Sedimentation patterns during the Precambrian: a unique record?

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Abstract

Although the similarities between depositional processes and products as well as the analogous controls on basin-filling and evolution appear to have enjoyed great uniformity throughout the sedimentary rock record, a noticeable distinction exists in the rates and intensities of a broad range of geological processes in the Precambrian epoch. This paper searches for distinctiveness in the Precambrian sedimentary record, both siliciclastic and carbonate, through an extensive, though not exhaustive, review of the relevant literature augmented by new observations. While differences in Precambrian deltaic, aeolian, glacial and possibly also lacustrine deposits and settings appear to have been small, their large-scale development was controlled largely by a combination of temporal and geodynamic influences, essentially of global compass. In this regard the onset of the supercontinent cycle and major perturbations in palaeo-atmospheric composition appear to have been significant. Marine environments provide a poor platform for Precambrian - Phanerozoic comparisons of sedimentation patterns, as those from the former period are preserved almost exclusively in epeiric settings, an environment essentially lacking on modern Earth. For the shallow marine carbonates, biological mediation of chemical sediment deposition changed radically from dominance by microbial biota in the Precambrian to a combination of metazons, protozoans and algae for the skeletal carbonates of the Phanerozoic. Despite it being widely recognized that Precambrian channel systems were braided in all environments (deltaic, tidal, alluvial, fluvial) as a consequence of the lack of vegetation and poor development of soils, the fluvial setting has some enigmatic aspects. Amongst these is evidence for ponding of muddy detritus in apparently sandstone bed-load dominated braided systems, with effects on local palaeoslopes which have resulted in unusual palaeohydraulic parameters for Precambrian fluvial systems. This is perhaps a field of research which holds greater promise when investigating sedimentation patterns prior to the Phanerozoic.

Keywords: Precambrian Eon; Sedimentation patterns; Process-product distinction; Siliciclastic-carbonate; Major environments; Features with Precambrian bias

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1. Introduction

Many readers might question the title of this paper: why do Precambrian sedimentation patterns even merit special attention at all? Examination of most widely accepted treatises on sedimentary geology shows little attention devoted to Precambrian sedimentation *per se*, and a preponderance of papers published in the leading journals deals with the Phanerozoic sedimentary record. Some of the reasons for this are merely historical, in that in the Northern Hemisphere (specifically in Europe and North America that until recently dominated scientific publishing in the field), Phanerozoic rocks are much more widespread, and easily accessible; the global hydrocarbon industry has long dominated employment of those concerned with the sedimentary record on Earth, and this commodity is largely obtained from Phanerozoic basins. It is entirely logical that most studies of aspects like sedimentary processes, depositional models, the facies themselves, and more recently, the concepts inherent in sequence stratigraphy (e.g., Catuneanu, 2006) have been based on Phanerozoic exposures – apart from mere availability, this record is better preserved and its generally good fossil record allows much greater detail in facies definition, and biostratigraphic subdivision and correlation.

However, seen from a purely chronological perspective, the Phanerozoic record would appear to be a poor candidate to base sedimentary studies upon: the Precambrian record comprises about 85% of Earth history, and as an added bonus accounts for approximately 75% of global mineral resources (e.g., large quantities of Au, Pt, Cr; most FE and Mn deposits; even some hydrocarbons, coal and graphite in the Neoproterozoic; Altermann and Corcoran, 2002b). The oldest indirect evidence of sedimentation (i.e., evidence for the presence of liquid water on Earth; e.g., Wilde et al., 2001; Bibikova, 2010) dates back to about 350 my after the formation of the planet (e.g., Eriksson et al., 2005); the oldest known supracrustal deposits, within the Isua greenstone belt of West Greenland, >3.8 Ga, include chert, banded iron formation (BIF), and lesser conglomerates and sandstones (e.g., Fedo et al., 2001). It may thus be assumed that a broad spectrum of sedimentation accompanied formation of the earliest continental crust and has continued ever since. It is obeisance to the principle of modern uniformitarianism (cf., simplified often to "the present is the key to the past"; e.g., Hallam, 1990) that has confirmed the belief by many that the Phanerozoic sedimentary record is fully adequate to understand the processes and products of global sedimentation throughout Earth history (see discussion in Donaldson et al., 2002).

Certainly, amongst the most important fundamental principles of sedimentation patterns over time must be the observation that Precambrian sedimentary lithologies and the sedimentary structures they bear, as well as their interpreted genetic processes and settings, almost all have modern equivalents (e.g., Eriksson et al., 1998, 2004b; Altermann and Corcoran, 2002a). Indeed, the similarities between the Precambrian and the Phanerozoic-Modern sedimentary records far outweigh their differences (e.g., Eriksson et al., 2005), and the range of basin types and their fills in both time periods show no meaningful deviations from a common thread. Nor do the inferred origins of basins reflect anything other than a common interaction of first-order plate tectonic and mantle thermal processes with second-order modification by eustasy and palaeoclimate (e.g., Eriksson et al., 2001a, 2004a; Bose et al., 2001). The second fundamental principle of sedimentation patterns over time is possibly of even greater importance: the main difference between Precambrian and younger sedimentary settings lay in the variability of the rates and intensities of processes controlling weathering, erosion, transport, deposition, lithification and diagenesis in the Precambrian (e.g., Donaldson et al., 2002; Eriksson et al., 2005). This principle can apply equally to fundamental controls on uplift, sediment supply and creation of accommodation, through variable rates of crustal growth, plate movement and heat flux from the mantle (e.g., Eriksson et al., 2004a, and references therein), but this is a topic beyond the focus of the current paper.

What is most noticeably different for the Precambrian time period is the lack of vegetation and the limited development of soils, which directly affected erosion and sediment supply rates; the lack of bioturbation, allowing better preservation of fine sedimentary detail (yet balanced against the greater likelihood of deformation and metamorphism); the greater penetration depth of light through seawater on marine shelves due to a lack of plankton, allowing carbonates to develop at greater photic depths; the pervasive growth of microbial mats within shallow water and even subaqueous terrestrial environments, affecting both clastic and chemical sediment deposition and even architecture (e.g., Donaldson et al., 2002; Eriksson et al., 1998, 2000, 2004a, 2005; Sarkar et al., 2005; Catuneanu, 2006; Schieber et al., 2007; and references therein). Precambrian river deposits were almost exclusively braided (e.g., Schumm, 1968; Cotter, 1978; Long, 1978; see, however, Jackson, 1978), although the influence of muddy sediment within active channel systems rather than forming the floodplains bounding channels might have been more important, with evidence for strongly episodic flow patterns and enhanced gradients being derived from detailed palaeohydrological studies of temporarily mud-carrying braided systems (e.g., Eriksson et al., 2008a, 2009a).

Aeolian ergs are absent prior to c. 1.8 Ga, although evidence for localized wind deposition is much older (c. 3.0-2.8 Ga); formation of large deserts might be related to the antiquity of the supercontinent cycle (Eriksson and Simpson, 1998; Simpson et al., 2004a). More uniform shelf circulation and storm patterns, a lack of preserved barrier island deposits and a predominance of preserved epeiric marine basins rather than open ocean basins typify the Precambrian record (e.g., Eriksson et al., 1998, 2004b and references therein), all of course predicated on the non-preservation of ocean crust older than Jurassic, implying that Precambrian marine deposits are mostly limited to preserved epeiric basins. Iron formations (IF) comprise a truly uniquely Precambrian lithology, whose origin remains the subject of much debate (e.g., Trendall, 2002); their origin has been related by many to evolution of palaeo-atmospheric compositions over time, although they may merely reflect basin formation at the approximate depths of the pycnocline on a planet already colonized by significant cyanobacterial colonies producing shallow oxygenated waters (e.g., Trendall and Blockley, 2004). Their character and possible origins would require a separate paper and they will thus not be considered in any detail in this contribution, which restricts itself essentially to siliciclastic and carbonate sedimentation systems.

The ongoing debate on atmospheric and hydrospheric chemical evolution during the Precambrian rests directly on study of the concomitant sedimentary record, with change from an original reducing atmosphere to a more oxidizing composition inherent in most views, although not all (e.g., Ohmoto, 2004, for a review). It is to address all of these aspects briefly that the present paper is presented, as a necessary background to better appreciate the complexities related to global correlation of accommodation curves (derived from the sedimentary record) across Precambrian cratons, which is the theme of this special issue. The Precambrian sedimentary record is addressed through separate sections for siliciclastic and carbonate deposits. All the major depositional environments are discussed under the clastic heading, in exclusion of only the glacial environment which is included at the end of the carbonate section, due to the inferred close relations between palaeo-atmospheric and palaeo-hydrospheric evolution, carbonate deposition (and specifically C isotopes found in "cap carbonates" associated with glacigenic deposits) and global-scale glaciation. Since distinction of the Precambrian sedimentation record is the primary issue, the paper also brings focus to some sedimentary features that have a clear Precambrian bias in occurrence.

2. Siliciclastic depositional systems

2.1. Introduction

As Precambrian crustal evolution progressed, potential source terrains for clastic sediments would have changed over time, as did the palaeo-atmospheric and –hydrospheric chemical environments. Together, these highly complex and interacting processes would have largely determined the composition of detrital sediments within the pre-Phanerozoic rock record, with additional controls being provided by sediment sorting and energy levels inherent in the transport and depositional processes themselves. It is inferred by a majority of scientists that the early Precambrian (cf. Archaean) was characterised by conditions best described as "weathering-aggressive" (Corcoran and Mueller, 2004). These would have been caused by the combination of high levels of heat, humidity, and greenhouse gases like CO_2 and CH_4 (e.g., Young, 1991; Kasting, 1993; Des Marais, 1994; Pavlov et al., 2001; Kasting and Siefert, 2002). As a result of this intense Archaean weathering regime and an absence of binding vegetation, labile minerals and unstable rock fragments would have been rapidly broken down to form clay minerals, which were swiftly separated by sediment transport and sorting agents from the predominant quartz and accessory heavy mineral grains more resistant to chemical breakdown (Corcoran and Mueller, 2004). By using the Chemical Index of Alteration (CIA; Nesbitt and Young, 1982), Condie et al. (2001) were able to demonstrate a gradual decline in weathering intensity over geological time, related to change in palaeo-atmospheric composition: average CIA values for Archaean shales of 80, declined to 75 in the Proterozoic and to 70 in Phanerozoic shales. Obviously, sediment recycling (estimated to be c. 90% in post-Archaean times by Veizer and Jansen, 1985) also needs to be taken into account here, but first cycle sediments can generally be recognised (although not exclusively) on the basis of textural immaturity (Corcoran and Mueller, 2004).

Another aspect to be addressed within Archaean clastic deposits is the so-called "quartz budget problem" (cf., Pettijohn, 1970, 1972). The basis of this paradox is the expectation that detrital deposits within Archaean greenstone terranes would have formed within a tectonic framework of great instability with concomitant rapid creation of accommodation related to inherently unstable basin-forming and subsidence mechanisms; as a result, highly immature compositions and textures should have formed (e.g., Donaldson and de Kemp, 1998). However, this is not the case, and numerous mature to even supermature quartz arenites are known from the Canadian Shield and many other ancient cratonic areas, where they show a relatively common spatial association with mafic volcanic rocks (Donaldson and de Kemp, 1998). The latter authors point out the associated problem of where these sediments might have been sourced, and deduce that phanerocrystalline granitoid rocks as well as recycling of existing sedimentary strata offer the best possible solutions. Once again, as in the previous paragraph, weathering-aggressive conditions are implicit, together with relatively high energy transport systems, as well as the likely necessity for prolonged periods of crustal stability within the overall active tectonic framework of Archaean crustal development (Donaldson and de Kemp, 1998).

Within the Proterozoic, particularly the Palaeoproterozoic (and assuming this to be a period of essentially post-cratonisation character), it could be assumed that the development of relatively stable cratons and probably also of the first supercontinents (e.g., Aspler and Chiarenzelli, 1998), allied to a still-reducing palaeo-atmosphere would have resulted in a detrital sedimentary record dominated by compositionally more mature sediments. This line of reasoning is complicated by the debate over the "Great Oxidation Event" at some time between about 2.35 and 2.0 Ga (e.g., Ohmoto, 2004; Lindsay and Brasier 2004; see also Eriksson et al., this volume) and the influence this may have had on the efficiency of the extant palaeo-atmosphere to break down source material. Some workers (e.g., Sheldon, 2006; Eriksson et al., 2009a) dispute any significant lessening of the greenhouse atmosphere for the c. 2.5-1.8 Ga period at all. Within this framework, it is interesting to examine, very briefly, the petrology of the clastic sediments of the

Pretoria Group (Transvaal Supergroup, Kaapvaal craton; c. 2.35-2.05 Ga) which fall almost exactly within this much debated time period.

The Pretoria Group comprises a maximum thickness of 6100 m of predominant alternating mudrocks and sandstones with lesser basaltic andesites, and subordinate conglomeratic and carbonate rocks (e.g., Eriksson et al., 2001b). A tectonic setting comprising two episodes of rifting followed by much longer periods of thermal subsidence is inferred (Eriksson et al., 2001b). Detailed petrographic studies of 170 sandstone thin sections around the basin provide the following proportions of rock types (after Pettijohn et al., 1972): quartz arenites (31%), lithic arenites (20.5%), arkosic arenites (21.5%), lithic wackes (14%), feldspathic wackes (9.5%) and quartz wackes (3.5%) (based on detailed data in Schreiber et al., 1991, 1992). Essentially thus, this well-studied Palaeoproterozoic example provides an extension of the Archaean "quartz budget" paradox – within a much less active tectonic setting and following cratonisation, more mature sandstones would be expected to have formed, but a contrary trend is, again observed. Proponents of the "Great Oxidation event" could easily ascribe this to significant change from a greenhouse atmosphere to a partially oxidising one, with concomitant decrease in weathering intensities; this would be supported by the possibly global c. 2.4-2.2 Ga glaciation event (e.g., Young, 2004). However, even under Phanerozoic-Modern atmospheric conditions, deposition of large quantities of feldspathic and lithic sandstones could argue in favour of rapid sedimentation processes rather than cold climatic conditions.

Although some workers do challenge any significant oxidation of the palaeo-atmosphere within the Palaeoproterozoic (e.g., Sheldon, 2006; Eriksson et al., 2009a), possible differences in the tempos and intensities of sedimentation systems may also have played a role in explaining the apparent conundrum. Eriksson et al. (2009a) suggest that strongly episodic yet high energy fluvial systems fed equally episodic marine coastal clastic systems on many of the cratons in the Palaeoproterozoic; they relate these intermittent yet high energy sediment supply and depositional events to the overall influence of a still strongly reducing greenhouse palaeoatmosphere. Comparison of Precambrian and Phanerozoic-Modern sedimentation rates is fraught with difficulty, for at least two reasons: the resolution of c. 1-5 my for radiometric dating of Precambrian rocks, and the very wide range in the more recent rates, varying from 10^1 to 10^6 Bubnoff units ($mmky^{-1}$) (e.g., Eriksson et al., 2004c). However, it is possible that such episodic sediment supply and rapid transportation and depositional events combined, in conjunction with generally slower rates of accommodation creation in the Palaeoproterozoic, may explain the sandstone composition conundrum in that part of the rock record, as can the importance of weathering-aggressive atmosphere composition account for the "quartz budget problem" in the Archaean (cf., Donaldson and de Kemp, 1998). Palaeo-atmospheric and palaeohydrospheric evolution will be further addressed under the carbonate sedimentation section of the paper.

2.2. Fluvial systems

A widely held consensus on Precambrian fluvial systems is that they were characterized by broad channel systems on very large braidplains, enhanced rates of channel migration, high discharge rates and flashy surface runoff, abundant bed-load, and that they lacked bank stability due to an absence of plants and to poorly-developed soils; braided systems would consequently have predominated (e.g., Schumm, 1968; Cotter, 1978; Long, 1978, 2004; Fuller, 1985; Els, 1990; Rainbird, 1992; Eriksson et al., 1998; see however, Jackson, 1978). Their scale, lack of bank stability and faster runoff rates would doubtless have made these river systems more susceptible to palaeoclimatic change, and concomitantly, ephemeral rivers probably occurred through a broader climatic range than Phanerozoic-Modern systems (Tirsgaard and Øxnevad, 1998). In general, models of bed-load dominated, sheet-like, braided alluvial deposits (cf. the Scott and Donjek types identified by Miall, 1977, 1978, 1996) are most often espoused for the Precambrian (Long, 2004). Identification of incontestable channel elements, or even channel margins, remains rare (Long, 2004).

Identified alluvial fan deposits are not common in the Precambrian record, their proximal positions affecting preservation (e.g., Els, 1998; as a caveat, however, see Mueller and Corcoran, 1998). The latter authors (also, 2001) point out that pyroclastic debris in Archaean greenstone belts commonly suffocated alluvial channels, leading to prominence of hyperconcentrated flood flow and sheetflood deposits in such systems. The weathering-aggressive atmosphere in the Archaean would have resulted in more muddy detritus entering river systems, also thus promoting such flood flow and sheetflood deposits (cf., Donaldson and de Kemp, 1998; Corcoran et al., 1998). Long (2004) points out that debris-flow deposits may thus well have occurred in non-fan fluvial systems as well (e.g., Buck and Minter, 1985; Pflüger and Seilacher, 1991; Eriksson et al., 2008a, 2009a). Despite the expected large volumes of muddy sediment, identified examples of high-sinuosity river deposits (e.g., Sweet, 1988; however, disputed when compared to Rainbird, 1992; see discussion in Long, 2004) are rare.

Discrimination of fan and fluvial braidplain deposits should logically best be achieved through the criteria outlined by Blair and McPherson (1994), particularly the "natural depositional gap" separating the gradients of fans (>0.026 m/m) and rivers (<0.007 m/m). However, palaeohydrological results from c. 2.05-1.8 Ga alluvial deposits from the Kaapvaal craton (discussed below) coincide almost exactly with this gap (e.g., Van der Neut and Eriksson, 1999; Eriksson et al., 2008a, 2009a). Architectural element analysis of Precambrian fluvial deposits has been relatively limited (e.g., Long, 2004 and references therein) as has been establishing sequence stratigraphic frameworks (e.g., Ramaekers and Catuneanu, 2004; Eriksson and Catuneanu, 2004a). Below we briefly discuss two examples of Precambrian river systems, from the Neoproterozoic Sonia Formation in India, and from the Palaeoproterozoic Waterberg Group of Kaapvaal.

Studies of the facies and architectural elements within a sequence stratigraphic framework in the fluvial interval at the base of the Neoproterozoic Sonia Sandstone (Jodhpur, Rajasthan, India) (Fig.1) identifies at least some reasons for Precambrian river channels shedding their usual mobile (i.e., braided) nature (Samanta et al., 2007, 2008; Samanta, 2008). Confined between a rhyolitic basement (~600 Ma) and a transgressive lag flooring a coastal interval above (Samanta, 2008; Sarkar et al., 2005, 2008), the three divisions of the Sonia fluvial interval are superposed one above the other and each is fining-upward, together giving rise to an overall fining-upward trend (Fig. 2; Sarkar et al., 2008). The middle division is, however, the coarsest, being pebbly in its lower part. Mudstone is completely absent in this division, but is present marginally in the lower and substantially in the upper divisions.

