.* (2014) have invoked that a short-lived atmospheric oxygenation event occurred about 600 Ma prior to the Great Oxidation Event. This probably took place because of the presence of organisms capable of oxygenic photosynthesis.

The BIF-bearing Malayagiri sequence unconformably overlies the ~2.8 Ga old Pallahara Gneiss (Saha, 1994) and the zircon grains from a dacite in the supracrustal sequence have a concordant age of 2806±6 Ma (Nelson *et al.*, 2014). Furthermore, the age of the youngest detrital zircon population from quartz-kyanite-muscovite schist is 2788±25 Ma (Chattopadhyay *et al.*, 2015). Chattopadhyay *et al.* (2015) concluded that the deposition of the Malayagiri supracrustals was between 2.79 and 2.42 Ga and the rocks were affected by three major metamorphic events at ~2.42 Ga, 0.98–0.94 Ga, and 0.57–0.54 Ga. Hence, the Malayagiri supracrustals are younger than Keonjhar Quartzite or Mahagiri Quartzite.

Kolhan Group

The main Kolhan basin, covering an area of ~600 km², extends from Chaibasa to Noamundi, and straddles the boundary between the IOG of the western belt and the Singhbhum Granite (Dunn, 1940) (Fig. 1). The Kolhan sequence unconformably overlies the Singhbhum Granite and the Iron Ore Group, and is the youngest stratigraphic unit in the Singhbhum craton.

Bandopadhyay and Sengupta (2004) suggested that the shale-dominated succession of the Kolhan sequence was deposited in continental rift setting. The lower red and purple sandstones with beds/lenses of conglomerates were deposited by shallow, ephemeral braided streams. The dominantly parallel-bedded sandstones with sheet-like geometry of the sandstone beds and rare channel-form structures imply that much of the sediments is a product of sheet floods, while the overlying thick and extensive shale was deposited in lacustrine environment. Petrography and geochemistry of the shales and CIA (Chemical Index of Alteration) values of the Kolhan siliciclastics, suggest deposition in plate interior, either at stable

continental margin or in intracratonic basin. Mukhopadhyay *et al.* (2006) have described the Kolhan succession as a siliciclastic shelf to deep-water carbonate platform succession. They suggested that the Kolhan succession represents a syn-rift to post-rift succession, which is equivalent to the terminal Mesoproterozoic carbonate platform successions developed in other parts of peninsular India. It represents an event of major transgression and relative sea level rise.

Ghosh *et al.* (2015) reported pervasive deformation features from the Kolhan basin and pointed out that the north-western boundary of the basin is marked by a major thrust fault, named as the Chaibasa thrust, with top-to south displacement along which older Archean succession was thrust over the younger Kolhan rocks.

Darjing Group

The region north of the Bonai Granite extending up to the Gangpur Group is occupied by a sequence of low-grade metamorphic rocks (Fig. 1) for which Mahalik (1987) proposed the name Darjing Group. Alternative stratigraphic groupings of the rocks of this region have been proposed by Sen (2001) and Naik (2001). The differences in interpretations are to be sorted out by detailed mapping and facies analysis of the sediments. The maximum age of the Darjing Group is constrained by the 3129 ± 10 Ma age of the youngest detrital zircon grains (Chakraborty *et al.*, 2019). It is intruded by the Tamperkola Granite having an age of 2809 ± 12 (Bandyopadhyay *et al.*, 2001). So the age of the Darjing Group is between 3.1 and 2.8 Ga.

Chakraborty *et al.* (2019) used the term Upper Bonai Group for the supracrustal sequence that includes the Darjing Group and extends up to the SBG pluton. Two major metamorphic episodes are recorded in the rocks – an earlier M_1 episode produced garnet₁ and staurolite, M_2 produced garnet₂ which grew at the expense of staurolite. The estimated P-T conditions are: M_1 -520°-580°C, 4.8 kb; M_2 -650°C, 7.4 kb. The Upper Bonai Group is intruded by garnet-bearing granites which are deformed.

Supracrustal Rocks with Debatable Stratigraphic Status

The low grade phyllites, tuffs and quartzites at the

NW corner and along the northern part of the western edge of the SBG pluton have a debatable stratigraphic status. Ghosh *et al.* (2015) inferred the presence of two thrust sheets within the phyllites and quartzites of this area (Fig. 7A). The upper boundary of this stack of thrust sheets is the Singhbhum Shear Zone (SSZ) of the North Singhbhum Mobile Belt (NSMB), and lower boundary is with the Kolhans. According to this interpretation, this group of siliciclastic rocks is older than the Dhanjoris. Bhattacharya and Mahapatra (2008) grouped the rocks unconformably lying over the northern and northwestern boundary of the Singhbhum Granite in two units, the Bistrampur Formation and the Nayadih Formation, which were thrust over by the Iron Ore Group and the Dhanjori Group in the north (Fig. 7B). This is an interesting hypothesis, but it needs to be confirmed by more detailed structural mapping over a larger region and geochronological work.

