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Research paper

Proterozoic sedimentation and volcanism in the Singhbhum crustal province, India and their implications

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Abstract

The Proterozoic volcano-sedimentary succession comprising successively younger Dhanjori, Chaibasa, Dhalbhum, Dalma and Chandil Formations of the Singhbhum crustal province, India records sedimentation and volcanism in a rapidly changing tectonic scenario. Cooling of the vast volume of Archaean Singhbhum granite possibly induced an isostatic readjustment. The associated tensional regime and deep-seated fractures controlled the formation of the Proterozoic Singhbhum basin.

The Dhanjori Formation unconformably overlies the Singhbhum granite and is entirely terrestrial, dominantly fluvial. At the base, the conglomerate deposits, coarse-grained sandstone, and shale represent the distal fringe of an alluvial fan complex. The rest of the formation including the volcaniclastic rocks consists almost entirely of fining-upward fluvial cycles. The base of the Dhanjori Formation is a sequence boundary and the formation itself represents a lowstand systems tract. The Dhanjori volcano-sedimentary succession displays evidence of having passed through passive, as well as, active phases of continental rifting with an increasingly important influence of volcanism on the sedimentation through time. The Chaibasa Formation sharply overlies the Dhanjori Formation and a transgressive lag demarcates their contact. The basal part of the Chaibasa Formation immediately overlying the transgressive lag deposit represents a transgressive systems tract while offshore shales that sharply overlie the shallow marine sandstones up-section represent marine flooding surfaces. The shallow marine subtidal sandstones bear an excellent record of sandwave migration and provide a rare opportunity to unlock the Late Palaeoproterozoic lunar orbital periodicities. Unlike the Chaibasa, the overlying Dhalbhum Formation is entirely terrestrial (fluvial-aeolian) indicating that the Chaibasa-Dhalbhum contact is a sequence boundary. The Dalmas, overlying the Dhalbhum Formation, represents concordant lava outpourings without any break in sedimentation; these lavas are genetically related to mantle plume upwelling in an intracontinental rift setting. The volcano-sedimentary package lying north of the Dalma volcanic belt (Chandil Formation) is of Mesoproterozoic age.

The entire Late Palaeoproterozoic volcano-sedimentary package displays post-depositional compressional deformation and greenschist to amphibolite facies metamorphism dated at ca. 1600 Ma, forming the so-called North Singhbhum fold belt. The volcano-sedimentary package lying south of the Dalma volcanic belt was pushed further south towards the Singhbhum granite batholith complex as a result of uplift related to the Dalma plume. The Singhbhum granite batholith acted as a rigid body. This

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gave rise to a compressional stress regime that induced shearing/thrusting at ca. 1600 Ma along the Singhbhum Shear Zone. The Dalma plume magmatism was possibly part of a ~1600 Ma global tectono-thermal event.

Keywords: Proterozoic; Singhbhum crustal province; Sedimentation; Volcanism; Palaeogeography; Sea level change

1. Introduction

Detailed sedimentological and stratigraphic analyses of Precambrian supracrustal rocks are often problematic because of a fragmentary sedimentary record, post-depositional deformation and metamorphism (cf. Eriksson et al., 1998, 2001; Myers, 2004; Basu et al., 1981; Devaney, 2004)). Notwithstanding this, it has been demonstrated that sedimentary structures in metasedimentary rocks can be recognized and a modern sedimentological and stratigraphic evaluation can be conducted in many cases (Eriksson et al., 1988, 1994, 1998, 2001, 2004; Cristie-Blick et al., 1988; Bose et al., 1997; Catuneanu and Eriksson, 1999; Mueller and Corcoran, 2001; Mazumder and Sarkar, 2004 and references therein). The Precambrian volcano-sedimentary succession of the Singhbhum crustal province of the South Indian Block (SIB; Fig. 1a) is one of the few in the world that records sedimentation and volcanism at the Archaean–Proterozoic boundary (Acharyya, 1993, 2003a,b; Sengupta et al., 1997; Eriksson et al., 1999; Mazumder et al., 2000; Sengupta and Mukhopadhaya, 2000; Mazumder, 2002, 2003; cf. Condie, 1997; Mazumder and Arima, 2004).

Mukhopadhaya (2001) recently presented a detailed account of the growth and the temporal evolution of the Archaean nucleus of Singhbhum (encompassing various granitoids and the Meso- to Neoarchaean Iron Ore Group (IOG) supracrustals; see also Chakraborty and Majumder, 1986; Dasgupta et al., 1992; Acharyya, 1993; Chakraborty, 1996, and Mazumder et al., 2000 for an overview of the geology of the IOG rocks and Archaean evolutionary history of the Singhbhum crustal province). Although an effort has been made to understand the Proterozoic volcano-sedimentary evolutionary processes and contemporary basin tectonics by many authors working in various parts of the Singhbhum crustal province (Mathur, 1960; Naha, 1959, 1961, 1965; Naha and Ghosh, 1960; Gaal, 1964; Bhattacharya and Bhattacharyya, 1970; Sarkar and Deb, 1971; Yellur, 1977; Mukhopadhaya, 1976, 1984, 1990, 1994; Bhattacharya and Dasgupta, 1979; Gupta et al., 1980, 1982; Banerjee, 1982; Ghosh and Lahiri, 1983; Sarkar, 1984; Chakraborti and Bose, 1985; Mukhopadhaya et al., 1990; Bhattacharya, 1991; Gupta and Basu, 1991, 2000; Sarkar et al., 1992; Bose, 1994, 2000; Blackburn and Srivastava, 1994; Saha, 1994; Virnave et al., 1994; Bose et al., 1997; Das, 1997; Mazumder, 1998, 2002; Bhattacharya and Bandyopadhaya, 1998; Mazumder et al., 2000; Mazumder and Sarkar, 2004), critical analysis of the Singhbhum Proterozoic supracrustals and a synthesis thereof, remain difficult (cf. Sengupta and Mukhopadhaya, 2000; Gupta and Basu, 2000; Zhao et al., 2002, 2003a,b; Mazumder, 2003; Acharyya, 2003a,b). This paper presents a stratigraphic and sedimentological analysis of the Singhbhum Proterozoic supracrustal rocks in terms of sea level change, palaeogeographic shifts, volcanism and contemporary basin tectonics, along with their implications.

2. Geological background

The Singhbhum crustal province, covering the Singhbhum district of Bihar (now Jharkhand) and a part of north Orissa, exposes a vast tract of Precambrian rocks occupying an area of approximately 50,000 km^2 (Fig. 1a, b). The lack of any fossil record except for some rare alleged organo-sedimentary structures, stromatolites, makes any attempt to determine temporal constraints of these rocks solely dependent on isotope geology (cf. Sengupta and Mukhopadhaya, 2000; Mukhopadhaya, 2001). Isotopic data from rocks of this crustal province, although meagre, indicate an age range from 3500 Ma to 1400 Ma (Sharma et al., 1994; Goswami et al., 1995; Mishra et al., 1999; Sengupta et al., 2000; Sengupta and Mukhopadhaya, 2000 and references therein). Earlier workers identified three distinct petrotectonic zones in the Singhbhum crustal province (Fig. 1b; Bose and Chakraborty, 1994; Sarkar et al., 1992; Blackburn and Srivastava, 1994). From south to north, these are: (1) the southern Archaean granite–greenstone terrain (Acharyya, 1993; Sengupta et al., 1997; Mukhopadhaya, 2001), widely referred to as the "Singhbhum Granite Craton"; (2) the almost 200 km long North Singhbhum Fold Belt (NSFB) comprising the Dhanjori, Chaibasa, Dhalbhum, Dalma and Chandil Formations (cf. Gupta and Basu, 1991, 2000; Acharyya, 2003a,b), and (3) the extensive granite-gneiss and migmatite terrain in the north, known as the Chottanagpur Gneissic complex (CGC; Fig. 1b).

A zone of sheared and deformed rocks (the Singhbhum Shear Zone, SSZ; cf. Mukhopadhaya et al., 1975; Mukhopadhaya, 1984; Ghosh et al., 1985;

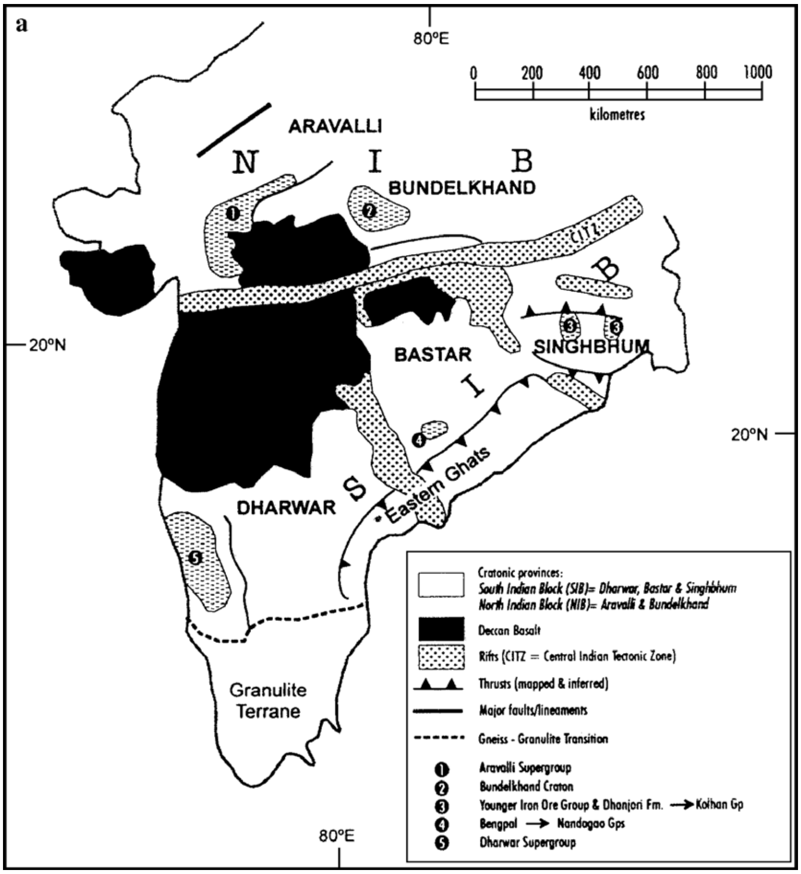


Fig. 1. (a) Map showing different crustal provinces constituting the Precambrian Indian craton (after Eriksson et al., 1999). The Singhbhum, Bastar and Dharwar crustal provinces together constitute the South Indian Block (SIB) and the Aravalli-Bundelkhand crustal province constitutes the North Indian Block (NIB). Disposition of some of the Late Archaean to Palaeoproterozoic litho-units has been shown. Interested readers may examine Eriksson et al. (1999) and Mazumder et al. (2000) for an overview of Late Archaean to Palaeoproterozoic Indian volcano-sedimentary successions. (b) Generalised geological map of the Singhbhum crustal province showing major Proterozoic litho-units (modified after Gupta et al., 1980; Sengupta et al., 2000). The Ongarbira volcano-sedimentary succession closely resembles the Dalma-Chandil succession. According to Mukhopadhaya et al. (1990) both the Ongarbira and Dalma volcanic rocks occupy comparable structural horizons and continuity between them is lost because of attenuation and faulting on the fold limb. (c) Revised Proterozoic stratigraphic succession of the Singhbhum crustal province. Poor sedimentological data and age constraints are the major impediments to place the Gangpur, Kolhan, Ongarbira, Simlipal and Jaggannathpur supracrustal rocks in this stratigraphic succession (see Section 3.1 for details); age data after Sengupta et al. (2000), Sengupta and Mukhopadhaya (2000) and Roy et al. (2002a,b).