An additional difference between the fluvial divisions also exists in the abundance of aeolian products in them, being substantial in the middle, restricted in the lower and subordinate in the upper division; the river system was apparently ephemeral in the middle, perennial in the upper and probably semi-perennial in the lower division. Recorded channel flow direction veers selectively from a predominant westward- to southwestward-oriented trend in the middle division. Apparently, the palaeoslope direction altered significantly between the divisions. The top eight metres of the middle division selectively bears soft sediment deformation structures, which are pervasive all over the 1800 km^2 study area. The role of tectonism in partitioning the Formation into its three component divisions is thus amply demonstrated.

Further differences exist between assemblages of fluvial architectural elements (Miall, 1985, 1996, 2006; Willis, 1997; Halbrook, 2001; Yu et al., 2002; Miall and Jones, 2003; Lumsdon-West and Plint, 2005; Alexander and Fielding, 2006; Halbrook et al., 2006; Labourdette and Jones, 2007) identified within the three divisions (Samanta, 2008). The Sandy Lateral Accretion element dominates over Downsream Accretion elements in the lower and the upper divisions of the fluvial interval of the Sonia Formation, while not a single example of the former was encountered in the middle division. There is little difference, however, in the abundance of the Small Channel element, except that it is relatively more tabular in geometry in the lower division. The Laminated Sand Sheet element is present in all three divisions, but in the form of levee facies, is developed only marginally in the lower division and profusely in the upper division, while being completely absent in the middle. The same is true for the Flood Plain element. Laminated Sand Sheet elements in the form of crevasse splay facies encased in either Flood Plain elements or Inclined Heterolithic Strata (IHS) elements, and separated from Sandy Lateral Accretion elements on the basis of geometry and structure, are exclusive for the upper division, especially its top part. The Sandy Bedform element is rare for all three divisions.

Surfaces bounding the three divisions are readily correlatable all over the study area and define three different aggradational valley cycles, palaoslope changing significantly between them. Taking the transgressive surface that terminates the fluvial interval of the Sonia Sandstone at its top as a near-horizontal datum plane, and correlating the division boundaries observed in vertical sections distributed along two near-orthogonal transects, makes the three-dimensional geometries of successive valley cycles amenable to reconstruction (Fig. 3). It is amply clear from the reconstruction that fluvial aggradation was initiated, presumably after the maximum fall of the river base profile and deepest incision (cf., Blum and Tornqvist, 2000), in a constricted valley cut into the basement. Rapid widening of the valley, apparent at the transition from the basal to the middle division, is attributable to base profile rise or to base profile fall at a slow rate (cf., Whipple et al., 1998; Strong et al., 2005). A sharp increase in sediment grain-size across the transition testifies to an increase in flow velocity and higher water discharge despite rapid expansion of the drainage area, and thus favours profile rise, but possibly with source uplift; the observed change in channel flow direction makes the tectonic effect apparent.

Flood plain amalgamation in the upper Sonia fluvial division is a definite indication for an enhanced rate of base profile rise. Rapid basin subsidence ushered in transgression of the sea, although the study area was inundated only later. Relative rise of base profile, nonetheless, induced a tendency for ponding and allowed mud to settle at the downstream end of the river valley in ever increasing amounts, until the sea flooded the entire area; the river channels, in the meantime, became fixed. Tidal influence expected during drowning of the river system is evident only perhaps in the HIS architectural elements, observed exclusively within the top part of the upper division.

It thus appears that Proterozoic river channels may have tended to become fixed in Falling Stage Systems Tracts (FSST) and at the beginning of Lowstand Systems Tracts (LST) when the river valleys are likely to be narrow and may be incised within hard bedrock. Otherwise, Proterozoic rivers were destined to be braided because of the inherent instability of their banks, devoid of vegetation. However, during the closure of the Lowstand Systems Tract also, with an enhanced rate of rise of relative sea level, the Proterozoic rivers tended to become fixed, meandering or even anastomosed because of bank stability gained through increased mud settling.

The c. 2.0-1.8 Ga (Walraven and Hattingh, 1993; Eglington and Armstrong, 2004; Hanson et al., 2004) Waterberg Group is amongst those units globally, containing the first red beds *sensu stricto* (reflecting free oxygen in the palaeo-atmosphere) and relatively large erg deposits (e.g. Eriksson and Cheney, 1992; Eriksson and Simpson, 1998). These sedimentary beds were laid down within two basins on the Kaapvaal craton (Fig. 4) and are dominated by fluvial deposits, with lesser desert, fan and lake sediments (Callaghan et al., 1991; Simpson et al., 2002, 2004b). Both basins are bound by fundamental Archaean structures within the Kaapvaal craton (Callaghan, 1991; Eriksson et al., 1996). The large Waterberg basin contains 11 formations and the small one, a single formation, the Wilgerivier; this later unit and the Mogalakwena Formation from the large (Main) basin have been studied in detail to try and unravel the nature of their fluvial deposits (Van der Neut and Eriksson, 1999; Bumby, 2000). The deposits of the c. 2.0 Ga Blouberg Formation, which is unconformably succeeded by Waterberg rocks, have also been studied in detail; all three units comprise analogous fluvial deposits (Eriksson et al., 2008a and references therein).

The Mogalakwena Formation comprises cyclically interbedded sheets (0.5-1.0 m thick, 100's of metres in extent in outcrop) of medium- to coarse-grained sandstone or granulestone, and matrix-supported, largely massive conglomerate sheets (Bumby, 2000). These cycles generally fine upwards, and are equated with architectural element CHS (major sandstone sheet; Miall, 1985; 1996). The sheets are cut locally by conglomeratic channel-forms (architectural element CH of Miall, 1985, 1996), 20-30 m across, and 2-5 m deep. Common trough cross-bedding with a consistent unimodal palaeocurrent trend characterizes the predominant sandstone-granulestone sheets (Bumby, 2000). A general braided fluvial model is inferred for these deposits, and for those of the Blouberg and Wilgerivier Formations (Eriksson et al., 2008a). Applying standard palaeohydrological methodology to these three Waterberg braided fluvial formations indicates palaeoslope estimates which coincide almost precisely with the observed “natural depositional gap” separating alluvial fan gradients (>0.026) and river slopes (<0.007; cf. Blair and McPherson, 1994) (Eriksson et al., 2006, 2008a) (Fig. 5).

Despite all three formations exhibiting typical low sinuosity-type fluvial facies and internally consistent and unimodal palaeocurrent patterns typical of braided river deposits, there are subordinate facies indicating sheetflow-sheetflood, debris-flow or density-modified grain flow deposition (Van der Neut at al., 1991; Bumby, 2000; Eriksson et al., 2008a). Eriksson et al. (2009a) postulate that temporary argillaceous sediment accumulations within these fluvial systems (related to aggressive greenhouse palaeo-atmospheric conditions providing more labile clay material within continental sedimentary systems) enabled higher palaeoslopes to develop locally, and also led to localized gravity-flow deposits within these braided systems. An analogue is provided by the Neoproterozoic Kuujjua Formation (Canadian Shield; Rainbird, 1992). Eriksson et al. (2009a) relate this strongly episodic style of fluvial sedimentation to either a continuation of a greenhouse palaeo-atmosphere at this time, or a more gradual or possibly even diachronous transition to a more oxygenated atmosphere than that implicit in the “Great Oxidation Event” (e.g., Holland, 2002) at some time between c. 2.35 and 2.0 Ga, at least for the Kaapvaal craton (see also, Voegelin et al., 2010, for comment on global redox changes and the Kaapvaal craton during the Neoarchaean-Palaeoproterozoic).

2.3. Aeolian systems

Wind ripple stratification is adequately diagnostic of aeolian processes to allow confident discrimination from aqueous sedimentation processes; supplementary diagnostic features include pin-stripe lamination, adhesion structures and coarse sand/granule ripples; together, such features equally allow identification of Precambrian aeolian deposits (e.g., Hunter, 1977, 1981; Kocurek and Dott, 1981; Kocurek and Fielder, 1982; Clemmensen and Abrahamsen, 1983; Fryberger et al., 1988, 1992; Clemmensen and Dam, 1993; Bose and Chakraborty, 1994; Eriksson and Simpson, 1998; Bose et al., 1999; Simpson et al., 2004a). Eriksson and Simpson (1998) distinguish various depositional settings or facies within the preserved Precambrian aeolian record: (1) *draa deposits*, normally large in extent and thickness, characterised by three orders of bounding surfaces (e.g., Brookfield, 1977); (2) non-recognition of the latter within inclined cross-strata comprised of combinations of wind-ripple, grainfall and grainflow deposits (Hunter, 1977; Kocurek and Dott, 1981), results in a classification as *dune deposits*; (3) *dune plinth deposits* are commonly associated with dune deposits as they occur at the base of dunes, but are distinguished by near-horizontal to horizontal wind-ripple stratification; (4) *sand sheet deposits* which are commonly comprised of coarser-grained (coarse-grained sand to granule calibres) low angle to horizontal inversely graded strata with adhesion structures (e.g., Fryberger et al., 1979). Sand sheet development is often linked to conditions which reduce dune formation, such as lower sand supply, high water table, surface cementation and binding (cf., vegetation) (Kocurek and Nielson, 1986); (5) *interdune deposits* occur between dunes and/or draas and typically include argillaceous sediments, a combination of aeolian and water-formed structures as well as evidence for desiccation.

In their review of Precambrian aeolianites, Eriksson and Simpson (1998) find that such deposits become relatively common in the rock record from c. 1.8 Ga onwards. In their list of twenty identified Precambrian aeolian deposits, only three (two members of the c. <2.4->2.1 Ga Kinga Formation, Canada, and the c. 2.1 Ga Deweras Formation, Zimbabwe) pre-date this c. 1.8 Ga age; most of the early deposits, including those prior to c. 1.8 Ga comprise dune and sand sheet types (Eriksson and Simpson, 1998). An association of these various aeolian deposits with either fluvial or shallow marine strandline facies is commonly observed, as is a rift-related tectonic setting (Eriksson and Simpson, 1998). Two more ancient windblown deposits have been recognised since this compilation: the Tamanduá Group in Brazil, and the Dhalbhum Formation in India (Simpson et al., 2004a).

The oldest recognised aeolian deposit is that within the <2.70 - > 2.42 Ga Tamanduá Group (Minas Supergroup, SE Brazil) (Babinski et al., 1995; Machado et al., 1996; Simpson et al., 2004a). These deposits occur within a rift setting, overlie continental deposits of predominantly braided fluvial character, and are about 2500 m thick; they are interpreted as a sand sheet facies, characterised by decimentre- to metre-thick trough cross-beds, with locally complex cross-stratification as well as interdune sediments (Martins-Neto and Costa, 1985; Simpson et al., 2004a). Within the Hearne crustal province of Canada, overlying inferred glacial deposits equivalent to the well-known Huronian Supergroup occurences is the Kinga Formation of the Hurwitz Group (e.g., Aspler and Chiarenzelli, 1998). The Maguse Member of this formation comprises both dune and sand sheet deposits, c. <2.4 - >2.1 Ga, associated with fluvial deposits within an overall coastal aeolian complex (Patterson and Heaman, 1991; Aspler et al., 1992; Heaman and Le Cheminant, 1993). The Whiterock Member, interpreted to be a lacustrine deposit, exhibits adhesion structures reflecting an aeolian genesis (Aspler et al., 1994).

The 2.3 Ga Dhalbhum Formation (Singhbhum craton, India) formed within an intracratonic rift basin, and is spatially associated with fluvial deposits; it includes a relatively thin (average 30 m) aeolian unit characterised by wind ripple laminae, translatent strata with both grainfall and grainflow laminae, isolated lenticular cross-bedded dune deposits (forming sets up to 40 cm thick) and adhesion structures (Mazumder et al., 2000, Simpson et al., 2004a). The latter authors note that the wind ripple laminae are stacked locally into sets of c. 20 cm thickness, separated by iron-stained, second-order surfaces that lack armour and which may have been non-erosional. The translatent strata are found locally associated with the aeolian ripples; the crinkled adhesion laminae occur as ~ 6cm thick sets. The aeolian features noted above commonly formed ~ 90 cm thick cycles from basal adhesion laminae – translatent/ripple laminae – isolated cross-bedded dune sets at the tops, cycles being bound by extensive approximately planar erosion surfaces (Simpson et al., 2004a). A relatively high water table (preventing draa evolution) and low relief provenence close to sea level are inferred; the ~ 90 cm thick drying-up cycles may have been influenced by palaeoclimatic fluctuations (Simpson et al., 2004a). The next younger known example of a Palaeoproterozoic aeolian deposit is from the c. 2.1 Ga Deweras Group in NW Zimbabwe that bears large scale cross-beds with inversely graded stratification, associated with evaporites (Master, 1991) within inferred wind ripple deposits (Eriksson and Simpson, 1998). A relatively low water table is again implied.

From about 1.8 Ga, large and well-developed aeolian deposits become common in the rock record: e.g., erg deposits of the c. 1.85-1.75 Ga Baker Lake and Wharton Groups (Dubawnt Supergroup, Canada); dune, sand sheet, interdune and dune plinth deposits of the c. 1.8-1.74 Ga Bottletree Formation (Upper Mount Guide Quartzite, Mount Isa Orogen, Australia); draa and dune deposits of the c. 1.66->1.60 Ga Hornby Bay Group and Thelon Formation (NW Canada)

(Eriksson and Simpson, 1998 and references therein). An excellent example of the first widespread aeolianites is given by the c. 2.0 – 1.85 Ga Makgabeng Formation (Waterberg Group, Kaapvaal craton; see Fig. 4), which is interpreted to be an erg deposit (Meinster and Tickell, 1975; Callaghan et al., 1991; Bumby, 2000; Simpson et al., 2002, 2004b). The Makgabeng Formation is preserved over an extent in excess of 100 km in its longest dimension, reaches up to c. 800 m in thickness, and is bounded above and below by essentially fluvial deposits (Bumby et al., 2001; Eriksson et al., 2000). Despite having formed within a fault-bounded depository (Bumby et al., 2001), its sediments locally prograded across at least one of the major bounding structural features (Bumby, 2000) suggesting thermal subsidence. The lower contact with underlying fluvial deposits is gradational (Callgahan et al., 1991), supporting the likelihood of palaeoclimatic deterioration; the upper contact is erosional (Bumby, 2000; Eriksson et al., 2000). By far the predominant facies within the Makgabeng palaeo-desert deposits is ascribed to aeolian origin, and comprises texturally and mineralogically mature fine-grained sandstones with either straight-crested or barchanoid dune cross-bedding (Fig. 6), with set thicknesses mostly between 1 and 3 m with a maximum recorded of 15 m (Eriksson et al., 2000). Within the best exposed area (Makgabeng plateau, where about 300 m of vertical succession can be studied), Makgabeng aeolianites exhibit a medial planar cross-bedded succession, sandwiched between barchanoid trough cross-bedded successions. Towards the top of this vertical aeolian transition, four interdune beds (40-110 cm thick) are exposed over 5 m of thickness and extend laterally for about 100m; these four lenticular beds exhibit planar laminations, wave, current and combined flow ripples, mudcracks superimposed on the ripples, possible evaporate casts, as well as muddy "roll-up" structures indicating desiccation and subsequent reworking of microbial mats by sheetfloods (Eriksson et al., 2000; Simpson et al., 2002). Locally, erosively-based massive sandstones are interbedded with the lower portions of aeolian cross-bed sets and dune toesets (Fig. 7), indicating genesis through mass-flow processes related to significant precipitation events; hyperconcentrated flows are inferred to have flowed down dune lee faces and eroded into them, with lobate deposits resulting on the dune plinths (Simpson et al., 2002). Several metres above the uppermost interdune bed, Simpson et al. (2004b) documented saline pan deposits, up to about 2 m in thickness, with facies stacking patterns suggesting flooding and desiccation cycles of variable lengths (estimated at months to hundreds of years). Although not large by current erg sizes, the Makgabeng counts as one of Earth's oldest such deposits, and exhibits most of the features to be observed in any Phanerozoic – Modern desert succession, with strong evidence for an interplay of temporally dominant aeolian and subordinate aqueous processes and products (Eriksson et al., 2000, Simpson et al., 2002, 2004a and b).

By the Neoproterozoic, ergs had become common within the Precambrian record, as for example in the 0.6 Ga Upper Bhander Sandstone. It is the topmost Member of the Bhander Formation of the Vindhyan Supergroup, best studied in Rampur Hills, skirting Maihar in central India (Fig. 1; Bose et al., 2001). Resting on the overall prograding marine shelf succession of the coeval Sirbu Shale, this Member is dominantly non-marine, marking the penultimate stage of sedimentation within an intracratonic sag basin (Bose et al., 1999, 2001; Sarkar et al., 2002a; Ray, 2006). Parasequences, separated by gently seaward dipping convex-upward erosive surfaces, comprise this Member. Individual parasequences bounded below and above by second-order superbounding surfaces, are upward-drying with interdune sand-sheet deposits passing upward into dune and draa deposits. The lower part of each parasequence is dominated by adhesion laminae, followed upward by translatent strata encasing solitary sets of dune laminae in the middle part and multiple dune-cross beds, locally made up of longitudinal dunes, in the top part (Bose et al., 1999). Thickness of parasequences, as well as thickness of draa deposits on top of some of the parasequences, tends to increase landward. The marine flooding surfaces defining individual parasequences, however, rapidly disappear in an inferred landward direction. Relatively thin fluvial and lacustrine beds intermittently disrupt the record of otherwise continuous aeolian deposition. As a whole, the major part of the Upper Bhander Sandstone represents an erg deposit with a seaward counterpart of marine supralittoral storm bed packages.

Explaining the temporal distribution of aeolian deposits in the Precambrian record is complex. Eriksson and Simpson (1998) discuss various explanations such as an absence of vegetation, crustal growth rates, tectonic setting, eustasy and local sea level changes, palaeoclimatic conditions, fluvial reworking and non-recognition. They note a c. 600 my gap between significant continental crustal growth rates and cratonisation (at. c. 2.4 Ga) and the development of widespread large-scale aeolianites globally at c. 1.8 Ga.

The oldest known evidence for aeolian sedimentary activity is from ventifacts found within the c. 3.0-2.8 Ga Witwatersrand basin, Kaapvaal craton (e.g., Minter, 1976). Prior to c. 1.8 Ga, no preserved erg deposits are known, and the relationship between aeolian and spatially associated fluvial deposits as well as a rift-related tectonic setting is striking. Although it is a contentious issue, supercontinentality may already have begun on Earth in the Neoarchaean (e.g., Aspler and Chiarenzelli, 1998; see discussion in Eriksson et al., this volume). If this is so, then the oldest known aeolianites, from the <2.70 - > 2.42 Ga Tamanduá Group are approximately coeval with earliest supercontinent development, and are followed by the Hurwitz Group and Dhalbhum aeolianites in the c. 2.45 - 2.1 Ga period (see discussions in Eriksson and Simpson, 1998; Simpson et al., 2004a). However, supercontinentality only appears to have become widespread on Earth by about 2.2 – 1.8 Ga (e.g., Eriksson et al., this volume), approximately coincident with widespread aeolianite development including the first ergs. The role of large continental land masses may thus have been critical. Also important was groundwater and fluctuations in the water table, in determining sand supply and preservation of aeolianites (Eriksson and Simpson, 1998).

