Available online at www.sciencedirect.com

SCIENCE DIRECT®

Precambrian Research 130 (2004) 267–287

Precambrian Research

www.elsevier.com/locate/precamres

Sedimentation history of the Palaeoproterozoic Dhanjori Formation, Singhbhum, eastern India

Rajat Mazumder[a,*], Subir Sarkar[b]

[a] Geological Institute, Yokohama National University, 79-7, Tokiwadai, Hodogaya ku, Yokohama 240-8501, Japan
[b] Department of Geological Sciences, Jadavpur University, Kolkata 700 032, India

Received 29 January 2003; accepted 10 December 2003

Abstract

The Palaeoproterozoic Dhanjori Formation, eastern India is a terrestrial (dominantly fluvial) volcano-sedimentary succession. Basal conglomerate, coarse-grained sandstone, and shale rests directly on the granite basement and represents the distal fringe of an alluvial fan complex. Sieve, sheet flood, mass flow and alluvial channel and overbank deposits comprise the fan segment. The rest of the formation is entirely constituted by fining upward fluvial cycles. An event of basin tilting and volcanic eruption intervened and resulted a second phase of Dhanjori sedimentation, although the general fluvial depositional framework remained unaltered. The two members of the Dhanjori Formation display different paleocurrent trends related to fluvial response to basin tilting. In the lower member spatial variability in paleocurrent directions occurs as a consequence of fan development. Unlike the first phase, the second phase of fluvial deposition incorporates profuse volcanics and accompanying volcaniclastic deposits. These volcanic and volcaniclastic rocks might have locally blocked the river courses resulting in short-lived lacustrine deposition. Composition of the volcanics varies upward from ultramafic to mafic. The Dhanjori volcanism took place in an intracontinental rift setting as is evident from the interbedded terrestrial deposits. Interbedded volcanic and volcaniclastic rocks in different stratigraphic levels suggest episodic volcanic eruption. A semiarid paleoclimate is proposed and is consistent with development of an alluvial fan comprised of coarse clastic deposits.

Keywords: Palaeoproterozoic; Fluvial; Alluvial fan; Volcanism; Continental rift; Singhbhum

1. Introduction

The lack of vegetation in the Precambrian would have enhanced flash flooding, rapid erosion, and low-sinuosity braided fluvial systems (e.g. Long, 1978, 2002; Eriksson et al., 1998). Frequent volcanism has a profound effect on fluvial depositional systems (Smith, 1991; Kataoka and Nakajo, 2002). For example, the sudden influx of voluminous volcanic and volcaniclastic material into a restricted basin can create resistant barriers and may have pronounced effect on the sediment dispersal patterns and depositional sequences (Kuenzi et al., 1979; Mueller et al., 1994). Although volcanic, pyroclastic, and sedimentological aspects of terrestrial depositional systems have been discussed by a number of authors (Smith, 1987, 1988; Smith and Swanson, 1987; Mueller et al., 1994; Altermann, 1996; Mueller and Corcoran, 2001; Mueller and Mortensen, 2002), the influence of allocyclic controls such as tectonism and volcanism on basin formation, sediment dispersal patterns and

* Corresponding author. Present address: Department of Geology, Asutosh College, Kolkata 700 026, India.
Tel.: +81(0)8031791213 (mobile).
E-mail address: mrajat2003@yahoo.com (R. Mazumder).

0301-9268/$ – see front matter © 2004 Elsevier B.V. All rights reserved.
doi:10.1016/j.precamres.2003.12.005

facies architecture have not been adequately addressed (cf. Sato and Amano, 1991; Corcoran et al., 1999; Kataoka and Nakajo, 2002). This paper describes facies characteristics, paleogeographic setting, and the sedimentation history of the Palaeoproterozoic Dhanjori Formation of the Singhbhum crustal province, India and highlights the volcanic and tectonic influences on sedimentation.

2. Geological background

The 2.1-Ga-old Dhanjori Formation (Roy et al., 2002) covers 800 sq. km area, and is composed of siliciclastic sedimentary rocks interlayered with mafic to ultramafic and rare acidic volcanic and volcaniclastic rocks deformed and metamorphosed to lower greenschist facies (Dunn and Dey, 1942; Iyenger and Alwar, 1965; Sarkar and Saha, 1962, 1977; Sarkar, 1984; Gupta et al., 1985; Basu, 1985; Singh, 1998; Singh and Nim, 1998). The Dhanjori Formation crops out along an east-west trending, northerly dipping broadly linear belt extending from Singpura in the southeast to Narwa-Pahar in the northwest (Fig. 1). The thickest section up to 1.25 km is in the central part of the study area across the Mosaboni-Barunia sector (Fig. 2). As one moves away from the central part, the volcano-sedimentary succession progressively becomes thinner and ultimately pinches out (Fig. 2). Dunn and Dey (1942), and Sarkar and Saha (1962, 1983) considered the Dhanjori Formation to be younger than the Chaibasa Formation mainly on the basis of relatively higher grade of metamorphism in the Chaibasa despite its stratigraphically higher position. In contrast, Sarkar and Deb (1971) and Mukhopadhaya (1976) stated that the Dhanjori-Chaibasa succession preserves normal stratigraphic order and the Dhanjori Formation is therefore older. The Dhanjori Formation directly overlies the Singhbhum Granite and is, in turn, overlain by mica schist-quartzite assemblages of the Chaibasa Formation (Bose et al., 1997; Mazumder et al., 2000). The contact between the two formations, defined by a conglomerate, has been interpreted as an unconformable contact (e.g. Dunn and Dey, 1942; Basu, 1985; Singh, 1998; Singh and Nim, 1998). Alternatively, Gupta et al. (1985) suggested that the Dhanjori Formation represented a lateral facies variation of the Chaibasa

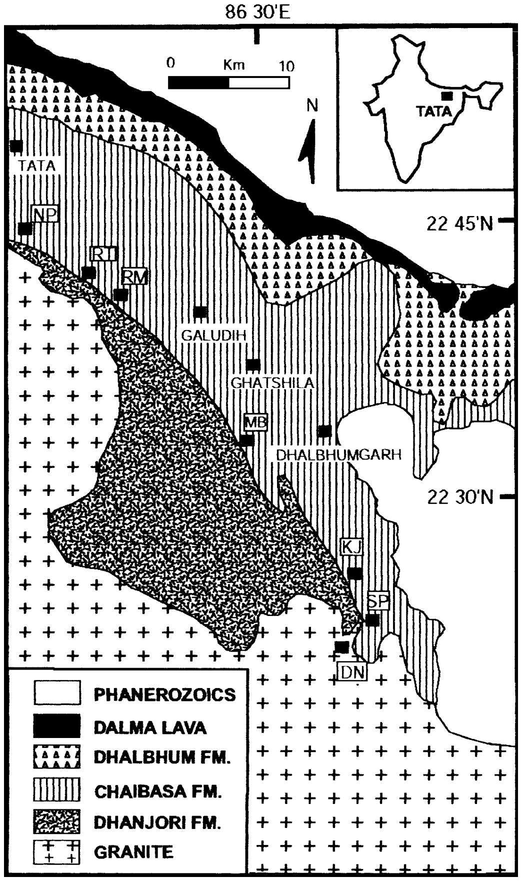

Fig. 1. Simplified geological map of the Dhanjori basin; study locations: DN, Dongadaha; KJ, Khejurdari; SP, Singpura; MB, Mosaboni; RM, Rakha Mines; RK, Rukmini Temple (Jadugorah); NP, Narwa Pahar (modified after Saha, 1994).

Formation and the two assemblages were telescoped during later deformation. Mukhopadhaya (1976), Bose (1994), Mazumder et al. (2000), and Mazumder (2003) have presented evidence for a continuous depositional history and normal order of superposition between the Dhanjori and Chaibasa formations.

The Dhanjori basin is flanked by the Singhbhum Shear Zone (SSZ) to its northern and eastern side and by the Iron Ore group of rocks in the northwest (Mukhopadhaya, 1984, 1988; Ghosh and Sengupta, 1987; not shown in Fig. 1; for details see Saha, 1994). Small linear granitic bodies (Arkasani granophyre and Soda granite) are present along the SSZ (cf. Saha, 1994). The Rb–Sr and lead isotopic ages (ca. 1600 Ma) obtained from these granites are inferred

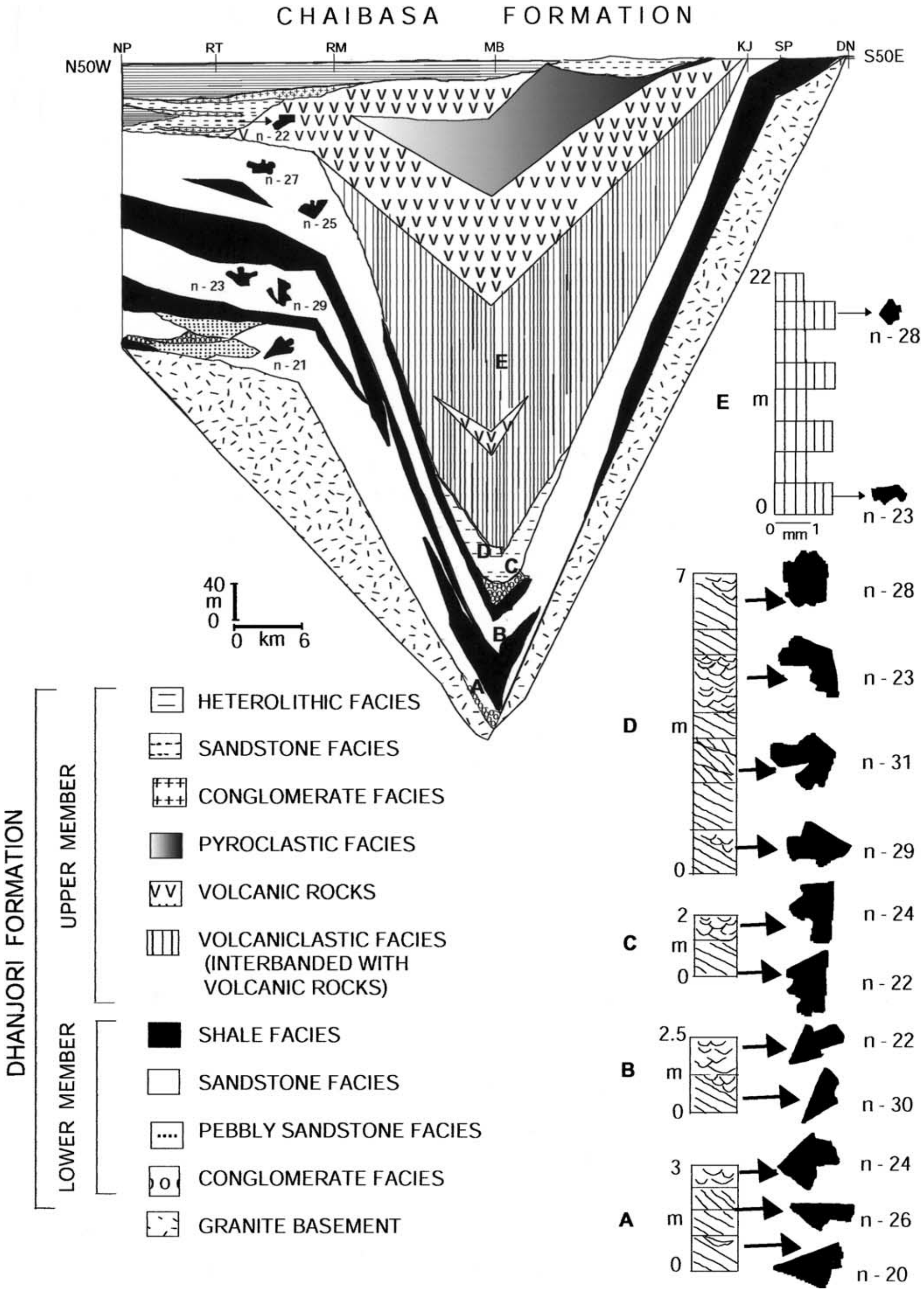

Fig. 2. Panel diagram showing lateral and vertical facies transition during Dhanjori sedimentation. Short sections showing interrelationship between various sandstone facies and temporal change in palaeocurrent direction are also shown. Study locations are marked in Fig. 1.

as the timing of metamorphism of the sediments and thrusting along the SSZ (cf. Sarkar et al., 1986; Sengupta and Mukhopadhaya, 2000). A comparable age of 1580 Ma has been obtained from uraninite samples collected from the SSZ (Krishna Rao et al., 1979). Sengupta and Mukhopadhaya (2000) suggested that at 1600 Ma isotopic re-homogenization occurred subsequent to the crystallization of these granites.