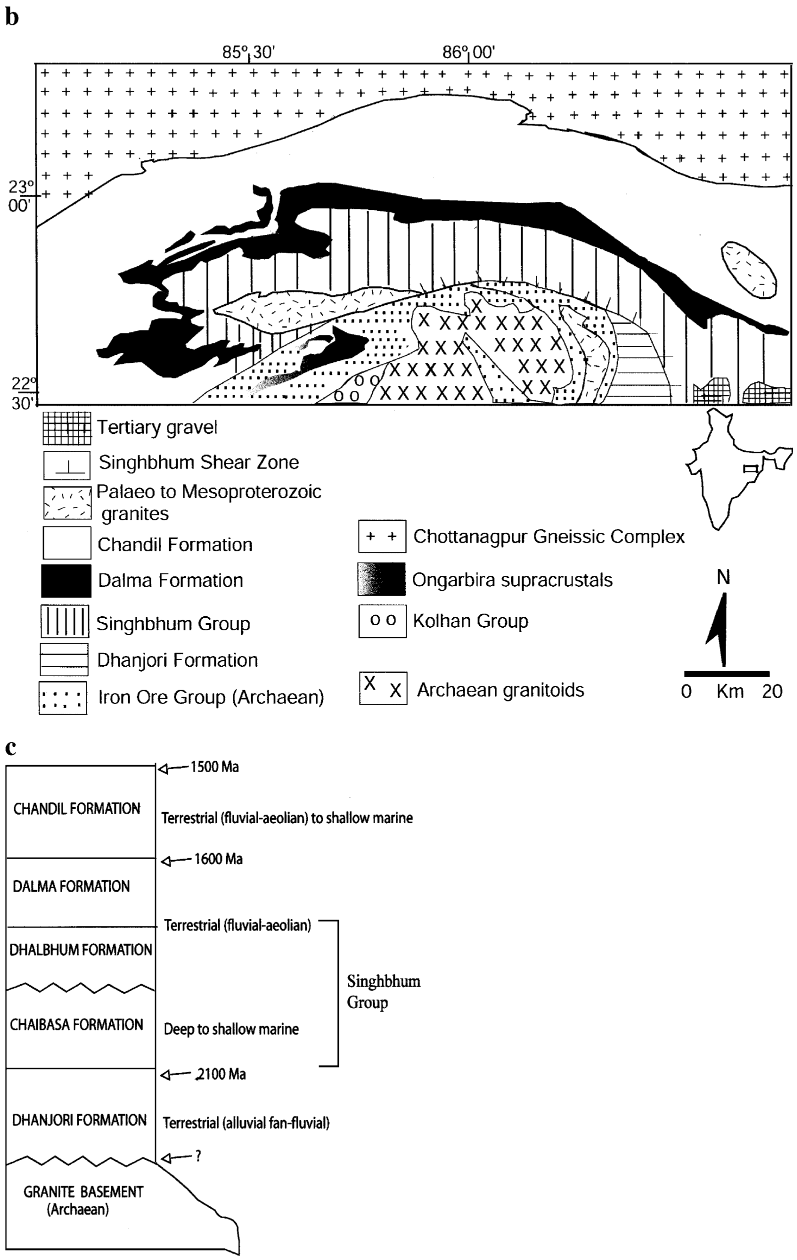


Fig. 1 (*continued*).

Ghosh and Sengupta, 1987; Saha, 1994; Sengupta and Mukhopadhaya, 2000; Joy and Saha, 2000) developed close to the contact of the oldest Proterozoic supracrustal (Dhanjori) belt with the Archaean nucleus. Small linear granitic bodies (Arkasani granophyre and Soda granite) are present along the SSZ (cf. Saha, 1994). The Rb–Sr whole rock age of about 1600 Ma obtained from these granites is inferred to reflect the age of metamorphism of the sediments and of thrusting along the SSZ (cf. Krishna Rao et al., 1979; Sarkar et al., 1986; Sengupta and Mukhopadhaya, 2000; Mazumder and Sarkar, 2004). Contrary to the earlier interpretation (Sarkar and Saha, 1962; Saha, 1994), it has been established that the SSZ does not mark the interface between the Singhbhum Archaean nucleus and Proterozoic NSFB (cf. Blackburn and Srivastava, 1994; Gupta and Basu, 2000). The Dalma volcanic rocks and the Chandil metavolcanic and volcaniclastic rocks to its north were extruded and were metamorphosed contemporaneously soon after ca. 1600 Ma (cf. Acharyya, 2003a,b). Recent studies establish that the Singhbhum Group of rocks (the Chaibasa and Dhalbhum Formations, Sarkar and Saha, 1962; Fig. 1c) and the Dhanjoris suffered a prominent metamorphism around 1600 Ma (Acharyya, 2003b).

3. Proterozoic supracrustals of the Singhbhum crustal province

3.1. Stratigraphic status

The Archaean protocontinental nucleus in the Singhbhum crustal province is girded by an elongated and arcuate belt of sedimentary and volcanic rocks, deformed and metamorphosed to greenschist (locally amphibolite) facies (Naha, 1965; Saha, 1994), and belonging to the Palaeo- to Mesoproterozoic era (Figs. 1b and 2; Sarkar and Saha, 1962; Sarkar et al., 1992; Sengupta and Mukhopadhaya, 2000; Acharyya, 2003a,b). This belt records a wide array of sedimentary and volcanic successions attributed to varying sea level, continental freeboard, palaeogeography, and tectonic regime (Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2003). The lower part of this volcano-sedimentary belt unconformably overlies the Archaean nucleus (Singhbhum Granite) and it is entirely siliciclastic and incorporates ultramafic–mafic–felsic volcanics and volcaniclastic rocks (Dhanjori Formation, whose minimum age is ~2001 Ma, Roy et al., 2002a; Figs. 2 and 3; age data from the lower member of the Dhanjori Formation is not yet available). This lower volcano-sedimentary Dhan jori succession is conformably overlain by the psammo-pelitic and pelitic rocks of the Singhbhum Group (Sarkar and Deb, 1971; Mukhopadhaya, 1976; Mazumder et al., 2000; Mazumder, 2003; Fig. 1c). The Singhbhum Group (Sarkar and Saha, 1962) is two-tiered and is made of the lower Chaibasa and the upper Dhalbhum Formations, separated by an unconformity (Fig.1b, c; cf. Eriksson et al., 1999; Mazumder et al., 2000). The Chaibasa Formation rests directly on the Archaean granitic basement at the southeastern corner of the study area and on the Dhanjori Formation elsewhere (Fig. 2).

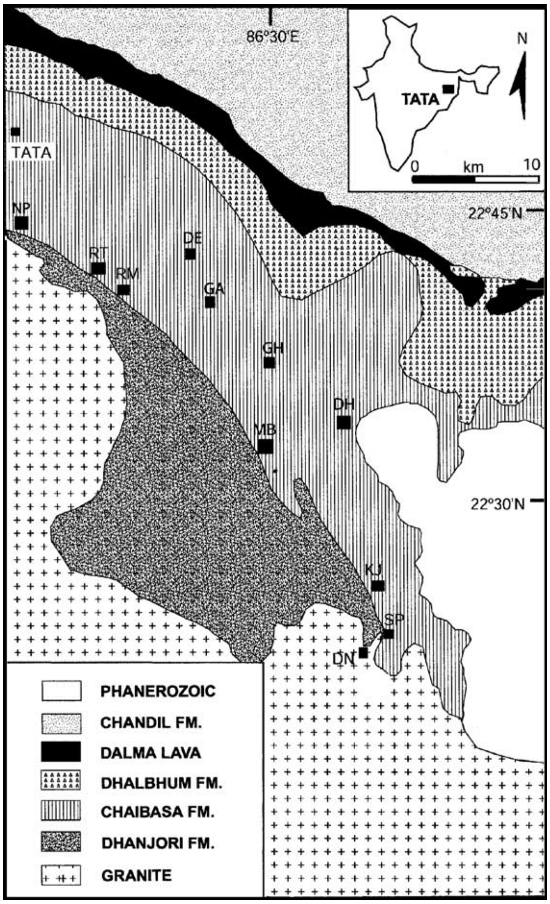


Fig. 2. Simplified geological map showing the disposition of the type Dhanjori basin and the stratigraphic relationships between the Singhbhum Group of rocks and the bounding formations (modified after Saha, 1994); study locations: NP—Narwa Pahar, RT—Rukmini Temple (Jadugoda), RM—Rakha Mines, DE—Deoli, GA—Galudih, MB—Mosaboni, GH—Ghatshila, DH—Dhalbhumgarh, KJ—Khejurdari, SP—Singpura, DN—Dongadaha.

The Dalmas are a dominantly volcanic and volcaniclastic succession with quartzites and phyllites as sedimentary interbeds (Yellur, 1977; Bhattacharya and Dasgupta, 1979; Gupta et al., 1980, 1982; Chakraborti, 1980; Chakraborti and Bose, 1985; Bose et al., 1989; Singh, 1997, 1998). Gupta et al. (1980;

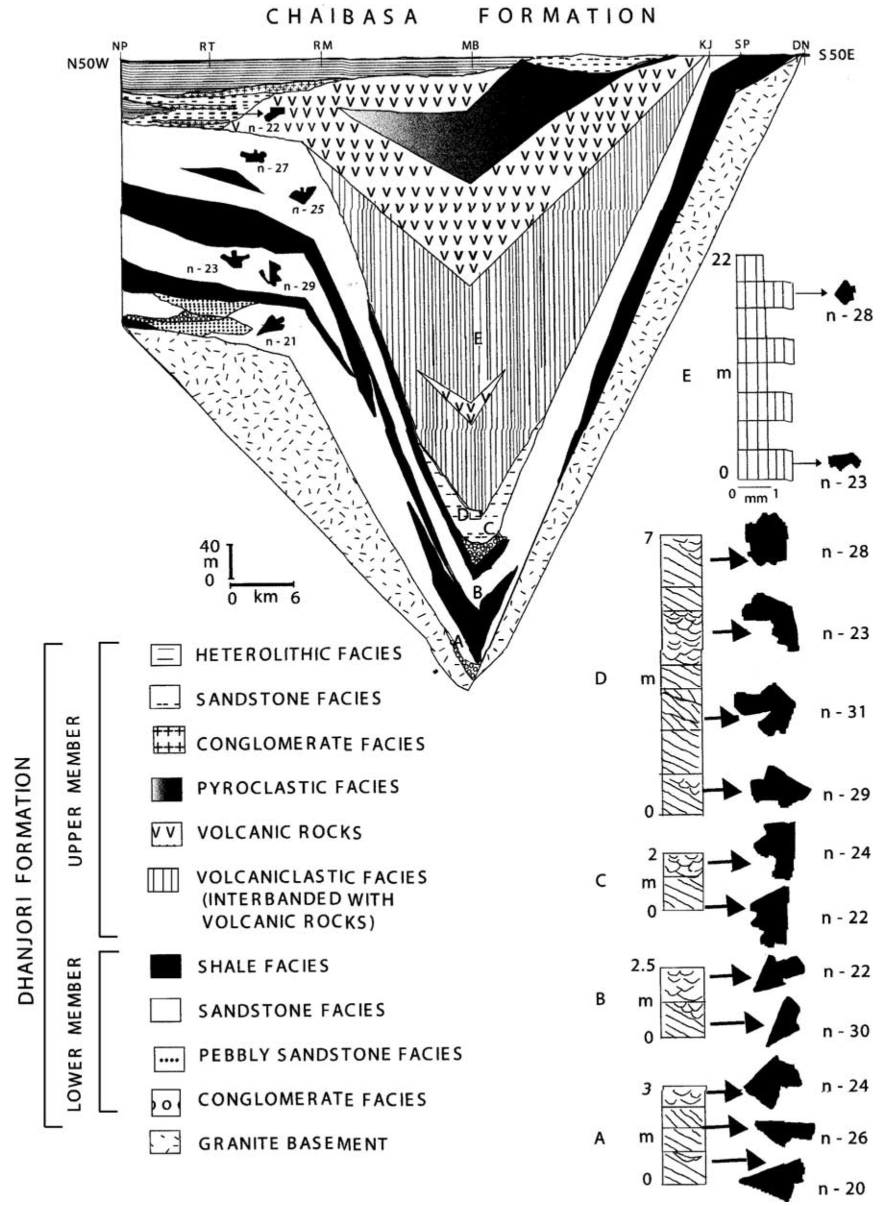


Fig. 3. Panel diagram showing lateral and vertical lithofacies transitions during Dhanjori sedimentation. Spatial and temporal changes in palaeocurrent directions have been shown (after Mazumder and Sarkar, 2004).

see also Gupta et al., 1982) subdivided the Dalma Group into the Lower and Upper Dalma Formations. These authors clearly stated that their "Lower Dalma was included earlier in the Iron Ore stage (Dunn, 1929; Dunn and Dey, 1942) and later in the Dhalbhum Formation (Sarkar and Saha, 1962)" (see Gupta et al., 1980, pp. 211–213 and their Fig. 2). The chloritic and tremolitic schists in the Lower Dalma Formation were considered as sheared epidiorite, and the major mafic and ultramafic bodies within the Lower Dalma were grouped together with the Upper Dalma metabasalts and designated as Dalma trap by earlier workers (see Gupta et al., 1980, p. 211; cf. Dunn and Dey, 1942; Sarkar and Saha, 1962). Singh (1998) suggested that the Dalma Group is made up of mafic to acidic pyroclastic deposits in the lower part and ultramafic–mafic lava flows in the upper part. He also assigned a separate stratigraphic status of formation to the upper Dalma lava flows (Singh, 1998, p. 388). The present author favours a two fold lithostratigraphic subdivision of the entire Dhalbhum-Dalma volcano-sedimentary assemblage with a lower Dhalbhum Formation and an upper Dalma Formation (Fig. 1c). The Dalma lava thus conformably overlies the Dhalbhum Formation (cf. Bhattacharya and Bhattacharyya, 1970; Gupta et al., 1980; Mazumder, 2003) and represents concordant volcanic outpouring without any significant break in sedimentation (cf. Bhattacharya and Bhattacharyya, 1970). The precise age of the Dalma volcanic rocks is unknown. An Rb–Sr whole rock isochron of the gabbro–pyroxenite intrusives into the Dalma volcanic rocks yields an age of 1619±38 Ma (Roy et al., 2002b). Based on very similar trace element and rare earth element characteristics of these intrusives and the Dalma mafic volcanic rocks, both are inferred to be comagmatic (Acharyya, 2003b, p. 18). It has been proposed that the Dalma volcanic rocks were metamorphosed and intruded by comagmatic gabbro–pyroxenite rocks around 1600 Ma (Acharyya, 2003a,b).