Ultramafic-Mafic-Anorthosite Suites

Ultramafic-mafic-anorthosite suites mostly occur along the eastern side of the Singhbhum Craton as a number linear discontinuous bodies extending from south of Haludpukur to Nuasahi and Sukinda (Fig. 1). Ultramafic rocks are also present at Jojohatu, south of the Ongarbira Volcanics and also east of the Bonai Granite (Fig. 1). The ultramafic rocks of Sukinda-Nuasahi are intrusive into the Singhbhum Granite and are intruded by the Mayurbhanj granite and granophyre (Saha, 1994). Augé *et al.* (2003) reported 3123 ± 7 , 3122 ± 5 and 3119 ± 6 Ma zircon ages from the gabbroic suite at Nuasahi, and 3205 ± 280 Ma Sm-Nd isochron age of the gabbroic suite. Anorthositic rocks are found near Gorumahisani and Nuasahi. At Sukinda and Nuasahi the intrusive body is a layered ultramafic body, with ultramafics in the lower part and gabbro with anorthosite in the upper part, but there is no direct age determination of the emplacement of ultramafics. The ultramafic rocks are picrite, lherzolite, orthopyroxenite, harzburgite, dunite and chromitite and characteristically show adcumulate texture. The parent magma is boninitic or siliceous high-Mg basalt (SHMB) (Khatun *et al.*, 2014; Mondal, 2009, Mondal *et al.*, 2019). The geochemistry and REE patterns suggest that the boninitic parental magma was generated by second-stage melting of a metasomatized depleted mantle source in a supra-subduction setting. The spidergrams of trace elements in gabbro samples

Fig. 7: (A) Geological map of the northwestern edge of the Singhbhum Granite pluton, showing the stacked thrust sheets over the Kolhan Group (after Ghosh *et al.*, 2015). (B) Interpreted stratigraphic set up at the northwestern corner of the Singhbhum Granite pluton (after Bhattacharya and Mahapatra, 2008)

show negative Nb-, Th-, Zr-, and Hf-anomalies indicating derivation from a depleted mantle (Mondal and Zhou, 2010). $[\text{Nd(t)}]$ value of $+2.63 \pm 0.88$ for the high-Mg gabbroic suite is consistent with an origin from a long-term depleted source region. The high Os concentrations (and IPGE-rich character in general) of the studied Archean chromitites are supportive of the idea of a high degree of melting of a previously depleted upper mantle source domain for

generating the parental melts of chromite-bearing ultramafic suites (Mondal *et al.*, 2004, 2006). This points to the presence of early Archean sub-continental lithospheric mantle (SCLM) domain beneath the SC which represents the stable residue after extensive melt extraction in the Archean. Os model age calculation from the chromites indicate that depletion from the primitive mantle to form the SCLM started as early as ~ 3.7 Ga, nearly 500 to 600 Ma earlier than the crystallization of chromite (Mondal *et al.*, 2007). This is consistent with the conclusions derived earlier from the isotopic data from granitoid rocks. Alternatively, the parent melts may have come from depleted zones surviving within the convective mantle. Mondal *et al.* (2007) hypothesized that the parental magma of the ultramafic suite was produced in a supra-subduction setting and intruded the volcano-sedimentary greenstone belts at shallower crustal levels. They envisioned a process of early subduction of sections of melt-depleted subchondritic oceanic lithospheric mantle with large degree melt extraction events and concomitant major crust building processes during middle-to late-Archean in the Singhbhum Craton. A supra-subduction setting has recently been suggested to explain the existence of boninite-like metabasalts in the 3.7 Ga old Isua Greenstone Belt in Western Greenland (Polat *et al.*, 2002; Polat and Frei, 2005). Mondal (2009) has proposed that similarity of the sub-continental lithospheric mantle characteristics of the Singhbhum Craton and the Zimbabwean craton suggest that South Africa and India were parts of the same supercontinent in early Archean (Myers, 1995).

Newer Dolerite Dyke Swarm

Spectacular dyke swarms called the Newer Dolerite traverse the Singhbhum Granite pluton and occasionally occur within the Bonai Granite, Kaptipada Granite and Mayurbhanj Granite. Such dyke swarms are interpreted to represent the plumbing systems of LIPs which are deeply eroded to expose the feeder dykes (Ernst, 2014). Hence, they are useful for identifying Large Igneous Provinces (LIPs) and for paleogeographic reconstruction (Srivastava *et al.*, 2016). The dominant trend of near vertical dykes within the Singhbhum Granite is NNE-SSW to NE-SW; subsidiary trends are NW-SE and E-W. The dimension of the dykes varies from a few tens of cm to one km in width and from a few hundred metres up to 20 km in length.