Palaeoclimatic changes and fluvial reworking are harder to evaluate. The likelihood of a humid and warm palaeoclimate up till about 2.3 Ga is agreed upon by most (e.g., Ohmoto, 2004), and this humid setting may have impaired aeolianite formation. The "Great Oxidation Event" at some time between about 2.35 and 2.0 Ga (Ohmoto, 2004) would not have done much to encourage dry conditions, but may have been counteracted by increasingly large continental masses, with dry interiors. However, some have questioned this oxidation event and find that the global greenhouse palaeo-atmosphere may have endured to at least 1.8 Ga (e.g., Sheldon, 2006; Eriksson et al., 2009a). In such a case, a rift setting may have led to locally favourable palaeoclimatic conditions for sand sheet and even erg formation within a globally warm yet humid palaeoclimatic regime. Strongly episodic fluvial flooding events may have typified continental aqueous sedimentation regimes in the c. 2.3-1.8 Ga period (cf., Eriksson et al., 2009a), and this likely strongly influenced fluvial reworking of aeolian deposits also.

2.4. Marine systems

Degassing of Earth's mantle provided water to the palaeo-atmosphere, which condensed to form the early oceans by about 4.0 Ga (c. 90% of current volume); early seawater was most likely acidic due to relatively high contents of dissolved CO_2 and other acidic components like H_2S and HCL (e.g., Condie, 1997). However, volcanic eruptions and recycling of seawater at mid-ocean ridges would relatively rapidly have resulted in a “modern” ocean composition (Condie, 1997).

The long and complex history of early plate tectonics (in whatever form; see for example, discussion by Eriksson and Catuneanu, 2004b) followed by a more modern style of plate interactions in the c. <2.0 Ga period and by the subsequent Neoproterozoic and Phanerozoic regime of Wilson cycles as supercontinents formed and dispersed, resulted in marine deposits marginal to and adjacent to Precambrian cratons and terranes not generally being preserved. Essentially thus, Precambrian marine deposits comprise a record of epeiric seas rather than the remnants of open ocean margins, shelves and deep sea settings (Eriksson et al., 1998; 2004d). The Precambrian marine record must therefore be treated with some caution due to this preservational bias towards epicontinental sea deposits. Their study is not made any easier by a relative lack of true modern examples (e.g., Hallam, 1981; Galloway and Hobday, 1983; see also the recent treatise on epeiric seas edited by Pratt and Holmden, 2008). It is also necessary to discriminate between epeiric seaways (with shelf-like portions, shelf-breaks and even deeper distal parts; strongly directional currents) and epeiric embayments (lacking shelf breaks and ocean-type currents) (Brenner, 1980; see also Eriksson et al., 1998; 2008b). The former resembled extant open ocean shelves (Brenner, 1980) and can be compared to shallow oceans with some confidence, while the epeiric embayments were much smaller and shallower (Eriksson et al., 2004d; 2008b).

The reduced shelf gradients of many epeiric seas would have tilted the balance in favour of tides instead of storms (and thus also waves) (Eriksson et al., 2008b); on shallow water platforms, waves dissipate rapidly, while tidal height generally increases (Pratt and James, 1986). This supposition is supported by the presence of extensive peritidal flat deposits on the shoaling portions of several inferred epeiric sea basins of Phanerozoic age (Swett et al., 1971; Reading, 1978; Pratt and James, 1986; Friedman et al., 1992). Another characteristic of the Precambrian epeiric seas was profuse high energy braid-delta systems along their coastlines (Els, 1998; Eriksson et al., 2008b). Away from the river mouths where tidal interaction was important (e.g., Eriksson et al., 1995), wave-dominated coastline segments were possibly typified by amalgamation of supralittoral storm beds (Sarkar et al., 2004).

The development of the large scale and tide-dominated Precambrian epeiric embayments, with gentle gradients towards open oceans was commonly closely associated with large braided fluvial systems draining craton interiors and which flowed directly into these shallow seas (e.g., Eriksson et al., 2008b). High denudation rates under earlier Precambrian palaeo-atmospheric compositions also played a role in these systems (Els, 1998). In broad terms, preserved Precambrian shelf (cf., shelf-like epeiric) deposits are commonly sandy and they tend to resemble closely their Phanerozoic-Modern equivalents (Jackson et al., 1990; Lindsey and Gaylord, 1992; Eriksson et al., 1998) but with some significant exceptions (Sarkar et al., 2005, 2008; Catuneanu and Eriksson, 2007). Open shelf deposits dominated by wave imprinted nearshore sandstones and by massive and laminated mudstone with subordinate amounts of storm–deposited graded siltstone and fine grained sandstone, are spectacular in their preservation of fine primary features because of a lack of bioturbation (e.g., Schieber, 1989; Chakraborty and Bose, 1992; Sarkar et al., 2002a). Precambrian shoreface deposits, like their younger counterparts, are characterised by highly mature, well sorted sandstone (Walker and Plint, 1992), with virtually no mudstone, although pebbles and pebble lags may occur (Harris and Eriksson, 1990; Eriksson et al., 1998). Though hummocky and swaley cross-strata, parallel laminae, trough and planar cross-bedding are as common as in Phanerozoic successions (e.g., Soegaard and

Eriksson, 1985; Bose et al., 1988; Jackson et al., 1990; Tirsgaard and Sønderholm,1997), Precambrian barrier island, wash-over fan (Eriksson, 1979) and lagoon deposits are only rarely discussed in the literature (e.g., Eriksson et al., 1998). However, tidal accentuation behind localized shoals has been recorded at places, such as in the Neoproterozoic Sonia Sandstone, India (Sarkar et al., 2008). The variation in sedimentation processes operating on Precambrian shelves (cf. epeiric seas) is elicited in the three examples discussed here, with the first two of Archaean and Palaeoproterozoic age highlighting considerable tidal influences and the third, of Neoproterozoic age, illustrating a dominant storm influence.

The c. 3.1-2.8 Ga Witwatersrand basin formed while composite terranes including greenstone belts were still accreting with the nucleus of the Kaapvaal craton (South Africa) and while widespread intrusion of granitoid rocks took place (De Wit et al., 1992; Robb and Meyer, 1995). As a result of accretion from north and west, the basin comprised a double (retroarc flexural) foreland depository, with the Witwatersrand Supergroup reflecting the foredeep and the partially correlated Pongola Supergroup the backbulge parts of a "greater Witwatersrand basin" (Catuneanu, 2001; Eriksson et al., 2009b) (Fig. 8). The weaker Kaapvaal continental lithosphere, inherent in the coeval evolution of both craton and supracrustal basin, allied to low angle subduction of the accreting terranes, gave a much shorter flexural wavelength, resulting in the forebulge remaining emergent throughout basin history (Catuneanu, 2004; Fig. 8). This emergent forebulge separated the foredeep and backbulge sub-basins, each characterized by a unique yet partly correlatable supergroup.

This greater Witwatersrand epeiric embayment can be traced for at least 400 km inland from the inferred craton margin and for at least 600 km along its inferred seaward margin (e.g., Eriksson et al., 1998). During an earlier phase of epeiric sedimentation, the c. 2970-2914 Ma

(Robb and Meyer, 1995) West Rand Group (Witwatersrand Supergroup; foredeep) sub-basin collected up to 7.5 km of mudrocks, texturally mature sandstones and minor lavas, thought to have been deposited within shoreline to distal shelf palaeoenvironments subject to strong tidal influences (Eriksson et al., 1981; Burke et al., 1986; Stanistreet and McCarthy, 1991; Beukes, 1996). In the correlated Mozaan Group (Pongola Supergroup; backbulge) sub-basin, a combination of wave and storm-deposited sediments and subordinate fluvial braidplain sediments accumulated during this early phase (Beukes and Cairncross, 1991), under higher energy marine conditions due to a craton-marginal position (Fig. 8) (Eriksson et al., 2008b). A later phase of greater Witwatersrand basin sedimentation was restricted to the foredeep sub-basin, where the >2894 – c. 2780 Ma (Robb and Meyer, 1995) Central Rand Group, comprising mainly braided fluvial sandstones and subordinate auriferous conglomerates, was laid down. Large braided channel systems are thought to have flowed directly into the epeiric embayment due to a lack of vegetation, and large braid-deltas probably formed at the shorelines (Els and Mayer, 1993, 1998) where tidal reworking was important.

The c. 2.66- 2.05 Ga Transvaal Supergroup (Kaapvaal craton; preserved in three basins: Transvaal, Griqualand West, Kanye – Fig. 9) comprises four stratigraphic intervals: (1) a set of lowermost "protobasinal" (a descriptive term) immature fault-bounded basin-fills, present only in the Transvaal basin; (2) thin fluvial sheet sandstones of the Black Reef Formation (Transvaal and Kanye basins only); (3) a thick carbonate-BIF platform succession (c. 2642-2432 Ma Chuniespoort-Ghaap-Taupone Groups) across all three preserved basins; (4) a thick succession of clastic sedimentary and lesser volcanic rocks, belonging to the c. 2.4-2.1 Ga Pretoria Group (Transvaal and Kanye basins); a clastic succession overlying #3 in the Griqualand West depository may be a correlate of the Pretoria, or might be approximately coeval with an estimated 80 my hiatus separating the chemical and clastic successions in the Transvaal basin (see discussion in Eriksson et al., 2001b; Catuneanu and Eriksson, 1999, 2002) (Fig. 10).

Three epeiric sea deposits are identified within the Transvaal succession. The first is that of the Chuniespoort-Ghaap-Taupone Groups (up to 2.5 km of stromatolitic carbonate rocks and 0.7 km of overlying BIF), interpreted to have formed within an epeiric seaway occupying a thermal sag basin which was at least 600,000 km^2 in size (Beukes, 1987; Altermann and Siegfried, 1997; Eriksson and Altermann, 1998; Catuneanu and Eriksson, 1999). Within this basin, water depths are estimated to have been between 40 and 80 m for the carbonates and over 100m for the BIF (Klein et al., 1987; Klein and Beukes, 1989). Within the succeeding Pretoria Group, two clastic epeiric embayment successions are inferred, for the Timeball Hill and for the Silverton-Magaliesberg formations (Fig. 10).

For the Timeball Hill embayment, preserved over an area of ≥500 x 300 km, rifting and subsequent thermal subsidence allowed advance of the sea onto the Kaapvaal craton (Catuneanu and Eriksson, 1999, 2002; Eriksson et al., 2001b). The preserved basin-fill (and its interpretation) comprises 5 facies associations: (1) basal carbonaceous mudrocks (anoxic suspension deposits); (2) grading up into sheets of laminated, graded mudrocks and (3) overlying sheets of laminated/cross-laminated siltstones and fine-grained sandstones (Te, Td and Tc subdivisions of low-density turbidity current systems; cf., Bouma, 1962; Lowe, 1982; Stow, 1986); (4) disconformably followed by sheets of mature cross-bedded sandstones (lower tidal flat), interbeded with lenses of immature sandstones and mudrocks (medial to upper tidal flat); (5) small lenses of coarse siltstone-very fine-grained sandstone (contourites; discussion in Eriksson et al., 2008b) occur within #'s 1 to 3 and as local wedges at the # 1 -2 facies association transition (Eriksson and Reczko, 1998). These genetic facies associations, the inferred tectonic setting, and thin stromatolitic carbonate interbeds within #'s 1 to 3 (thereby suggesting photic water depths of c. 90-100m in the absence of planktonic fossils), together suggest a relatively deep water embayment model (Fig. 11) for the Timeball Hill Formation sea (Eriksson and Reczko, 1998). It is postulated that as the Timeball Hill basin gradually filled with thick sheetlike delta-fed distal turbidite sediments, shallow water conditions followed, allowing progradation of uppermost sandy tidal flat deposits across the regressing basin, the latter fed by braid-deltas distal to proximal braided river systems advancing into the shrinking embayment from the west (Eriksson and Reczko, 1998; Eriksson et al., 2008b).

The younger Silverton-Magaliesberg clastic embayment is thought to have formed within a similar tectonic setting, and was of comparable dimensions to the Timeball Hill depository, but its inferred sedimentation systems were more akin to those of a modern passive margin-shelf setting (Eriksson et al., 2002, 2004d; 2008b). Up to 2 km of Silverton Formation basin-fill comprises thin basal braid-delta sandy deposits, overlain by a thick succession of argillaceous facies (inferred sub-storm wave base pelagic deposits in transitional and offshore mud belts) which were locally reworked by offshore storms (Eriksson et al., 2002; Fig. 11). These argillaceous deposits are interpreted as fluvial muds which bypassed a high energy coastal sand belt (the Magaliesberg Formation; Fig. 11) (Eriksson et al., 2002, 2008b). The latter formation reflects ephemeral braid-delta systems which provided clastic sediment to the high energy sandy peritidal flats characteristic of the Magaliesberg-Silverton embayment coastline (Eriksson et al., 1995). Detailed modeling of this palaeoenvironment by Parizot et al. (2005) suggests that braid-delta sediment supply exceeded tidal reworking capacity, thus promoting progradation of the Magaliesberg sandy deposits over the muddy-silty shelf-like deposits of the Silverton (Figs. 10

and 11) (Catuneanu and Eriksson, 1999; Eriksson et al., 2001b). Comparing the two models in Fig. 11 shows a much lower inferred palaeoslope for the Silverton-Magaliesberg embayment.

Detailed studies of ripple marks and cyanobacterial mat features, extensively preserved within the Magaliesberg sandstones, indicate a paucity of high energy swash-formed ripples, limited wave heights (c. 1.5-23.5 cm; average 7 cm), and shallow coastal water depths above tidal surfaces (c. 7-152 cm; average 31 cm), with tidal conditions estimated to have been meso- to macrotidal (Eriksson et al., 1995; 2004d, 2008b; Parizot et al., 2005). A tide-dominated coastline lacking effective wave action is thus postulated, similar in many ways to coastal conditions inferred for the Witwatersrand epeiric foredeep discussed above.

The Neoproterozoic Sirbu Shale (Ray, 2006), Vindhyan Supergoup in central India (Fig. 1) has its basal ~7.2 m constituted by mixed lithologies, both siliciclastic and carbonate, of lagoonal origin. The rest of the formation, up to 185 m thick, originated in a shelf setting through overall progradation following a transgression of the sea (Sarkar et al., 2002b). These authors identified six shelf facies, five of them maintaining gradational contacts between themselves and recurring in occurrence. The sixth facies is non-recurring, inordinately coarser and thicker than all other facies and has sharp lower and upper contacts. Wave-formed features are common to all facies. Among the first five facies, four are dominated by shale that encases sheets of well sorted siltstone or sandstone characterized by current structures at the sole, wave structures and overall grading within sheets, and with wave ripples on the tops in most cases. The fifth facies is sandstone with thinner interbedding of sandy siltstone. Current-formed sole structures, internal wave features, overall grading and bed-top ripples are present in this facies too, only larger in scale. All five facies were apparently deposited on a shelf dominated by seasonal storms. These five facies were genetically related, but differing in palaeogeographies, which varied from near storm wave base to above fair-weather wave base. The single occurrence of the incongruent sixth facies represents an extraordinarily strong wave-formed deposit, possibly induced by a seismic event.

The structures at the soles of the coarser grained beds of the first five facies often belong to multiple generations. Systematic generation-wise vector analysis of these structures reveals the operation of shore-parallel geostrophic flows, as documented in storm-affected modern seas (Walker, 1984; Niedorroda et al., 1985). It is apparent that geostrophic flow increasingly dominated over offshore directed flow with the passing of the peak storm stage and with increasing distance from the shore (Chakraborty, 2002; Sarkar et al., 2002b).

2.5. Deltaic systems

Deltas in the Precambrian, both Archaean and Proterozoic, are well documented, but it is not easy to distinguish their constituent subenvironments. The distinction between ideally channelized and fining-upward delta plain deposits, and non-channelized and coarsening-upward progradational delta front deposits is often too subtle to pin-point in non-fossiliferous Precambrian successions. The difficulty in discriminating subenvironments thus stems mainly from the poorly preserved evidence of biogenic activity extant in Precambrian time. General similarity in terms of facies associations, sedimentary structures (non-biogenic) and lithological distributions, nevertheless, indicates general similarity in terms of depositional environments and processes between deltas of Precambrian and Phanerozoic times (e.g., overview in Eriksson et al., 1998). Thus, Precambrian deltaic sequences are characterized by the same overall coarsening-upward vertical profiles as observed in Phanerozoic-modern deltas, showing a transition from the muddier facies of the prodelta into the sandier facies of the delta front and distributary mouth bar subenvironments, conformably or unconformably overlain by the delta plain facies. Distributary channels on Precambrian delta plains presumably had poorly consolidated banks and consequently had a preference for braiding in consistency with the general consensus for Precambrian rivers in this regard (Schumm, 1968; Long, 2004). Their deposits exhibit the most diverse lateral facies variation and include a variety of sedimentary structures corresponding to different flow regimes, conspicuous erosional surfaces and wide palaeocurrent variations. Precambrian delta front sediments, on the other hand, show an overall similarity but with localised gravity-flow products and soft-sediment deformation structures. The prodelta facies also include frequent deformational structures, such as convolute bedding, diapiric features and loading structures. The presence of such deformational features attests to higher slope of the depositional palaeosurface and high sedimentation rates. Conglomerates, as channel lag deposits, are quite common at the base of the tidal channels or distributary channel-fills. Petrographically, the sandstones are somewhat immature and can be characterized as quartz arenites and subarkoses with lithoclasts such as chert fragments. One major difference between the Precambrian and Phanerozoic delta system lies in the stupendous thickness of stacked deltaic lobes. For example, the Basnaering delta complex of northern Norway has a maximum thickness of 3500 m, and thickness of the delta front – mouth bar – delta plain succession exceeds 400 m (Siedlecka et al., 1989); thickness of the delta front – mouth bar – delta plain succession in the Archaean Moodies Group, South Africa, similarly exceeds 400 m (Eriksson, 1979). The other significant difference is in comparative textural immaturity of the Precambrian deposits. These two differences can be accounted for by the high subsidence rates, steep topographic slopes and sharp reliefs of a tectonically active rift-related basin, very much characteristic of the

Precambrian depositories (Eriksson et al., 1998). A high rate of sediment supply under efficient alluvial transportation in pre-vegetational time could also have facilitated acquisition of huge thicknesses (Eriksson et al., 1998).