The metasedimentary and metavolcanic package sandwiched between the Dalma volcanic belt to the south and the Chottanagpur Gneissic complex to the north is made up of metapelites, intercalated with lenses of quartzites, metabasic, volcaniclastic and minor carbonate rocks (Bhattacharya, 1992; Bose, 1994, his Fig. 1, sub-basin II association; Ray et al., 1996). Earlier workers considered them as part of the Singhbhum Group of rocks (Dunn, 1929; Sarkar and Saha, 1962; Saha, 1994; Sarkar, 1995). Geological and geophysical investigations of Bhattacharya and Bhattacharyya (1970), however, clearly show that the metasedimentary package lying north of the Dalma volcanic belt is younger than the Dalma volcanic rocks. This post-Dalma belt of metasedimentary and metavolcanic rocks is termed the Chandil Formation by Ray et al. (1996; cf.Acharyya, 2003b). An Rb–Sr age of 1487±34 Ma has been determined recently from the acid tuffs of the Chandil Formation (cf. Sengupta et al., 2000; Sengupta and Mukhopadhaya, 2000). In spite of the metamorphic overprints, it is argued to be the age of eruption/consolidation of the acid volcanics (cf. Sengupta et al., 2000; Sengupta and Mukhopadhaya, 2000); however, Acharyya (2003b) finds that this age is most likely that of a metamorphic event. The oval cluster of granite exposures comprising the Kuilapal Granite (cf. Ghosh, 1963; Saha, 1994) exposed within the Chandil supracrustals has yielded an Rb–Sr whole rock isochron age of 1638±38 Ma (Sengupta et al., 1994), which, according to Acharyya (2003b), fixes the upper age limit of the Chandil volcanic rocks. A continuous history of deposition and normal superposition of the litho-units across this metasedimentary pile (Chandil Formation, Ray et al., 1996) has been inferred from the preserved sedimentary structures (Bhattacharya, 1992; author's own unpublished data). Thus, the Chandil Formation is of Mesoproterozoic age.

Krishnan (1937) defined the Gangpur Series as a group of calcareous psammopelitic and manganiferous metasediments exposed further west of the structural closure of the Dalma rocks (Acharyya, 2003b, his Fig. 3). Sedimentological descriptions of the Gangpur rocks are very poor and their stratigraphic status and deformation history is not well established (cf. Dunn and Dey, 1942; Banerjee, 1967; Chaudhuri and Pal, 1983; Ramachandran and Raju, 1982; Pandey et al., 1998; Naik, 2001; Acharyya, 2003b). A model Pb age of ~1.66 Ga (Viswakarma and Ulabhaje, 1991) determined from the associated ore deposits (galena) has been interpreted as that of the syngenetic evolution of the ore and broadly representative of the age of Gangpur sedimentation (Acharyya, 2003b).

Supracrustals of tentative Proterozoic age occur south of the SSZ as well (Fig. 1b). These include Ongarbira metasedimentary and metavolcanic rocks (Gupta et al., 1981; Mukhopadhaya and Dutta, 1983; Mukhopadhaya et al., 1990; Blackburn and Srivastava, 1994; Singh, 1998; Acharyya, 2003b; Fig. 1b), the Kolhan Group (Ghosh and Chatterjee, 1994; Saha, 1994; Singh, 1998; Mukhopadhaya, 2001; Fig. 1b), Simlipal metavolcanic rocks (Sarkar and Saha, 1962; Saha, 1994; not shown in Fig. 1b), and Jagannathpur mafic lava (Sarkar and Saha, 1962; Banerjee, 1982; Saha, 1994; not shown in Fig. 1b). The Ongarbira volcano-sedimentary succession rests unconformably on a basement topographic low between the Chakradharpur Granite and Singhbhum Granite (Blackburn and Srivastava, 1994; Acharyya, 2003b). Mafic (tholeiitic basalts)–ultramafic (pyroxenite) volcanic rocks, felsic volcaniclastics (tuffaceous metasediments) and low-grade pelites constitute the Ongarbira volcano-sedimentary succession (Gupta et al., 1981; Mukhopadhaya et al., 1990; Blackburn and Srivastava, 1994; Chattopadhaya and Ray, 1997; Acharyya, 2003b). The entire Ongarbira volcano-sedimentary succession closely resembles the Dalma-Chandil succession and it has been claimed that both the Ongarbira and Dalma volcanic rocks occupy comparable structural horizons, and that continuity between them is lost because of attenuation and faulting on the fold limb (cf. Mukhopadhaya et al., 1990; see also Acharyya, 2003b, pp. 22–23). The Kolhans (basal conglomerate–sandstone assemblage overlain by phyllitic shale and impersistent limestone) are almost undeformed and unmetamorphosed and unconformably overlie the Singhbhum Granite and the IOG (cf. Saha, 1994; Mukhopadhaya, 2001). The Simlipal and Jagannathpur volcano-sedimentary successions have been correlated with the Proterozoic Dhanjori volcano-sedimentary succession solely based on lithostratigraphy (cf. Sarkar and Saha, 1962; Sarkar and Saha, 1977; Goodwin, 1996, p. 338). Sedimentological and sequence stratigraphic analysis is essential to resolve the controversy regarding the depositional and tectonic setting of these Proterozoic litho-units (cf. Blackburn and Srivastava, 1994; Sarkar, 1996; Gupta and Basu, 2000; Mazumder et al., 2000; Acharyya, 2003b).

3.2. Sedimentology of the Proterozoic litho-units

3.2.1. Dhanjori Formation

The 2.1 Ga (Roy et al., 2002a) Dhanjori Formation comprises siliciclastic sedimentary rocks interlayered with ultramafic to mafic (felsic at places) volcanic and volcaniclastic rocks, deformed and metamorphosed to greenschist facies (Dunn and Dey, 1942; Iyenger and Alwar, 1965; Sarkar, 1984; Gupta et al., 1985; Basu, 1985; Saha, 1994; Singh, 1998). The Dhanjori Formation is made up of two members: phyllites, quartzites and thin conglomerate comprise the lower member, whereas volcanic and volcaniclastic rocks along with some quartzites and phyllites are important components of the upper member (Fig. 3; Gupta et al., 1985; Mazumder, 2002; Mazumder and Sarkar, 2004). There is considerable change in the dip of beds (~35° to ~60°) between the two members, with higher dips recorded by the upper member. The lower member is overall fining-upward, having conglomerates at its base. The upper member is the volcano-sedimentary segment of the Dhanjori Formation (Mazumder and Sarkar, 2004).

The sandstone bodies appear as broadly lenticular with widely variable thickness, ranging up to 30 m. The sandstones are, generally, medium- to coarse-grained, even locally granule-rich, and are poorly sorted; they have a matrix content around 10–12% but sometimes the matrix content becomes >15% (Fig. 4a, b). Grains, where they retain their primary boundaries, appear subangular to subrounded. Quartz is the most dominant mineral (more than 50% by volume). In

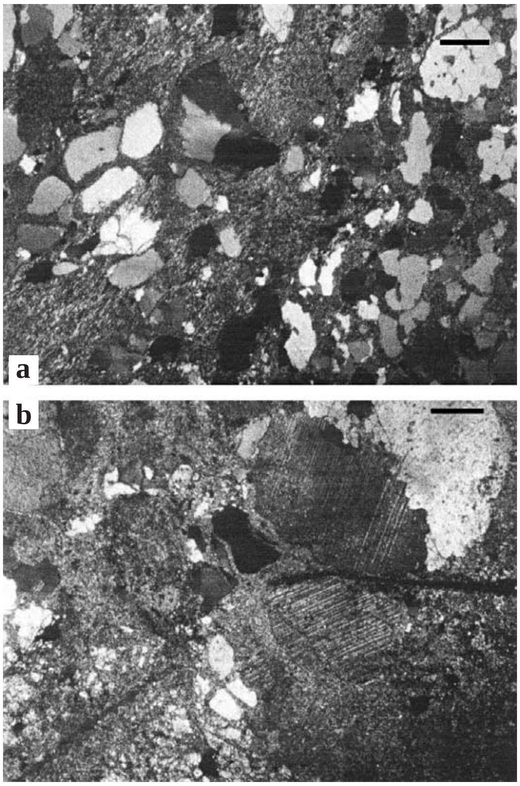


Fig. 4. Photomicrographs of Dhanjori sandstone near Rukmini Temple: (a) poorly sorted and texturally immature Dhanjori sandstone. Note the dominance of angular quartz and rock fragments (under crossed polars; bar length 0.2 mm). (b) Coarse-grained, feldspar-rich Dhanjori sandstone (under crossed polars; bar length 0.2 mm).

addition, there are feldspars, both potassic and calcic, and lithic fragments occur in fair proportion. Plagioclase feldspars are often altered to an assemblage of clay and micaceous (sericite) minerals (Fig. 4b). The lithic components are mostly represented by quartzite, granitic and basic volcanic rock fragments. Granitic rock fragments, always rare, remain confined to the basal sandstones of the Lower Member (Mazumder, 2002). The proportion of volcanic rock fragments, however, is much higher in sandstones of the Upper Member. The coarser sandstone beds of the Upper Member also contain siltstone and shale fragments, albeit in very low frequency. Sandstone compositions vary from arkose to feldspathic arenite to lithic arenite/wacke, depending on relative proportion and dominant species of feldspar (cf. Pettijohn, 1975; Bose, 1994; Mazumder and Sarkar, 2004; Mazumder, 2004a). Recently, Mazumder and Sarkar (2004) presented a detailed account of the facies analysis and the mode of Dhanjori sequence building.