The dykes are mostly made up of quartz dolerite, and locally of norite, peridotite and lamprophyre. Rarely acid rocks, such as, micropegmatitic dolerite, microgranite, xenolithic granophyre and leucogranophyre are also associated with the dolerite along the margins of the dykes.

Recently, Pb-Pb baddeleyite ages of 1766 ± 1 and 1765 ± 1 Ma were obtained for the WNW'ly trending swarm (Shankar *et al.*, 2014). Dykes of this trend distinctly cut across dykes of other orientations and possibly represent one of the youngest swarms in the region. Coeval mafic magmatism has been found in North China Craton, Kimberley Basin in Australia, Rio de Plata Craton, Uruguay and Brazilian craton. This may point to the assembly of supercontinent Columbia. Kumar *et al.* (2017) recognized three distinct trends in the Newer Dolerite dyke swarm: NNE-SSW, NW-SE and WNW-ESE. Based on cross-cutting field relations the NNE'ly trending swarm appears to be the oldest. Six dykes from this swarm have yielded Pb-Pb baddeleyite ages ranging between 2764 Ma and 2760 Ma, with a weighted mean age of 2762 ± 2 Ma. Baddeleyite from the NE-SW to ENE-WSW trending Kaptipada swarm has been dated by Srivastava *et al.* (2016, 2019). They give a weighted mean of 2251 ± 2 Ma for the $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the samples.

The mean paleomagnetic pole position obtained from dykes of the NNE-SSW trending swarm by Kumar *et al.* (2017) is 14°N , 78°E . This is comparable to one of the pole positions obtained earlier by Verma and Prasad (1974). The similarity of determined paleolatitudes and the broad contemporaneity of the 2762, 2800, 2752 Ma magmatic events in the Singhbhum Craton, the Pilbara Craton of Australia, and the Kaapvaal Craton of South Africa led Kumar *et al.* (2017) to suggest that these three cratons were assembled together in the Neoarchean time to form the Vaalbara Supercraton.

Based on field relations and U-Pb dating the Newer Dolerite dykes have been grouped by Srivastava *et al.* (2019) in seven distinct swarms. These range in age from ~ 2.80 Ga to 1.76 Ga, and include the following:

1. Keshargaria swarm - ~ 2.8 Ga, NE-SW trending.
2. Ghatgaon swarm - ~ 2.76 - 2.75 Ga, NNE-SSW

to NE-SW trending, most extensive swarm.

3. Keonjhar swarm - Between 2.76 and 2.26 Ga, early-Paleoproterozoic, NNE-WSW to E-W trending, radiating pattern.
4. Kaptipada swarm - ~ 2.26 Ga, NE-SW to ENE-WSW trending, radiating pattern.
5. Bhagamunda swarm - Between 2.26 and 1.77 Ga, middle-Paleoproterozoic, NNW-SSE to NW-SE trending, radiating pattern.
6. Pipilia swarm - ~ 1.77 Ga, WNW-ESE trending.
7. Barigaon swarm - < 1.77 Ga, late Paleoproterozoic, N-S to NNE-SSW trending, radiating pattern.

Except for the Pipilia swarm which has similar age as the Dalma Volcanics of the NSMB, none of the other dyke swarm can be linked to a specific episode of mafic magmatic activity within the Singhbhum Craton. The Kaptipada swarm (2.26 Ga) and the Pipilia swarm (1.76 Ga) are closely matched with two swarms of comparable age in the Dharwar Craton, and Srivastava *et al.* (2019) have tentatively identified two swarms of similar ages in the Bastar Craton. They have, therefore, concluded that by at least 2.26 Ga the Dharwar, Bastar and Singhbhum cratons were joined together, and two LIPS existed at the two above mentioned times covering all the three cratons. They further suggested that the trends of 2.26 Ga dykes in the three cratons display a radiating pattern and they converge to define a possible plume centre at the northeastern end of the Dharwar-Bastar join. The 1.76 Ga swarms in the three cratons, on the other hand, display parallel trends. They had also correlated the NNE-SSW trending Ghatgaon swarm (2.76-2.75 Ga) of the Singhbhum Craton with the WNW-ESE trending Keshkal swarm (2.7 Ga) of the Bastar Craton, and the dyke trends converge to a possible plume centre situated SE of the Bastar Craton. However, in the absence of paleomagnetic data, these reconstructions would remain speculative.

Window on the Hadean and the Eoarchean Earth

Paleoarchean ages dominate among the determined ages from different granitoid units making up the SC. But detrital zircon grains in the supracrustal rocks

and xenocrystic zircon grains in the granitoid gneisses provide information about the existence of older crust. The existence of Eoarchean continental crust was first inferred by Goswami *et al.* (1995) from Eoarchean ages of 3628±72, 3591±64, and 3583±50 Ma from cores of detrital zircon grains.