The marginal marine environment was also characterized by the more common presence of braid-delta systems during the Precambrian than in Phanerozoic time (Nemec and Steel, 1988). Braid-delta deposits tend to show both fluvial and wave/tide influences, reflecting complex modification of alluvial deposits in the shore zone. In Precambrian time, rapid chemical weathering (Corcoran et al., 1998; Donaldson and de Kemp, 1998) and intense erosion in the absence of vegetation promoted the development of large braided river systems characterized by high discharge rates. Such braided river systems debouching directly into gently sloping marine shelves of Precambrian epeiric seas, commonly produced extensive braid-deltas (Els, 1998).

2.6. Lacustrine systems

Lacustrine deposits, whether Precambrian or Phanerozoic, can be explained as analogous to products of small-scale ocean margins, with less intense wave processes (Galloway and Hobday, 1983). Except for significant lunar tides and marine swells, wind-induced waves in lakes are almost the same as in seas (Friedman et al., 1992). Hence the distinction clearly lies in the magnitude of the wave-action simulating the processes of open marine microtidal coasts (Galloway and Hobday, 1983). In the Phanerozoic record, lake deposits can easily be identified by palaeontological studies, supported further by travertine, tufa, trona and magadiite (a sodium-silicate mineral) occurrences (e.g., Reeves, 1968; Hardie et al., 1978; Hallam, 1981; Eugster, 1986; Ordonez et al., 1991). In contrast, Precambrian lake deposits are hard to identify chiefly due to the absence of preserved palaeontological evidence. Furthermore, evaporites of Precambrian time are generally represented by pseudomorphs, which do not point to any certain origin (Donnelly and Crick, 1988). The common cyclicity of lake deposits, reflecting climatic variability, is also analogous to many marine mesosequences (Friedman et al., 1992), and stromatolites tend to be common in both environments (Hallam, 1981). The discrimination of Precambrian lacustrine deposits from shallow marine and epeiric sediments is thus subtle and problematic (Hallam, 1981; Miall, 1984; Montes et al., 1985; Winston, 1986; Donnelly and Crick, 1988; Schieber, 1998; Eriksson et al., 1998).

2.7. Microbial mat influence on clastic sedimentation

One of the most spectacular and unique features of the Precambrian, particularly the Proterozoic siliciclastic depositional systems, was the prolific growth of microbial mats on almost all wet sediment surfaces within the photic zone. In the absence of grazers and burrowers there were few environmental barriers to microbial mat growth and in the absence of plankton, light penetration extended the photic zone deeper in Precambrian seas or lakes (Walter and Heys, 1985). Microbiota appeared on earth before 3.8 Ga, and likely proliferated and flourished on most damp sediment surfaces during Proterozoic time (Schopf, 1999; Hagadorn and Bottjer, 1999; Eriksson et al., 2000; Schieber, 2007). The documented siliciclastic microbial mat record extends back 2.9 Ga (Noffke et al., 2006). While stromatolites and microbial laminites abound in Precambrian carbonates, mat features in their siliciclastic counterparts are far more subtle and provide challenges in their recognition. Although laboratory experiment suggests microbial mat growth needs several weeks of non-burial (Gerdes and Klenke, 2003, 2007), thin mat does develop on siliciclastic beaches or tidal flats within ebb intervals of 8 to 12 hours (Eriksson et al., 2009a).

Microbial mats are tough leathery clusters of micro-organisms embedded in extracellular polymeric substances (EPS) that stabilise clastic sediment surfaces (Schieber et al., 2007). The cohesiveness imparted to the sediment grains reduces erodibility of sediment, even that of granular sand (Schieber, 1998, 1999, 2004; Gehling, 1999; Pflüger, 1999; Gerdes et al., 2000; Noffke et al., 2001, 2002, 2003; Sarkar et al., 2004, 2005, 2006, 2008). A relatively high degree of preservation of bedforms, including those with little preservation potential, is thus a distinctive feature of shallow marine Proterozoic siliciclastic formations (Sarkar et al., 2008). More significant distinction between Proterozoic and Phanerozoic siliciclastic formations, in terms of sequence-building, is also likely to arise from this mat-induced reduced erodibility of sediment. Consequent reduction of sediment budget and the low gradient of epeiric sea floors of the Proterozoic could have encouraged vertical stacking of highstand systems tracts, restricting the intervening records of transgressions merely to transgressive lags (Sarkar et al., 2005, 2008; Catuneanu and Eriksson, 2007).

Depending on the role microbial mats played in their formation, the resultant sedimentary structures can be grouped into three categories:

1) Mat layer or mat ground: structures that represent the extensive mat grounds themselves, either preserved intact or deformed;
2) Mat-induced: structures that owe their origin to microbial mats; without presence of mats they are not expected to form in the first place;
3) Mat-protected: for these structures microbial mat is necessary not for generation but for preservation. In other words, the formative process of these structures is not

dependent on microbial mat growth, but being too delicate, their preservation needs stabilization, either through microbial mat growth or by cementation that is generally delayed in siliciclastics (Sarkar et al., 2008).

A genetic classification of mat-related structures would perceivably be more meaningful, notwithstanding the fact that similar features often draw diverse genetic interpretations. Among the three main genetic schemes in existence, Gerdes et al. (2000) on the basis of their work on modern siliciclastic coastlines recognise six genetic factors: (i) intrinsic biofactors; (ii) biological response to physical disturbances; (iii) trapping and binding of detritus; (iv) mechanical deformation of biologically stabilized sediment surface; (v) post-burial process; (vi) bioturbation. Maintaining the same approach, but fitting them into the list of primary (physical) sedimentary structures of Pettijohn and Potter (1964), Noffke et al. (2001) grouped microbial mat-induced structures (MISS) into two classes, (A) on bedding planes, and (B) within beds; both are divided into several subclasses:

Class A:

a) levelled depositional;
b) microbial mat chips;
c) erosional remnants and pockets;
d) multidirectional or palimpsest ripples;
e) mat curls and shrinkage cracks.

Class B:

a) sponge pore fabrics, gas domes, fenestrae structures;
b) sinoidal laminae;
c) oriented grains, benthic ooids;

d) biolaminites, mat-layer-bound grain sizes.

In a more insightful genetic scheme, Schieber (2004) classified microbial mat-related structures on the basis of the processes involved, such as growth, metabolism, destruction, decay, and diagenesis. His scheme appreciates the difference in the nature of proxy records of past proliferation of microbial mats between sandstones and mudstones (Schieber et al., 2007). A glimpse into the spectrum of Precambrian microbial mat-related structures in sandstones and mudstones is exemplified by the Neoproterozoic Sonia Sandstone, belonging to the Jodhpur Group of the Marwar Supergroup, western India, and by the Palaeo- to Neoproterozoic Vindhyan shales in central India (Fig. 1).

Microbial mat-related structures in sandstone are exemplified here from the Neoproterozoic Sonia Sandstone to immediate north of Jodhpur City, Rajasthan (Fig. 1B) not only because the array is wide, but also because it includes rare examples of mat ground preserved intact (Fig. 12). The Sonia Sandstone Formation, almost entirely arenitic, rests on a rhyolitic basement (Pareek, 1984; Chauhan, 1999; Sarkar et al., 2005, 2008). U-Pb dates for the youngest detrital zircon population of the Jodhpur Group which includes the Sonia Sandstone at its base, fixes the maximum age of the Formation at c. 800 Ma (Malone et al., 2008). The 150m thick Sonia Sandstone is topped by an unconformity, and all the mat-related structures described so far from it, occur within its uppermost 60m thick coastal marine interval (Chauhan, 1999; Chauhan and Ram, 1999; Sarkar et al., 2005, 2008).

Reddish iron-encrusted veneers on bed surfaces that incorporate platy minerals in random or imbricated orientation, and occasional pyrite enrichment in subjacent carbonaceous laminae, support past proliferation of microbial mats (Fig. 13). The pyrite may point to anoxic conditions developed beneath the inferred mats by sulphur-reducing bacteria (Sarkar et al., 2008 and references therein). The mat grounds include many features: some disc-like forms with non-erosional bases and corrugated tops, marked by radiating chains of spindles, as if stromatolites grew horizontally (Fig. 12). Extensive mat grounds also include: crumpled sheets bearing minute drag folds (Fig. 14), and wrinkle structures (cf., Hagadorn and Bottjer, 1997; Sarkar et al., 2008). The observed variation in mat-related structures is widest for the mat-induced category, and the most common elements within this category are cracks and ridges in varied patterns and in varied relationships with primary, physically-formed structures. Considering the well sorted nature of the sandstone, the cracks evince unexpected cohesiveness that is most readily attributable to the EPS associated with microbial mats (cf., Parizot et al., 2005; Pflüger and Greese, 1996). The variation also includes small bulges possibly reflecting trapping of fluid underneath the mats, and small craters surrounded by radiating cracks, inferred to have been created by fluid piercing through the mats (cf., Gerdes et al., 1993; Pflüger, 1999; Sarkar et al., 2004, 2006, 2008). The mat-protected category includes setulfs (Friedman and Sanders, 1974; Bottjer and Hagadorn, 2007; Sarkar et al., 2008), patchy (Noffke et al., 2001; Bouougri and Porada, 2002; Sarkar et al., 2004, 2008) and mat-protected ripples (Sarkar et al., 2006, 2008), preserved intact despite being subjected to subsequent reworking by high energy flows.

Microbial mats in the Sonia Sandstone did not remain confined to high littoral-supralittoral palaeoenvironments like most of their modern counterparts, but also extended to the shallow neritic zone. Some of the mat-related structures show a palaeoenvironmental preference in occurrence (Fig. 2). For example, mat grounds retaining intact primary growth structures show distinct preference for the transition between the high littoral and supralittoral palaeoenvironments. It is possible that beach rock formation helped their retention. In contrast, wrinkle structure shows little palaeoenvironmental preference, although Banerjee and

Jeevankumar (2005) described the occurrence of wrinkle structures exclusively from a neritic palaeoenvironment within the Palaeoproterozoic Koldaha Shale, central India. The only example of crumpled sheets in the Sonia Sandstone, however, belongs to the high littoral palaeoenvironment; strong flow shear presumably created the drag folds. Bulges, mat-protected ripples and patchy ripples are straddled across the high littoral and supralittoral palaeoenvironmental transition. Mat-protected setulfs were found in numbers, but exclusively within sedimentary rocks ascribed to the low supralittoral palaeoenvironment. Cracks and ridges, apparently polygenetic, are distributed almost all through the studied marine succession.

Microbial mat features are comparatively more subtle in shales because of the more compactible nature of the sediment (Schieber, 2007) and examples have been drawn here from two shales, viz. the Rampur Shale and the Bijoygarh Shale, belonging to the Vindhyan Supergroup, exposed around Rampur and Amjhor (Fig. 1; Schieber et al., 2007). The former shale was deposited at the beginning of the Mesoproterozoic period, and the latter at the early part of the Neoproterozoic in a deep offshore setting where sedimentation had been episodic. Both the shales are generally dark in colour, but are striped with shade variations and have significantly elevated organic carbon contents. The most conspicuous feature in them is carbonaceous laminae whose wavy-crinkly nature relates them to microbial mats and distinguishes them from planar laminae in non-mat carbonaceous shales. Swarms of the wavy-crinkly carbonaceous laminae characteristically give rise to an anastomosing pattern; individually they may have frayed edges and together they may give rise to false cross-strata. Another significant feature is abundant occurrence of minute shale fragments that are readily outlined in lamina-parallel polished sections and bed-normal thin sections. Some of the fragments bear disparately coarser detrital particles evincing rafting by mat fragments (cf., Fagerstrom, 1967; Olsen et al., 1978; Schieber, 1999). Some carbonaceous laminae in the Bijoygarh Shale reveal early diagenetic pyrite concentrations in an anastomosing pattern that are readily attributable to mat decay. In the same association there are clusters of phosphatic spheroids that fill in and accrete around spherical structures of a few microns diameter. These spheroids resemble those reported from the Neoproterozoic of China that apparently formed by phosphate precipitation on bacterial cell walls (Xiao and Knoll, 1999).

Widespread microbial mat growth on Precambrian sediment surfaces, whether carbonate or siliciclastic, is widely accepted (e.g., Schieber et al., 2007; Sarkar et al., 2008). Lack of bioagronomy must thus have led to general intrastratal preservation of organic carbon. Shallow water sediments of Precambrian age, both carbonates and siliciclastics, therefore had a hydrocarbon source potential almost as good as that of deep water anoxic sediments. Prolific growth of microbial mats required a slow rate of net sedimentation, and widespread epeiric seas during the Proterozoic had been especially conducive to such conditions. Development of transgressive systems tracts was in consequence severely impaired and vertical stacking of highstand systems tracts may have been preferred in consequence (Sarkar et al., 2005; Catuneanu and Eriksson, 2007).

3. Carbonate depositional systems

3.1. Introduction

Carbonates have been a feature of the sedimentary record since early Archaean times, and have been associated typically with microbialites and allochems such as stromatoclasts, ooids and peloids. Carbonates became increasingly widespread and abundant through the Neoarchaean and into the Palaeoproterozoic, through the Mesoproterozoic and peaked in the Neoproterozoic, followed by a sharp decline of dolostone through the Phanerozoic (Fig. 15), when limestone continued to increase due to the rise of shelly faunas. The dominant mineralogy of carbonates has thus changed over time, from early Archaean iron-rich dolomites, to well-ordered true dolomites from the Neoarchaean through to the Mesoproterozoic, with gradually increasing proportions of limestone throughout that time.

An overwhelming preponderance of abiological silica deposition over carbonate rocks existed from the Archaean until the mid-Mesoproterozoic, but declined thereafter and became insignificant by the Neoproterozoic (Fig. 15). Most explanations for this involve the geotectonic setting of deposits within active plate margins (greenstone belts), which led to prevalence of hydrothermal, silica-rich fluids and to deposition of cherts. In contrast, the carbonates are usually interpreted as passive plate margin deposits, and here silica domination led to syn- and post-depositional silicification of clastic and carbonate sediments, often with relict minerals and replacement structures of former carbonates (Lowe, 1999). However Palaeoproterozoic cherts associated with iron formations (e.g., the 1.88 Ga Gunflint Formation, northwestern Lake Superior region) often appear to have formed largely by direct silica precipitation at or just below the seabed, and lack ghosts or inclusions of carbonate precursors. They may provide evidence that global oceanic silica concentrations were higher during the Palaeoproterozoic era than at later times. In Mesoproterozoic and Neoproterozoic strata, early diagenetic silicification was largely restricted to peritidal environments, with cherts typically occurring as nodules or discontinuous beds within carbonate deposits that have similar depositional textures. These cherts formed primarily by carbonate replacement with subsidiary primary silica precipitation

(Maliva et al., 2005). Younger Precambrian active plate margin sedimentary rocks do not display such a dominance of chert, supposedly because of lower geothermal gradients and weaker hydrothermal activity. Another factor however was the high proportion of CO_2 in the earlier Precambrian atmosphere. This would have produced acidic oceans, favouring silica precipitation; carbonates would have been precipitated only where alkalinity was raised, e.g. in the ambient waters mediated by the metabolic processes of microbialite-building benthic microbial communities. The pattern of carbonate abundance was mirrored to a large extent by an increase in stromatolite diversity through most of the Precambrian (cf., Riding, 2006; Awramik and Sprinkle, 1999), perhaps revealing an early, strong microbial link with carbonate precipitation.

The main locus of carbonate precipitation in the Early Archaean was in association with hydrothermal systems (e.g., the 3.5 Ga Dresser Formation of the Warrawoona Group, Western Australia). A sudden rise in carbonate accumulation at 3.0 Ga could have involved the emergence of cyanobacteria able to calcify and to construct stromatolitic reefs in varying environments (Altermann et al., 2006). The oldest preserved stromatolitic carbonate platform is the 3.0 Ga White Mfolozi Formation of the Pongola Supergroup of South Africa (Altermann et al., 2006). It has been argued that the emergence of stable continental shelves permitted the accumulation of thick carbonate deposits over time (Grotzinger, 1989, 1994). By the Neoarchaean, thick shallow-water shelfal sequences were deposited on a number of cratons (e.g., Fennoscandian, Superior, Kaapvaal, São Francisco, Amazon, Zimbabwe, Dharwar and Congo). Carbonate platforms in which most of the essential Phanerozoic features can be recognized were well developed by the Neoarchaean/Palaeoproterozoic, including ramps and rimmed shelves (e.g., Grotzinger, 1989). The Archaean-Proterozoic boundary represents a major episodic change in the growth of the continental crust – and also a jump in the number of stromatolite taxa

(Awramik and Sprinkle, 1999). The appearance of widespread, kilometres-thick, stromatolitic carbonate platforms such as the exceptionally well preserved 2650-2500 Ma Campbellrand-Malmani platform of Kaapvaal, South Africa, and the mechanisms of biocalcification and carbonate precipitation are highly debated, and are discussed below.

In the Precambrian, stromatolites accreted in basinal to shallow subtidal and supratidal environments, but it is in reefs (or biostromes) that they achieved their greatest morphological diversity, with a wide range of growth forms, including branched, columnar, domal and planar, often interbedded with oolites, peloids and intraformational breccias. The number of recognized columnar stromatolite forms rises to a maximum in the Mesoproterozoic, before dropping sharply in the Neoproterozoic, with relatively few in the Cambrian (e.g., Tucker et al., 1990). An interplay of biological and environmental factors produced widespread, diverse and abundant forms that have varied over time, with particular forms restricted to certain time intervals. As such, they have been used for stratigraphic purposes, though the environmental control on morphology renders a binomial taxonomy untenable; instead microstructural variation has been used to apply biostratigraphic markers, though with similar constraints. However, stromatolites can be used with some confidence as environmental indicators (e.g., Altermann, 2008).

Despite their abundance, mechanisms of non-skeletal carbonate precipitation are poorly understood. The contrast between carbonate platforms dominated by microbialitic deposits in the Precambrian and those dominated by algal and invertebrate remains in the Phanerozoic records a significant change in the major processes of carbonate production and sediment accumulation. However, many questions remain: (1) were Precambrian carbonates largely mediated by microbial metabolic processes?; (2) are Precambrian stromatolites witnesses to the early evolution of life, preserving fossil microbial remains as old as ~3.5 Ga?; (3) or might they simply have been chemical precipitates from carbonate oversaturated oceans?; (4) do modern stromatolites provide a process analogue for carbonate precipitation in the Precambrian? Detailed field-based studies of ancient carbonates from around the world provide some answers, as detailed in the next section.

3.2. Carbonates: genetic patterns through early Precambrian time

In the Early Archaean, marine carbonates were deposited as minor components of greenstone belts during periods of local tectonic stability. Although relatively scarce, these rocks are important because they record chemical and isotopic informaton about the composition of contemporary oceans. The earliest examples date from ~3.5 Ga and include the ~3.5 Ga Warrawoona Group of the Pilbara region, Western Australia, and the ~3.3 Ga Barberton Greeenstone Belt of the Swaziland Supergroup of South Africa, with stromatolites present in both. The Barberton stromatolites (Byerly et al., 1986), which are developed in thin cherty units interbedded with komatiitic lavas in the uppermost part of the Onverwacht Group, and stromatolites of the Warrawoona Group, both feature small, domal forms. Massive black and banded carbonaceous cherts, and carbonaceous laminites, are also common in early Archaean greenstone belt volcanic sequences.