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 (Gupta et al., 1985; Mazumder, 2002). In addition, there is a considerable change in the dip of beds (~35° to ~60°) between the two members, with steeper dips measured in the upper member. The lower member has conglomerate at its base and displays overall fining upward trends. The upper member is the volcano-sedimentary portion of the Dhanjori Formation. Volcaniclastic rocks at the base of the upper member are mostly ultramafic in composition. In the upper part of the Upper member, basaltic lava flows alternate with the volcaniclastic rocks (Fig. 2; Mazumder, 2002). Recently minor acidic lava flows and pyroclastic rocks have been reported from near the top of the upper Member (Singh, 1998, p. 384–385).

3. Petrography

Although sedimentary rocks in the Dhanjori Formation have undergone deformation and lower greenschist facies metamorphism, primary textural features are locally well preserved. Constituent grains have sutured contacts from pressure welding. Because of good preservation of textural characteristics the prefix 'meta-' has been omitted and the precursor rock names have been used. Here we concentrate on general petrographic character of the sandstones.

The sandstones are generally poorly sorted, medium-to coarse-grained with local granule-rich layers. Sandstone matrix content is locally variable and sometimes higher than 15%; however, typically it is around 10% (Fig. 3a). Where detrital grains retain their primary boundaries they appear subangular to subrounded. Quartz is the predominant mineral (>50% by volume); however, potassium feldspar, plagioclase,

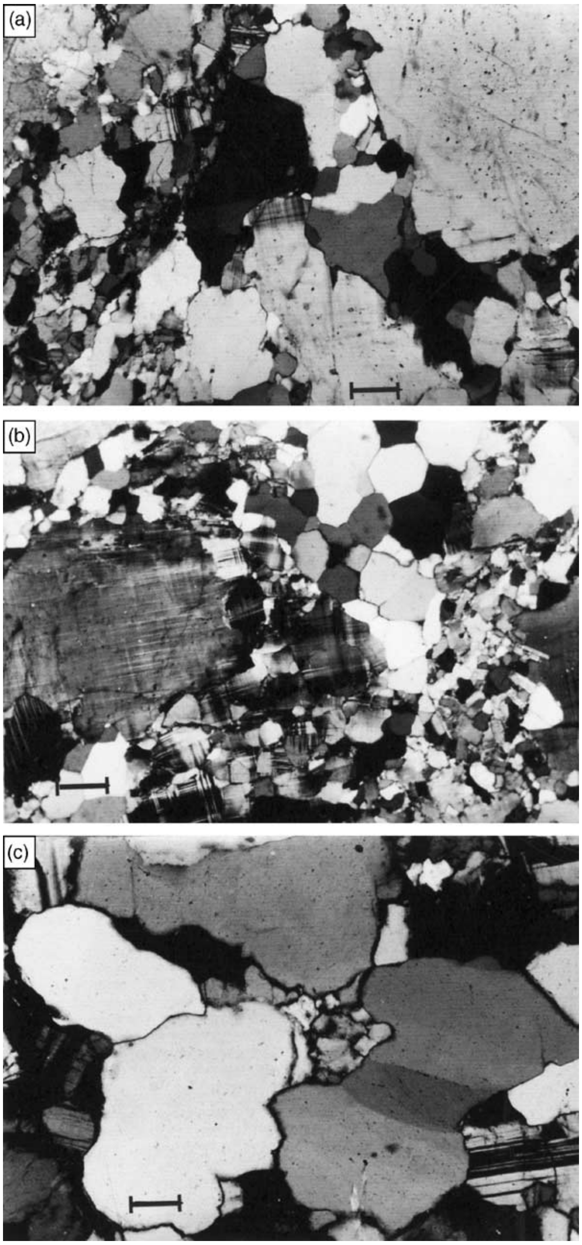

Fig. 3. (a) Coarse-grained, poorly sorted sandstone (bar length 0.2 mm). (b) Poorly sorted feldspathic sandstone; note dominance of alkali feldspar (microcline; bar length 0.2 mm). (c) Coarse-grained sandstone with lithic fragment and plagioclase feldspar (bar length 0.2 mm).

and lithic fragments occur in fair proportion (Fig. 3b and c). Plagioclase feldspars are often altered to an assemblage of clay and micaceous (sericite) minerals. Lithic grains are mostly quartzite and basic volcanic rock fragments. Granitic rock fragments, always rare, remain confined to the basal sandstones of the lower member. 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 litharenite/lithic wacke depending on relative proportion and dominant species of feldspars and matrix content (cf. Pettijohn, 1975; see also Bose, 1994; Mazumder, 2002).

4. Facies analysis

Sedimentary units have been divided into assemblages according to grain-sizes, and facies represent subunits of the assemblages (Table 1). The two members of the Dhanjori Formation are similar in terms of their facies character, although significant differences do exist. Because of the large-scale similarity the facies constituents of both members are described together, with emphasis on important differences.

4.1. Conglomeratic assemblage

The conglomerates (up to 4 m thick) are found as localized lens concentrated only at the bases of the two members (Fig. 2). Clasts are mostly of vein quartz, quartzite and rarely of shale, siltstone, granitic rocks (Fig. 4a). The size of the clasts varies by composition: quartzite (up to 3 cm), vein quartz (>1.5 cm) and other rock fragments (>0.5 cm) dominate the clast population. The shape of the pebbles varies from spherical, elliptical to tabular but all are generally well abraded along their edges. The assemblage can be subdivided in five facies.

4.1.1. Clast-supported sheet conglomerate facies (facies A)

Facies A is a clast-supported conglomerate (up to 2 m thick) with a sandy matrix filling the interstitial spaces (Fig. 4b). It is sheet-like, but broadly lenticular in geometry, and is traceable laterally for distances of 10 m within the exposure limit. Average bed-thickness is about 20 cm. Vertical amalgamation of facies units measures up to 55 cm, although generally the beds are separated by laminated pebbly sandstones with scouring at the base. Both lower and upper contacts of conglomerate units are sharp. The clasts are relatively well sorted and have an average length of about 7 cm. In places clasts show a moderately developed bedding parallel alignment. In limited bedding plane exposures they are also found oriented roughly parallel to each other.

These conglomerates are inferred as products of sheet flow, because of their sheet-like geometry, sorting, medium size, local bed-parallel alignment of clasts and interstratification with laminated pebbly sandstones (Hogg, 1982; Blair and McPherson, 1994; Blair, 1999a,b). Such sheet flood deposits have been described from modern (Van de Kemp, 1973; Nemec and Postma, 1993; Blair and McPherson, 1994) and ancient successions (Mueller et al., 1994; Dillard et al., 1999; Martins-Neto, 2000, his Fig. 8b).

4.1.2. Convex-up matrix-supported conglomerate facies (facies B)

Facies B are matrix-supported, lenticular conglomerates up to 2.5 m thick and 4.5 m wide. Their top surfaces are more convex than their bases. The clasts are poorly sorted and range in size up to 18 cm. The matrix is silty and sandy mud. Clast orientation is chaotic (Fig. 4c) and local reverse grading is discernable. The upper bed surface is often irregular because of clast protrusion. However, both the base and top of the conglomerate beds are sharp. Bed-thickness ranges up to 50 cm.

Poor sorting, chaotic clast orientation, clast protrusion above bed surfaces, convex-up body geometry, and local reverse grading suggest that these conglomerate bodies are likely by debris flow products (Johnson, 1970; Nemec et al., 1980; Sohn et al., 1999; Blair, 1999b,c; Martins-Neto, 2000, his Fig. 8a; Dasgupta, 2003).

4.1.3. Convex-up matrix to clast-supported conglomerate facies (facies C)

Facies C conglomerates are up to 4 m thick and are bi-convex in shape. The maximum bed thickness observed is 11 cm. Their width in outcrop is 65 cm.

Table 1
Characteristics of the sedimentary and volcaniclastic lithofacies in the Dhanjori Formation

Lithofacies	Facies constituents/characteristics	Process	Inferred setting
Conglomerate lithofacies assemblage	Clast-supported sheet conglomerate (facies A)	Sheet flow	Alluvial fan
	Convex-up matrix-supported conglomerate (facies B)	Debris flow	Alluvial fan and oversteepened fluvial channel
	Convex-up matrix to clast-supported conglomerate (facies C)	Sieve deposit	Alluvial fan
	Stratified conglomerate (facies D)	Deposited from high velocity traction current	Fluvial channel
	Pebbly sandstone (facies E)	Deposited from waning flash floods	Fluvial channel
Sandstone lithofacies assemblage	Trough cross-bedded coarse-grained sandstone (facies F)	Larger troughs are product of mid-channel bar migration; smaller troughs were formed by bedform migration over the mid-channel bars	Fluvial bank-attached bar
	Tabular cross-stratified coarse-grained sandstone (facies G)	Bank-attached fluvial bar migration	Fluvial bank-attached bar
	Down-dip cross-stratified sandstone (facies H)	Abandoning of bank attached bars during water level rise	Abandoned fluvial bank attached bar
	Trough cross-bedded fine-grained sandstone (facies I)	Large ripple migration	Sandflat
Heterolithic facies	Fine-grained sandstone-siltstone interbedding	Silt grade particles with red cutans are wind deflated; deposited in a restricted water body; sandstone interbeds are formed by episodic hyperpycnal underflows	Lacustrine
Shale facies	Grey shale with siltstone interbeds, having sheet-like geometry	Suspension sedimentation	Floodplain
Volcaniclastic facies	Ultramafic to mafic volcaniclastic rocks with angular to subrounded clasts, generally cross-stratified; confined within the younger member	Fluvially reworked volcanic materials	Shallow fluvial channel-fills
Pyroclastic facies	Mafic-ultramafic (rarely acidic) tuff and agglomerates containing flattened lapillis; generally massive but occasionally shows dune like bedforms having high angle foresets; the acid pyroclastics exhibit normal grading	Pyroclastic flow and surge	Generally fluvial; the acid tuff and agglomerates are of air-fall origin (Singh, 1998, p. 385)

The conglomerate is almost clast-supported at the base, but rapidly becomes matrix-supported upward (Fig. 4d). The matrix is sandy. The majority of clasts lie parallel to bedding however, some are subvertical. The clast population is poorly sorted, but is distinctly smaller (<5 cm) in size than clasts in facies B. The beds are internally massive except for normal grading transitioning rapidly upwards into the pebbly sandstone.

The convex-up geometry and general massiveness of beds, poor sorting of clasts and their local subvertical orientation suggest very rapid sedimentation. The facies is inferred as a sieve deposit (cf. Einsele, 1992; Collinson, 1996; Mazumder, 2002). Water from the sediment-laden flow probably percolated down rapidly. The upward increase in matrix content suggests that much of the sand matrix infiltrated into the coarser sediment.