Poor sediment sorting, compositional immaturity, lenticular body geometry and section-wide unimodal palaeocurrent pattern suggest that the Dhanjori sandstones are fluvial deposits (cf. Miall, 1996; Eriksson et al., 1998; Mazumder and Sarkar, 2004; Mazumder, 2004a). The palaeoslope was broadly to the north (Mazumder, 2002). The conglomerates form localized patches at the base of the two members (Fig. 5a). The conglomeratic assemblage at the base of the Lower member appears to belong to alluvial fans (cf. Blair and McPherson, 1994; Mazumder and Sarkar, 2004). The conglomerates are, however, not the sole constituents of the inferred fan deposits because the finer facies (sandstones) occur between the conglomerates as well (Fig. 5b). The conglomerates formed as mass flow as well as traction current deposits within and without channels (Mazumder and Sarkar, 2004). The conglomerate–sandstone assemblage at the base of the lower member thus represents the distal fringe of an alluvial fan deposit (cf. Blair and McPherson, 1994). The Upper Member does not include any sheet flood and sieve deposits and is constituted merely by channel and mass flow deposits. This clearly suggests steepening of the depositional surface and the increase in dip of beds from the Lower to the Upper Member corroborates this interpretation. Interestingly, sedimentary clasts are present only in

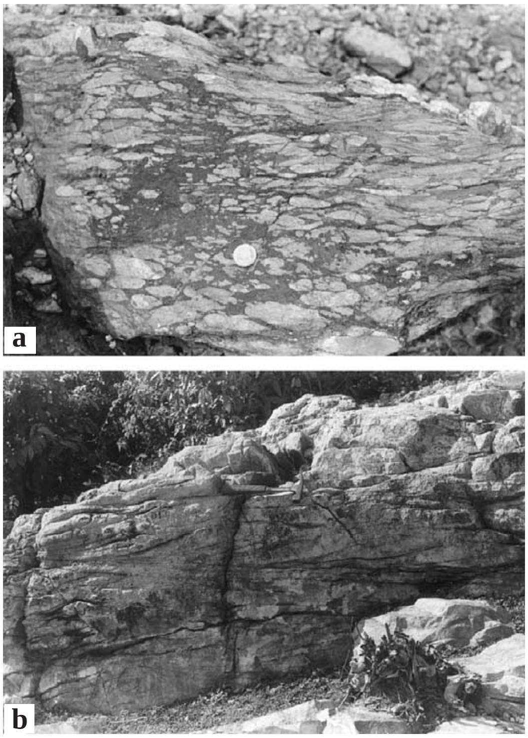


Fig. 5. (a) Clast-supported Dhanjori conglomerate (coin diameter 2.5 cm) near Rakha Mines. Pebbles are flattened as a consequence of post-depositional deformation. (b) Cross-stratified Dhanjori sandstone (hammer length 42 cm) near Barunia.

the conglomerate assemblages at the base of the Upper Member (Mazumder, 2002; Mazumder and Sarkar, 2004).

The Dhanjori volcanic rocks include ultramafic to mafic (Gupta et al., 1985; Gupta and Basu, 1991, 2000) and rarely acidic lava flows and are mostly confined to the Upper Member (cf. Singh, 1998, p. 385). Gupta et al. (1985) and Gupta and Basu (2000) suspected that part of the phyllites/schists in the basal part of the Dhanjori succession possibly represents felsic tuff/ash beds (Gupta and Basu, 2000, p. 199), but they have not provided any photomicrograph. The MgO content of the ultramafic rocks varies from 25% to 13% (cf. Gupta et al., 1985). The ultramafic rocks have a komatiitic affinity, both in terms of chemistry and texture (microspinifex texture, see Gupta et al., 1980, their Fig. 5; Gupta et al., 1985; Gupta and Basu, 2000). These rocks are now largely composed of actinolite, hornblende, relict pyroxene, olivine along with epidote and chlorite (Fig. 6). On the contrary, the

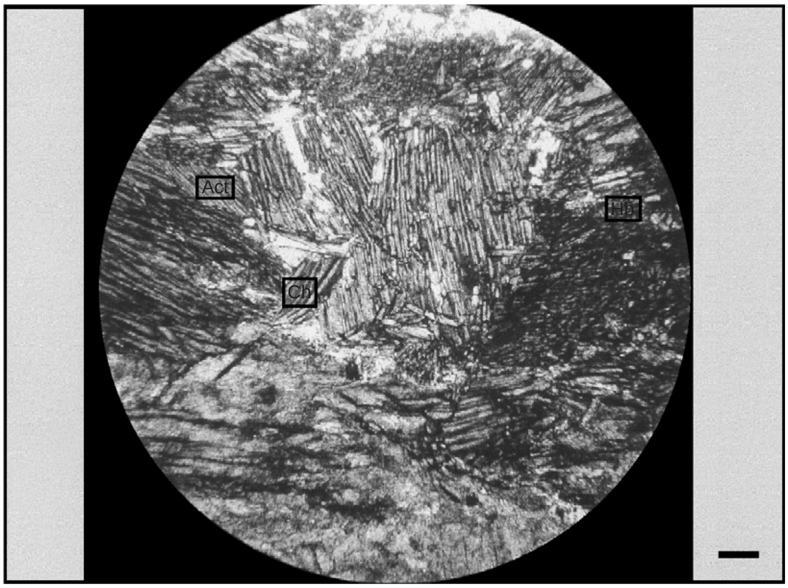


Fig. 6. Photomicrograph of the Dhanjori volcanic rock; note the presence of actinolite, hornblende and chlorite (under crossed polars, bar length 0.1 mm).

volcanic rocks constituting the upper part of the Upper Member are mostly basaltic in composition. They are fine- to coarse-grained, vesicular, amygdaloidal, and rarely pillowed. Associated with these volcanic rocks are thick volcaniclastic and pyroclastic rocks (cf. Fisher and Schminke, 1984). The volcaniclastic facies is generally cross-stratified. Close correspondence between its cross-strata orientation and the fluvial palaeocurrents inferred from the associated sandstones strongly supports reworking of volcanic materials (Mazumder, 2002; Mazumder and Sarkar, 2004). The terrestrial depositional setting of the pyroclastic facies is constrained by the bounding sandstone–siltstone and volcaniclastic lithofacies of fluvial origin (cf. Cas and Wright, 1987, p. 270; Fritz and Howells, 1991; Mueller et al., 2000; Mazumder and Sarkar, 2004).

3.2.2. Chaibasa Formation

The 6 to 8 km thick siliciclastic Chaibasa Formation rests directly on the granitic basement at the south-eastern corner of the study area and on the Dhanjori Formation elsewhere (Fig. 2; Bose et al., 1997). Lithologically it is characterized by the interbedding of sandstones, shales and a heterolithic (very fine sandstone/siltstone–mudstone) facies in different scales (Fig. 7a, b) with minor mafic volcanic rocks. Its lower surface is demarcated by a thin (one or two pebbles thick) conglomerate made up, predominantly, of rounded vein quartz pebbles (Bose et al., 1997, p. 78; Fig. 8a). It is around 14 cm thick with pebbles of average diameter 13 cm, and is reduced to 3 cm thickness with pebbles of 1 cm diameter 2 km away from the southeastern extremity. The conglomerate is clast-supported, the interstitial spaces being filled by sand-sized matrix (Fig. 8a). The Chaibasa sandstone is very fine- to fine-grained, generally well sorted, compositionally as well as texturally mature, and is interbedded with the heterolithic and shale facies. The sandstone facies units are 5 to 40 m thick (average thickness 20 m). Beds constituting the sandstone facies units are up to 2.5 m thick, and lenticular or wedge-shaped (Bose et al., 1997, their Table 2). The majority of the sandstone beds are pervasively tabular cross-stratified (Fig. 8b) with spectacular intraset cyclic variation in stratification style (Bose et al., 1997; Sarkar et al., 1999; Mazumder, 2000). Others are partly or completely massive or graded, and a few beds are planar laminated. Locally the sandstone bed surfaces preserve wave ripples (Mazumder, 1999). Bose et al. (1997) presented a detailed account of the facies characteristics and the mode of Chaibasa sequence building.

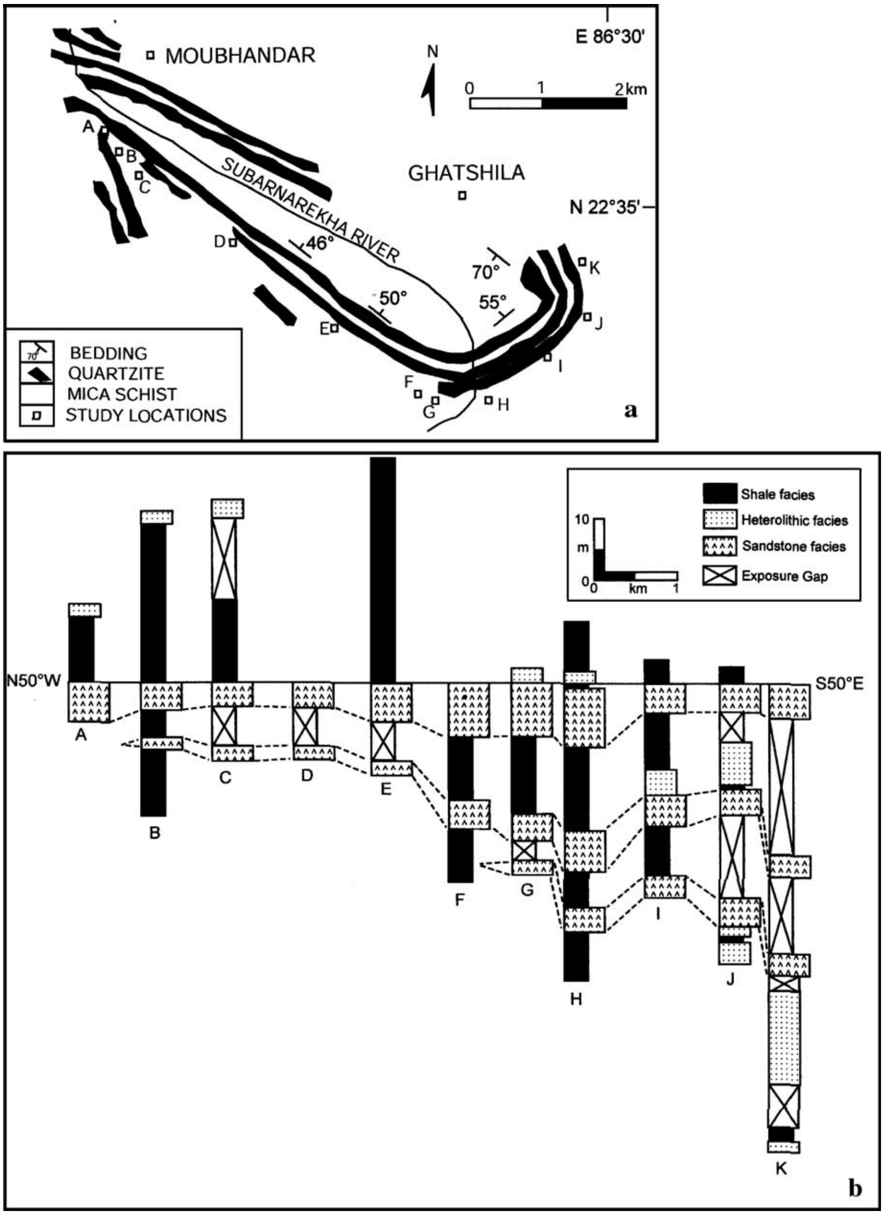


Fig. 7. (a) Geological map showing litho-units (Chaibasa Formation) in and around Ghatshila (modified after Naha, 1965). (b) Panel diagram showing lateral and vertical lithofacies transitions of the Chaibasa sedimentation; study locations are marked in (a) (modified after Bose et al., 1997).

Fine grain-size, good sorting, mineralogical as well as textural maturity, local bipolar-bimodal sediment dispersal pattern, and characteristic rhythmicity in the cross-stratification foreset thickness variation (thick-thin alternation, De Boer et al., 1989; Fig. 9a, b) are indicative of the tidal origin of the Chaibasa sandstones (Bose et al., 1997; cf. Eriksson et al., 1998; Eriksson and Simpson, 2004). The palaeoslope was broadly to the north (Bose et al., 1997). Presence of wave ripples implies sand deposition above the wave base (cf. Mazumder, 1998, 1999). The invariable presence of a double mud drape (Fig.