The 3.5 Ga Dresser Formation is part of the mostly volcanic Warrawoona Group in the North Pole Dome area of the Pilbara craton. It comprises interbedded chert and baryte units, carbonates and pillow basalt. Silicified during the late Eocene to Oligocene, the surface outcrops of layered chert actually represent the altered equivalents of bedded carbonate rocks at depth (van Kranendonk et al., 2006). The carbonates, primarily ferroan dolomites, were deposited both within a volcanic caldera that was dominated by hydrothermal processes and continued into shallow water and intermittently exposed conditions with bedded carbonates, often with ripple forms (Fig. 16) where stromatolites flourished. These include coniform, domical and stratiform varieties with wrinkly laminations. At the North Pole locality, wrinkly stratiform stromatolites pass laterally into smooth domical forms (Fig. 17). Black- and red-weathering surface outcrops of wrinkly stromatolitic laminates are the surface-altered equivalents of laminated pyrite at depth, as seen in fresh drillcore: textural evidence shows that the pyrite replaced sedimentary carbonate (van Kranendonk et al., 2008). Ueno et al. (2008) found quadruple sulphur isotope evidence for sulphate reducing bacteria (SRB) in cherts of the Dresser Formation.

The 3.43 Ga Strelley Pool Chert (SPC) is widespread in the upper part of the Warrawoona Group and ranges from 15 to 30m in thickness. It consists of grey or white, iron-rich, dolomitic carbonate or silicified carbonate interbedded with units of quartz, possibly after barite, with primary chert units. The carbonates can be traced laterally into the silicified intervals providing clear evidence of replacement. The carbonate and silicified carbonate intervals are generally laminated. Stromatolitic structures are found at well defined intervals of 1-2 m thickness in carbonate and silicified carbonates lower in the formation.

A different stromatolite morphology is developed in the SPC, where small crested domes and conical stromatolites are typical. At the Shaw River (Trendall locality), stromatolites comprise coniform columns and bumps (‘egg-boxes’) in stratiform, generally isopachous stromatolitic layers, deposited under shallow, low-energy, subaqueous conditions, with occasional evidence of erosion around the base of some steeper domes. The angular, coniform structures at the Trendall locality show greater uniformity of laminae when compared with inter-stromatolite areas, indicating that the latter were subjected to more variable depositional conditions. Steep-sided cones show slopes that are higher than the angle of sediment repose, while tracing the structures along strike shows that they grade into domal and undulatory forms with irregular spacings, both vertically and laterally, arguing against slumping or lateral compression as a cause. Undulatory, domical and coniform laminae often show considerable vertical continuity (Fig. 18) within specific horizons, indicating irregular growth consistent with competitive growth under slightly varying environmental and biochemical conditions. These combined factors favour a biogenic rather than a physico-chemical origin, although the biogenicity of Archaean stromatolites has been challenged in favour of a hydrothermal or diagenetic origin (e.g., Lowe, 1994; Grotzinger and Rothmans, 1996; Brasier et al., 2004).

Allwood et al. (2006) investigated stromatolite morphologies in the SPC in the context of their palaeoenvironmental setting, across more than 10 km of relatively continuous outcrop where the formation is best preserved. They argued that the SPC can best be interpreted as a fossil microbial carbonate platform based on a number of criteria: the inferred palaeoenvironmental setting, the combined attributes of the stromatolites, stromatolite distribution within the palaeoenvironment, and their similarity to younger, microbially mediated peritidal carbonates. They undertook the first morphotype-specific analysis of the structures within their palaeoenvironment and refute abiogenic hypotheses for their formation, arguing that the diversity, complexity and environmental associations of the stromatolites describe patterns that compare with similar settings throughout Earth's history that reflect the presence of organisms.

Partially silicified dolomitic stromatolites occur in the ~3.0 Ga White Mfolozi section of the Pongola Supergroup, South Africa. Composite columnar stromatolitic bioherms 0.7 - 1.6 m high and 0.4 - 1.0 m in diameter formed along the margins of a tidal channel, with flat stratiform and small domical stromatolites growing in low energy tidal flat environments (Altermann, 2008). Conical stromatolites accreted in high-energy, coarse-grained carbonate sand along the bottom of the tidal channel, and have been linked to the activities of filamentous, oxygen-producing, photoautotrophic cyanobacteria (e.g., Beukes and Lowe, 2006).

Greenstones containing carbonates at ~2.8 Ga include the Abitibi, Yellowknife, Wabigoon (Steep Rock Lake), Michipicoten and Uchi belts of Canada and the "Upper Greenstones" of Zimbabwe. The Zimbabwean Belingwe Greenstone Belt, 2.7 Ga, is one of the best preserved and least deformed of all Archaean greenstone successions. Carbon and sulphur stable isotope data from Belingwe stromatolitic carbonates indicate that rubisco-based oxygenic photosynthesis was operating by at least 2.7 Ga, while data from shales indicate a strong methanogenic signal (Grassineau et al., 2002).

The ~2.8 Ga Steep Rock Group of northwest Ontario is one of the world's thickest Archaean carbonate platform successions (Kusky and Hudleston, 1999). Deposited unconformably over the Marmion Complex, a 3001-2928 Ma gneissic terrane, it comprises biogenic and oolitic limestones, dolostones, micrites, and karst breccias capped by a thick palaeosol developed between and over karst towers. The presence of aragonite fans, herringbone calcite, and rare gypsum molds points to the carbonate platform having experienced hypersaline depositional conditions, and suggests that solid sulphates were able to form, possibly replaced by carbonate on a large scale (Gandin et al., 2005; Gandin and Wright, 2007).

Stromatolites of the ~2.7 Ga Beechy Lake Group of the Yellowknife Supergroup form lenses within volcanic breccias at the margins of felsic domes. Bioherms of low, wavy-laminated dolomitic mounds up to 2 m thick extend laterally for hundreds of metres, characteristically forming thin encrustations on breccia blocks. These units typically contain millimetre-scale layers of fine volcanic ash at regular intervals, testifying to periodic explosive eruptions during the development of isolated microbial communities around areas of fumarolic (or hydrothermal) activity near active volcanic domes along the shallow flanks of an emergent stratovolcano (Lambert, 1998).

The Archaean Kaapvaal craton, one of the best-preserved, includes the Campbellrand-Malmani platformal carbonates, which persisted for some 80 my, from 2588±6 Ma to at least 2516±4 Ma (Altermann and Nelson, 1998). Similarly extensive and well-preserved is the 2.6 to 2.5 Ga old Carrawine Formation of the Pilbara craton (Nelson et al., 1999). The deposition of the Campbellrand Subgroup occurred on a shallow marine platform, and comprises basinal non-stromatolitic, laminated carbonate and shale with minor chert, iron formation and mafic tuff beds, with abundant shallow water stromatolitic carbonates, some of which may be replacive after evaporites (Gandin et al., 2005; Gandin and Wright, 2007).

3.3. Relationships between carbonate rocks, carbonate mineralogy, seawater chemistry, microbial mediation and physico-chemical precipitation

The significant differences between Precambrian and Phanerozoic carbonates have been attributed in large part to ‘controlled biomineralization’ which was either unknown or uncommon prior to the Phanerozoic (e.g., Lowenstam, 1981; Wright and Oren, 2005). Precambrian carbonates generally comprise a range of microbialites, particularly stromatolites with other facies, such as oolitic and peloidal grainstones, and although conventional interpretations link their genesis to chemical precipitation directly from seawater, their abundance and diversity together with the lack of unequivocal evidence for large-scale marine carbonate precipitation raises the question of a genetic association between carbonates and microbially-mediated processes. In particular, the widespread belief that "with higher carbon dioxide levels, Precambrian seawater was more supersaturated with respect to $CaCO_3$ than Phanerozoic seawater and carbonates were precipitated easily" (e.g., Tucker, 1992) does not fit with the observation that higher CO_2 levels increase ocean acidity and make it more difficult for some organisms to secrete calcium carbonate. Similarly, speculation that warming of the oceans and enhanced biological uptake at that time drew down oceanic CO_2 so that alkalinity levels in the ocean were sufficient to counteract dissolution (Karhu and Holland, 1996) conflict with observations that increased CO_2 leads to lower pH and therefore an increase in oceanic acidity that inhibits $CaCO_3$ precipitation (e.g., Andersson et al., 2003; Wright and Oren, 2005).

Seawater saturation state is often cited as the principal factor determining carbonate precipitation during the Precambrian (e.g., Grotzinger, 1994; Sumner, 1997; Sumner and Grotzinger, 2000). In modern oceans, the minimum thermodynamical requirement for calcite to precipitate is exceeded by a factor of almost 5 (> 3 for aragonite) (Ridgewell and Zeebe, 2005). Yet despite supersaturation of surface sea water with calcite, aragonite and dolomite, spontaneous inorganic precipitation of these phases from sea water does not normally follow (e.g., Leeder, 1982; Wright, 2000; Morse et al., 2003; Wright and Oren, 2005), while experiments have shown that homogeneous nucleation does not occur in sea water solutions until saturation levels of calcite reach > ~20 - 25 (Morse and He, 1993).

However, these are not the only factors operating: kinetic factors can strongly affect the behaviour of ions in saline solution, including the formation of ion pairs or complexes, and hydration shells (e.g., Slaughter and Hill, 1991; Wright and Oren, 2005). In saline solutions, ion pairs form due to short-range interactions of adjacent ions, attracted by coulombic forces. This complexing reduces the ions' activities below their modalities, making precipitation of carbonate minerals unlikely. For example more than 90% of total CO_3^{2-} is complexed with hydrated metal cations, chiefly magnesium (Garrels and Christ, 1965; Wright, 2000; Wright and Oren, 2005). The presence of other ions in solution thus shields Ca^{2+} and CO_3^{2-} ions from interacting and precipitating.

Although carbonate precipitation occurs as cements and coatings in the modern marine environment, it is primarily under direct metabolic control, associated with the activities of metazoa. No microbes have been found to be "obligate calcifiers" (Riding, 1982; Riding and Liang, 2005). So we should ask: in the absence of metazoans, how can the kinetic inhibitors to marine carbonate production be overcome? There is abundant evidence that these kinetic inhibitors can be overcome on a large scale through microbial mediation (e.g., Nadson, 1928; Slaughter and Hill, 1991; Robbins and Blackwelder, 1992; Robbins et al., 1996; Warthmann et al., 2000; Wright, 2000; Wright and Altermann, 2000; Wright and Wacey, 2004, 2005; Wright and Oren, 2005). Numerous different bacterial species have previously been detected and assumed to be associated with natural carbonate precipitates from diverse environments. The primary role of bacteria in the precipitation process has subsequently been ascribed to their ability to create an alkaline environment (high pH and [DIC] increase) through various metabolic processes (Castanier et al., 1999; Douglas and Beveridge, 1998).

Late Archaean and Palaeoproterozoic strata are suggested to record extreme marine supersaturation maintained by the inhibiting effect of Fe^{2+} on calcite nucleation and growth (Sumner and Grotzinger, 1996), leading to the "*in situ*" precipitation of widespread aragonite botryoids and crystal "fans" as well as calcite (Sumner and Grotzinger, 2000). But these aragonite botryoids and crystal fans, up to 50 cm tall in the ~2.6 Ga Campbellrand-Malmani carbonates of the Transvaal basin of South Africa, accreted while expanding upwards from (carbonate) bedding surfaces and stromatolites – if this was an inorganic process unassociated with microbial processes, why did these aragonite deposits not appear on the surfaces of different facies and lithologies? Why do the observed features occur only in carbonate? Similarly, Simonson et al. (1993) considered that crystal fans in the coeval carbonates of the Carawine Dolomite of the Hamersley Basin, Australia, once possibly joined with the Transvaal basin, were originally aragonite, but found it "worrisome" that the crystal morphologies lacked the "distinctive square-tipped terminations" of radiating aragonite crystals.

Hardie (2003) and Gandin et al. (2005) have argued convincingly that the fans (Fig. 19) were not originally aragonite, but gypsum. Gypsum domes of the Upper Miocene Solfifera Series of Sicily (Hardie and Eugster, 1971), are “astonishingly similar” to the crystal fans, indicating that a sulphate precursor is entirely possible (Hardie, 2003). In the Gamaohaan Formation of South Africa, certain features and textures, including carbonate fans and “herringbone calcite”, indicate the former presence of evaporites, now replaced by carbonate, and that these evaporites were thick at times and also occurred at particular, laterally extensive horizons. Gandin et al. (2005) and Gandin and Wright (2007) argue that calcitization of the vanished but once laterally-extensive evaporites was driven by bacterial sulphate reduction of solid sulphate in association with organic diagenesis and pyrite precipitation within platform-wide microbialites and sapropels. Large-scale microbial mediation of ambient waters across a shallow to emergent platform is thought to have raised carbonate alkalinity and removed kinetic inhibitors to carbonate formation.

Shallow-water and sabkha evaporites previously formed in settings much larger than any found today, in vast expanses of evaporitic lagoons and mudflats reaching > 100,000’s km^2 in extent, over which brine depths were at most a few metres. Examples include the 90 m thick Lower Cretaceous Ferry Lake Anhydrite deposited in a shallow lagoon 260 km wide (Loucks and Longman, 1982), and the 100 m thick, Permian San Andres Formation (Palo Duro Basin) in which individual anhydrite beds extend some 26,000 km^2 with only minor changes in thickness and facies (Fracasso and Hovorka, 1986).

The low preservation potential of Precambrian solid sulphate can be related in part to bacterial sulphate reduction within the microbially-dominated ecosystems. Evidence for the former presence of solid sulphate in shallow Neoarchaean seas includes pseudomorphs after selenite, also recorded from the contemporaneous Carawine Dolomite of Australia (Winhusen, 2001), together with rock fabrics and textures typical of evaporite dissolution. Importantly, sulphur isotopes of pyrite samples from the Cambellrand carbonates show a wide range of values indicating biogenic fractionation of sulphate, a signature also seen in the Neoarchaean Belingwe Greenstone Belt of Zimbabwe (Grassineau et al., 2001), and the Mount McRae and Jeerinah shales of Western Australia (e.g., Kakegawa et al., 1998).

3.4. Chemical versus microbial genesis

The evidence leads us inevitably to a discussion of the role and scale of microbial processes in taphonomic evolution and carbonate production through time, and of carbonate precipitation itself: there is a clear distinction between overwhelmingly biogenic carbonate precipitation throughout the Phanerozoic, and unresolved processes of carbonate precipitation in the Precambrian. In other words, despite supersaturation of surface sea water with calcite, aragonite and dolomite, inorganic precipitation of these phases from the water column does not normally follow (e.g., Leeder, 1982).

Although studies of the early evolution of life and its contribution to the Archaean sedimentary record suffer from a scarcity of preserved microbial remains (Altermann, 2004), there is abundant evidence that the first several billion years of life on Earth was microbial, and Precambrian carbonates are well-known for organo-sedimentary structures such as stromatolites. Proterozoic stromatolites have yielded well-preserved, mostly silicified microfossils that strongly resemble present-day cyanobacteria (e.g., Bitter Springs [Australia], Chickan, Draken, Gunflint and Sukhaya Tunguska formations). The range and combinations of features of fossil stromatolites are often found in living stromatolites, and are difficult, if not impossible, to explain by inorganic processes. Morphology also remains a valid criterion to indicate biogenicity (e.g., Allwood et al. 2006).

In the Precambrian and for much of the Phanerozoic, dolomite [$CaMg(CO_3)_2$] was typically more abundant than limestone - an observation generally known as "the dolomite problem." It remains a source of controversy in sedimentary geology (McKenzie, 1991; Vasconcelos and McKenzie, 1997, 2000; Wright, 1997, 2000; Burns et al., 2000; Wright and Altermann, 2000). A number of kinetic barriers to dolomite precipitation have been shown to operate in the marine environment: (1) the disproportionate distribution of the component ions of dolomite; (2) the low concentration and even lower activity of the CO_3^{2-} ion; (3) the high enthalpy of hydration of the Mg^{2+} and Ca^{2+} ions; (4) the presence of SO_4^{2-} ions and the formation of ion pairs. Evidence has been accumulating that SRB are directly involved both in the initial formation and in the diagenetic development of dolomite. SRB were found associated with dolomite concretions, while other types of bacteria as well as archaea dominated in the surrounding sediment. Lipid biomarkers and 16S rRNA sequences characteristic of SRB were found in samples from 15 and 40 cm below the sediment/water interface, depths corresponding with ages of about 500 and 2,000 years (Mauclaire et al., 2002). Dolomite has been produced in low temperature laboratory experiments using SRB from modern dolomitic sediments (e.g., Vasconcelos et al., 1995; Warthmann et al., 2000; Wright and Wacey, 2004, 2005). The key to precipitation of sedimentary, low-temperature dolomite is the removal of the kinetic barriers by microbial mediation. Bacterial sulphate reduction (BSR) can thus provide a "process analogue" for the formation of dolomite through biosphere–hydrosphere–lithosphere interactions, wherever such conditions prevailed in the past. However, it is not always possible to demonstrate a direct link between BSR and calcite precipitation: in laboratory modelling experiments, using a medium mimicking presumed Precambrian seawater chemistry in the presence of the *Desulfovibrio desulfuricans*, dead or metabolically inactive *Desulfovibrio* cultures apparently stimulated calcite formation more than active cultures (Bosnak and Newman, 2003).

BSR was much more prevalent in the Precambrian, when microbial communities dominated the environment, generating the potential for major changes in ambient water chemistry in ecosystems across extensive epeiric seas and marine shelves (Wright, 2000; Wright and Altermann, 2000; Altermann et al., 2006; Eriksson et al., 2009a). Shen and Buick (2004) provide evidence from $\delta^{34}S$ values of microscopic pyrites in former gypsum crystals in the ~3.47 Ga North Pole barite deposit of northwestern Australia for the oldest example of microbial sulphate reduction and the earliest indication of a specific microbial metabolism.

Thin section studies show that sedimentary carbonate was the earliest mineral precipitate at several different levels in the Dresser Formation, while trace element analysis of carbonate from the lower part of the Dresser Formation indicates ankerite precipitation from seawater under anoxic conditions (Garcia-Ruiz et al., 2003; Van Kranendonk et al., 2003). Foriel et al. (2004) used fluid inclusions from quartz in lava escape tubes of overlying pillow basalts to show a strong similarity to modern seawater, apart from a higher concentration of dissolved salts. Does this mean that the carbonate was a direct seawater precipitate? This is a debatable subject. Large volumes of sulphate were present in the Dresser Formation, as evidenced by the presence of primary barite, with SRB present (Shen and Buick, 2004), indicating strongly that carbonate precipitation occurred alongside microbial mediation of ambient waters.