The oldest reported Hadean ages from the SC are the concordant ages (~4.03 to ~4.24 Ga) of xenocrysts of zircons in the OMTG (~3.4 Ga old) from near Champua (Chaudhuri *et al.*, 2018); these have thin rims with Eoarchean (~3.8 to ~3.9 Ga) ages (Table 1). The cores of other xenocrysts yield Eoarchean crystallization ages (~3.7 to ~3.6 Ga). The Hadean xenocrysts have nonradiogenic $^{176}\text{Hf}/^{177}\text{Hf}$ initial compositions (0.27995 ± 0.0009 to 0.28001 ± 0.0007), and the $\epsilon_{\text{Hf}(t)}$ values range from -2.5 to -5.2, suggesting that they originated from Hadean enriched reservoir (Fig. 5). Time integrated $\epsilon_{\text{Hf}(t)}$ compositional array of the Hadean xenocrysts indicates that a mafic protolith with $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of ~0.019 was reworked during ~4.2-4.0 Ga. The isotopic data suggest that the separation of such an enriched reservoir from chondritic mantle took place at 4.5 ± 0.19 Ga, that is, soon after (after ~40 Ma) the accretion of the Earth. This mafic source may represent an enriched residual mafic crust generated from a partially solidified magma ocean. More radiogenic yet subchondritic $^{176}\text{Hf}/^{177}\text{Hf}$ values for some other ~3.6 to ~3.7 Ga old zircon grains from OMTG have values like $\epsilon_{\text{Hf}(t)} = -1.5$ to -6.6 , and ~3.4 Ga zircons have $\epsilon_{\text{Hf}(t)} = -0.4$ to -3.7 . ~3.4 and ~3.3 Ga old zircons from the Singhbhum Granite have $\epsilon_{\text{Hf}(t)} = -0.3$ to -4.7 . These have been interpreted by Chaudhuri *et al.* (2018) to indicate that the enriched Hadean reservoir subsequently underwent mixing with depleted-mantle-derived juvenile magma during Eoarchean to Paleoproterozoic.

Sreenivas *et al.* (2019) reported Eoarchean detrital zircons (3950-3601 Ma crystallization age) from Mesoproterozoic (~3.0 Ga) Mahagiri Quartzite in the southeastern part of the SC (Table 1). Their T_{DM} ranges from 4.51 to 4.15 Ga. They also show nonradiogenic $\epsilon_{\text{Hf}(t)}$ values ranging from -3.37 ± 0.5 to -9.75 ± 0.4 suggesting an enriched source. The authors interpret that the Eoarchean zircon grains represent crust generated by recycling of Hadean felsic crust formed at ~4.3 to 4.2 Ga and ~3.95 Ga (Fig. 5).

Miller *et al.* (2018) carried out U-Pb age determinations of detrital zircon grains in modern river sediments of the Baitarini River basin that is wholly contained within the SC. The data for grains which are less than 5% discordant are principally grouped in three clusters: 3.62-3.55 (5%), 3.50-3.22 (87%) and 3.10-3.06 Ga (6%). One grain (1.3% discordant) shows a Hadean age of 4015 ± 9 Ma (Table 1). Single stage Lu-Hf model ages of these grains range from 4.29 to 3.10 Ga. Since Lu-Hf model age is an estimate of the time when the protolith of the rock left the uniform chondritic reservoir the above model ages suggest the existence of Hadean, Eoarchean and younger crust. The $\epsilon_{\text{Hf}(t)}$ values show a range; 67% values are positive scattered above and below the depleted mantle growth curve, and 33% values are negative. This indicates contribution from both depleted mantle and enriched crustal sources. The Hadean grain (4.015 Ga) had an $\epsilon_{\text{Hf}(t)}$ value of -5.3 suggesting an enriched, likely crustal, influence on the parent magma.

All the data sources have yielded similar results and point to the same conclusions. Collectively, the data show a contiguous spread of zircon ages from 4.2 to 3.6 Ga (Fig. 5). Therefore, there is ample evidence of early mantle depletion and formation of enriched crust during the Hadean and Eoarchean time in the Singhbhum craton. Olierook *et al.* (2019) have drawn attention to the interesting fact that 4.24 Ga and 4.03-4.01 Ga ages obtained in Singhbhum are similar to two peaks in the global Hadean zircon record (Griffin *et al.*, 2014; Marchi *et al.*, 2014).