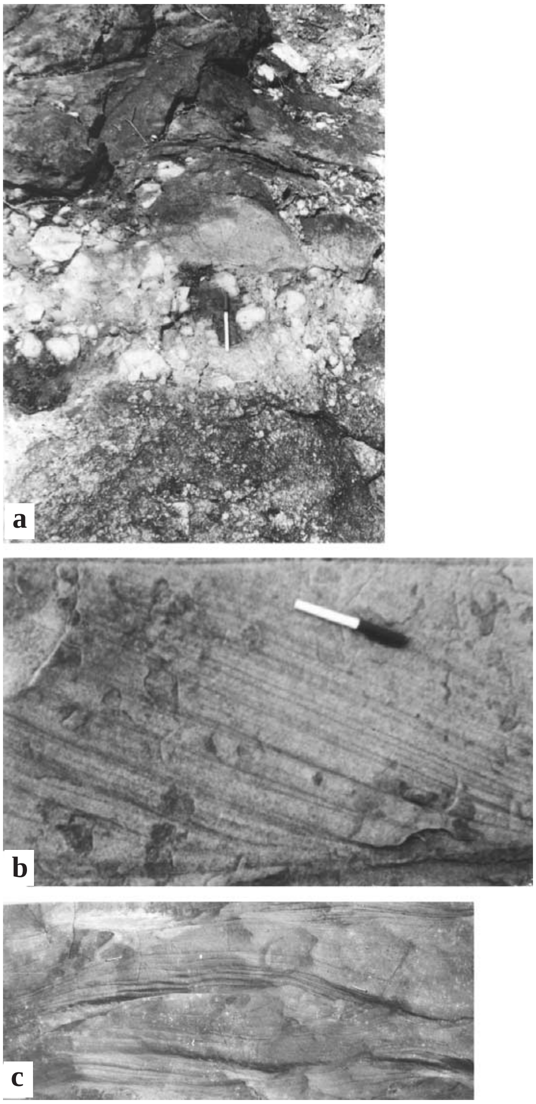


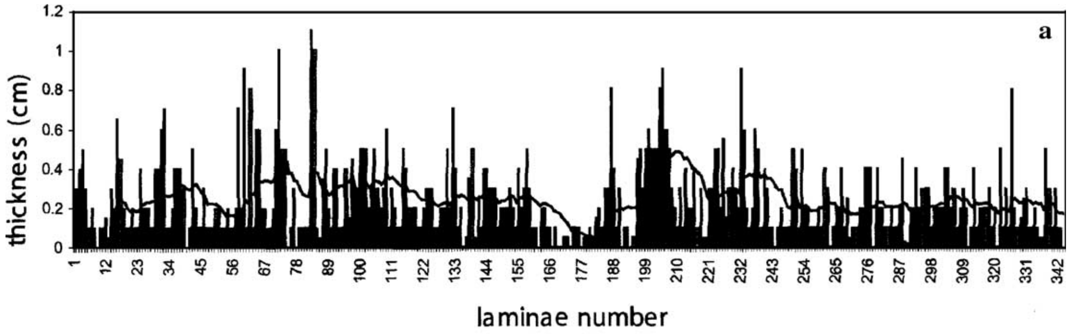
Fig. 8. (a) A clast-supported sheet conglomerate occurring along the contact of the Singhbhum Granite (SG) and the cross-bedded Chaibasa sandstone (CS) (matchstick length 4.2 cm; southwest of Singpura, Fig. 2). Bose et al. (1997) interpreted this sheet conglomerate as a transgressive lag deposit. (b) Fine-grained, cross-stratified Chaibasa sandstone near Harindungri, Ghatshila, with characteristic thick-thin alternation in foreset laminae thickness; note also double mud drape (after Bose et al., 1997; pen length 12 cm). (c) Hummocky-cross-stratification within the Chaibasa heterolithic facies (after Bose et al., 1997; matchstick length 4.2 cm) near Harindungri, Ghatshila.

8b) further indicates that the sandstones formed in a subtidal setting (cf. Visser, 1980; Eriksson and Simpson, 2000). Characteristic thick-thin alternation in sandstone foreset thickness reveals the dominantly semidiurnal nature of the palaeotidal system (cf. Williams, 1989, 2000; De Boer et al., 1989; Bose et al., 1997; Mazumder, 2002). Laminae counting (Fig. 9a) in combination with spectral analysis of foreset thickness data (Fig. 9b) further indicate that there were about 32 lunar days in a synodic (full Moon to full Moon) month (Mazumder, 2004b; Mazumder and Arima, 2005). The heterolithic facies contain profuse wave generated structures including hummocky cross-stratification (Fig. 8c) and numerous slides and slumps implying deposition on a relatively steeper slope in a setting between the fair-weather and storm wave base. The shale facies formed in a deeper offshore setting below the storm wave base (Bose et al., 1997; Mazumder, 2002). Alternatively, Bhattacharya (1991) and Bhattacharya and Bandyopadhaya (1998) proposed that the shale facies formed in an intertidal setting. The sheet conglomerate occurring at the base of the Chaibasa Formation represents a transgressive lag deposit (Bose et al., 1997; cf. Cattaneo and Steel, 2003).

3.2.3. Dhalbhum Formation

The 2 to 4 km thick Dhalbhum Formation (Fig. 2) unconformably overlies the Chaibasa Formation (Fig. 1c; cf. Eriksson et al., 1999; Mazumder et al., 2000). Lithologically, the Dhalbhums are represented by a thick pile of alternating phyllites and quartzites, and ferruginous shale, towards the base of the unit. Mafic/ultramafic intrusive bodies (peridotites and serpentinites) occur widely within this basal sedimentary sequence (Gupta et al., 1980). The upper parts of the Dhalbhum Formation are characterized by extensive development of tuffaceous rocks (Gupta et al., 1980, 1982). The Dhalbhum Formation also contains a number of thin lava flows of variable petrochemical types (tholeiites, feldspathoidal basalts, and basaltic komatiites), interlayered with the sediments and tuffaceous members (Gupta et al., 1980, p. 217).

Sarkar and Saha (1962; see also Sarkar, 1995) determined the boundary between the Chaibasa and Dhalbhum Formations on the basis of combined LANDSAT imagery and field data, although they could not correlate the boundary with any structural discontinuity in the field (see also Naha, 1965, pp. 45–46). A drastic change in terms of texture and primary structures of sandstones as well as in palae-



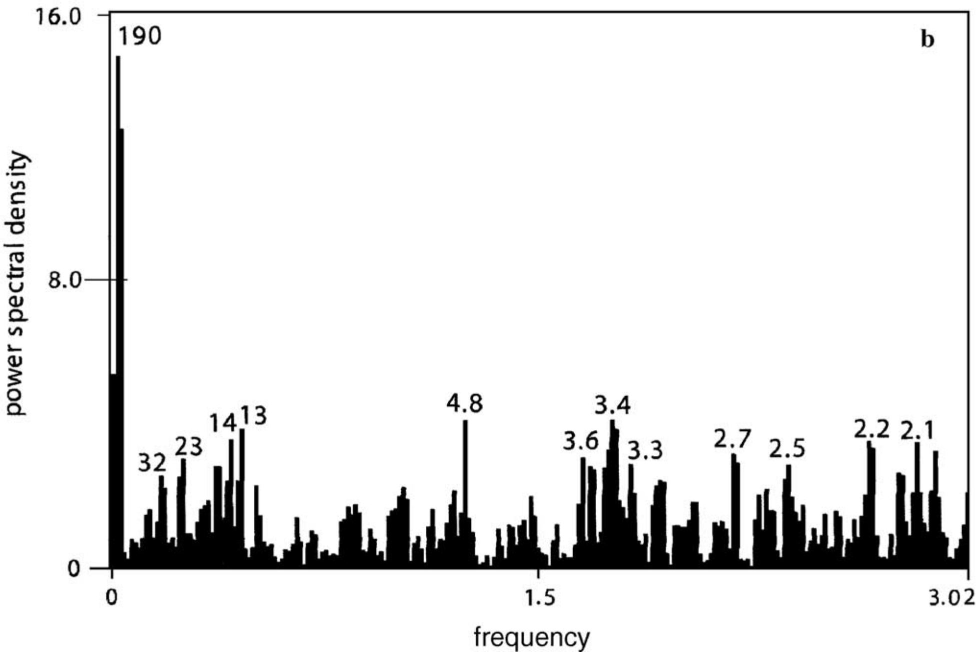


Fig. 9. (a) Plot of Chaibasa sandstone foreset laminae thickness vs. laminae number. The laminae thickness data set has been statistically tested following the methodology of De Boer et al. (1989), revealing a significant semidiurnal signal with 27–30 laminae per neap-spring cycle (Poppe L. De Boer, personal communication, 2003). (b) Fast Fourier Transformation (FFT) of the Chaibasa tidal rhythmite data (shown in (a)) following the methodology of Archer et al. (1991); spectral periods of 32, and around 2 laminae/cycle strongly corroborate the inferred semidiurnal nature of the Chaibasa palaeotidal system. For details see Mazumder (2004b).

ocurrent direction has, however, been used to differentiate the Dhalbhum Formation from the underlying Chaibasa Formation (cf. Naha and Ghosh, 1960; Mazumder et al., 2000; Mazumder, 2002, 2003).

Unlike the Chaibasa sandstones, the Dhalbhum sandstones are medium- to coarse-grained, poorly sorted and feldspar-rich (Fig. 10a). Compositionally they are subarkoses to arkoses, containing a high proportion of quartz and alkali feldspar with some plagioclase feldspar (Fig. 10a). Rock fragments are generally rare. The sandstone units are made up of a number of fining-upward cycles. The sandstone bed geometry is irregularly wedge-shaped although lenticular beds are not uncommon. In places, sinuous-crested linguoid bedforms are well preserved on the bedding planes. The sandstones are either planar-laminated, tabular and/or trough cross-stratified, or massive (Fig. 10b,c). Downcurrent transition from dune to upper-stage plane beds is common (Fig. 10d). Massive sandstone beds are often amalgamated. They

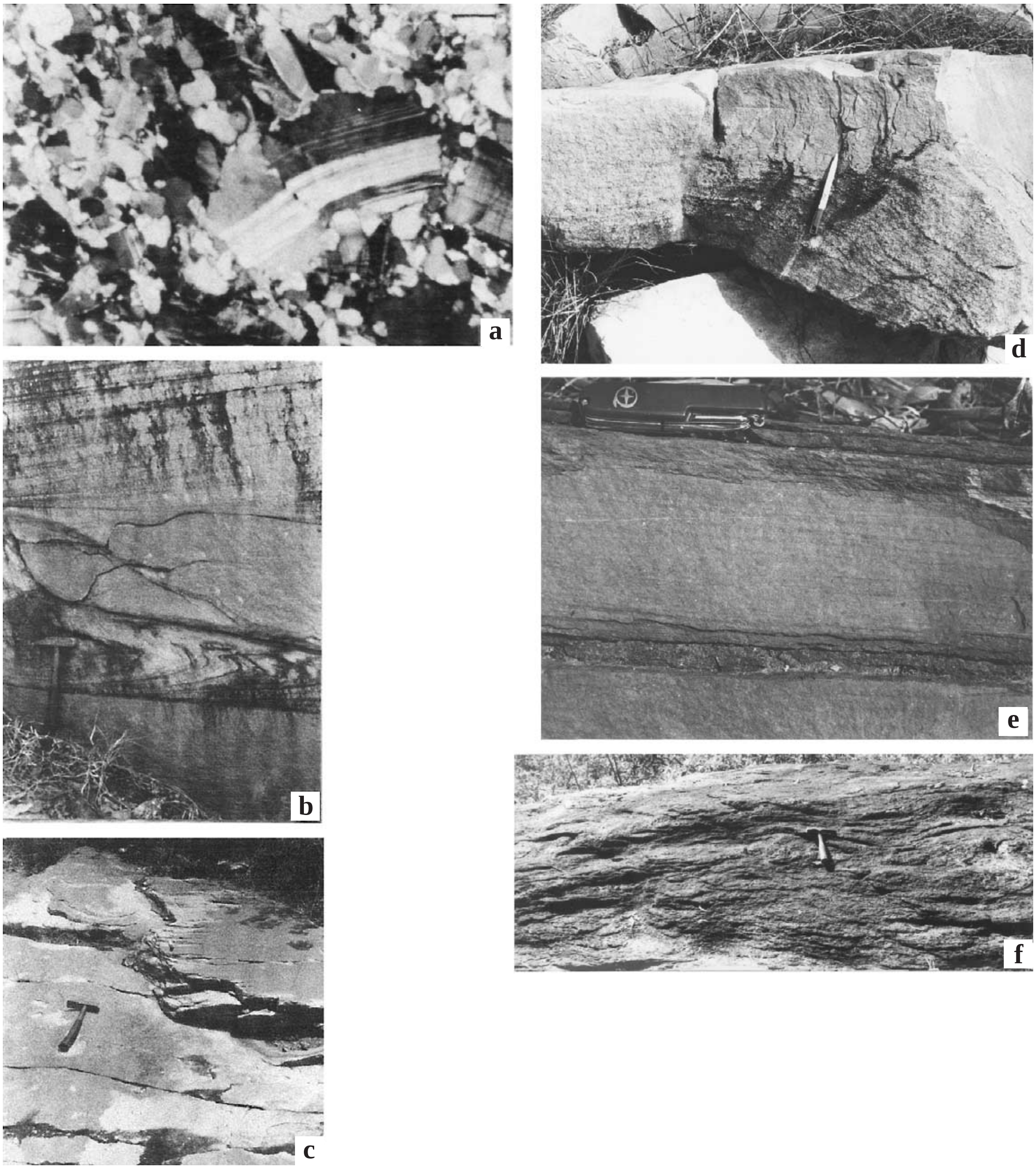


Fig. 10. (a) Photomicrograph (under crossed polars) of coarse-grained, texturally as well as compositionally immature Dhalbhum sandstone; note the presence of alkali feldspar and plagioclase (bar length 0.2 mm). (b) Large-scale (hammer length 42 cm) planar cross-bedded Dhalbhum sandstone; note that emplacement of massive sandstone bed results in penecontemporaneous overturning of the underlying foreset planes. (c) Trough cross-bedded Dhalbhum sandstone (hammer length 42 cm). (d) Dune to upper stage plane bed transition within the Dhalbhum sandstone (pen length 14 cm). (e) Translatent strata within fine-grained Dhalbhum sandstone (knife length 10 cm). (f) Massive Dhalbhum shale (hammer length 42 cm).