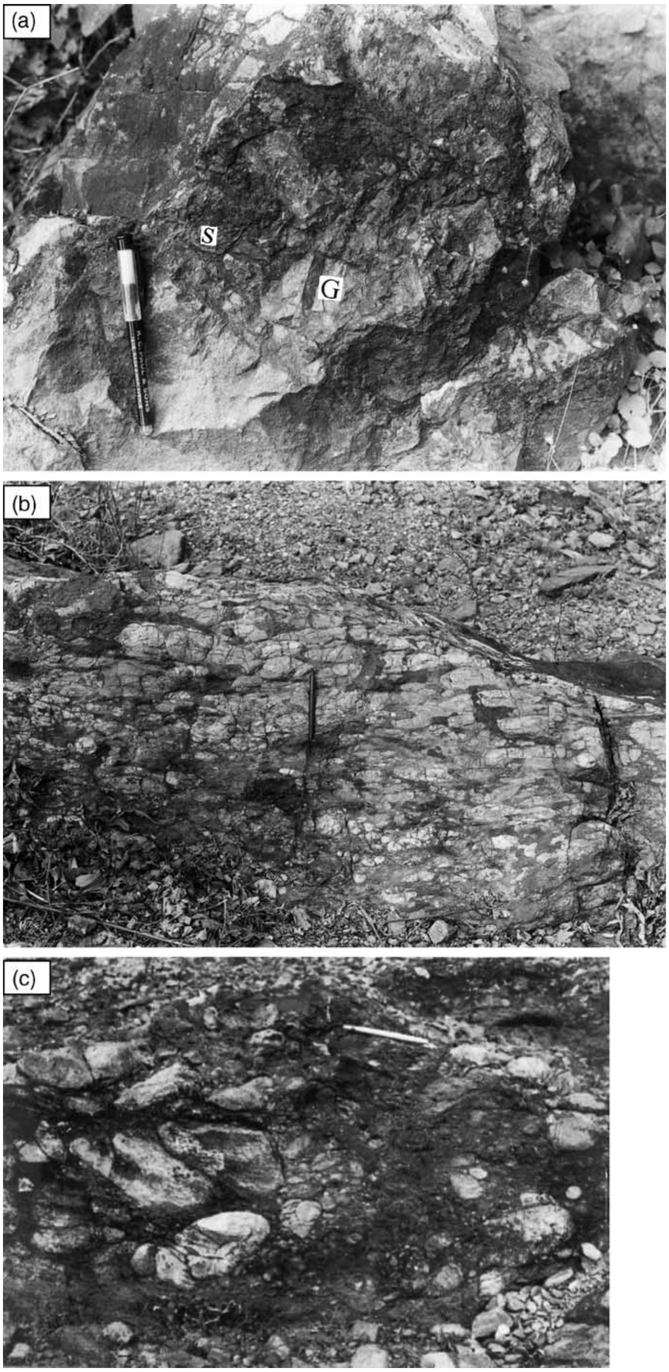

Fig. 4. (a) Conglomerate with granite (G) and shale (S) clasts (pen length 12 cm). (b) Clast-supported sheet conglomerate body (facies A; pen length 12 cm). (c) Conglomerate with poorly sorted clast population; note chaotic clast orientation (facies B; pen length is 12 cm). (d) Clast-supported conglomerate rapidly turning matrix-supported upward; note normal grading lower left corner of the photograph (facies C; pen length is 12 cm). (e) Clast-supported conglomerate with moderately sorted pebbles; note lenticular geometry (facies D; scale length 10 cm). (f) Massive pebbly sandstone associated with facies D (facies E; coin diameter 1.5 cm).

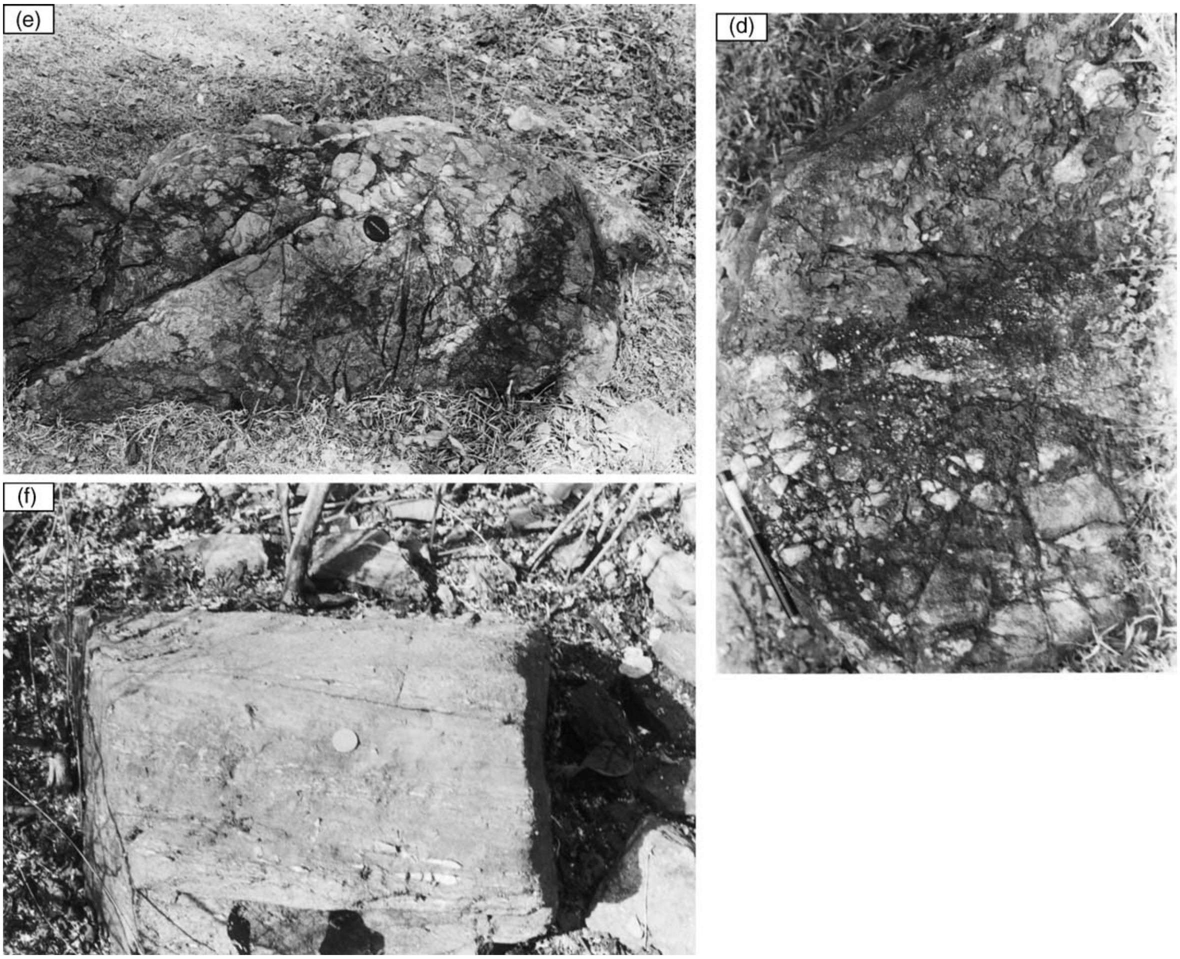

Fig. 4. (*Continued*).

4.1.4. Stratified conglomerate facies (facies D)

Facies D is comprised of conglomerate up to 2.5 m thick with crude planar stratification. Channels have a concave-up base and a flattened top (Fig. 4e). Bed-thickness are up to 48 cm, however, amalgamation results in unit thickness are up to 90 cm. The width of the conglomerate measured in outcrop is up to 5.4 m, however, the actual width was often considerably more. The conglomerate is generally clast-supported (Fig. 4e). The clasts are moderately sorted, ranging in length up to 14 cm. This facies is always overlain by pebbly sandstone confined within the same channels. Lower and upper contacts of the conglomerate are always sharp.

The characteristics of this facies are consistent with deposition in fluvial channels from high velocity flows (cf. Miall, 1988; Collinson, 1996; Mazumder, 2002).

4.1.5. Pebbly conglomerate/sandstone facies (facies E)

This facies is more commonly a sandstone (Fig. 4f) and becomes a true conglomerate only rarely when its pebble content exceeds 30% by volume (cf. Pettijohn, 1975). Nevertheless, this facies is considered within the conglomeratic assemblage because of the fact that it occurs only in association with the conglomerates, particularly at the top of the assemblage found at the base of both two members. Characteristically it is massive, but locally may be planar laminated. A sandy top generally occurs in association with the stratified conglomerate as channel-fills.

In view of its association with channel deposits, this facies is considered a bankward deposit of the same channels. Alternatively, deposition during the waning stage of flash floods is also possible (cf. Collinson, 1996; Mazumder, 2002).

4.1.6. Summary

As mentioned above, the conglomeratic assemblage occurs as patches and only at the base of both the members of the Dhanjori Formation. The sieve deposits clearly point to subaerial deposition (cf. Blair and McPherson, 1994; Collinson, 1996; Mazumder, 2002). The assemblage at the base of the formation thus appears to belong to alluvial fans (Blair and McPherson, 1994; Collinson, 1996; Blair, 1999a,b). The conglomerates are understandably not the sole constituents of the fan deposit because the finer facies that are described below also occur between the conglomerate facies (Blair and McPherson, 1994). The conglomeratic bodies formed by mass flow, as well as, via traction currents. The assemblage probably represents the distal fringe of the alluvial fan below the intersection level of ground water with the fan surface (Collinson, 1996). Martins-Neto (2000, his Fig. 8a and b and Table 2) reported similar association of mass-flow conglomerates and clast-supported stratified conglomerates of traction current origin (Natureza tectonosequence) from the Palaeo/Mesoproterozoic Espinhaco basin, southeastern Brazil; he interpreted the association as having been deposited in an alluvial fan and braided stream depositional system.

The conglomeratic assemblage at the base of the younger member (Fig. 2) does not include any sheet flood and sieve deposits and is constituted by channel and mass flow deposits. The presence of mass flow conglomerates indicates a steeper dip of depositional surface. This is consistent with the greater dips in bedding in the younger member. Interestingly, sedimentary clasts are present only at the base of the younger member (Mazumder, 2002).

4.2. Sandstone assemblage

The sandstones appear broadly lenticular with variable thickness ranging up to 30 m (Fig. 2). Planar or broadly undulating master erosion surfaces (MES) occur at various stratigraphic levels within the sandstones. Between any two MES the sandstone segments fine upwards. Thickness of these segments is widely variable, both laterally and vertically, with a maximum of 10 m. Facies succession within all such segments is similar, although not all the sandstone facies are present everywhere. The various facies constituents of the sandstone assemblages are described below.

4.2.1. Trough cross-bedded coarse-grained sandstone facies (facies F)

This is very poorly sorted, granular sandstone (Fig. 5a) of lenticular geometry that immediately overlies the MES. Internally, a bimodal size population of trough cross-sets is characteristic. The minimum set thickness of the larger population is 15 cm, whereas the maximum set thickness of the smaller population is 10 cm. The smaller troughs remain confined to individual larger sets converging towards axes of the latter. Maximum dip of the larger foresets is about 7° and that of the smaller ones is about 15°. Although directional variability is relatively greater in case of the smaller sets, both the kinds of cross-strata are oriented in same direction in same outcrop or section. Their orientation, however, differs between outcrops and more markedly between the two members (Fig. 2). Facies unit thickness measures up to 3 m; maximum outcrop width is 7 m.