Evolutionary History and Geodynamic Considerations

There is an intense debate on the geodynamics of crust formation during Hadean and Archean times (Hawkesworth *et al.*, 2010, 2017; Griffin *et al.*, 2014). Basically there are two contending hypotheses; (1) plume-driven, and (2) subduction-driven development. Some hold the view that some form of subduction and plate tectonics was an important process contributing to crust formation and it started from Eoarchean and even Hadean times (Polat *et al.*, 1998, 2002, 2016; Nutman *et al.*, 2007; Reimink *et al.*, 2014; Kaczmarek *et al.*, 2016). Others suggest that plume-related processes were mainly responsible for the Archean crust formation (Puchtel *et al.*, 1998;

Polat *et al.*, 2006; van Kranendonk, 2010; Moyen, 2011; Johnson *et al.*, 2014, 2017; van Kranendonk *et al.*, 2004, 2015). For the Singhbhum Craton also a number of models have been proposed. These include: plume origin (Sharma *et al.*, 1994), repeated diapiric rise of felsic magma, punctuated by sedimentation in basins peripheral to the domical central part of the craton (Nelson *et al.*, 2014), crustal melting in response to plume driven magmatic underplating in an oceanic plateau setting (Dey, 2017), and island arc type accretionary setting (Saha *et al.*, 2004; Mukhopadhyay *et al.*, 2008, 2012; Tait *et al.*, 2011; Manikyamba *et al.*, 2015; Singh *et al.*, 2016).

The evidence for the existence of Hadean and Eoarchean continental crust that is >500 Ma older than the oldest preserved continental crust in the Singhbhum Craton has already been discussed. However, the felsic rocks of such ancient age are completely reworked and recycled attesting to a highly unstable tectonic regime. Hafnium isotopic data from zircons in Jack Hills supracrustal rocks (Yilgarn Craton) (Blichert-Toft and Albarede, 2008; Kemp *et al.*, 2010), the Acasta Gneisses (Slave Craton) (Iizuka *et al.*, 2009; Bauer *et al.*, 2017), Greenland (Fisher and Vervoort, 2018), the Isua Supracrustal Belt (Greenland) (Bennet *et al.*, 1993; Caro *et al.*, 2006) and the Nuvvuagittuq Supracrustal Belt (Canada) (O'Neil *et al.*, 2008, 2012), provide evidence of early mantle depletion and existence of enriched crust during Hadean to Eoarchean. In the Singhbhum Craton too evidence of early mantle depletion and existence of an enriched reservoir during Hadean and Eoarchean come from zircon grains older than 3500 Ma; these display exclusively subchondritic (negative) $\epsilon_{\text{Hf}(t)}$ values (Fig. 5), signifying the existence of non-chondritic enriched reservoir as early as ~4.2 Ga. It is hence inferred that the source reservoirs of the rocks with Hadean and Eoarchean zircon grains had separated from the chondritic material quite early in the Earth's history.

Chaudhuri *et al.* (2018) proposed that in the $[\text{Hf}/\text{t}]$ – Age diagram a straight line (isotopic growth curve) corresponding to $^{176}\text{Lu}/^{177}\text{Hf} = 0.019$ fits the plots of the Hadean zircon grains (Fig. 5). This corresponds to an enriched source area like a 'mafic protocrust' that separated from the chondritic reservoir at ~4.5 Ga, soon after (~40 Ma) the accretion of the Earth. They envisaged that the enriched residual mafic

magma was generated in an essentially non-plate tectonic way from a partially solidified magma ocean analogous to KREEP (K, REE, P) rich basalt below a lunar anorthositic crust (Kemp *et al.*, 2010). It has also been suggested that the primordial crust that covered the Hadean Earth when the magma ocean solidified was made up of komatiites and related rocks (Santosh *et al.*, 2017, and references therein). Hadean and Eoarchean zircon crystallized from minor silicic melts produced by differentiation or remelting of this pre-~4.2 Ga juvenile mafic protocrust. The presence of Paleoproterozoic granites with subchondritic $\epsilon_{\text{Hf}(t)}$ values (Fig. 5) signifies that this enriched reservoir was sustained till the Paleoproterozoic.

For the Eoarchean zircons from the SC, the Hf isotopic ratios are more radiogenic but still subchondritic. While Chaudhuri *et al.* (2018) are of the view that the second stage of melt generation during Eoarchean was from the mafic protocrust (enriched source) modified by its interaction with juvenile (more radiogenic) mantle melt, Sreenivas *et al.* (2019) proposed that the plots of Hadean and Eoarchean zircons fit into two broad trajectories with $^{176}\text{Lu}/^{177}\text{Hf} = 0.008$, and these correspond to TTG-like sources that formed at ~4.3–4.2 Ga and ~3.95 Ga respectively (Fig. 5); that is, according to them the Eoarchean zircons represent crust generated by recycling of Hadean felsic crust. There is no unequivocal evidence to choose between the two hypotheses. Sreenivas *et al.* (2019) remarked that the continuous involvement of the enriched reservoirs between 4.2 and 3.6 Ga is not consistent with magma generation by plate tectonic processes.