often occur on top of large-scale tabular cross-stratified sandstones and the underlying cross-beds are frequently overturned (Fig. 10b; cf. Allen and Banks, 1972). Palaeocurrent directions are broadly southward and unimodal (cf. Naha and Ghosh, 1960; author's own unpublished data). In the Duarshini-Asanpani sector (south of the Dalma volcanic belt on the Galudih-Kunchia road; not shown in Fig. 2), there occur some fine-grained, well-sorted sandstones (see Simpson et al., 2004, their Figs. 7.6–2b, c) bearing very low-angle stratification (Fig. 10e). Unlike the Chaibasa shale, the Dhalbhum shales (Fig. 10f) are chlorite- and/or amphibole (tremolite–actinolite series)-rich and are generally devoid of primary structures, except for uncommon thin (up to 3 cm) plane or ripple laminated fine sandstone interbeds. The high-magnesian vitric tuffs constituting the upper part of the Dhalbhum Formation are now represented by tremolitic schists characterized by a partially devitrified glassy matrix (Gupta et al., 1980, 1982). The tuffaceous pile also contains mafic to ultramafic concordant volcano-plutonic igneous bodies (Gupta et al., 1980, p. 213).

Compositional immaturity, coarser grain size, poor sorting, fining-upward cycles, coupled with a unimodal palaeocurrent pattern suggest that the Dhalbhum sandstones are of fluvial origin (Eriksson et al., 1998, 1999; Mazumder et al., 2000; Mazumder, 2003). Frequent overturning of the cross-strata possibly took place as a consequence of variable discharge, suggesting thereby a braided character for the fluvial depositional system (cf. Miall, 1977, 1996; Collinson and Thompson, 1989; Collinson, 1996). In close association with the sandstones of inferred fluvial origin, the Dhalbhum shale was likely a flood-plain deposit. Very low angle stratification (Fig. 10e), preserved within relatively fine-grained well-sorted sandstones on top of coarse-grained sandstones, can be compared to translatent strata (cf. Kocurek, 1996), implying possible aeolian reworking of fluvial sediments (Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2003; Simpson et al., 2004, their Figs. 7.6–2b, c). The Dhalbhum sedimentation therefore took place entirely in a terrestrial depositional regime (Mazumder, 2003); the broadly southward-orientated palaeocurrents indicate a palaeoslope towards the south (cf. Naha and Ghosh, 1960; Bose, 1994, p. 336).

The high-magnesian vitric tuff along with its comagmatic intrusives show quench textures and have chemistry closely comparable to the peridotitic-pyroxenitic komatiites of Archaean greenstone belts (Gupta et al., 1980, their Table 1 and Fig. 12). Gupta et al (1980, p. 225) interpreted these vitric tuffs and related intrusives as Proterozoic komatiitic magma. The mafic–ultramafic volcanism took place in an entirely terrestrial setting as is evident from the interbedded sedimentary rocks (Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2003).

3.2.4. Dalma Formation

The Dalma Formation conformably overlies the Dhalbhum Formation and is represented by a thick sequence of mafic–ultramafic volcanic rocks with lenses of basic agglomerates (Gupta et al., 1980; Chakraborti, 1980; Chakraborti and Bose, 1985; Bose, 1994; Singh, 1997, 1998; Sengupta et al., 2000). Pillow structures are common within the basalts along lower stratigraphic levels. The basalts are massive, highly compact, vesicular (Fig. 11a) and metamorphosed to greenschist facies. Petrographically, these metabasalts exhibit fine- to medium-grained, subophitic to intergranular or intersertal textures and are composed of actinolite, chlorite, epidote, clinozoisite and plagioclase, with subordinate amounts of quartz and hornblende (Gupta et al., 1980; Chakraborti and Bose, 1985). The agglomerates are coarse-grained, characterized by angular basaltic fragments embedded in a lava matrix (Fig. 11b; cf. Bhattacharya and Dasgupta, 1979) and exhibit interlayering with the basalts (Gupta et al., 1980). In places, the volcaniclastic rocks exhibit planar cross-stratification (Fig. 11c; see also Gupta et al., 1982).

Earlier Dunn and Dey (1942) and De (1964) proposed a continental origin for the Dalma mafic volcanic rocks. Naha and Ghosh (1960) however, postulated an island arc environment for the Dalma basaltic volcanism. Gupta et al. (1980) proposed that distension of the protocontinental crust led to the formation of the Proterozoic Chaibasa basin, and that rising geotherms associated with more or less complete mantle melting produced the Dalma ultramafic–mafic volcanic rocks (see also Mukhopadhaya, 1984, 1994). Bose et al. (1989) and Bose (1994) suggested that the Dalma basalts are significantly depleted in incompatible elements (particularly LREE), compara-

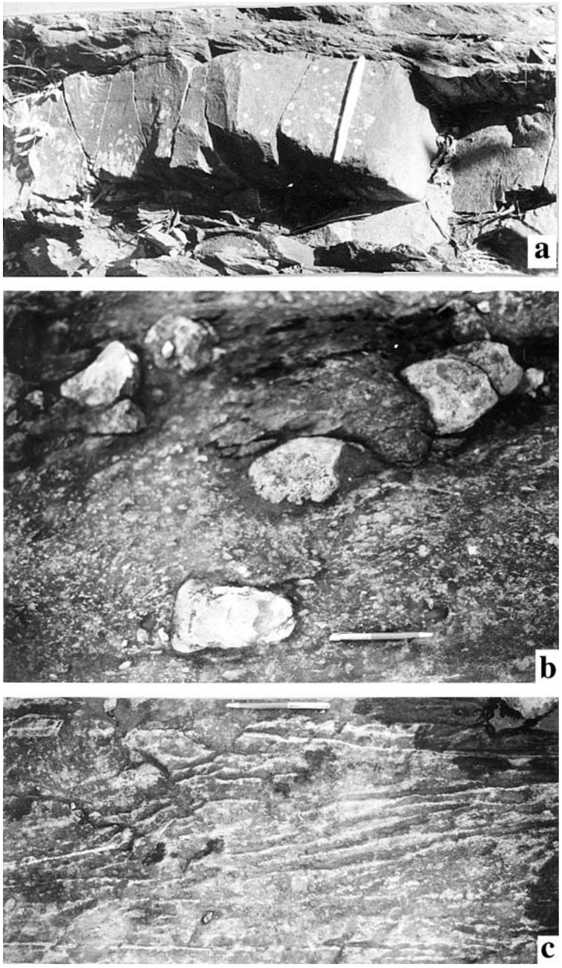


Fig. 11. (a) Vesicular Dalma basalt (pen length 14 cm). (b) Dalma volcaniclastic rock with embedded angular basaltic fragments (pen length 13 cm). (c) Planar cross-stratification within the Dalma volcaniclastic rocks (pen length 13 cm).

ble to MORB, and thus are a close chemical analogue of basalts formed in a back-arc basin. Bose (1994, p. 329) questioned the validity of the ensialic evolutionary models of Dalma volcanism and its genetic relationship to mantle plume upwelling as proposed by Gupta et al. (1980) and Mukhopadhaya (1984), on the grounds that convincing sedimentary history, supporting geochemical evidence and magmatic characterization are lacking. Acharyya (2003b) suggested that an asymmetric distribution of the volcanic rocks, with the development of the acidic volcanic rock-dominated Chandil belt to the north of the Dalma volcanic belt, as well as the absence of 1600–1500 Ma old charnockites in the southern parts of the CGC belt flanking the Dalma volcanics, does not support the postulate of Roy et al. (2002b) that the Dalma volcanic rocks represent Mid-Proterozoic (~1600 Ma) global plume-related thermal uplift. Roy et al. (2002b, p. 144), however, clearly stated that the positive ΔNb values of most of the magmatic units of the Dalma volcanic belt are indicative of their mantle plume origin. These authors (see also Roy, 1998) also proposed that a thick envelope of hot N-MORB-like upper mantle and entrained heterogeneity at the plume head possibly gave rise to depleted MORB-like REE and isotopic imprints with some LILE enrichment (Roy et al., 2002b, p. 144). Interestingly, the underlying Dhalbhum Formation represents entirely terrestrial (fluvial-aeolian) deposits (Mazumder et al., 2000; Mazumder, 2003; Simpson et al., 2004; see Section 3.2.3). Singh (1997) reported peraluminous rhyolitic to dacitic tuffs from the Chandil Formation that conformably overlies the Dalma Formation (cf. Bhattacharya and Bhattacharyya, 1970; Mazumder, 2003) and proposed an aeolian origin for these tuffs (Singh, 1997; 1998, p. 390). Pillow structures indicate quiescent subaerial flows locally entering streams or fluvial channels (cf. Dillard et al., 1999). The Dalma volcano-sedimentary assemblage is thus bounded by entirely terrestrial deposits that, in turn, support its development in a mantle plume activated intercontinental rift setting (cf. Gupta et al., 1982; Mukhopadhaya, 1990, 1994; Eriksson et al., 1999; Mazumder et al., 2000; Roy et al., 2002b; Mazumder, 2003; elaborated below).

3.2.5. Chandil Formation

The metasedimentary and metavolcanic rocks lying between the Dalma volcanic belt and the CGC (Fig. 1b) are of Mesoproterozoic age (~1500 Ma, Rb–Sr whole rock isochron age; see Sengupta et al., 2000; Sengupta and Mukhopadhaya, 2000) and include quartzites, mica schists, carbonaceous phyllite, weakly metamorphosed acidic volcanic and volcaniclastic rocks (including vitric and lithic tuffs), and amphibolites (Bhattacharya, 1992; Bose, 1994; Ray et al., 1996; Singh, 1997, 1998; Sengupta et al., 2000; Acharyya, 2003b). The different litho-units occur as east–west trending bands without any stratigraphic inversion. This volcano-sedimentary succession is markedly different from the Late Palaeoproterozoic Singhbhum Group (cf. Dunn and Dey, 1942; Bose, 1994; Ray et al., 1996; Sengupta et al., 2000; Gupta and Basu, 2000). Ray et al. (1996) assigned the status of formation to these rocks and included them within the Chandil Formation.

The sandstones appear broadly lenticular with variable thickness ranging up to 20 m and are medium- to coarse-grained, poorly sorted and compositionally and texturally immature. Matrix content is variable, typically around 10% but sometimes >15%. Planar or broadly undulating master erosion surfaces (MES) occur at various stratigraphic levels within the sandstones. Between any two MES's the sandstone segments fine upwards (see also Singh, 1998). The sandstones are generally trough cross-stratified (Fig. 12a) and interbanded with shales (schist). In places, large channel-fills are well preserved (Fig. 12b). Palaeocurrent trend is unimodal and northward (author's own unpublished data). However, sandstones occurring close to the CGC in and around Barabazar (23°03′N:86°21′E) are fine-grained, texturally mature and bear large scale tabular cross-stratification (maximum set thickness 1 m) with characteristic double mud drape (cf. Visser, 1980; Bose et al., 1997; Fig. 13). Thin cross-laminations, and flaser-wavy-lenticular bedding are well preserved in psammopelitic rhythmites around Lawa (23°01′N: 86°00′E) (Gupta and Basu, 2000, p. 205). The shale facies is generally devoid of primary structures except for cm-thick planar and/or ripple laminated fine-grained sandstone interbands.