Poor sediment sorting, lenticular geometry and section-wise unimodal cross-strata orientation suggest this facies to be a fluvial deposit (Miall, 1996; Collinson, 1996; Eriksson et al., 1998). The low dip of the larger foresets, their occurrences invariably on the MESs and coarser grain-size suggest that the facies possibly formed as mid-channel bars (Smith, 1970; Rust and Jones, 1978; Bose and Chakraborty, 1994; Chakraborty, 1995). The smaller troughs are probably generated by megaripples that migrated from the stoss face down the low gradient lee face of the bars (Collinson and Thompson, 1989; Collinson, 1996). Bar morphology exerted a secondary control on migration of the smaller dunes/ripples, since smaller troughs converge towards axes of the larger troughs.

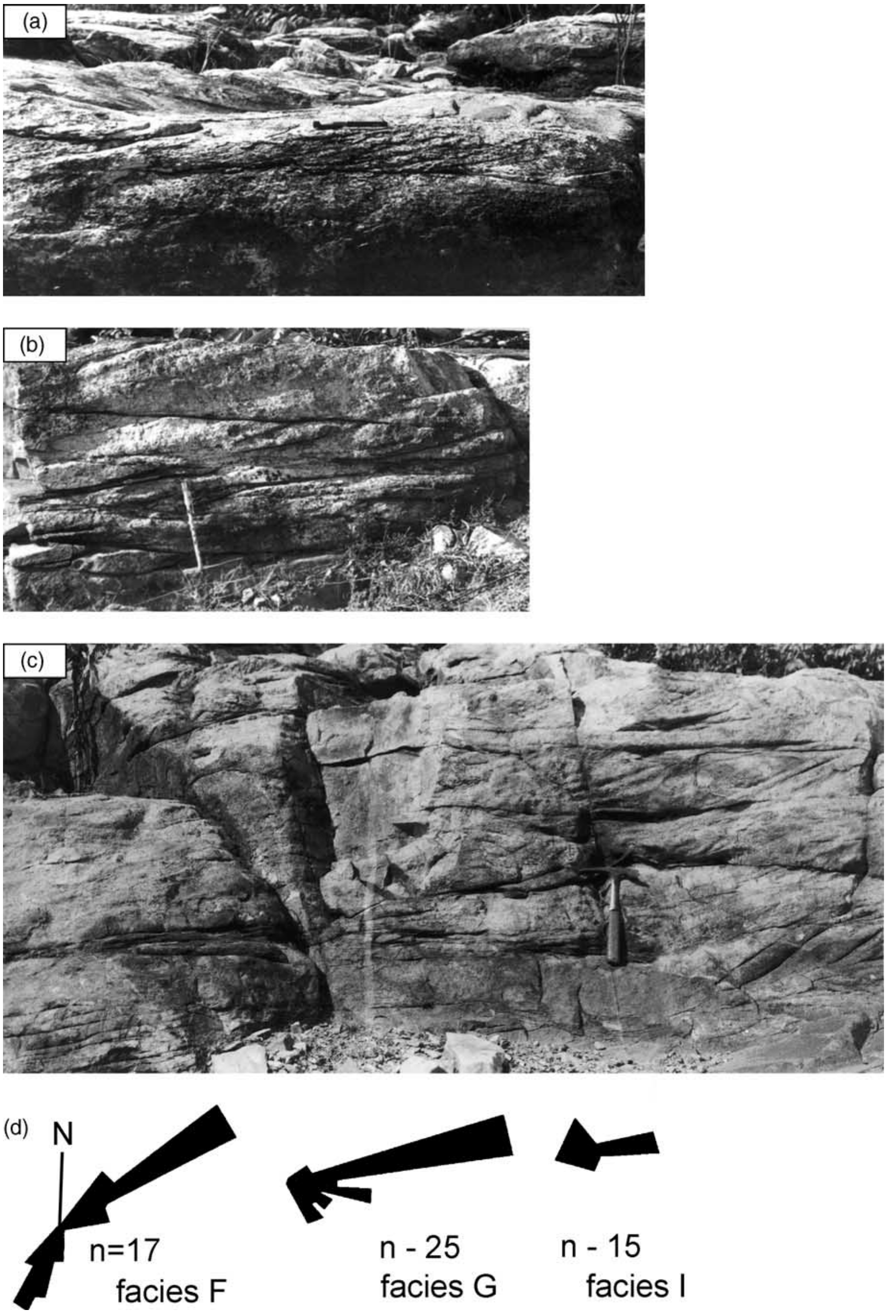

Fig. 5. (a) Trough cross-stratified sandstone (facies F) overlying MES (pen length is 12 cm). (b) Tabular cross-stratified sandstone (facies G) (pen length is 13 cm). (c) Facies G; note that the foreset dip decreases downcurrent (hammer length is 42 cm). (d) Paleocurrent rose showing cross-strata orientation between facies F, G, and I. (e) Down-dip cross-stratified sandstone facies (facies H) (hammer length 42 cm). (f) Trough cross-stratified sandstone facies (facies I) occurring on top of facies G (pen length 12 cm).

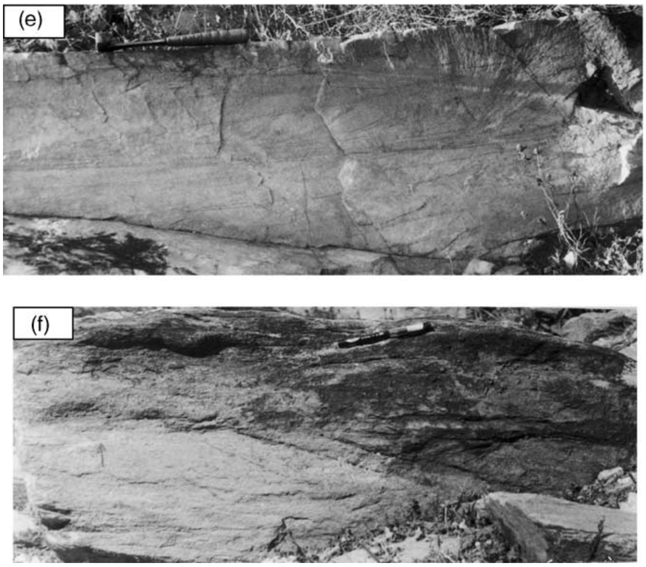

Fig. 5. (*Continued*).

4.2.2. Tabular cross-stratified coarse-grained sandstone facies (facies G)

This is a poorly sorted coarse-grained sandstone characterized internally by solitary sets of planar tabular cross-strata locally having short asymptotic toes (Fig. 5b). The maximum cross-set thickness is 27 cm. Notably, this facies also invariably occurs on the MESs. On many instances, the foreset dip decreases downcurrent (Fig. 5c). The facies generally forms convex-up bodies whose width is up to 5 m. At places, there are scours on top of the larger sets. The scour width is variable and ranges up to 75 cm, while their maximum depth is about 12 cm. The scours are generally filled by concave-up laminae roughly conforming the scour base. The scour-fill sediment is similar to that form the tabular cross-strata. On other instances, the fill is silty shale and almost massive. Cross-strata orientation is consistent within individual outcrops or sections, but differs between sections and more significantly between Members (Fig. 2). Facies F generally follows up-section.

This facies G possibly formed as bank-attached fluvial bars (cf. Chakraborty, 1995; Collinson, 1996; Mazumder, 2002). Towards the river bank the bar height was reduced. Occasionally scouring took place on the bar top (Cant and Walker, 1976; Walker, 1978). While most of the scours were filled up as soon as they formed, some may have remained open and subsequently filled with silty mud. The similarity of cross-strata orientation between the facies F and G (Fig. 5d) belonging to the same Member suggests down-channel orientation of these tabular cross-strata. Downcurrent decrease in foreset slope, noted on certain instances, possibly reflects rise in water level and gradual abandoning of the bars (Blodgett and Stanley, 1980; Kirk, 1983; Bose and Chakraborty, 1994; Chakraborty, 1995).

4.2.3. Down-dip cross-stratified sandstone facies (facies H)

This sandstone facies is relatively fine-grained, and internally characterized by down-dip cross-strata (Fig. 5e; cf. Banks, 1973; Miall, 1988; Kelley and Olsen, 1993; Bose and Chakraborty, 1994). The larger foresets have tabular geometry, and maximum dip of about 10° and smaller foresets make an angle of about 18° with them. This facies invariably rests on the facies G, or in places, on a MES. Both the larger and smaller cross-strata in an association dip consistently in the same direction (Fig. 2). At the upper surface a MES or shale, overlies this facies. Both large and small cross-strata maintain the same orientation as the closely associated cross-strata in facies G.

This facies is thought to have evolved from abandoning of bank-attached bars during water level rise (cf. Blodgett and Stanley, 1980; Kirk, 1983; Bose and Chakraborty, 1994). Ripples formed on the bar stoss presumably migrated down the bar lee as gradient of the latter decreased. The facies thus formed on top, as well as, between the bank-attached bars (Blodgett and Stanley, 1980; Kirk, 1983; Bose and Chakraborty, 1994; Chakraborty, 1995; Collinson, 1996).

4.2.4. Trough cross-bedded fine-grained sandstone facies (facies I)

This facies is finer grained than the facies H and internally characterized by small cross-sets having set thickness up to 8 cm (Fig. 5f). The facies units are wedge-shaped, range in thickness up to 42 cm, and generally overlie tabular cross-sets. They underlie MESs or the shale facies. Orientation of these troughs has a significant difference with that of the associated bimodal (in size) trough (facies F) or the large-scale tabular cross-strata (facies G) (Fig. 5d).

These troughs presumably formed by migration of large ripples. Marked variation in orientation of these ripples with associated bars suggests that this facies might have developed as sand flats outside the channels during flooding (cf. Chakraborty, 1995; Mazumder, 2002).

4.3. Heterolithic facies (facies J)

This facies is characterized by fine-grained sandstone-siltstone interbedding (Fig. 6a). The maximum facies thickness is 10 m and on regional reconstruction, its geometry appears to be lenticular (Fig. 2). The beds are sheet-like or tabular in geometry. The dark colored beds, appearing massive on exposed surfaces, are of fine-grained siltstone whereas the lighter colored beds are of fine-grained sandstone. In the dark colored beds constituent terrigenous grains generally have a ferruginous coating. Locally there are scours (average width 22 cm, depth 5 cm; Fig. 6b). Siltstone fills in these scours. The fine-grained sandstone beds also look massive, but have small-scale ripple laminae at their top. This facies has been encountered only in the western part of the study area and in the younger Member.

This facies probably formed in a restricted water body, possibly a lake. Silt particles with red cutans around them (Folk, 1971) were possibly trapped within a lake during wind deflation. The sandstone interbeds might have formed by episodic hyperpycnal underflows. Paucity of features and exposures, however, renders detailed interpretation of the origin of this facies difficult.

4.4. Shale facies

This facies is represented by grey shale with siltstone interbeds. Maximum thickness of this facies is up to 10 m. The siltstone beds have variable thickness ranging up to 2 cm and have sheet-like geometry. Internally, these beds are characterized by parallel or ripple laminations. Locally there are scours at the base of the siltstone beds (Fig. 6c). Bedding plane exposures are extremely rare and sheared limiting physical evidence of subaerial exposure, if it occurred, during sedimentation. In regional reconstructions the facies units appear lenticular, possibly because of erosional contacts with all other facies (Fig. 2). In sandstone-dominated sections this facies occurs immediately under the MESs. This facies is found distributed throughout the formation from bottom to top.

In close association with fluvial sediments, this facies is likely to be of flood plain origin (cf. Collinson, 1996; Mazumder, 2002).