High heat production in the Archean time probably led to more rapid mantle convection which was responsible for quick recycling of the Hadean protocrust preventing its preservation. Hence the mafic Hadean protocrust in the Singhbhum craton has not survived and has been recycled and assimilated into younger voluminous TTG magma. The amphibolite enclaves within the OMTG and the SBG may represent the remnants of the oldest mafic protocrust or these enclaves could represent a modified mafic component developed from interactions between ancient enriched reservoir and mantle-derived juvenile mafic magma and were preserved as melting residuum of the mafic protolith from which granitoid magma was generated. A definite conclusion

will have to wait till we get more geochemical and isotopic data from the enclaves. Because of the quick assimilation or recycling into the mantle of the Hadean protocrust it has been suggested that the early depleted mantle was likely to be transient (Bennet *et al.*, 1993; Zeh *et al.*, 2008).

From ~3.6-3.5 Ga a shift in the isotopic composition of Hf is noticed, marked by upward excursion of $\epsilon_{\text{Hf}(t)}$ plots towards suprachondritic values (Fig. 5). This signifies that the early enriched reservoir underwent modifications through serial mixing with depleted mantle-derived juvenile magma. A large number of samples of Paleoproterozoic zircons have suprachondritic $^{176}\text{Hf}/^{177}\text{Hf}$ values which attests to the importance of magma generation from a depleted mantle reservoir at this time (Dey *et al.*, 2017). Therefore, a major depletion of the mantle and concomitant extensive continental crust formation started in the SC from 3.5 Ga. A separate depleted reservoir which was probably complementary to the ancient enriched reservoir existed under the Singhbhum craton in the Paleoproterozoic. Chaudhuri *et al.* (2018) suggested that the ~3.4 Ga komatiites of the eastern IOG belt with $\epsilon_{\text{Nd}(t)} = +2$ to $+4$ are the most plausible direct representatives of depleted lower mantle below the SC.

The evidence for Paleoproterozoic and Eoproterozoic mantle depletion also comes from the study of IOG mafic-ultramafic rocks and the chromitites of Sukinda and Nausahi (Mondal, 2009 and references therein). Mondal *et al.* (2007) concluded on the basis of Sm-Nd isotopic composition that the high-Mg mafic rocks of the IOG in Sukinda and Nausahi and the OMG amphibolites are nearly contemporaneous. Their $+\text{ve } \epsilon_{\text{Nd}(t)}$ ratios suggest a long-term depleted source region. They proposed suprasubduction setting for the generation of the parental magma that produced the ultramafic suite. The $^{187}\text{Os}/^{188}\text{Os}$ ratios of the Sukinda and Nausahi chromitites are nonradiogenic and on the basis of the isotopic data the authors suggested that a subcontinental lithospheric mantle (SCLM) started to form beneath the Singhbhum craton in early Archean (~3.7 Ga) through depletion of a primitive mantle and grew through the Archean eon. An alternative possibility is that the depleted mantle region could have remained within the convecting mantle for a few hundred million years before being added to the SCLM beneath the Singhbhum Craton.

During the Paleoproterozoic period the central part of the Singhbhum Craton was built up through successive pulses of granitic magmatism. Felsic magmatism at 3.52 Ga, 3.47-3.43 Ga, and 3.40-3.35 Ga produced the TTG-type suite of the OMTG. The ages indicate that some parts of OMTG are older than the OMG and some are younger intrusions (cf. Roy and Bhattacharya, 2012). Younger pulses at 3.32-3.35 Ga, and 3.31-3.28 Ga produced the voluminous granitic magma that engulfed the OMTG and produced the composite pluton of Singhbhum Granite. This was also a period of major metamorphism in the terrain (3.34-3.26 Ga). Younger tectonothermal events also affected the craton.

On the basis of the geochemical parameters and the $+\text{ve } \epsilon_{\text{Hf}(t)}$ values of the SBG samples studied by them Dey *et al.* (2017) concluded that the 3471 Ma-old tonalite was produced by shallow melting of a juvenile mafic source, derived from a depleted mantle, and the 3347 Ma-old granite was produced by high-temperature melting of a heterogeneous juvenile source of tonalites and mafic rocks at low crustal depths and the younger 3304 Ma-old granite was produced by high-pressure melting of tonalite-dominated source with short crustal residence. There was no direct involvement of the mantle in the origin of any of these granitoid rocks. Regarding tectonic setting Dey *et al.* (2017) proposed crustal reworking driven by episodic plume-related mafic-ultramafic magma underplating and intraplate in an oceanic plateau setting as a model for the origin of the Singhbhum granitoids. Melting of mafic crust at different depths within the thick oceanic crust led to variation in composition of the different pulses of felsic magmatism. Bédard (2006) has discussed how successive plume activities may build up a thick volcanic crust which in its lower part is metamorphosed into amphibolite. Successive underplating of magma melts the amphibolite at the base of the crust to produce successive generations of tonalite melt. The older generations of tonalites are extensively remobilized and also contribute to the younger tonalites.