Mass microbial colonisation across extensive Neoarchaean epeiric seas witnessed the microbiogeochemical transformation of the Earth's hydrosphere, atmosphere and biosphere. The consequences for a reducing ocean would have been the progressive oxidation of the major dissolved species in surface seawater, most notably of reduced sulphur and iron. Cyanobacterial photosynthetic oxidation of surface seawater drove formation of aqueous sulphate and permitted the precipitation of extensive evaporites in restricted basins. The first dramatic explosion of carbonate precipitation can be related to intense bacterial sulphate reduction in association with anoxic organic diagenesis and pyrite formation within the decaying interiors of microbialites and in sapropels.

It has been argued that stromatolites were originally formed largely through *in situ* precipitation of laminae during Archaean and older Proterozoic times, but that younger Proterozoic stromatolites grew largely through the accretion of carbonate sediments, most likely through the physical process of microbial trapping and binding (Grotzinger and Knoll, 1999). Historically, early mineralisation of microbialites has been attributed to abiotic submarine cementation, or to calcification of cyanobacterial sheaths induced by photosynthesis (e.g., Logan, 1961; Monty, 1976; Dill et al., 1986). However, the kinetic barriers to dolomite formation also apply to a lesser extent to abiotic calcium carbonate precipitation, and these can similarly be overcome by microbial mediation (e.g., Wright and Oren, 2005 and references therein). Recent work has shown that bacterial sulphate reduction is associated with *in situ* carbonate precipitation in many modern settings (e.g., Canfield and Raiswell, 1991; Hendry, 1993; Reid et al., 2000; Walter et al., 1993; Visscher et al., 1998, 2000; Wright, 1999, 2000; van Lith et al., 2003; Wright and Wacey, 2004, 2005).

The link between bacterial sulphate reduction and calcification has been investigated in modern marine stromatolites of Exuma Sound, Bahamas. Visscher and co-workers showed a direct link between sulphate reduction and aragonite precipitation, with cumulative lithified laminae formed by microbial activity near the sediment surface (Reid et al., 1995, 2000; Golubic and Browne, 1996; Macintyre et al., 1996; Feldman and Mackenzie, 1998). Field and experimental work by Reid et al. (2000) has also shown that the growth of modern stromatolites in the Bahamas is achieved through a dynamic balance between *in situ* carbonate precipitation associated with SRB, and sedimentation by trapping and binding of grains, each process being closely related to a different microbial community depending on environmental dynamics.

In situ precipitation was shown to occur in continuous surface layers of exopolymer secreted by cyanobacteria, associated with both aerobic and anaerobic microbial consortia. SRB are an important component of the surface layer, accounting for up to 40% of carbon consumption by the community, despite the presence of oxygen at the surface. Distinct bands of SRB activity were found, located exactly at the lithified micritic horizons. High rates of sulphate reduction associated with degradation of exopolymers coincided with aragonite precipitation in the stromatolitic surface crusts, while radiolabelled organic matter revealed elemental exchange between bacteria and aragonite needles. Carbonate precipitation associated with exopolymers may also explain the rarity of microbial fossils in ancient microbialites.

The association between cyanobacteria and SRB as key components within the microbial communities responsible for the construction of the Bahamian stromatolites has huge implications for the understanding of carbonate precipitation not only in ancient microbialites, but also for the accumulation of thick platformal carbonates and for non-skeletal sedimentary carbonate precipitation throughout the geologic record.

3.5. Molar tooth structure

One feature distinctive of Precambrian carbonates, particularly those of shelf origin, is molar tooth structure (MTS), a crack-fill cement of unique character. It is almost exclusive to Mesoproterozoic and early Neoproterozoic times (900-600Ma), preferably the latter (Meng and Ge, 2002). The cement mass is unzoned, being constituted by uniform-sized sucrosic carbonate crystals, calcite or dolomite, generally sheet-like in three-dimensional geometry, conspicuous when subvertical in attitude, wispy in cross-section, often ptygmatically folded, and laminae in host sediment are differentially compacted around them (Fig. 20). The last two features as well as evidence for transport of fragments clearly indicate early generation of the cement. MTS can, however, be horizontal, spindle-shaped and ribbon- or blob-shaped also.

Widely different views have been proposed for the origin of MTS over the last half century, but the structure still remains enigmatic. It has been thought by some to be derived from non-carbonate precursors, such as algal structures (Smith, 1968; O'Connor, 1972; Moussine-Pouchkine and Bertrand-Sarfati, 1997) or evaporite minerals (Eby, 1977). The majority opinion, however, does not question its primary carbonate composition, considers it not as a replacement product but as crack-filling cement; these researchers, nevertheless differ widely with regard to the mechanism for crack generation. Tectonic fracturing (Daly, 1912; Cowan and James, 1992; Smith and Winston, 1997; Bishop and Sumner, 2006), earthquake induced dewatering (Fairchild et al., 1997; Pratt, 1998b), desiccation or synaeresis (Bell, 1966; Horodyski, 1976, 1983; Young and Long, 1977; Hofmann, 1985; Beukes, 1987; Knoll and Swett, 1990; Calver and Baillie, 1990; Demicco and Hardie, 1994; Liu et al., 2005) and fluid pressure (Dix and Mullins, 1987; Desrochers and Al-Aasm, 1993; Mozley and Burns, 1993; Furniss et al., 1994, 1997; James et al., 1998; Kuznetsov, 2003; Pope et al., 2003; Marshall and Anglin, 2004; Pollock et al., 2006) have all been suggested in this respect.

None of the processes mentioned above, however is exclusive to the Meso-Neoproterozoic period and the age preference of the structure remains unexplained. Shields (2002) held maximum accentuation in ionic concentration of Ca^{+2} and CO_3^{-2} in Neoproterozoic seawater to be responsible for this preference. Frank and Lyons (1998) supported a unique genetic combination of $CaCO_3$ saturation and redox conditions, in shallow marine waters in the absence of biotic agronomy. Marshall and Anglin (2004) postulated that destabilization of CO_2–clathrate by some mechanism, such as seismic shocks, could have created the MTS, and explained the preferential Proterozoic chronology by stabilization of the clathrate in the sediment column because of higher partial pressure of CO_2 in the extant atmosphere. Meng and Ge (2002) contemplated a relation between MTS and prevalent bioforms transitional to Phanerozoic multicellular forms, and suggested that related micro-bioelectromagnetic waves encouraged $CaCO_3$ precipitation within cracks induced by CO_2 overpressuring. Pollock et al. (2006) proposed intimate links between crack formation and concomitant microspar precipitation and the decomposition of sedimentary organic matter in the presence of supersaturated seawater; they produced a variety of crack morphologies within unconsolidated mud under gas pressure in the laboratory. Proliferation of sulphate reducing bacteria and methanogens in the Proterozoic time period could have promoted carbonate precipitation within cracks (e.g., Vasconcelas and McKenzie, 1997; Wright, 1999; van Lith et al., 2003; Roberts et al., 2004; Wright and Oren, 2005; Wright and Wacey, 2004, 2005).

Two generations of MTS coexisting within the Neoproterozoic Bhander Limestone of the Vindhyan Supergroup in the area around Maihar, central India and within the Mangurda Limestone of the Pranhita-Godavari Valley Basin, in the area north of Adilabad, South India (Fig. 1) provide a rare insight into MTS formation. In these limestone formations, vertical MTS cut across the horizontal MTS indicating their later generation. Differential compaction of host sediment around the vertical group, nevertheless indicates only a short time gap between the two generations, with both belonging to early diagenesis. In both limestones the preferred lithology for MTS occurrence is ribbon limestone facies characterized by frequent alternations between light coloured carbonate lutite and darker coloured very fine-grained marl laminae. Cracks are abundant at certain stratigraphic levels; with the majority of them being parallel to bedding the limestone appears almost shredded, an appearance that has been attributed to synsedimentary earthquakes (Fig. 21; Coniglio, 1986; Chakraborty, 1995). Primary laminae of the host sediment as well as the bed-parallel crack-fill MTS are differentially compacted against the vertical, relatively less common, second generation MTS.

It is apparent that these Indian examples of MTS, even those of the second generation, formed before consolidation of the host sediment. Cementation, which is generally rapid in carbonates, was relatively delayed in this case, for the host sediment. Although this fact argues against the common belief that concentration of Ca^{+2} and CO_3^{-2} ions had been high in Neoproterozoic sea water (Grotzinger and Kasting, 1993; Kaufman and Knoll, 1995; Frank and Lyons, 1998; Shields, 2002), the delayed cementation can alternatively be attributed to preponderance (possibly localised) of inhibitors to $CaCO_3$ precipitation, like Mg^{2+}, Fe^{2+}, PO_4^{3-} or SO_4^{2-} ions. Both the Bhander Limestone and the Mangurda Limestone contain pseudomorphs after gypsum (Fig. 22). The former also contains barite within second generation MTS, and is furthermore encased by red beds themselves containing pseudomorphs after gypsum and halite (Bose et al., 2001). High concentration of inhibitors like SO_4^{-2} in the concerned seawater is thus a logical conclusion. Sediment pore fluid could have become progressively enriched in salts which were more difficult to precipitate as it was squeezed out from the host sediment and entered the first generation MTS cracks, and then moved from there into the MTS cracks of the second generation. Evaporite mineral alteration from gypsum in the host sediment to barite in the second generation MTS amply supports such a postulated course of pore fluid evolution.

Similarity between the MTS of the two different generations (attitudes) is overwhelming, both being constituted by clear microspar crystals, with crystal size being comparatively more uniform in the case of the second generation MTS; boundary-parallel crystal-size zonation, i.e. drusy growth, a common characteristic of void-fill cement, is absent in both the cases. Crystals constituting the MTS are largely bright, while those in the host sediment appear to be predominantly dull under cathode luminescence. Bright crystals, nonetheless, are also found scattered within the host sediment in the immediate vicinity of the MTS, decreasing in frequency of occurrence away from them (Fig. 23) (Amieux, 1982). This outwardly diminishing halo of bright crystals around the MTS suggests that the pore fluid that traveled mostly along the cracks also permeated the sediment beyond the crack walls. Crack formation might have preceded the fluid flow or could be tied up with it, being triggered by a common mechanism, such as earthquake seiches. Preferred occurrence of MTS within the shredded part of the same ribbon limestone strongly corroborates the possibility of earthquake seiches.

The $\delta^{18}O$ values in (apparently least altered) micro-drilled samples from both first and second generation MTS as well as from the sediment hosting them within the Bhander Limestone and the Mangurda Limestone, are all moderately negative, while the corresponding $\delta^{13}C$ values are all positive. The latter tend to be comparatively lower in the host sediment than in the MTS, this contrast being relatively more profoundly expressed in the Mangurda Limestone samples (Fig. 24). $^{87}Sr/^{86}Sr$ values, on the other hand, have very limited ranges in the MTS samples, although they are comparatively lower in the Mangurda Limestone than in the Bhander Limestone. The $^{87}Sr/^{86}Sr$ ratios in the sediments that host the MTS are, however highly variable, greater variation and significantly higher values being found in the Mangurda Limestone; a variable fluid-rock ratio is inferred. Plotting of $^{87}Sr/^{86}Sr$ values of the samples against their $\delta^{13}C$ values clearly separates the MTS and the sediment that hosts them (Fig. 24). The $\delta^{18}O$ and $\delta^{13}C$ values decrease with diagenetic alteration (Choquette, 1968; Veizer et al., 1992; Kaufman and Knoll, 1995; Kah, 2000; Pope et al., 2003) and the positive values of the latter isotope, in contrast to the negative values of the former, as noted in the two studied limestones, are consistent with greater stability of $\delta^{13}C$ during diagenesis (Hudson, 1977). Comparatively higher and widely variable $^{87}Sr/^{86}Sr$ values in the host sediment further corroborate a distinctly higher degree of diagenetic alteration, because diagenetic fluid is dominantly influenced by ^{87}Rb that is derived from silicate minerals and changes into ^{87}Sr. The MTS evidently largely escaped this alteration (Fairchild et al., 2000; Shields, 2002) and their $^{87}Sr/^{86}Sr$ ratios (which are as low as c. 0.706) may reasonably be the result of accentuated hydrothermal activity (Veizer and Compston, 1976; Veizer et al., 1983;

Jacobsen and Kaufman, 1999). This differential response to diagenesis between the host sediment and the MTS could possibly be due to variation in primary mineralogy, i.e. more stable carbonate in the MTS, such as low Mg-calcite or dolomite.

A progressive increase of salinity in residual pore fluid as it passed from the host sediment to the first and then to the second generation of MTS cracks could have increasingly facilitated precipitation of carbonate and also favoured the mineral phase of lower diagenetic potential, viz. dolomite or low-Mg calcite, in lieu of aragonite and high Mg calcite. This would explain why the host sediment remained uncemented while precipitation took place within the cracks. With progressive increase in soluble salt concentration, the pore fluid could have become gel-like, with resultant precipitation therefrom becoming unzoned, and without much variation in crystal size.

Extensive development of shallow hypersaline water bodies is inferred for the epeiric seas that developed widely during the Proterozoic, especially in its later stages. It is likely that pore-fluid pressure itself created the cracks and that depressurisation in consequence of crack-opening could have facilitated carbonate precipitation as well; crack-generation and -filling thus resulted from a single mechanism, pore fluid overpressuring. Earthquake seiches, in turn, could have been responsible for inducing the pore fluid overpressuring. The combination of earthquakes and extensive development of quiet stratified epeiric sea bodies apparently encouraged MTS formation selectively in Meso- and Neoproterozoic times, but more preferably in the Neoproterozoic.

3.6. Carbonate rocks, global palaeoclimate change, and glaciation

Although glacigenic deposits are essentially clastic rather than chemical in their nature, they appear to be genetically closely linked to isotopic values determined from directly associated carbonate deposits reflecting significant changes in global climates and possibly also palaeo-atmospheric conditions; hence they are discussed here rather than in the earlier, clastic part of the paper. The placing of this section also serves to emphasize that the clastic and chemical sedimentary record of the Precambrian needs to be studied holistically. Additionally, of all the clastic deposits, those ascribed to glacigenic depositional systems show the smallest divergence from Phanerozoic equivalents (both the deposits and the inferred palaeoenvironmental processes), while maintaining the most complex array of possible geneses, all seemingly related to major global changes which occurred more than once and at specific times in Earth history rather than either changing gradually, or having been subject to major change at a single, specific time (e.g., Eriksson et al., 1998; their Table 1; references therein).

Widespread negative $\delta^{13}C$ excursions in carbonate rocks appear to have been related to major glaciations (e.g., Kaufman et al., 1991, 1997; Frimmel et al., 2002), while in contrast, an enrichment of ^{13}C may be linked to preferential fixation of ^{12}C during photosynthesis in stagnant environments (e.g., Schidlowski et al., 1976; Hayes et al., 1999; Rothman et al., 2003). At the beginning and end of the Proterozoic Eon (2500–543 Ma), Earth experienced a period of glacial conditions, with evidence for large continental ice sheets that may have extended to sea level at low latitudes (e.g., Evans et al., 1997; Williams and Schmidt, 1997; Schmidt and Williams, 1999; Sohl et al., 1999; Kempf et al., 2000). These two major glaciogenic intervals show some striking similarities: both coincide with supercontinent rifting; both feature “cap carbonates” - regionally persistent, continuous, thin intervals of limestone and/or dolostone that sharply overlie glacial or related deposits, on almost every continent, even in regions otherwise lacking carbonate strata, and these cap carbonates are characteristically depleted in ^{13}C.

The oldest known mid-latitude glaciation, recorded in the Pongola Supergroup (Kaapvaal craton) diamictite, occurred at 2.9 Ga (Nhleko, 2003). At least three discrete intervals of glacial activity punctuate the Early Palaeoproterozoic geological record between about 2.45 and 2.22 Ga (e.g., Hambrey and Harland, 1981) and are believed to be associated with assembly and rifting of a Late Archaean supercontinent ("Kenorland"; Williams et al., 1991). Of these, the best-known and best-preserved belong to the Huronian Supergroup of Canada and the Transvaal Supergroup of southern Africa. Although three discrete episodes of glaciation have been recorded in several widely spaced sedimentary successions in North America, there are very few carbonates in this interval (Young et al., 1998; Bekker et al., 2005).

Glaciogenic units in the Transvaal Supergroup are found within two of the three separate preservational basins (Fig. 9), with a prominent diamictite (Makganyene Formation; Griqualand West basin; age poorly constrained at <2.43 Ga and >~2.22 Ga; Cornell et al., 1996) in the one, and three discrete, but thin and laterally restricted diamictites within the second (the two lower units occur in the Duitschland Formation and both have associated carbonates; age of formation <2.43 Ga and >2.23 Ga, Hannah et al., 2004; the upper unit is in the Upper Timeball Hill Formation, <2.23 Ga and >2.22 Ga; all in the Transvaal basin). The lower Duitschland cap carbonate was deposited above glacial diamictite on an open-marine margin during oceanic transgression and records $\delta^{13}C$ depletion in carbonates with values ranging from −3.7‰ to 0.1‰, as well as ^{13}C-enrichment in organic matter (Bekker et al., 2001). The consistently negative $\delta^{13}C$ values of these cap carbonates contrast with the younger Duitschland carbonates (above the second diamictite, which is apparently non-glacial) above a notable sequence boundary, which are significantly enriched in ^{13}C up to +10.1‰. Palaeomagnetic data from the Transvaal Supergroup shows that the glacial diamictites were deposited within about 11º of the equator at 2.22 Ga (Evans et al., 1997), indicating a near-global glaciation. In North America, the three Huronian diamictites have no reliable palaeolatitudinal constraints, and only one is overlain by a cap carbonate, which recorded negative δ^{13}Ccarb, and enriched δ^{13}Corg (Kopp et al., 2005).

The Palaeoproterozoic glaciations were followed by a sharp reversal in the C-isotope signatures of carbonates globally, marked by a positive $\delta^{13}C$ carbon isotope excursion, with $\delta^{13}C$ values typically ranging from +7 to +12 ‰, known as the 2.2–2.1 Ga Lomagundi-Jatuli event (e.g., Schidlowski et al., 1975; Galimov et al., 1975; Schidlowski and Todt, 1998). This Palaeoproterozoic $\delta^{13}C$ positive anomaly has a global character, with fractionation between inorganic and organic carbon in ^{13}C-enriched carbonates close to 30 ‰, and represents a major perturbation of the global carbon cycle in geological history. The carbon isotope excursion began earlier than 2.22 Ga in the Fennoscandian and Canadian shields. Sedimentary successions with an age between 2.2 and 2.1 Ga contain evidence for warm arid climate including pseudomorphs after sulphate and halite, magnesite and red beds. They are devoid of BIFs. The magnitude and duration of the unique Lomagundi-Jatuli Event is documented in back-arc basins formed during ocean closure and subduction. The data imply that isotopically light carbon from organic-rich, passive margin sediments accumulated during subduction into the lower crust and mantle (Melezhik et al., 2007). Carbon isotope values are strongly correlated with environmental setting: playa and sabkha stromatolitic dolostones are most enriched, whereas those from intertidal settings exhibit lower $\delta^{13}C$ values (Melezhik et al., 2005). These facies-dependent trends suggest that the global δ^{13}Ccarb excursion was amplified by up to 8‰ by local bio-environmental factors. The Lomagundi-Jatuli carbon isotope excursion ended between 2.11 and

2.06 Ga. Carbon isotope values of carbonates deposited shortly afterwards range between 0 and +3 ‰. The end of the excursion coincided with deposition of Mn-carbonates, phosphorites, BIFs, organic-rich shales, the oldest fossils of possible eukaryotic origin, and a decrease in stromatolite abundance.