The banded tuffaceous rocks (Fig. 14a) constitute a significant portion of the Chandil Formation. The banding is defined by subtle colour variations and also by differential weathering characteristics implying thereby presence of compositional variation between bands (cf. Sengupta et al., 2000, pp. 44–45; see also Ray et al., 1996, their Fig. 2). The tuffaceous rocks are porphyritic in nature with essentially quartzo-feldspathic fine-grained groundmass. The phenocrysts are made up of equidimensional biotite and quartz and rarely euhedral garnet (Singh, 1997; Ray et al., 1996, their Fig. 3; Sengupta et al., 2000). These rocks

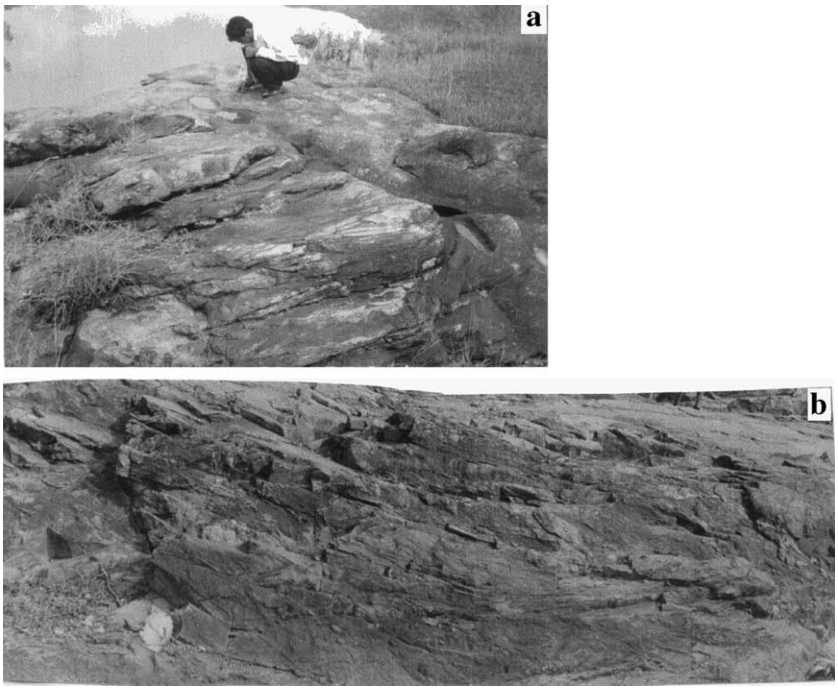


Fig. 12. (a) Cross-bedded Chandil sandstone near Bandwan. (b) Large-scale (hammer length 42 cm) trough cross-bedded Chandil sandstone, north of Chandil.

Fig. 13. Large-scale (set thickness 1 m) planar cross-bedded, fine-grained shallow marine Chandil sandstone near Barabazar; note double mud drape.

characteristically show spherulites (Singh, 1997, his Fig. 4), accretionary lapilli (Ray et al., 1996, their Figs. 3–5), and pele's tears (Singh, 1997, 1998). Chemically, these tuffaceous rocks range from rhyolite to dacite and have major- and trace element composition comparable with ash flow tuffs from other parts of the world (Ray et al., 1996). These tuffaceous rocks represent an acid magmatic event contemporaneous with deposition of fine-grained clastic sediments, and their petrographic, geochemical and isotopic character indicates consolidation immediately after eruption (Sengupta et al., 2000).

Poor sediment sorting, lenticular geometry and unimodal cross-strata orientation in combination with compositional and textural immaturity suggest that the Chandil sandstones are fluvial deposits (cf. Miall, 1996; Eriksson et al., 1998 and references therein). Associated shales are therefore inferred as of flood-plain origin. Singh (1997, 1998) proposed an aeolian origin for the acidic tuffs. However the sandstones occurring close to the CGC (Fig. 13) are definitely of marine origin. The presence of a double mud drape (Fig. 13) further indicates that these sandstones formed in a subtidal setting (cf. Visser, 1980; Eriksson and Simpson, 2000, 2004). Gupta (Gupta and Basu, 2000, p. 205) suggested that the psammopelitic rhythmites bearing flaser-wavy-lenticular bedding around Lawa area were formed in an intertidal to subtidal setting. The mode of the Chandil sequence building is not clearly understood at present and detailed sedimentary facies and sequence stratigraphic analysis remains to be done.

An ellipsoidal granite gneissic body, known as the Kuilapal Granite, occurs in the eastern part of the belt (Fig. 1b). Compositionally it ranges from trondhjemite through granodiorite to adamalite to granite proper (Dunn and Dey, 1942; Saha, 1994). It was originally described as a gneissic body with evidences of injection of gneissic material within the surrounding mica schist along the schistosity (Dunn and Dey, 1942). Subsequently it has been established that the gneissic rocks were formed synchronously with the later phase of deformation of surrounding metapelites (Ghosh, 1963). A number of minor intrusive bodies of alkaline affinity occur between Sushina (22°57′N:

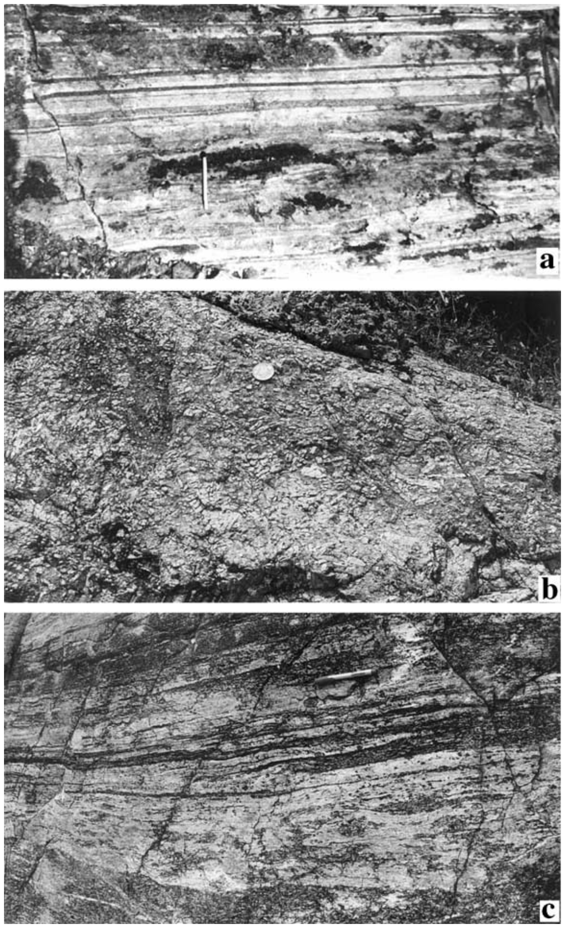


Fig. 14. (a) Felsic Chandil volcaniclastic rocks with compositional banding (matchstick length 4.2 cm), north of Chandil. (b) Post orogenic nepheline syenite with spectacular development of albite laths, near Sushina (maximum length 12 cm; coin diameter 2.5 cm). (c) Alternate dark (biotite, alkali pyroxene and amphibole-rich) and light (albite-rich) coloured compositional banding within the nepheline syenite near Sushina.

86°37′E) and Balarampur within the Chandil Formation (cf. Bhattacharya and Chaudhuri, 1986; Bhattacharya, 1992, his Fig. 1). Compositionally they range from syenite (Fig. 14b) to alkali granite. The major mineral constituents of the Sushina syenitic rocks include albite, perthite, Fe-rich biotite, alkali pyroxene (acmite), and feldspathoids (Bhattacharya and Chaudhuri, 1986; Bhattacharya, 1992; author's unpublished data). Emplacement age of these alkaline rocks is unknown.

4. Sea level change, palaeogeographic shift and contemporary basin tectonics: discussion and conclusions

During the emplacement of the Singhbhum granite (Saha, 1994) the crust in the Singhbhum crustal province was ~48 km thick (Mazumder et al., 2000; see alsoBhattacharya and Shalivahan, 2002; Shalivahan and Bhattacharya, 2002; Mazumder and Sarkar, 2004) at ca. 3100 Ma. Calculations following the method suggested by Condie and Potts (1969) and Condie (1973), from Rb–Sr distribution and the silica normalized K_2O content of the Dhanjori volcanic rocks, suggest that the crust was 15 to 20 km thick (Mazumder et al., 2000, their Fig. 4b, and Table 1) at ca. 2100 Ma (Roy et al., 2002a). Cooling down of the vast volume of Singhbhum granite may have induced an isostatic readjustment. The associated tensional regime and deep-seated fractures controlled the formation of the Dhanjori basin (Roy et al., 2002a; Mazumder, 2002; Mazumder and Sarkar, 2004). Roy et al. (2002a) reported an age of 2072±106 Ma (MSWD 1.56) based on Sm–Nd isotopic studies of the Upper Dhanjori basaltic to gabbroic rocks from the Rakha Mines area (Figs. 2 and 3). Lack of precise age data is the major impediment to infer the time of basin formation and initiation of Dhanjori sedimentation.

The Dhanjori Formation is entirely terrestrial, dominantly fluvial. Mineralogically and texturally immature clastics form the major components. There are coarser grained remnants of an alluvial fan complex at the base of the formation, whereas the rest of the unit is almost entirely constituted by fining-upward fluvial cycles (Mazumder, 2002; Mazumder and Sarkar, 2004). An event of basin tilting and volcanic eruption intervened and resulted in a second phase of Dhanjori sedimentation, although the general fluvial depositional framework remained unaltered (Mazumder and Sarkar, 2004, p. 283; Fig. 3). The two members of the Dhanjori Formation display different palaeocurrent trends related to fluvial response to basin tilting (Fig. 3; see Mazumder and Sarkar, 2004). In the lower member spatial variability in palaeocurrent directions occurred as a consequence of fan development. Unlike the first phase, the second phase of fluvial deposition incorporates profuse volcanics and accompanying pyroclastic and epiclastic deposits. These volcanic and volcaniclastic rocks might have locally blocked the river courses resulting in short-lived lacustrine deposition (see Mazumder and Sarkar, 2004, pp. 278–279). Composition of the volcanic rocks varies upward from ultramafic to mafic (Gupta et al., 1985; Gupta and Basu, 1991, 2000; Mazumder and Sarkar, 2004). The Dhanjori volcanism took place in an intracontinental rift setting as is evident from the interbedded terrestrial deposits. Interbedded volcanic and volcaniclastic rocks at different stratigraphic levels suggest episodic volcanic eruption and an increasingly important influence of volcanism on sedimentation (Mazumder and Sarkar, 2004).

The Dhanjori sedimentation ended with a major transgression (Bose et al., 1997; Mazumder et al., 2000). As a consequence, the overlying Chaibasa sea over-stepped the Dhanjori Formation and onlapped the granitic basement at the southeastern corner of the study area (Fig. 2). Relative sea level rise was a prerequisite and that was possibly due to basin subsidence. The transgressive lag (Fig. 7a; cf. Cattaneo and Steel, 2003) at the contact between the two formations thins and fines westward, i.e., seaward (cf. Bose et al., 1997; Mazumder, 2002). In the lower part of the Chaibasa Formation around Singpura (Fig. 2) relatively poorly sorted tide-affected nearshore sandstones pass westward into offshore shales (Mazumder, 2002). The upward facies transition from sandstone to shale in the middle part of the Chaibasa Formation in the Ghatshila-Moubhandar sector (Figs. 2 and 7a,b) is generally gradational because of intervention of the heterolithic facies. In significant contrast, the upward transition is sharp in the absence of the heterolithic facies (Bose et al., 1997, their Fig. 2a, b; Mazumder, 2002). The depositional cycles defined by successive such sharp transitions are thus separated from each other by marine flooding surfaces (offshore shale). The sandstone facies units developed gradationally over the heterolithic facies, have progradational bedforms predominant at their bases, and aggradational bedforms at their tops (Bose et al., 1997; Mazumder, 2002). A low rate of rise in relative sea level was apparently followed by a relatively enhanced rate of rise. Then, with the sharp upward transition from the sandstone directly to the shale facies, a further enhancement in the rate of rise in relative sea level is inferred (Mazumder, 2002). The Chaibasa succession thus seems to be overall transgressive, but was intermittently punctuated by short-term lowstands (Mazumder et al., 2000). These punctuations were possibly owed to a fluctuating rate of the overall transgression (cf. Cattaneo and Steel, 2003). Although the shale facies of the Ghatshila-Dhalbhumgarh sector are devoid of wave imprint and evidence for subaerial exposure (Fig. 2; see Bose et al., 1997, their Fig. 1, pp. 74–78), the shale facies of the Galudih-Deoli sector and those lying further north bear superimposed ripples and desiccation cracks (Fig. 2; Bhattacharya, 1991, his Figs. 6, 12). This implies that the Upper Chaibasa shale facies is of intertidal origin (Bhattacharya, 1991; Bhattacharya and Bandyopadhaya, 1998; cf. Eriksson, 1979; Reineck and Singh, 1980; Eriksson et al., 1994, pp. 57–58), indicating thereby progressive shallowing of the palaeogeographic setting during the terminal phase of Chaibasa sedimentation. This is further corroborated by regional ground gravity surveys over the Precambrian terrain of eastern India (Verma et al., 1978, 1984, 1996; see also Gupta and Basu, 2000, pp. 211–218, for a review). Analysis of the gravity field over north Singhbhum clearly shows two prominent gravity low areas (−36 m. Gal) on the Deoli and Galudih, in contrast to a prominent gravity high (−15 to −20 m. Gal) displayed by Proterozoic supracrustals to the north of the Singhbhum Shear Zone, indicating thereby basement highs and relatively smaller thickness of supracrustal cover there (see Gupta and Basu, 2000, p. 213, their Fig. 16a).