4.5. Cross-stratified volcaniclastic facies

This facies is rich in volcanic clasts and generally cross-stratified. The cross-strata are mostly of planar geometry and of variable set thickness, ranging up to 36 cm. The clasts are generally sub rounded and their diameter is variable from 0.14 to 25 cm (Fig. 7a). The beds form thinning and fining-upward hemicycles with average thickness about 0.9 m (Fig. 2). The maximum bed thickness measured is about 35 cm. Cross-stratification orientation broadly matches palaeocurrent direction derived from the associated sandstone facies (Fig. 2). Vertical stacking of the hemicycles produce large facies unit thicknesses up to 65 m. This facies is however, interbanded with the volcanic rocks, and restricted to the younger member (Fig. 2).

Close correspondence between its cross-strata orientation and fluvial paleocurrent strongly favors that this facies possibly represents reworked volcanic

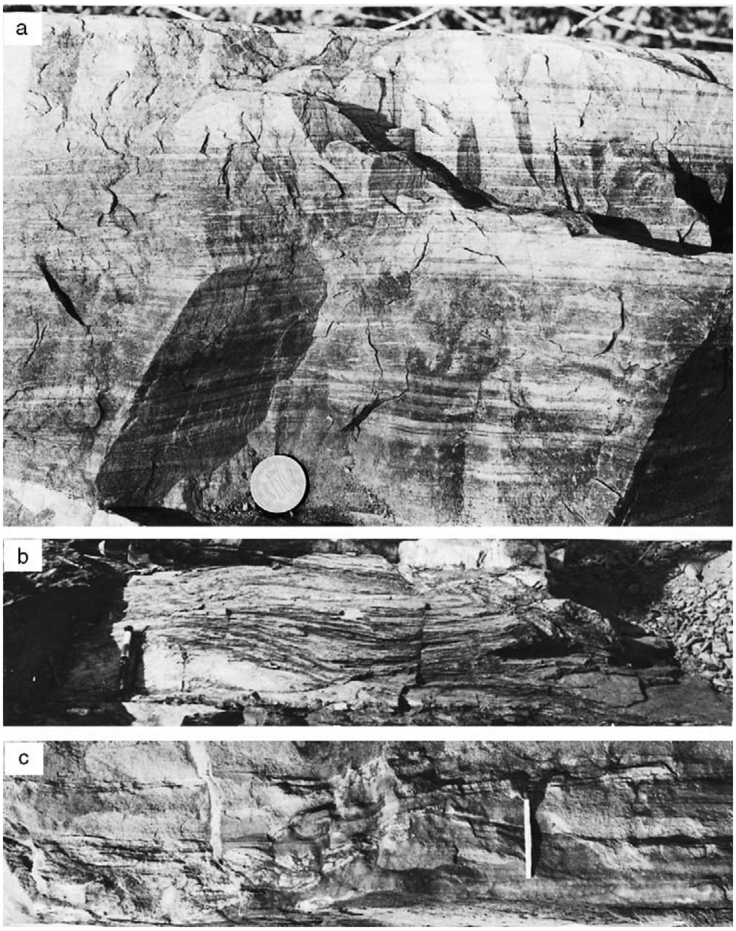

Fig. 6. (a) Dark and light color fine sandstone-siltstone interbanding within the heterolithic facies (facies J; coin diameter 1.5 cm). (b) Scours within facies J (pen length 12 cm). (c) Scours at the basal part of the siltstone beds within the Grey shale facies (match stick length 4.2 cm).

materials (Fig. 2; Mazumder, 2002). Most likely the thinning and fining upward hemicycles developed through filling of shallow channels.

4.6. Pyroclastic facies

This facies is represented by tuff and agglomerate and is confined within the younger member (Fig. 2). It generally occurs immediately above or in close lateral contact with the volcanic rocks. This facies comprises a lenticular body with maximum thickness up to 75 m and is wedge-shape in the direction of transport (Fig. 2). Maximum bed thickness is 3 m. The facies is generally massive, but occasionally shows low angle backsets, as well as, relatively high angle foresets (Fig. 7b). Clasts (bombs) are angular, irregular in shape (Fig. 7c) and poorly sorted. The lower level tuffs and agglomerates are mafic to ultramafic in composition, whereas those of the upper part are silicic (cf. Singh, 1998). Vitric tuffs (Fig. 7d) in the central part of the study area are overlain by basaltic lava flows (Fig. 2). Acid agglomerate contains flattened silicic lapillis set in a rhyolitic matrix and grades upward in to finely laminated lapilli tuff having maximum diameter around 4 mm (Singh, 1998, his Fig. 3). Thickness of individual agglomerate bands is up to 3.5 m.

Vertical (cross-sectional) distribution, internal structure and textural characteristics suggest that this facies possibly represents pyroclastic flow and surge (cf. Fisher and Schmincke, 1984, p. 107–109; Cas

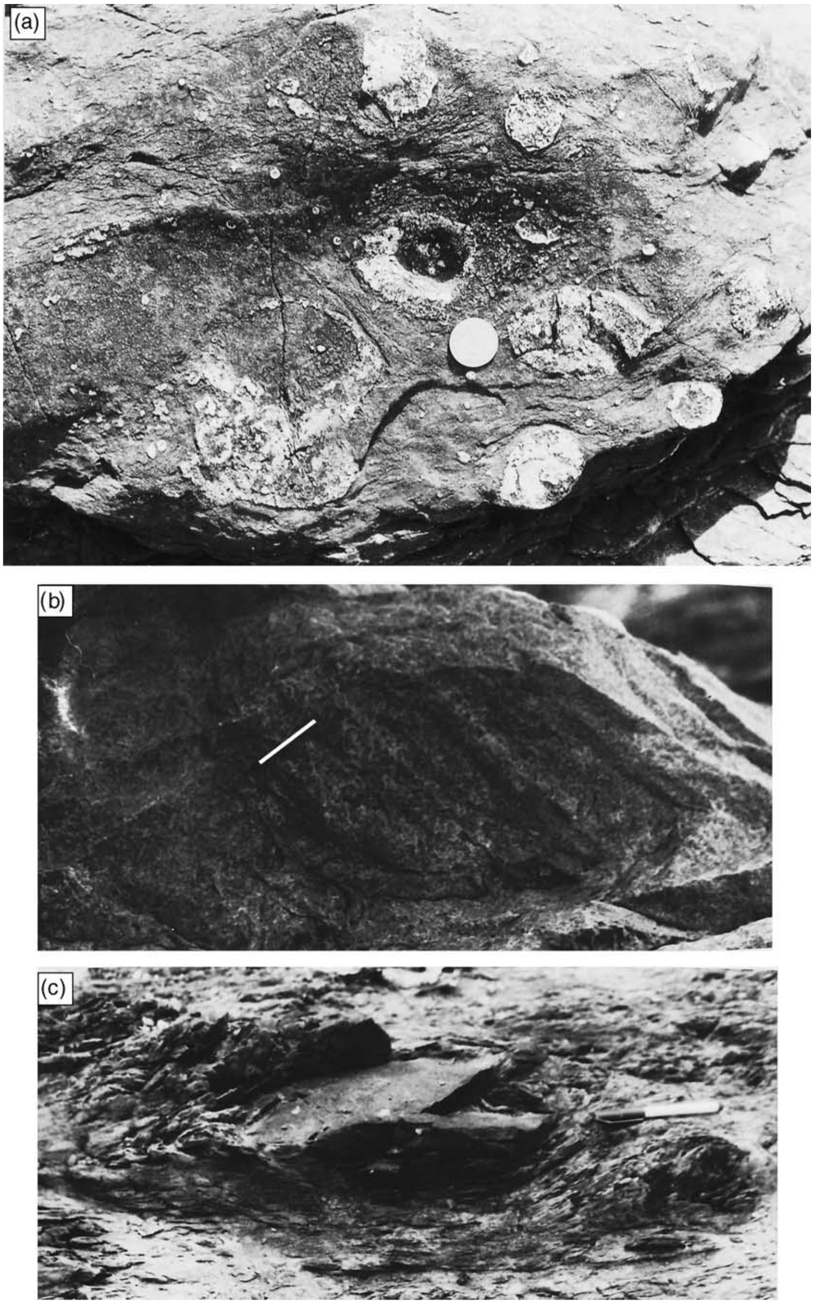

Fig. 7. (a) Rounded volcanic clasts within the volcaniclastic facies (pen length is 13 cm). (b) Generally massive but locally cross-bedded (high angle foresets) pyroclastic facies associated with the volcanic rocks (match stick length 4.2 cm). (c) Angular basaltic bomb within the pyroclastic facies (pen length 12 cm). (d) Vitric tuff containing poorly sorted angular clasts (bar length is 1 mm).

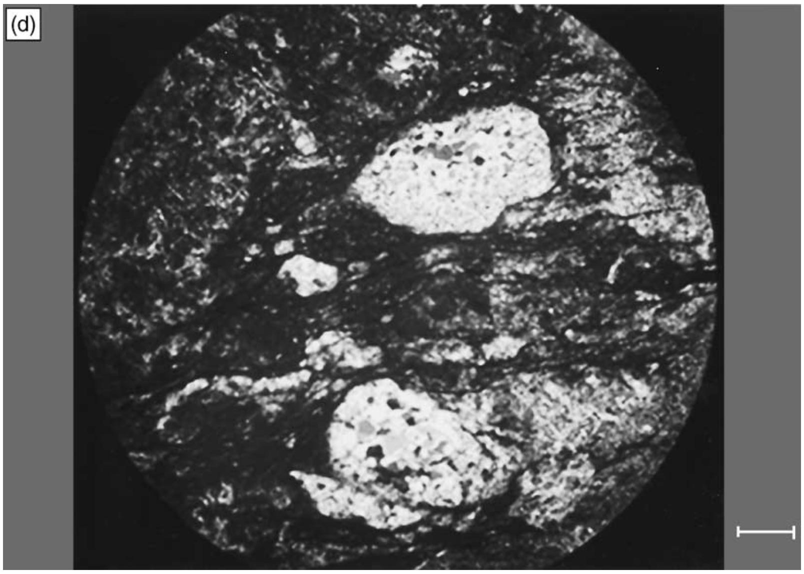

Fig. 7. (*Continued*).

and Wright, 1987; Bull and Cas, 2000). The terrestrial depositional setting of this facies is constrained by the bounding sandstone-siltstone and volcaniclastic lithofacies of fluvial origin (Fig. 2; cf. Cas and Wright, 1987, p. 270; Fritz and Howells, 1991; Mueller et al., 2000). The acid agglomerates and associated tuffs have been interpreted as air-fall deposits (Singh, 1998, p. 385).

5. Volcanic rocks

The Dhanjori volcanic rocks include ultramafic to mafic and rarely acidic (Singh, 1998, p. 385) lava flows and are confined to the Upper member (Fig. 2; Gupta et al., 1985; Roy et al., 2002; Mazumder, 2002). They are mostly massive but show excellent flow banding (Fig. 8a) and vesicular structure (Fig. 8b) in places. The volcanic rocks (including the volcaniclastics) are thickest (~1 km) in the central part of the study area (Fig. 2). Most of the ultramafic rocks (peridotitic komatiite as well as peridotites; Fig. 9a and b) are confined to the lower level of the Upper member but some are interlayered with the sandstones (Gupta et al., 1985; Roy et al., 2002; Mazumder, 2002). The MgO content of the ultramafic rocks varies from 25 to 13% (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; Majumder, 1996; Roy et al., 2002; cf. Le Bas, 2000, 2001). These rocks are largely composed of actinolite, hornblende, relict pyroxenes, olivines along with epidote and chlorite.