In contrast to the plume hypothesis, the geochemical signatures within a 3290 Ma-old weakly peraluminous TTG body within the SBG pluton led Tait *et al.* (2011) to infer a volcanic arc paleotectonic setting at a crustal margin. Furthermore, the

geochemistry of the Mg-rich OMG volcanics with boninitic affinity point to their genesis in intra-oceanic subduction setting. From the geochemistry of sedimentary rocks sourced from OMG-OMTG Saha *et al.* (2004) inferred that the OMG-OMTG formed a subduction-related arc complex at ~3.3 Ga. The OMG and the IOG in the three belts are broadly contemporaneous or formed within a time interval of about 100 Ma (between 3.5-3.4 Ga). The IOG belts are thought to have formed in a supra-subduction setting (Basu *et al.*, 2008; Mukhopadhyay *et al.*, 2012; Singh *et al.*, 2015, 2017), with the deposition of the supracrustals variously regarded to be in rifted arc, back-arc or fore-arc basins.

Saha *et al.* (2004) concluded on the basis of REE abundances and Sm-Nd isotopic characteristics that the sandstones from the Birtola Formation of the Darjiling Group were derived from the OMG amphibolites and OMTG tonalites. They pointed out that the strong Nb-Ta depletion relative to the primitive mantle suggests the sandstones to be derived from subduction-related magmatic arc sources. The low Nb/Ta and high Zr/Sm ratios of these sandstones are identical to those for the Archean tonalite-trondhjemites that require, based on recent trace-element partitioning results, their protoliths to have formed by subduction melting of low magnesium amphibolites or metamorphosed hydrous basalts. Saha *et al.* (2004) inferred that subduction-accretion processes operated during the mid-Archean in the Singhbhum Craton. For the mid-Archean Dhanjori Group, De *et al.* (2015) suggested a sedimentary provenance in a felsic Andean-type magmatic arc.

Malangtoli Volcanics which are regarded as the basal unit of the eastern IOG sequence are characterized as arc basalt-boninite-high magnesian andesite-Nb enriched basalt (Singh *et al.*, 2016, 2017). Their geochemical attributes reflect different stages of subduction and genesis of the volcanics in forearc to rifted backarc setting in a region proximal to active continental margin.

From geochemical considerations Mukhopadhyay *et al.* (2008) suggested that the southern IOG belt is akin to an ophiolite formed in a supra-subduction setting. They suggested a non-plume source for the volcanics and their formation in an extending oceanic-arc-forearc geodynamic setting.

Geochemistry of the southern BIFs suggests their deposition in close proximity to spreading centres that might have developed over a rifted arc.

The geochemical characters of the komatiites in the eastern IOG were interpreted by Chaudhuri *et al.* (2015) to indicate mantle plume activity in an oceanic/arc setting. Later, Chaudhuri *et al.* (2017) on the basis of an enlarged data set suggested geochemical affinity of the Singhbhum komatiites with other non-arc, plume related oceanic basalts pointing to their emplacement in oceanic tectonic settings.

In summary, the Singhbhum Craton is built up multiple pulses of discrete mantle plume events, episodic plume-related mafic-ultramafic magma underplating and intraplating, and repeated crustal reworking to generate the felsic magmas (Olierook *et al.*, 2019). These mantle plumes probably operated in a stagnant-lid tectonic regime at least in the Hadean and Eoarchean times, but appear to be punctuated by discrete subduction zone events as witnessed by the deposition of several volcano-sedimentary sequences in supra-subduction setting. The Singhbhum Craton possibly records a transition from plume-driven stagnant lid tectonics to active subduction-related tectonics at ~3.5 Ga (Sreenivas *et al.*, 2019). While the OMG supracrustals were incorporated within the intrusive granitic bodies and were co-deformed and co-metamorphosed at ~3.3-3.2 Ga, the peripheral IOG basins were accreted to the OMTG-SBG composite pluton. The timing of this accretion is not precisely defined (Prabhakar and Bhattacharya, 2013). The time of major metamorphism at 3.34-3.26 Ga is at present the best estimate.