Soft sediment deformation structures (cf. Maltman, 1984, 1994) are abundant in the Chaibasa Formation, in short but laterally persistent selective stratigraphic intervals separated by undeformed intervals, disregarding any lithological bias (Bose et al., 1997; Bhattacharya and Bandyopadhaya, 1998; Mazumder, 2002). These include pillows, multilobate convolute structures (Fig. 15a), slump folds (Fig. 15b), slump scars, (Fig. 15c), penecontemporaneous thrusts (Fig. 15d), and overturned cross-bedding (Fig. 15e). Although these structures are by themselves not diagnostic of any specific triggering mechanism, their abundance clearly suggests deposition in a tectonically active basin (Bhattacharya and Bandyopadhaya, 1998; Mazumder, 2002). Almost all the structures except overturned cross-bedding occur in the facies deposited below the storm wave base, for which a triggering mechanism other than an earthquake is very unlikely (Mazumder, 2002). Bhattacharya and Bandyopadhaya (1998) interpreted some of these structures as seismites (cf. Seilacher, 1984). Paucity and abundance of soft sediment deformation structures in the lower and upper part of the Chaibasa Formation, respectively, suggest that the marine shelf became considerably narrower in the later phase of Chaibasa depositional history (Mazumder, 2002). Destabilized shelf clastics, consequently, frequently slumped down the slope and, on liquefaction, gave rise to fine-grained mass flows into the basin.

The upward transition from the marine Chaibasa Formation to the terrestrial (fluvial-aeolian) Dhalbhum Formation indicates relative sea level fall and thus, subaerial exposure of the depositional surface (i.e., higher continental freeboard, cf. Wise, 1972; Eriksson, 1999). The Chaibasa-Dhalbhum contact therefore represents a sequence boundary (Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2003; cf. Posamantier and Vail, 1988). From magnetotelluric studies, Shalivahan and Bhattacharya (2002, pp. 1261–1262) estimated that the granitic crust is ca. 23 km thick in the northwestern part of the Singhbhum craton, towards the Dalma volcanic belt, as compared to ca. 38 km thick granitic crust at the southeastern end. The crustal thinning has been attributed to a much younger (~1600 Ma) plume magmatism in the Dalma volcanic belt (Roy et al., 2002b; Shalivahan and Bhattacharya, 2002). As Kent (1991) pointed out, the initial thermal input of the plume results in gentle doming and low-extension factor rifting, possibly accompanied by subsequent thinning of the subcontinental lithosphere. Such upliftment (doming) as a consequence of mantle plume upwelling may allow the establishment of a well-defined drainage network and sediment dispersal

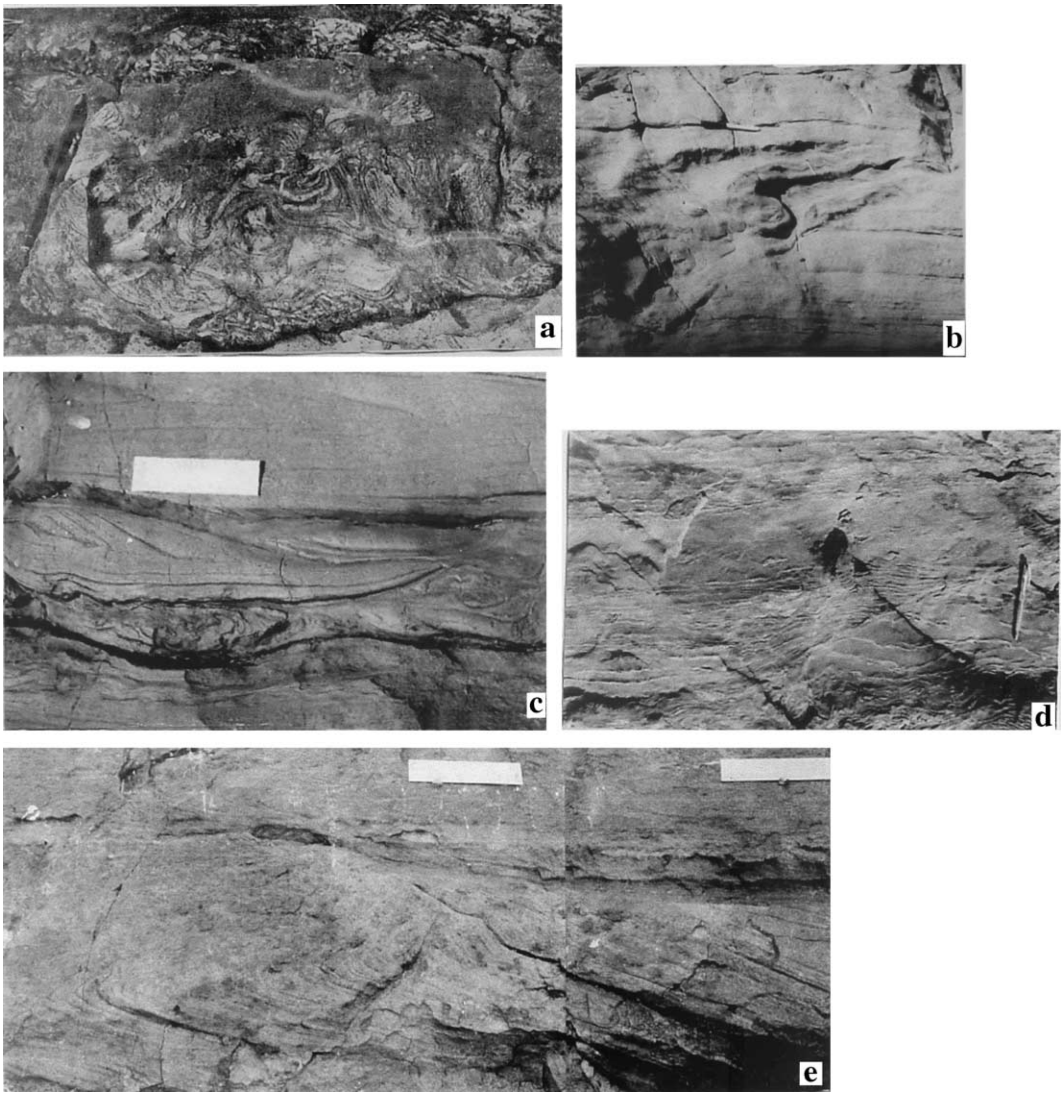


Fig. 15. (a) Multilobate convolute structure within the deep sea Chaibasa shale (pencil length 15 cm). (b) Slump fold within the Chaibasa heterolithic facies (after Bose et al., 1997; pen length 12 cm). (c) Slump scar within the Chaibasa heterolithic facies (scale length 16 cm). (d) Penecontemporaneous intrastratal thrust within the deep sea Chaibasa shale (pen length 13 cm). (e) Overturned cross-bedding within the fine-grained Chaibasa sandstone (scale length 16 cm).

pattern (cf. Cox, 1989; Kent, 1991). Progressive shallowing of the palaeogeographic setting from offshore deep marine (Bose et al., 1997) to shallow intertidal (Bhattacharya, 1991; Bhattacharya and Bandyopadhaya, 1998) during the terminal phase of Chaibasa sedimentation, and its transition to the Dhalbhum terrestrial (fluvial-aeolian, Mazumder, 2003; Simpson et al., 2004) depositional setting, indicate that such gentle crustal doming indeed took place prior to the commencement of volcanism (Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2002; cf. Rainbird, 1993; He et al., 2003; Mazumder, 2004a). As a consequence, the continental freeboard (elevation of the continent above the mean sea level, cf. Wise, 1972; Eriksson, 1999) was higher. Interestingly, the palaeocurrent pattern as determined by Naha and Ghosh (1960; see also Bose, 1994, p. 336) from the Dhalbhum Formation close to the southern flank of the Dalma volcanic belt, is southward-oriented, in contrast to the northward palaeocurrent trend of the underlying Chaibasa (south of Dalma Formation, cf. Naha, 1961; Bose et al., 1997) and overlying Chandil sandstones (north of the Dalma Formation, cf. Dunn and Dey, 1942, p. 331; Bhattacharya, 1992; authors own unpublished data). It must be noted that bimodal volcanism (basalt–rhyolite) is a common plume-related product (cf. Hatton, 1995; Eriksson et al., 2002; P.G. Eriksson, personal communication, 2004) and therefore occurrence of acidic volcanic and volcaniclastic rocks within the Dalma volcanic belt and further north (Chandil Volcanics) is not inconsistent with the plume model as argued by Bose (1994, p. 329). Available sedimentological, stratigraphic, geochemical and geophysical data thus strongly support that the Dalma volcanism was a consequence of mantle plume upwelling in a continental rift setting (cf. Kent, 1991).

Zhao et al. (2002) postulated that amalgamation of Archaean to Palaeoproterozoic cratonic blocks took place during a global collisional orogeny at 2100–1800 Ma. It has also been proposed that the collision between the Southern and Northern Indian cratonic blocks (cf. Eriksson et al., 1999; Mazumder et al., 2000) resulting in the formation of the Indian craton took place during a poorly defined period between 2100 and 1700 Ma as part of this global collisional event (Zhao et al., 2002, p. 141; Zhao et al., 2003a). Contrary to the global collisional event at ca. 2100–1800 Ma, the Singhbhum Late Palaeoproterozoic volcano-sedimentary successions document their generation in an intracontinental extensional tectonic regime (Gupta et al., 1980; Roy et al., 2002a; Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2002; Mazumder and Sarkar, 2004). The ~2100 Ma Dhanjori volcanism took place in an intracontinental rift setting as is clear from the interbedded terrestrial deposits (Mazumder and Sarkar, 2004). The Mesoproterozoic (~1600 Ma) Dalma plume was possibly part of a global thermal perturbation that affected pre-1600 Ma landmasses during which lithospheric thinning, sedimentation, and crustal melting/anatexis took place in all major cratonic provinces including Singhbhum (cf. Roy et al., 2002b and references therein). The entire rift-related volcano-sedimentary succession south of the Dalma volcanic belt and comprising the Dhalbhum, Chaibasa and Dhanjori Formations extended further southward towards the Archaean Singhbhum Granite batholith complex as a consequence of uplift associated with the Dalma plume at ca. 1600 Ma. The Singhbhum Granite batholith acted as a rigid body (cf. Sengupta and Mukhopadhaya, 2000) and gave rise to a compressional stress regime. Inversion of continental rift basins during compression is common (cf. Ziegler et al., 1995; Hansen and Nielsen, 2003; Turner and Williams, in press). As a consequence of this ~1600 Ma compression, the Late Palaeoproterozoic volcano-sedimentary successions were deformed and metamorphosed, and shearing/thrusting took place along the pre-existing weak zone close to the Singhbhum Granite complex (SSZ). Sengupta and Mukhopadhaya (2000, p. 55) suggested that the end of the thrusting event postdates the eruption/consolidation age of the Dalma-Chandil volcanogenic sequence at ca. 1500 Ma. The silica-undersaturated intrusive synites (Fig. 14b) occurring to the north of the Dalma volcanic belt were possibly derived by fractionation of the alkaline magma component of the Dalma volcanism (Bhattacharya, 1992). Absence of any pervasive planar or linear fabric (except excellent compositional banding, Fig. 14c) in these rocks, and truncation of the structural planes in the surrounding metasedimentary country rocks against the contact of these intrusive bodies, indicate that they were emplaced in an anorogenic setting (cf. Bhattacharya, 1993).

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