On the contrary, the 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 (Singh, 1998, p. 385). The amygdules are mostly composed of quartz and/or chlorite and rarely calcite (see also Alvi and Raza, 1992). The rocks contain augite and plagioclase laths. Occasionally olivine is present in the fine-grained plagioclase and augite groundmass. Unlike the ultramafic rocks, these basaltic rocks are unaltered except for the occasional chloritization of pyroxenes. Isotopic analyses of these basalts yield a Palaeoproterozoic Sm–Nd isochron age of 2072 ± 106 Ma (MSWD = 1.56; Roy et al., 2002). In addition, there are a few small bodies of coarse-grained gabbro and peridotites cutting the younger member (Gupta et al., 1985; Gupta, 1998; senior authors unpublished data).

5.1. Geochemistry

Though the Dhanjori volcanic rocks have undergone some amount of alteration during low-grade

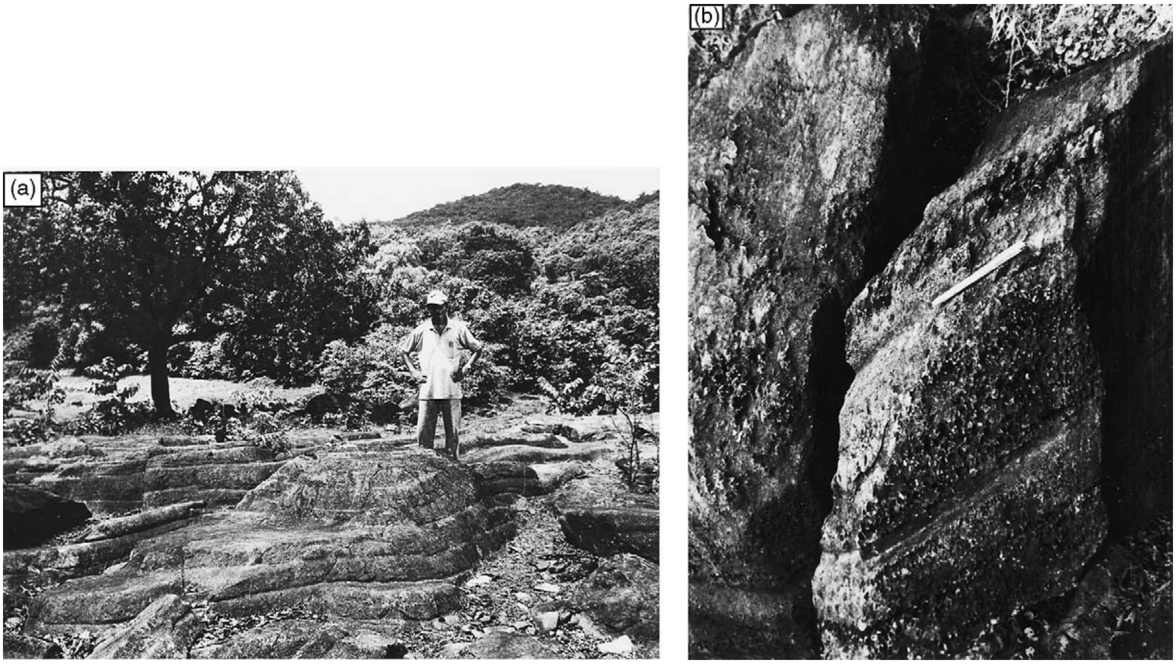

Fig. 8. (a) Flow banding within Dhanjori basalt. (b) Vesicular structure within Dhanjori basalt (pen length is 14 cm).

metamorphism, low Fe_2O_3/FeO ratios, moderate loss on ignition, and rare instances of serpentinization suggest limited chemical alteration (Gupta et al., 1985; see also Alvi and Raza, 1992; Saha, 1994; Roy et al., 2002). The MgO–CaO–Al_2O_3 plot (Fig. 9a) shows that the ultramafic rocks are chemically peridotitic komatiites (see also Gupta et al., 1985; Saha, 1994; Gupta, 1998). The basalts are low K-tholeiites (Sarkar and Deb, 1971; Sarkar et al., 1992). These ultramafic and mafic volcanic rocks are comparable with those of the Barberton and Munro Township (Fig. 9b; Viljoen and Viljoen, 1969; Arndt et al., 1977; Gupta et al., 1985; Saha, 1994). The high MgO content (>18%) of the ultramafic rocks, along with local presence of microspinifex texture, indicates that these are komatiite sensu-stricto (Viljoen and Viljoen, 1969; Brooks and Hart, 1974; Arndt et al., 1977; Gupta et al., 1985; Majumder, 1996).

Although komatiites and komatiitic basalts are characteristic of Archaean Greenstone belts (Arndt and Nesbitt, 1982; Sporule et al., 2002, and references therein), they are also reported from Proterozoic volcano-sedimentary successions (Hynes and Francis, 1982; Park et al., 1984; Puchtel et al., 1997; Kalsbeek and Manatschal, 1999; Zhou et al., 2000). The high MgO content of the ultramafics implies that they are genetically related to the Proterozoic mantle plume upwelling (Campbell and Griffiths, 1992; Campbell, 1998; Eriksson et al., 1999; Mazumder et al., 2000; Roy et al., 2002). Within the lower part of the Upper Member (Fig. 2) an olivine basalt with varying proportion of Al_2O_3 and MgO occurs (Gupta et al., 1985). Samples of this volcanic unit fall between the komatiitic and tholeiitic field in the MgO–CaO–Al_2O_3 plot (Fig. 9a and b).

6. Volcano-tectonic influences on sedimentation

The emplacement of Singhbhum granite (Saha, 1994) suggests the crust in the Singhbhum crustal province was ∼48 km thick (Mazumder et al., 2000; see also Bhattacharya and Shalivahan, 2002a,b) 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 reveals that the crust was 15–20 km thick (Mazumder et al.,

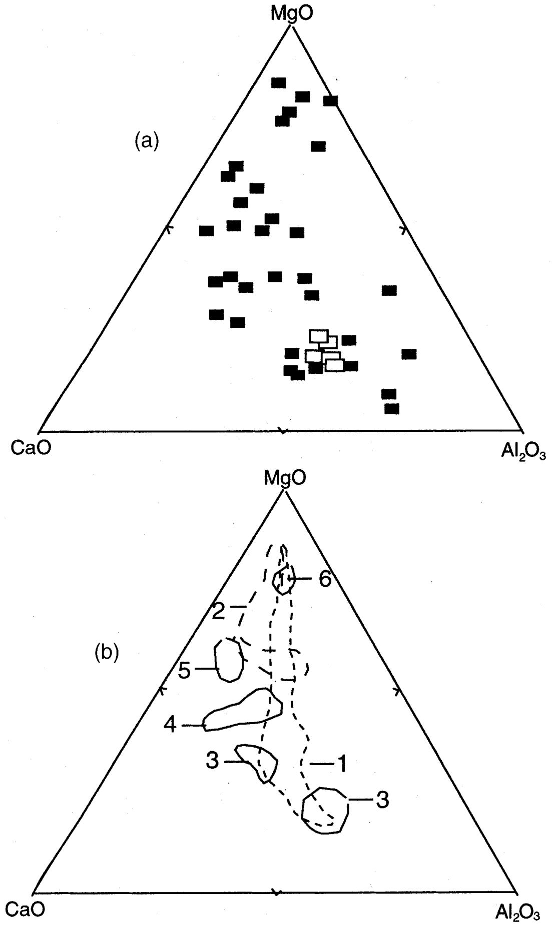

Fig. 9. (a) MgO–CaO–Al_2O_3 plot of Dhanjori volcanics. Data source: Gupta et al., 1985; Sarkar et al., 1992 (filled rectangle); Mazumder, 2002 (open rectangle). (b) The same plot depicting fields of different volcanic units worldover; 1. Munro Township (Pyke et al., 1973); 2. Barberton (Viljoen and Viljoen, 1969); 3. Upper Dhanjori basalts; 4. Dhanjori basaltic komatiite; 5. Dhanjori komatiitic peridotite-pyroxenite; 6. Dhanjori ultramafic volcaniclastics (modified after Sarkar et al., 1992).

2000, their Fig. 4b and Table 1) at ca. 2100 Ma (Roy et al., 2002). Cooling down of the vast volume of 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., 2002; Mazumder, 2002).

The basal conglomerate deposits, coarse-grained sandstone, and shale at the base of the Dhanjori Formation rest unconformably on the granitic basement and represent the distal fringe of an alluvial fan complex. Sieve, sheet flood, mass flow, fluvial channel, and overbank deposits comprise this fan segment. The proximal part of the fan lay further southward. Subsequent erosion presumably removed the fan and exhumed the granitic basement. During uplift the fan fluvial system reestablished itself by southward migration. The relatively consistent paleocurrent data from fluvial deposits in individual sections suggests the river channels incised on the fan surface were braided. Presence of sand flats on the riverbanks and mid-channel bars corroborate this contention. However, the bank-attached bars suggest a moderate degree of meandering. Such a scenario is, nonetheless, likely in vegitationless Palaeoproterozoic depositional setting. Considerable variation in fluvial paleocurrent direction between sections at, more or less, same stratigraphic interval is possibly due to fanning of the river system. The paleocurrent pattern changed drastically in the Upper Member because of change in paleoslope (cf. Willis, 2000; Yoshida, 2000; Miall and Arush, 2001) while volcanic outpourings further complicated the drainage pattern. The paleoclimate was likely semiarid to support fan growth during the deposition of coarse clastic sediments. The preservation and abundance of unstable rock fragments further corroborates this contention.

Dhanjori sedimentation apparently took place entirely in the terrestrial regime. Preservation of thick terrestrial sedimentary rocks (Fig. 2) requires a considerable rise in the base level. Aside from the effects of volcanism, the controls on deposition were largely the same in both Members constituting the Dhanjori Formation. Despite this, the significant change in bedding dips coinciding with sharp diversion in fluvial current direction between the two members indicates basin tilting (Mazumder, 2002). The change of a broadly northeasterly paleoslope towards west also coincides with the commencement of the volcanic phase (Mazumder, 2002). Although the lower member maintains a roughly uniform thickness laterally, the upper member is, in contrast, vastly thicker at the central part of the study area proximal to volcanism. Differentially subsidence occurred as a consequence of the distribution and thickness of volcanic rocks. An inferred isolated lacustrine deposit in the upper member owes its origin to damming of a river by volcanics (cf. Kuenzi et al., 1979; Mueller et al., 1994).

The Dhanjori volcanism took place in an intracontinental rift setting as is evident from the interbedded terrestrial deposits (Eriksson et al., 1999; Mazumder et al., 2000; Mazumder, 2002, 2003). Compositionally immature terrigenous clastics and bimodal volcanics corroborate this interpretation (cf. Wilson, 1993; Condie, 1997). Local pillow structure indicates quiescent subaerial flows locally entering streams, lakes, or ponds (cf. Dillard et al., 1999, p. 117). Interestingly, these volcanic and volcaniclastic rocks are confined to the upper member (Fig. 2). The lower member is entirely made up of clastic sedimentary rocks. This implies that the initiation of rifting was not a consequence of convective upwelling in the mantle (cf. Sengor and Burke, 1978; Sengor, 1995; Condie, 1997; Ziegler and Cloetingh, 2004). The presence of Dhanjori basalts including komatiites and komatiitic basalts within the upper member implies subsequent upwelling of the mantle plume and decompressional melting following crustal extension (Eriksson et al., 1999; Mazumder et al., 2000; Roy et al., 2002; cf. Miall, 2002). Interbedded volcanics and volcaniclastics in different stratigraphic level within the upper member, however, indicate episodic volcanic eruption during sedimentation. Thus, the Dhanjori volcano-sedimentary succession displays evidence of having passed through passive, as well as, active phases of continental rifting and increasingly important role of the volcanism on sedimentation through time (cf. Sengor, 1995; Leeder, 1995; Miall, 2002; Ziegler and Cloetingh, 2003).