It is generally believed that the Earth was once covered by a magma ocean (Elkins-Tanton, 2012). When the magma ocean solidified the Hadean Earth was covered by a mafic protocrust. This formed the lid and stagnant-lid (or single-lid) tectonics was likely to be the common mechanism at this stage (Stern, 2018). Bédard (2018) has discussed that the stagnant-lid behaviour was likely to be periodically disturbed because of perturbations in the Earth's heat generation/loss balance, eventually triggering mantle overturns. Archean basalts were derived from fertile mantle in overturn upwelling zones (OUZOs), which were analogous to, but larger than and more long-lived than, post-Archean plumes. Early cratons/

continents probably formed above OUZO as large volumes of basalt and komatiite were delivered for protracted periods by the mantle upwells (?plumes), allowing basal crustal anatexis to generate tonalite melt, along with the complementary generation of eclogite to pyroxenite restites (Bédard, 2006), periodic restite delamination, and coupled development of continental crust and sub-continental lithospheric mantle. Periodic mixing and homogenization hindered the development of depleted mantle. Only after the establishment of true negative buoyancy-driven subduction, leading to the sequestration of the subducted slab at the core-mantle boundary, could the long-lasting depleted mantle develop (Bédard, 2018). When this transition took place in geological time is a debatable point.

Stern (2018) discussed the conditions to be satisfied for plate tectonics to operate. One of the important conditions is that the lithosphere must be denser than the underlying asthenosphere; another one is that the subducting slab must be strong enough to remain intact in subduction zones. In Archean times, the relative thickness of oceanic crust (buoyant) is likely to be more and the relative thickness of lithospheric mantle (dense) less, so that the density of total lithosphere is likely to be less than the critical density necessary for sinking. Moreover, as the Archean mantle was hotter it would be weaker. The other important condition is the presence or development of zones of lithospheric weakness that are ~1000 km long or longer. Large bolide impact may cause such zones of weakness in early Earth. Numerical modelling, taking these points into consideration, shows that subduction-driven plate tectonics could not have occurred in the Archean; or, if subduction did occur, it would be short-lived and of limited length (Stern, 2018, and references therein). In the early stage of Earth history, plate tectonic and single lid tectonic regimes may have alternated. This might have been the scenario at the Singhbhum Craton during Eoarchean,

Gerya *et al.* (2015) and Davaille *et al.* (2017) showed through thermo-mechanical modelling and experiments that subduction may be plume-induced. The arrival of a sufficiently large, hot and long-lived mantle plume could render strong and dense negatively buoyant oceanic lithosphere weak enough to produce new self-sustaining subduction zones,

which marked the beginning of plate tectonics. However, only the most energetic mantle plumes with large plume head and considerable temperature excess were able to sufficiently weaken the lithosphere enough to initiate subduction. The formation of such plume-driven oceanic plateaus in the Archean eon (Dey *et al.*, 2017) could have triggered the onset of subduction and plate tectonics.

These points have to be taken into consideration in building up a geodynamic model for evolution of the Singhbhum Craton. In particular, it is to be critically assessed whether the supra-subduction settings inferred for the supracrustal belts on the basis of geochemical attributes were short-lived and followed by further plume activities in a stagnant-lid set-up. A viable geodynamic model for the evolution of the Singhbhum Craton tightly constrained by geological, geochemical, isotopic and geochronological data is yet to come up.

The Singhbhum Craton had stabilized by ~3.1-3.0 Ga. By this time it had emerged as a landmass so that paleosols could form on the surface and terrestrial siliciclastic sediments (Keonjhar Quartzite, Mahagiri Quartzite etc.) could be deposited on top of them. The craton had cooled and had become sufficiently rigid to permit the formation of the intracontinental rift basin in which the Dhanjori sediments and volcanics were laid down. The continental freeboard was at its maximum at this time (Bandopadhyay *et al.*, 2010; Mazumder *et al.*, 2012a, 2012b). The anorogenic Mayurbhanj Granite (~3.1 Ga) and the slightly younger Tamparkola Granite (~2.8 Ga) and Pallahara Gneiss (~2.8 Ga) are the last major granite magmatisms in the craton. The rift-controlled sedimentation, volcanism and plutonism along the margins of the craton may be tied up with the early stage of basin development in the North Singhbhum Mobile Belt bordering the craton. The Chaibasa Formation in the NSMB overlies the Dhanjoris with structural parallelism, but probably with a hiatus (Mukhopadhyay, 1976; Mazumder *et al.*, 2012a). This represents a major marine transgression marking the beginning of the main stage of sedimentation and basin development in the NSMB. We speculate that the Kolhan sedimentation may also represent a marine incursion along the western edge of the craton. But the age of the Kolhans is not yet precisely defined, so this idea would remain in the realm of speculation.

The formation of different sets of Newer Dolerite dyke swarms from 2.76 Ga to <1.77 Ga marks the end-Archean tensional regime in the craton, which may also be related to pre-orogenic stage of basin development in the NSMB. The Singhbhum craton remained a stable block since then and subsequent activity shifted to the Mesoproterozoic orogenic belt (NSMB) which bordered the craton in the north, east

and west. In the south, the Eastern Ghat Belt was sutured to the Singhbhum craton probably during Grenvillian orogeny or later during Pan-African orogeny. At its waning stage, a major shear zone (Eastern Indian Tectonic Zone) cut across the eastern boundary of the craton and the Mesoproterozoic orogenic belt.

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