The Mesoproterozoic appears to have been a time of relative environmental stability, and carbonate successions usually have $\delta^{13}C_{PDB}$ values close to 0‰, similar to Phanerozoic successions (Veizer et al., 1980; Buick et al., 1995; Kah et al., 1999). The Mesoproterozoic was also a time of great stromatolite abundance and diversity. While Mesoproterozoic carbonates are known to have carbon isotope values close to 0‰ $\delta^{13}C$, Neoproterozoic rocks may have values ranging from -12 to +13‰, or greater (e.g., Knoll et al., 1986; Magaritz et al., 1986; Kaufman and Knoll, 1995; Iyer et al., 1995; Buick et al., 1998; Santos et al., 2000). Large negative excursions are associated with transgressive "cap carbonates", typically of dolostone and found above diamictites, reflecting deglaciation and a supposed paradox where glacial deposits, and carbonates assumed to require a warm climate, are in close contact. Higher $\delta^{13}C$ values are usually associated with burial of ^{12}C-rich organic matter.

The widespread glaciations of the Cryogenian (850 - 630 Ma) have been associated with large-scale carbon isotope perturbations, the "Snowball Earth" hypothesis, and the rise of metazoan life in the Ediacaran period (Hoffman et al., 1998; Knoll and Caroll, 1999; Hoffman and Schrag, 2002; Knoll, 2003; Chen et al., 2004; Condon et al., 2005; Halverson et al., 2005). The "Snowball Earth" glaciation may have extended to very low latitudes, possibly right to the equator (Kirschvink, 1992; Hoffman and Schrag, 2002). The Cryogenian Period (850-630 Ma) includes the Sturtian Glaciation (~720-700 Ma) and the Marinoan ice ages, ~650 to ~630 Ma (Martin et al., 2000; Smith, 2009). Kilner et al. (2005) report palaeomagnetic data from the

Neoproterozoic Huqf Supergroup of Oman that yield a palaeopole that places the Muscat region of Oman at a latitude of 13º in the late Neoproterozoic, providing direct evidence that both glacial and overlying cap carbonate units were deposited in the tropics.

The Precambrian-Cambrian boundary marks a significant change in the nature and mineralogy of carbonate sediments, intimately associated with the emergence and diversification of metazoans. The preservation potential of the earliest shells was poor, since they were thin walled and often of aragonite, and therefore highly susceptible to destructive neomorphosis, dolomitisation or phosphatisation. Nevertheless, these changes brought an end to the deposition of large-scale Precambrian-style carbonates associated with microbial processes, and heralded the beginning of a new and very different era of carbonate sedimentation patterns. There is a clear dichotomy between skeletal carbonate precipitation associated primarily with Metazoa, algae and Protozoa throughout the Phanerozoic, and largely unresolved processes of carbonate precipitation in the Precambrian.

4. Discussion

This paper provides an overview of Precambrian sedimentation systems through examination of the siliciclastic and carbonate rock records, and highlights differences compared to their Phanerozoic counterparts. Critical appreciation of some features of proven Precambrian bias serves to highlight what are in essence relatively small differences.

Precambrian rivers tended to be braided in the absence of any vegetation and with poorly developed soils (e.g., Schumm, 1968; Cotter, 1978; Long, 1978, 2004; Fuller, 1985; Els, 1990; Rainbird, 1992); however, channel incision into the bedrock during fall and overbank mud deposition during rise of the base profile favoured fixed channels, as shown by the fluvial interval at the base of the Neoproterozoic Sonia Formation in India (e.g., Samanta et al., 2007, 2008; Samanta, 2008). Although it can logically be assumed that climate was more effectual in controlling erosion and deposition in the absence of vegetation, the two unconformities that intercept the Sonia fluvial interval apparently owed their genesis largely to tectonism. Significant deviation in palaeocurrent direction across them and also selective occurrence of water escape structures under the relatively younger unconformity all over the study area support this suggestion (Samanta, 2008; Sarkar et al., 2005, 2008). The accentuated impact of climate change could also have made Precambrian rivers flashier, more often ephemeral (e.g., Tirsgaard and Øxnevad, 1998; Mueller and Corcoran, 2001; Eriksson et al., 2009a). Steeper gradients than usually found in rivers, estimated from three Palaeoproterozoic river deposits in the Waterberg Group, South Africa (Fig. 5) suggest occasional and localised steepening of depositional slope because of enhanced production (under high atmospheric CO_2) and localized deposition of clay during flash floods (Eriksson et al., 2006, 2008a).

In the absence of vegetation the aeolian regime should purportedly have been far more extensive during the Precambrian than in the Phanerozoic. Notwithstanding this, large erg deposits only became commonplace since ~1.8 Ga, possibly reflecting small cratons and a strong oceanic (humidity) influence on palaeoclimatic regimes prior to that (Eriksson and Simpson, 1998). Growth of large continental masses could thus have been critical for extensive development of aeolian ergs (Eriksson and Simpson, 1998) and supercontinentality may only have become widespread on Earth by about 2.2 – 1.8 Ga (e.g., Eriksson et al., this volume). Widespread rifting after c. 1.8 Ga, plausibly related to the assembly of supercontinent Laurentia and concomitant orogenesis (cf., Hoffman, 1988) may have enhanced the rate of creation of the all important preservation space for Precambrian aeolianites. Nevertheless, water table fluctuation must have been a prime factor controlling supply of sand and preservation of aeolinites, as it is now also.

The sea water composition seemingly changed from its likely initial acidic to subsequent alkaline state from the Precambrian to the Phanerozoic (section 3). The temporal change in sea water composition is best manifested in the exclusive occurrence of giant botryoids of aragonite (Fig. 19), metres-thick Mg-calcite beds as well as the rare occurrence of evaporitic sulphate in the Archaean (section 3.2 and references therein). Either low sulphate concentration or a high bicarbonate : carbonate ratio in the early sea water is implied. The Precambrian marine record is dominated by deposits of epicontinental seas, deep marine deposits being very scarce and then commonly associated with inferred ophiolitic successions, of sometimes controversial veracity (e.g., Eriksson et al., 2008b). Abundant peritidal deposits manifest sedimentation in epeiric embayments, while some storm successions document development of epeiric seaways. Vertical amalgamation of supralittoral marine storm beds resulted from the low gradient of the epeiric sea coasts (Sarkar et al., 2005). The low shelf gradient and wide proliferation of microbial mat, a typical Precambrian phenomenon, combined to restrict the epeiric sedimentation budget and that, in turn, may have favoured vertical stacking of highstand systems tracts at the expense of transgressive systems tracts (Sarkar et al., 2005; Catuneanu and Eriksson, 2007). Along the coastlines of epeiric seas, large braided fluvial systems debouched, and braid-deltas were likely to have been common (Nemec and Steel, 1988; Els, 1998); a close association of braided river systems and epeiric coastline facies is often observed in Precambrian shallow marine successions and may even have been a diagnostic feature (as postulated by Eriksson et al., 2008b) for an apparently largely exclusively Precambrian environmental setting. The multiple shallow marine formations cited here from Africa and India range in age from the Archaean to the Neoproterozoic and depict the entire range of variation from tide domination to wave supremacy.

With regard to deltaic systems little difference in terms of depositional environments and processes is inferred between Precambrian and Phanerozoic examples. The Precambrian delta deposits may, nonetheless, be incomparably thicker ranging up to 3,500m; high sedimentation rate induced by aggressive weathering (Corcoran et al., 1998; Donaldson and de Kemp, 1998) and lack of vegetation was likely the prime cause for such overthickening (e.g., Eriksson et al., 1998 and references therein). Abundance of soft sediment deformation structures not only in the delta slope deposits, but also in the prodelta deposits, suggests steepness of the depositional surface; rift-related subsidence can account for this relatively steeper palaeoslope as well as the accommodation space needed to explain the unusually large thickness of many Precambrian delta deposits (cf., Eriksson, 1979; Siedlecka et al., 1989). These factors in conjunction with bed-load dominance and lack of bank stability in the rivers can also explain the common occurrence of braid-deltas in the Precambrian record. Lack of channel stability also renders distinction between delta plain and delta platform difficult within Precambrian delta deposits, with flows appearing non-channelized in both settings.

An essentially unique feature of the Precambrian period was the widespread growth on most subaqueous clastic sediment surfaces within the photic zone, of prolific microbial mats and their often spectacularly preserved proxy features (Walter and Heys, 1985; Schopf, 1999; Hagadorn and Bottjer, 1999; Schieber et al., 2007). Even within the c. 1.8 Ga Waterberg Group desert in Kaapvaal, microbial mats flourished in playa settings before being destroyed in flash flood reworking of these shallow aqueous settings (e.g., Eriksson et al., 2000). The oldest known clastic mat features go back to 2.9 Ga (Noffke et al., 2006). Although often subtle in their preservation, particularly within the mudrocks, mats played an important role in binding sediment, and possibly even in altering sediment dynamics and stacking patterns at the local scale (e.g., Sarkar et al., 2005); in the Phanerozoic, they were still present (and are still so today) but suffer from commonly excessive destruction and poor preservation potential through the action of metazoan grazers (e.g., Schieber, 1998). The growth of microbial mats along coastlines required several weeks of non-burial, as shown by laboratory experiments (e.g., Gerdes and Klenke, 2003, 2007), and this implies an episodic sedimentation regime along both coasts and also possibly in braided river systems debouching into epeiric seas (e.g., Eriksson et al., 2009a). The latter postulate supports the indication of episodic fluvial systems in the Palaeoproterozoic derived from palaeohydrological calculations, and related to a long-lived greenhouse palaeo-atmosphere by Eriksson et al. (2009a).

While modern uniformitarianism ("the present is the key to the past"; e.g., Hallam, 1990) may reasonably be applied to the clastic sedimentation systems discussed in this paper, bearing in mind the important caveat that it is the variation in rates and intensities rather than in the processes and products that distinguishes the Precambrian (e.g., Donaldson et al., 2002), the carbonate sedimentary record is less amenable to this panacean principle. The latter observation can be ascribed largely to the limitations in our understanding of abiotic carbonate precipitation processes, and to the ongoing debate on the chemical composition and attributes of ancient seawater.

There is no unequivocal empirical evidence that calcium carbonate or dolomite precipitates directly from modern seawater (e.g., Leeder, 1982; Wright, 2000; Morse et al., 2003; Wright and Oren, 2005), and it has been suggested that kinetic inhibitors to carbonate precipitation are especially effective in saline waters. However, there is abundant evidence that these inhibitors can be overcome through microbial mediation (e.g., Slaughter and Hill, 1991; Robbins and Blackwelder, 1992; Robbins et al., 1996; Warthmann et al., 2000; Wright, 2000; Wright and Altermann, 2000; Wright and Wacey, 2004, 2005; Wright and Oren, 2005). Modern marine carbonate production is almost exclusively biologically mediated, but in the Precambrian, skeletal biotas were unknown and prokaryotes (and from c. 2.1 Ga possibly, eukaryotes also) dominated every ecosystem. The question that arises therefore is: to what extent was carbonate precipitation microbially-mediated in the Precambrian? The answer has been sought through theoretical models and assumptions about physico-chemical processes, but real progress has only been achieved through detailed observation of modern processes and ecosystems, and experiments that realistically simulate past microbiogeochemical environments (sections 3.3 and 3.4). While microbially induced carbonate precipitation has been demonstrated in many, well-documented experiments, the scale at which it operates in the natural environment has yet to be quantified, but is likely to have been very significant.

Microbial consortia often exhibit an ability to change the chemistry of a solution and to control pH at the microscale, passively or actively. This leads to oversaturationof Ca^{2+} and CO_3^{-2} ions and to the removal of kinetic inhibitors to carbonate precipitation, like sulphate or phosphate. The kinetic barriers of low carbonate ion activity, ion hydration and ion complexing, especially in saline waters, inhibit spontaneous carbonate mineral precipitation from saturated solutions but oxygenic photosynthesis and sulphate reduction by sulphate-reducing bacteria can overcome these natural barriers. Sulphate in seawater tends to form pairs with Ca^{2+} and Mg^{2+} ions, particularly the latter (Garrels and Christ, 1965; Wright, 2000; Wright and Oren, 2005). The removal of sulphate reduces complexing, raises carbonate alkalinity, and along with pyrite formation, enhances carbonate precipitation. Cyanobacteria can store Ca^{2+} and Mg^{2+} ions in organic envelopes and precipitate carbonates within their sheaths and extracellular polymeric substances, thus triggering sedimentary carbonate production. In the Precambrian, organic influence was restricted to microbes.

The morphology, facies associations and arrangement of architectural elements in Archaean stromatolites have been used to argue for a biological origin of stromatolitic lamination preserved in Archaean cherts and carbonates (e.g., Allwood et al., 2006). The observed behaviour of *in situ* laminae accretion and sediment trapping and binding cannot be explained by abiogenic carbonate precipitation from saturated seawater. The increasing complexity of Precambrian stromatolitic structures over time provides additional evidence for biologic control on stromatolite formation, and supports a strong argument that biogenic stromatolites and microbial mats were present at ~3.47 Ga, before becoming increasingly widespread, abundant and diverse through the Proterozoic (sections 3.1 and 3.2). Current evidence thus suggests that Archaean carbonate sedimentation resulted from a combination of hydrothermal and biochemical processes. The latter processes probably led to the first formation of low-temperature carbonates on the limited areas of continental shelf-type environments. Although atmospheric CO_2 was much higher than today, giving rise to acidic oceans, biochemical processes in shallow water microbiota mediated ambient water composition to produce alkaline, reducing conditions in which the kinetic barriers to carbonate precipitation were removed, forming the first carbonate deposits. Thus low-temperature carbonate precipitation was closely linked to earliest life, and as benthic microbiota spread across developing continental shelves and epeiric seas in the form of stromatolites and other microbialites, acted as major repositories for carbonates. Interactions between micro-organisms and their environments have generated a variety of facies and microfacies in platformal carbonates, both by *in situ* precipitation and by sediment trapping and binding. The interplay of cyanobacteria and heterotrophic bacteria has thus been the major contributor to carbonate precipitation for at least the last 2 billion years of Precambrian history.

Creation of vast shallow epeiric seas during the Proterozoic, especially during its later part, made the shallow marine basinal environment more conducive to development of the factors stated above and thereby accentuated the possibility of direct precipitation of carbonates from sea water. In consequence, molar-tooth cracks that probably owed their origin to overpressuring of pore fluids were filled immediately by carbonate cement crystals that acquired a comparatively more uniform size and a stable mineralogy, and an unzoned nature within cracks of a comparatively later generation.

5. Conclusions

The range of variation that the Earth-system has undergone through time is prodigious, and this must have impacted on the physical-chemical-biological milieu of sedimentation. This has imposed certain temporal controls on the development of specific settings and sedimentation systems; as an example, prior to c. 1.8 Ga wind deposits of limited scale did develop, but large scale ergs only formed once large land masses resulted following the formation of supercontinents such as Laurentia or Kalahari after c. 2.2 - 1.8 Ga. Glaciation of apparently limited scale is known from the Archaean, and two global glaciation events in the Palaeoproterozoic and Neoproterozoic also underline the importance of chronological control on global-scale events, yet this should be seen against the backdrop of time-independent action of apparently uniformitarian processes and facies-scale sedimentary products. Precambrian deltas closely resemble their Phanerozoic counterparts, except for a strong influence of tectonism on apparently enhanced subsidence and sedimentation rates in early Precambrian examples prior to formation of large continental land masses; the braided nature of distributary channels reflects the influence of an absence of vegetation on all channel systems, deltaic, alluvial, fluvial and even tidal drainage networks (e.g., Eriksson et al., 1998). Precambrian lake deposits are not widely reported, and then with some bias towards identification of arid or semi-arid examples (related by some schools of thought to ancient greenhouse palaeo-atmospheric regimes); however, distinction of freshwater lake deposits from low energy epeiric shoreline settings remains difficult.

While the sedimentation processes and products within glacial, desert, delta, possibly also lake settings strongly resemble their Phanerozoic counterparts, and whereas inferred controls on basin evolution and –filling appear equally comparable in a majority of cases (e.g., Eriksson et al., 2005, 2007 and references therein), preservation of shallow marine environments and the characteristics of Precambrian fluvial sediments are less straightforward. As a caveat, it is of course conceivable that many distinctions between Precambrian and Phaerozoic-Modern sedimentation patterns are subtle and remain masked by the overall similitude at the facies to basinal scales. At this point, it is also pertinent to repeat the dictum of Donaldson et al. (2002) that it is really the rates and intensities of processes rather than the nature of the processes that distinguishes the Precambrian period.

Precambrian clastic and carbonate marine deposits are essentially a reflection of a rather unique setting of epeiric seas (both embayments, and the seaways that more strongly resemble shelfal environments), formed due to continental crustal growth processes and rates, especially from the Neoarchaean onwards. Marine deposits outside of this milieu are generally associated with highly deformed inferred ophiolitic successions, some of which are regarded by at least portions of the scientific community as controversial. True epeiric sea analogues of any scale are lacking in the current geological record (e.g., Hallam, 1981; Galloway and Hobday, 1983; Pratt and Holmden, 2008), which does not help in modelling of perceived ancient Precambrian clastic equivalents (such as the examples shown in Fig. 11; see also Eriksson et al., 2008b). Microbial mats flourished in the clastic settings of these shallow epeiric basins, and their role in lithofacies (e.g., section 2.7 of this paper) to facies tract (e.g., Sarkar et al., 2005) scale development were likely more pronounced in the Precambrian epoch as opposed to the Phanerozoic, when their role was much reduced through metazoan grazing activity.

Deposition of shallow marine carbonate sediments supports a very distinct difference between Precambrian examples, where biotic mediation to overcome kinetic factors inhibiting direct precipitation of carbonates from seawater was microbial, and the Phanerozoic when skeletal carbonate precipitation was largely mediated through metazoans, algae and protozoans. Molar tooth structure appears to be a unique Precambrian feature, particularly for the Meso-Neoproterozoic. However a large degree of uncertainty still hangs over modelling of Precambrian carbonate genesis and is also directly related to ongoing debate and research into palaeo-atmospheric and palaeo-hydrospheric evolution during this epoch.