Acknowledgements

Professors Pradip K. Bose, Wulf Mueller, Bryan Krapez, Jeff Chiarenzelli, Pat Eriksson and Kenneth Eriksson made numerous comments and suggestions that were most helpful during revision of this manuscript. Abhik Kundu provided assistance in the field. Financial support for this research came from the University Grants Commission, Government of India through the research grant PSW/00-01/045 [ERO] and Japan Society for the Promotion of Sciences (JSPS) through a Post-Doctoral Fellowship to the senior author. Prof. Makoto Arima inspired the senior author. The Geological Institute, Yokohama National University and the Department of Geology, Asutosh College provided necessary infrastructural facilities. This paper is dedicated to Late Prof. Sukomol K. Chanda.

References

Altermann, W., 1996. Sedimentology, geochemistry and palaeogeographic implications of volcanic rocks in the Upper Archaean Campbell Group, western Kaapval craton South Africa. Precambrian Res. 79, 73–100.

Alvi, S.H., Raza, M., 1992. Geochemical evidence for the volcanic arc setting of the Dhanjori volcanics, Singhbhum craton, eastern India. Geol. Mag. 129, 337–348.

Arndt, N.T., Naldrett, A.J., Pyke, D.R., 1977. Komatiitic and iron rich tholeiitic lavas from Munro Township, northeast Ontario. J. Petrol. 18, 310–369.

Arndt, N.T., Nesbitt, R.W., 1982. Geochemistry of Munro township basalts. In: Arndt, N.T., Nisbet, E. (Eds.), Komatiites. George Allen and Unwin, London, pp. 309–329.

Banks, N.L., 1973. The origin and significance of some downcurrent-dipping cross-stratified sets. J. Sediment. Petrol. 43, 423–427.

Basu, A., 1985. Structure and stratigraphy in and around S.E.—part of Dhanjori Basin, Singhbhum, Bihar. Rec. Geol. Surv. India 113, 59–67.

Bhattacharya B.B., Shalivahan, 2002a. The electric Moho underneath Eastern Indian craton. Geophys. Res. Lett. 29 (10), 14/1–14/4.

Bhattacharya B.B., Shalivahan, 2002b. Implications of novel results about Moho from magnetotelluric studies. Curr. Sci. 83 (10), 1259–1264.

Blair, T.C., 1999a. Sedimentary processes and facies of the waterlaid Anvil Spring canyon alluvial fan, Death valley, California. Sedimentology 46, 913–940.

Blair, T.C., 1999b. Cause of dominance by sheetflood versus debris-flow processes on two adjoining alluvial fans, Death Valley, California. Sedimentology 46, 1015–1028.

Blair, T.C., 1999c. Sedimentology of debris flow dominated warm-Spring Canyon alluvial fan, Death Valley, California. Sedimentology 46, 941–965.

Blair, T.C., McPherson, J.G., 1994. Alluvial fans and their natural distinction from rivers based on morphology, hydraulic processes, sedimentary processes, and facies assemblages. J. Sediment Res. A64, 450–489.

Blodgett, R.H., Stanley, K.O., 1980. Stratification, bedforms and discharge relations of the Platte Braided river system, Nebraska. J. Sediment Petrol. 60, 139–148.

Bose, M.K., 1994. Sedimentation pattern and tectonic evolution of the Proterozoic Singhbhum Basin in the eastern Indian shield. Tectonophysics 231, 325–346.

Bose, P.K., Chakraborty, P.P., 1994. Marine to Fluvial transition: Proterozoic Upper Rewa Sandstone, Maihar, India. Sediment. Geol. 89, 285–302.

Bose, P.K., Mazumder, R., Sarkar, S., 1997. Tidal sandwaves and related storm deposits in the transgressive Protoproterozoic Chaibasa Formation, India. Precambrian Res. 84, 63–81.

Brooks, C., Hart, S.R., 1974. On the significance of komatiite. Geology 2, 107–110.

Bull, S.W., Cas, R.A.F., 2000. Distinguishing base-surge deposits and volcaniclastic fluviatile sediments: an ancient example from the Lower Devonian Snowy River volcanics. Sedimentology 47, 87–98.

Campbell, I.H., 1998. The Mantle's chemical structure: insights from the melting products of Mantle Plume. In: Jackson, I. (Ed.), The Earth's Mantle. Cambridge University Press, pp. 259–310.

Campbell, I.H., Griffiths, R.W., 1992. The changing nature of mantle hotspots through time: implications for the chemical evolution of the mantle. J. Geol. 92, 497–523.

Cant, D.J., Walker, R.G., 1976. Development of a braided-fluvial facies model for the Devonian Battery Point Sandstone, Quebec. Can. J. Earth Sci. 13, 102–119.

Cas, R.A.F., Wright, J.V., 1987. Volcanic Successions. Allen and Unwin, London, 528 p.

Chakraborty, P.P., 1995. Facies and sequence development in some Late Proterozoic Formations in Son-Valley, India with some clues for basin evolution. Unpublished Ph.D. thesis, Jadavpur University, Kolkata, 104 p.

Collinson, J.D., 1996. Alluvial sediments. In: Reading, H.G. (Ed.), Sedimentary Environments: Process, Facies and Stratigraphy, 3rd ed. Blackwell Science, Oxford, pp. 37–82.

Collinson, J.D., Thompson, D.B., 1989. Sedimentary Structures. Allen Unwin, London, 194 p.

Condie, K.C., 1973. Archaean Magmatism and crustal thickening. Geol. Soc. Am. Bull. 84, 2981–2992.

Condie, K.C., 1997. Plate Tectonics and Crustal Evolution, 4th ed. Butterworth-Heinmann, Oxford, 282 p.

Condie, K.C., Potts, M.J., 1969. Calc-alkaline volcanism and the thickness of the early Precambrian crust in North America. Can. J. Earth Sci. 6, 1179–1184.

Corcoran, P.L., Mueller, W.U., Padgham, W.A., 1999. Influence of tectonism and climate on lithofacies distribution and sandstone and conglomerate composition in the Archaean Beaulieu Rapids Formation, Northwest Territories, Canada. Precambrian Res. 94, 175–204.

Dasgupta, P., 2003. Sediment gravity flow—the conceptual problems. Earth Sci. Rev. 62, 265–281.

Dillard, K.A., Simpson, E.L., Noto, R.C., Wizevich, M., 1999. Characterization of fluvial deposits interbedded with flood basalts, Neoproterozoic Catcotin Formation, central Applachians, USA. Precambrian Res. 97, 115–134.

Dunn, J.A., Dey, A.K., 1942. The geology and petrology of eastern Singhbhum and surrounding areas. Mem. Geol. Surv. India 69 (2), 281–456.

Einsele, G., 1992. Sedimentary Basins: Evolution, Facies, and Sediment Budget. Springer-Verlag, Berlin, 628 p.

Eriksson, P.G., Condie, K.C., Trisgaard, H., Muller, W., Altermann, W., Catunean, O., Chiarenzali, J., 1998. Precambrian Clastic sedimentation systems. Sediment. Geol. 120, 5–53.

Eriksson, P.G., Mazumder, R., Sarkar, S., Bose, P.K., Altermann, W., Van Der Merwee, R., 1999. The 2.7–2.0 Ga volcano-sedimentary record of Africa, India and Australia: evidence for global and local changes in sea level and continental freeboard. Precambrian Res. 97, 269–302.

Fisher, R.V., Schminke, H.V., 1984. Pyroclastic Rocks. Springer-Verlag, Berlin.

Folk, R.L., 1971. Longitudinal dunes of the northwestern edge of Simpson desert, Northern Territory, Australia: geomorphology and grainsize relationships. Sedimentology 16, 5–54.

Fritz, W.J., Howells, W., 1991. A shallow marine volcaniclastic facies model: an example from sedimentary rocks bounding the subaqueously welded Ordovician Garth Tuff, North Wales, UK. Sediment. Geol. 74, 217–240.

Ghosh, S.K., Sengupta, S., 1987. Progressive development of structures in a ductile shear zone. J. Struct. Geol. 9, 277–287.

Gupta, A., Basu, A., Ghosh, P.K., 1980. The Proterozoic ultramafic and Mafic Lavas and tuffs of the Dalma greenstone belt, Singhbhum, eastern India. Can. J. Earth Sci. 17, 210–231.

Gupta, A., Basu, A., Singh, S.K., 1985. Stratigraphy and petrochemistry of Dhanjori Greenstone belt, Eastern India. Quart. J. Geol. Min. Met. Soc. India 57, 248–263.

Gupta, A.K., 1998. Igneous Rocks. Allied Publishers Ltd., New Delhi, 690 p.

Hogg, S.E., 1982. Sheetfloods, sheetwash, sheetflow, or? Earth Sci. Rev. 18, 59–76.

Hynes, A., Francis, D., 1982. Komatiitic basalts of the Cape Smith Fold Belt, New Quebec, Canada. In: Arndt, N.T., Nisbet, E.G. (Eds.), Komatiites. Allen and Unwin, London, pp. 159–170.

Iyenger, S.V.P., Alwar, M.A., 1965. The Dhanjori eugesyncline and its bearing on stratigraphy of Singhbhum, Keonjhar and Mayurbhanj districts. D.N. Wadia Commem. vol. Min. Geol. Met. Inst. India 138–162.

Johnson, A.M., 1970. Physical Processes in Geology. Freeman, San Francisco, CA.

Kalsbeek, F., Manatschal, G., 1999. Geochemistry and tectonic significance of peridotitic and metakomatiitic rocks from the Ussuit area, Nagssugtoqidian orogen, West Greenland. Precambrian Res. 94, 101–120.

Kataoka, K., Nakajo, T., 2002. Volcaniclastic resedimentation in distal fluvial basins induced by large-volume explosive volcanism: the Ebisutoge-Fukuda tephra, Plio-Pleistocene boundary, central Japan. Sedimentology 49, 319–334.

Kelley, S.B., Olsen, H., 1993. Terminal fans—a review with special reference to Devonian examples. Sediment. Geol. 88, 339–374.

Kirk, M., 1983. Bar development in a fluvial sandstone (west phalian "A"), Scotland. Sedimentology 30, 727–742.

Krishna Rao, N., Agarwal, S.K., Rao, G.V.U., 1979. Lead isotopic ratio of uraninites and the age of uranium mineralisation in Singhbhum Shear Zone, Bihar. J. Geol. Soc. India 20, 124–127.

Kuenzi, W.D., Horst, O.H., McGehee, R.V., 1979. Effect of volcanic activity on fluvial-deltaic sedimentation in a modern arc-trench gap, southwestern Guatemala. Geol. Soc. Am. Bull. 90, 827–838.

Le Bas, M.J., 2000. IUGS Reclassification of the high-Mg and Picritic Volcanic Rocks. J. Petrol. 41, 1467–1470.

Le Bas, M.J., 2001. Reply to Comment by Kerr and Arndt. J. Petrol. 42, 2173–2174.

Leeder, M.R., 1995. Continental rifts and proto-oceanic rift troughs. In: Busby, C., Ingersoll, R. (Eds.), Tectonics of Sedimentary Basins. Blackwell Science, Boston, pp. 119–148.

Long, D.G.F., 1978. Proterozoic stream deposits: some problems of recognition and interpretation of ancient fluvial systems. In:

Miall, A.D. (Ed.), Fluvial Sedimentology. Can. Soc. Pet. Geol. Mem. 5, pp. 313–341.

Long, D.G.F., 2002. Aspects of Late Palaeoproterozoic fluvial style: the Uairen Formation, Roraima Supergroup, Venezuela International Assoc. Sedimentologists Spec. Publ. 33, 323–338.

Majumder, S.K., 1996. Precambrian geology of Peninsular India. Indian Min. 50, 139–174.

Martins-Neto, M.A., 2000. Tectonics and sedimentation in a paleo/mesoproterozoic rift-sag basin (Espinhaco basin, southeastern Brazil). Precambrian Res. 103, 147–173.

Mazumder, R., 2002. Sedimentation history of the Dhanjori and Chaibasa formations, eastern India and its implications. Unpublished Ph.D. thesis, Jadavpur University, 119 p.

Mazumder, R., 2003. Correlations between the Eastern Block of the North China Craton and the South Indian Block of the Indian Shield: an Archaean to Palaeoproterozoic link—comment. Precambrian Res. 127, 379–380.

Mazumder, R., Bose, P.K., Sarkar, S., 2000. A commentary on the tectono-sedimentary record of the pre-2.0 Ga continental growth of India vis a vis a pre-Gondwana Afro-Indian supercontinent. J. African Earth Sci. 30, 201–217.

Miall, A.D., 1988. Architectural elements and bounding surfaces in fluvial deposits: anatomy of the Kaneyta Formation (Lower Jurassic), southwest Colorado. Sediment. Geol. 55, 233–262.

Miall, A.D., 1996. The Geology of Fluvial Deposits: Sedimentary Facies, Basin Analysis and Petroleum Geology. Springer, Heidelberg.

Miall, A.D., 2002. An introduction to rift basins and their sediments. Sediment. Geol. 147, 3–8.

Miall, A.D., Arush, M., 2001. Cryptic sequence boundaries in braided fluvial successions. Sedimentology 48, 971–985.

Mukhopadhaya, D., 1976. Precambrian stratigraphy of Singhbhum—the problems and prospect. Ind. J. Earth Sci. 3, 208–219.

Mukhopadhaya, D., 1984. The Singhbhum Shear Zone and it's place in the Evolution of the Precambrian mobile belt of North Singhbhum. In: Proceedings of the Seminar on Crustal Evolution of the Indian Shield and its Bearing on Metallogeny. Indian J. Earth Sci., Seminar volume, pp. 205–212.

Mukhopadhaya, D., 1988. Precambrian of Eastern Indian Shield—perspective of the problems. Mem. Geol. Soc. India 8, 1–12.

Mueller, W.U., Corcoran, P.L., 2001. Volcano-sedimentary processes operating on a marginal continental arc: the Archean Raquette Lake Formation, Slave Province, Canada. Sedimentary Geol. 141/142, 169–204.

Mueller, W., Donaldson, J.A., Doucet, P., 1994. Volcanic and tectono-plutonic influences on sedimentation in the Archaean Kirkland Basin, Abitibi greenstone belt, Canada. Precambrian Res. 68, 201–230.

Mueller, W.U., Garde, A.A., Stendal, H., 2000. Shallow-water, eruption-fed, mafic pyroclastic deposits along a Paleoproterozoic coastline: Kangerluluk volcano-sedimentary sequence, southeast Greenland. Precambrian Res. 101, 163–192.

Mueller, W.U., Mortensen, J.K., 2002. Age constraints and characteristics of subaqueous volcanic construction, the Archean Hunter Mine Group, Abitibi greenstone belt. Precambrian Res. 115, 119–152.

Nemec, W., Porebski, S.J., Steel, R.J., 1980. Texture and structure of resedimented conglomerates: examples from Ksiaz Formation (Famennian-Tournaisian), southwestern Poland. Sedimentology 27, 519–538.

Nemec, W., Postma, G., 1993. Quarternary alluvial fans in southwestern Crete: sedimentation processes and geomorphic evolution. In: Marzo, M., Puigdefabreagas, C. (Eds.), Alluvial Sedimentation. Int. Assoc. Sediment. Special Publ. 17, pp. 235–276.

Park, A.F., Bowes, D.R., Halden, N.M., Koistinen, T.J., 1984. Tectonic evolution at an Early Proterozoic continental margin: the Svecokarelides of eastern Finland. J. Geodynamics 1, 359–386.

Pettijohn, F.J., 1975. Sedimentary Rocks, 3rd ed. Harper International, 628 p.

Puchtel, I.S., Haase, K.M., Hofmann, A.W., Chauvel, C., Kulikov, V.S., Garbe-Schonberg, C.D., Newchin, A.A., 1997. Petrology and geochemistry of crustally contaminated komatiitic basalts from the Vetreny Belt, Southeastern Baltic Shields: evidence for an early Proterozoic mantle plume beneath rifting Archaean continental lithosphere. Geochim. Cosmochim. Acta 61, 1205–1222.

Pyke, D.R., Naldrett, A.J., Eckstrand, O.R., 1973. Archaean ultramafic flows in Munro Township, Ontario. Geol. Soc. Am. Bull. 84, 955–978.

Roy, A., Sarkar, A., Jeyakumar, S., Aggrawal, S.K., Ebihara, M., 2002. Sm–Nd age and mantle characteristics of the Dhanjori volcanic rocks, Eastern India. Geochem. J. 36, 503–518.

Rust, B.R., Jones, B.G., 1978. The Hawkesbury sandstone south of Sydney Australia: Triassic analogue for the deposit of a large braided river. J. Sediment. Petrol. 57, 222–233.

Saha, A.K., 1994. Crustal evolution of Singhbhum-North Orissa, Eastern India. Geol. Soc. India Mem. 27, 341.

Sarkar, S.C., 1984. Geology and ore mineralisation of the Singhbhum copper–uranium belt, Eastern India. Jadavpur University, Calcutta, 263 p.

Sarkar, S.C., Deb, M., 1971. Dhanjori basalts and some related rocks. Quart. J. Geol. Min. Met. Soc. India 43, 29–37.

Sarkar, S.C., Gupta, A., Basu, A., 1992. North Singhbhum Proterozoic Mobile Belt, Eastern India: its character, evolution and metallogeny. In: Sarkar, S.C. (Ed.), Metallogeny Related to Tectonics of the Proterozoic Mobile Belts. Oxford & IBH, New Delhi, pp. 271–305.

Sarkar, S.N., Ghosh, D.K., Lambert, St., R.J., 1986. Rubidium–strontium and lead isotopic studies of the Soda granites from Musaboni area, Singhbhum Copper Belt. In: Sarkar, S.N. (Ed.), Geology and Geochemistry of Sulphide Ore Bodies and Associated Rocks in Musaboni and Rakha Mines Sections in the Singhbhum Copper Belt. Diamond Jubilee Monograph, Indian School of Mines (Dhanbad), pp. 101–110.

Sarkar, S.N., Saha, A.K., 1962. A revision of the Precambrian stratigraphy and tectonics of Singhbhum and adjacent region. Quart. J. Geol. Min. Metal. Soc. India 34, 97–136.

Sarkar, S.N., Saha, A.K., 1977. The present status of the Precambrian stratigraphy, tectonics, and geochronology of Singhbhum-Keonjhar-Mayurbhanj region, eastern India. Indian J. Earth Sci. (S. Ray vol.), 37–65.

Sarkar, S.N., Saha, A.K., 1983. Structure and tectonics of the Singhbhum– Orissa Iron Ore Craton, eastern India. In: Saha, A.K. (Ed.), Recent Researches in Geology: Structure and Tectonics of Precambrian Rocks. Hindustan Publ. Corp., New Delhi, pp. 1–25.

Sato, H., Amano, K., 1991. Relationships between tectonics, volcanism, sedimentation and basin development, Late Cenozoic, central part of Northern Honsu, Japan. Sediment. Geol. 74, 323–343.

Sengor, A.M.C., 1995. Sedimentation and tectonics of fossil rifts. In: Busby, C.J., Ingersoll, R.V. (Eds.), Tectonics of Sedimentary Basins. Blackwell, Oxford, pp. 53–117.

Sengor, A.M.C., Burke, K., 1978. Relative timing of rifting and volcanism on earth and its tectonic implications. Geophys. Res. Lett. 5, 419–421.

Sengupta, S., Mukhopadhyay, P.K., 2000. Sequence of Precambrian events in the eastern Indian craton. In: Proceedings of the International Symposium, Precambrian Crust in Eastern and Central India, IGCP 368, Bhubaneswar, pp. 49–56.

Singh, S.P., 1998. Precambrian Stratigraphy of Bihar. In: Paliwal, B.S. (Ed.), The Indian Precambrian. Scientific Publ., Jodhpur, pp. 376–408.

Singh, S.P., Nim, S.P., 1998. Stratigraphy of the Lower Proterozoic Sequence of Patharchakri-Mosaboni Area, Singhbhum, Bihar. Indian Min. 52, 1–14.

Smith, G.A., 1987. The influence of explosive volcanism on fluvial Sedimentation: the Deshutes Formation (Neogene) in central Oregon. J. Sediment. Petrol. 57, 613–629.

Smith, G.A., 1988. Sedimentology of proximal to distal volcaniclastics dispersed across an active foldbelt: Ellensburg Formation (Late Miocene), central Washington. Sedimentology 35, 953–977.

Smith, G.A., 1991. Facies sequences and geometries in continental volcaniclastic sediments. In: Fisher, R.V., Smith, G.A. (Eds.), Sedimentation in Volcanic Settings. SEPM Spec. Publ. 45, 109–121.

Smith, R.D., Swanson, F.J., 1987. Sediment routing in a small drainage Basin in the blast zone at Mount St. Helens, Washington, USA. Geomorphology 1, 1–13.

Smith, N.D., 1970. The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks: North-central Appalachian. Geol. Soc. Am. Bull. 81, 2993–3041.

Sohn, Y.K., Rhee, C.W., Kim, B.C., 1999. Debris flow and hyperconcentrated flood-flow deposits in an alluvial fan, northwestern part of Cretaceous Yongdong Basin, central Korea. J. Geol. 107, 111–132.

Sporule, R.A., Lesher, C.M., Ayer, J.A., Thurston, P.C., Herzberg, C.T., 2002. Spatial and temporal variations in the geochemistry of komatiites and komatiitic basalts in the Abitibi greenstone belt. Precambrian Res. 115, 153–186.

Viljoen, T.A., Viljoen, R.P., 1969. Evidence for the existence of a mobile extrusive peridotitic magma from Komati Formation of the Onverwacht group: Upper mantle Project. Geol. Soc. S. Africa Sp. Publ. 2, 87–112.

Van de Kemp, P.C., 1973. Holocene continental sedimentation in the Salton basin, California: a reconnaissance. Geol. Soc. Am. Bull. 84, 827–848.

Walker, R.G., 1978. Facies models. Geo. Sci. Canada Rep. Ser. 1, 171–188.

Willis, A.J., 2000. Tectonic control of nested sequence architecture in the Sego Sandstone, Neslen Formation and Upper Castlegate Sandstone (Upper Cretaceous), Sevier Foreland Basin, Uttah, USA. Sediment. Geol. 136, 277–317.

Wilson, M., 1993. Magmatism and the geodynamics of basin formation. Sediment. Geol. 86, 5–29.

Yoshida, S., 2000. Sequence stratigraphy and facies architecture of the upper Blackhawk Formation and the Lower Castlegate Sandstone (Upper Cretaceous), Book Cliffs, Uttah, USA. Sediment. Geol. 136, 239–276.

Zhou, M., Zhao, T., Malpas, J., Sun, M., 2000. Crustal-contaminated komatiitic basalts in Southern China: products of a Proterozoic mantle plume beneath the Yangtze Block. Precambrian Res. 103, 175–189.

Ziegler, P.A., Cloetingh, S., 2003. Dynamic processes controlling evolution of rifted basins. Earth Sci. Rev., in press.