But, amongst the clastic settings, it is perhaps in the fluvial depositional environment that the greatest potential lies in future study of possibly unique features in Precambrian sedimentation patterns. Their predominantly braided and sandy bed-load style is well researched and equally well ascribable to the lack of vegetation allied to poorly developed soils (e.g., Schumm, 1968; Cotter, 1978; Long, 1978, 2004; Fuller, 1985; Els, 1990; Eriksson et al., 1998). However, temporary and localised ponding of muddy sediment appears to have been a relatively important characteristic of at least some such systems (cf. the Indian and South African examples discussed in section 2.2 of this paper; see also, Rainbird, 1992; Eriksson et al., 2006, 2008a). In the case of the examples from the Waterberg Group (South Africa) palaeohydrological data suggest that enhanced palaeoslopes developed locally as a result of ponded muddy sediment, and Eriksson et al. (2009a) relate this to an overall greenhouse palaeo-atmosphere and concomitant strongly episodic (in terms of rainfall) palaeoclimate which may have dominated the Kaapvaal craton from c. 2.3 – 1.8 Ga. Fluvial systems thus have the potential to also impact understanding of the larger issues in Precambrian sedimentation patterns, such as evolution of the extant atmosphere, cratonic and supercontinental geodynamics, palaeoclimates, and the role of microbial organisms. Microbial mats also strongly support episodic channelised flows and distinctly punctuated sedimentation along epeiric coastlines, particularly where large braided fluvial systems debouched into them (cf., Eriksson et al., 2009a).

A distinct potential of future research thus exists in sifting out the commonly subtle evidence of distinctions in sedimentation milieus between the Precambrian and the Phanerozoic. While it might be overly ambitious to claim that an improved comprehension of the evolutionary course of the Earth system as a whole may also eventuate, it might be reasonable to suppose that, at the very least, Precambrian sedimentary studies can contribute to such an enhanced understanding of the planet's history through c. 85% of its history.

Acknowledgements

All the authors express gratitude to their respective institutions for infrastructural facilities and research support. SS acknowledges field grants from the CSIR, Government of India and the INSA-DFG for providing a fellowship to work with WA in Munich on molar tooth structure.

PGE thanks both the University of Pretoria and the National Research Foundation of South Africa for funding. Thino Rajab is acknowledged for drafting many of the figures.

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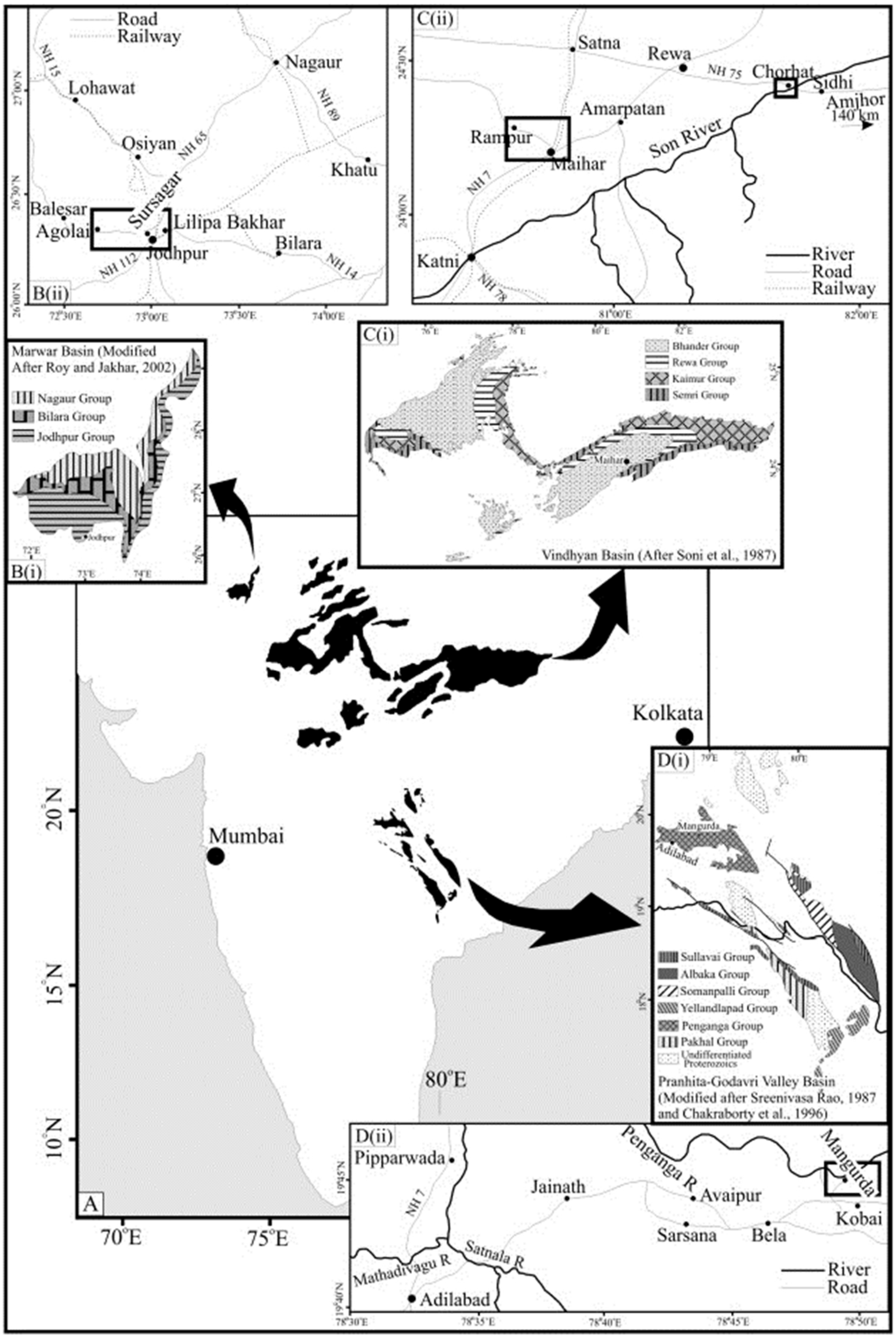


Figure 1 (A) Locations of Indian Palaeoproterozoic basins discussed in this paper. Geological sketch maps and locality maps for: (B) the Marwar basin; (C) the Vindhyan basin; and (D) the Pranhita-Godavari basin.

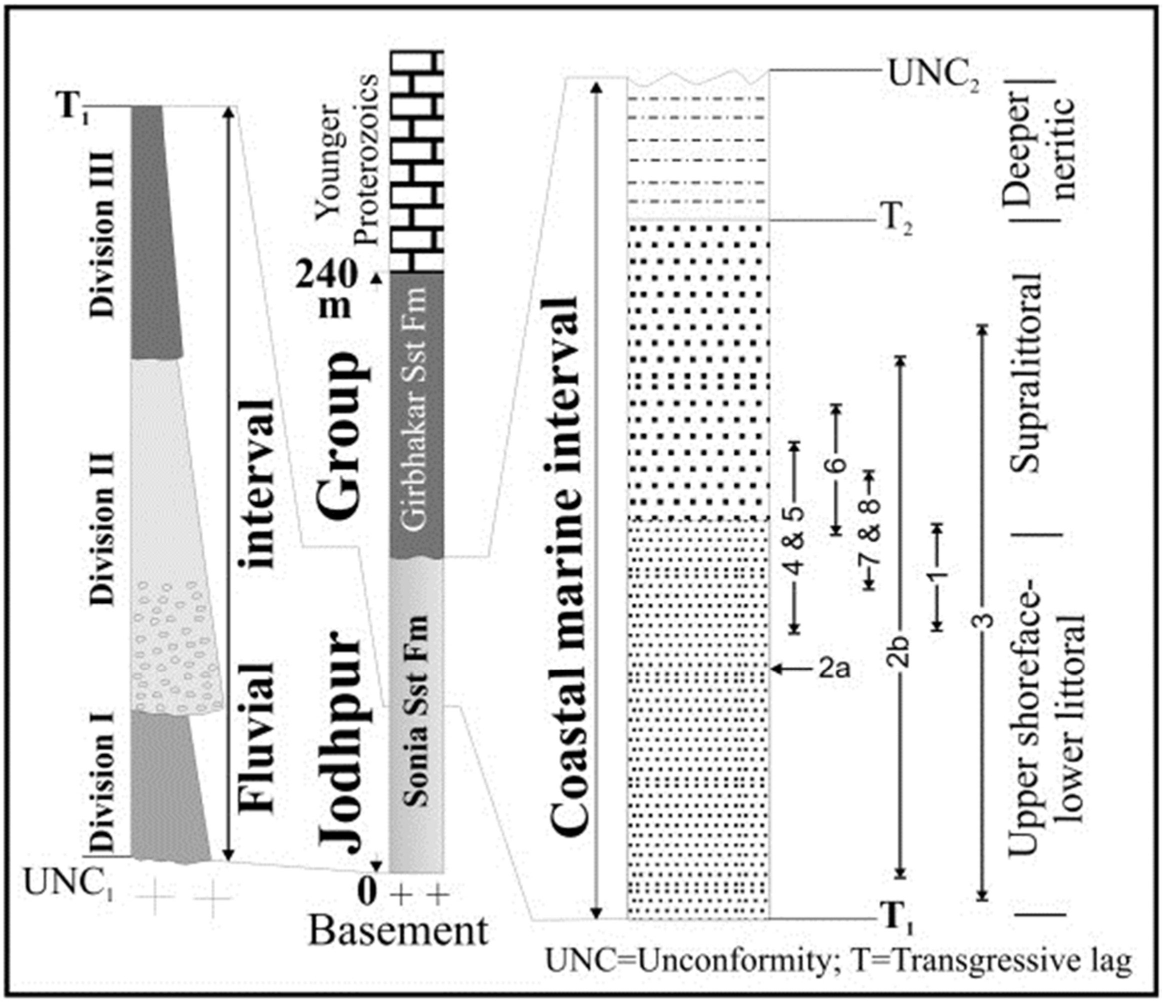


Figure 2 Temporal shift in palaeogeography-palaeoenvironment inferred for the Sonia Sandstone at the base of the Jodhpur Group, Rajasthan, India (see Fig. 1B for location). Note three divisions in the basal fluvial interval (left-hand column) and vertical variation in inferred marine palaeoenvironment (right-hand column). The numbers adjacent to the right-hand column denote detailed categories of microbial mat features, as discussed in the text (Section 2.7): 1, 2a and 2b = mat-layer structures (category 1 in Section 2.7): 1 = mat-layer discoidal; 2a = mat-layer crumpled; 2b = mat-layer wrinkled structures. 3, 4 and 5 = mat-induced structures (category 2 in Section 2.7): 3 = mat-induced cracks along ripple crests, mat-induced surface cracks and mat-induced surface ridges; 4 and 5 = mat-induced bulges. 6, 7 and 8 = mat-protected structures (category 3 in Section 2.7): 6 = mat-protected setulfs; 7 = mat-protected ripples; 8 = mat-protected patchy ripples. Details are given in Sarkar et al. (2008).

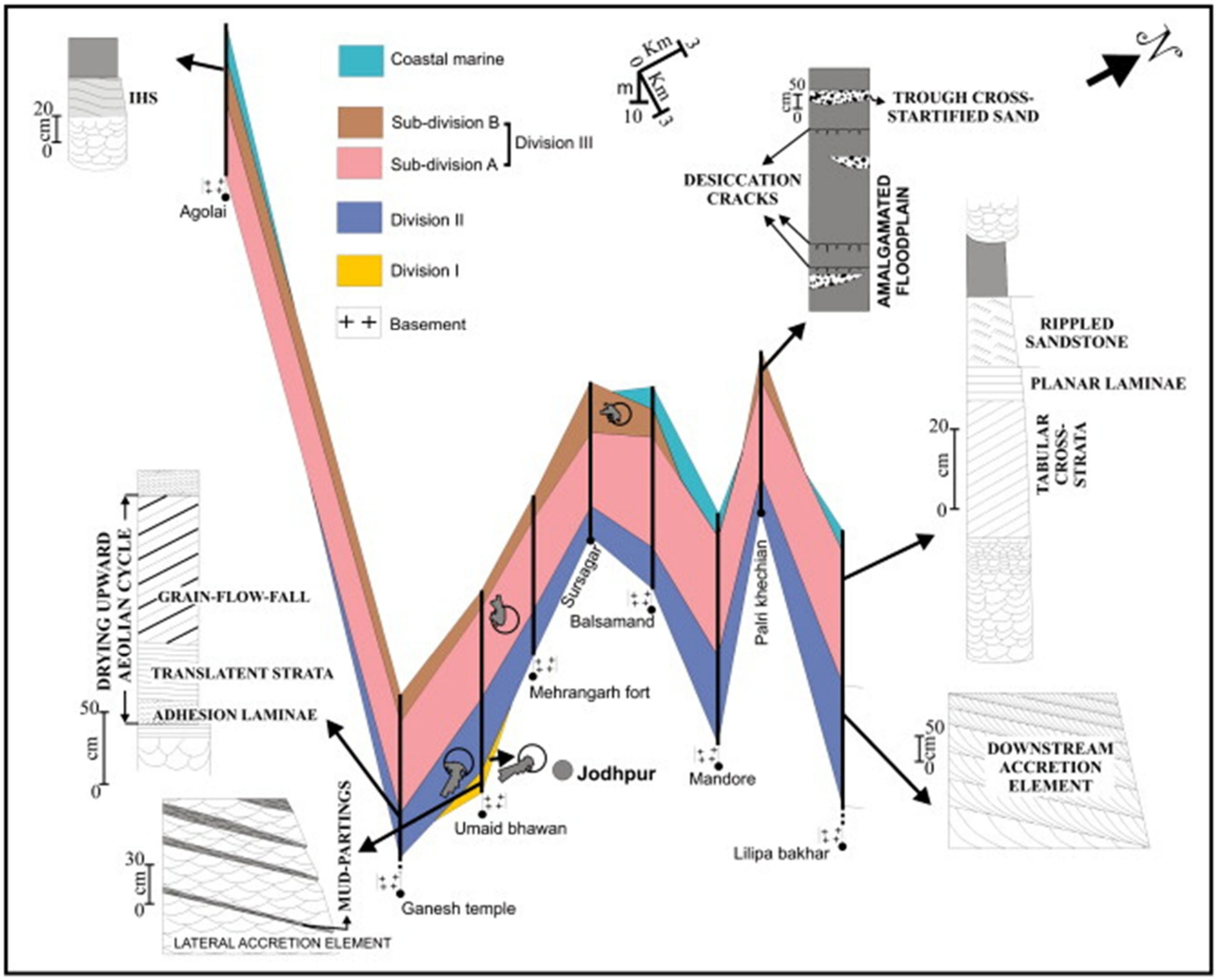


Figure 3 Fence diagram depicting lateral thickness variation of the three divisions of the fluvial interval at the base of the Sonia Sandstone, taking the base of the succeeding coastal marine interval as the datum plane (centre). The fluvial interval is bounded below by rhyolite basement and by the coastal interval above. Note palaeocurrent variation between the three divisions. Also note characteristic sedimentary features of the three divisions shown in the individual profiles.

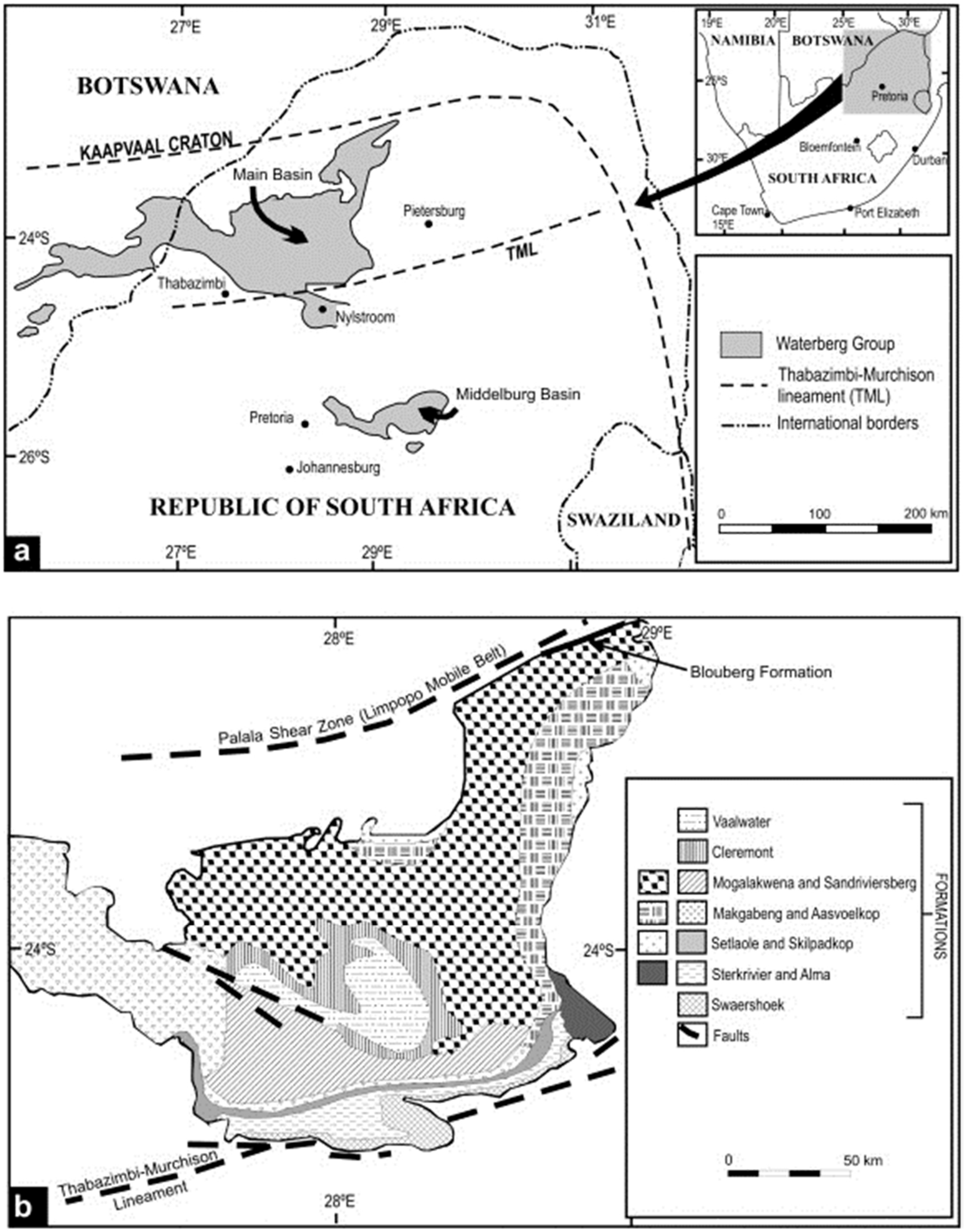


Figure 4 The Waterberg Group, preserved within South Africa in a Main basin and the smaller Middelberg basin (a) Sketch map of locations of these two basins, and their relation to major crustal architectural features of the Kaapvaal craton (b) Geological sketch map of the Main basin; note synsedimentary Vaalwater fault in the SW of the basin, and three pairs of correlated Formations: Setlaole–Skilpadkop, Makgabeng–Aasvoelkop, and Mogalakwena–Sandriviersberg. Modified after Callaghan et al. (1991) and Eriksson et al. (2006).

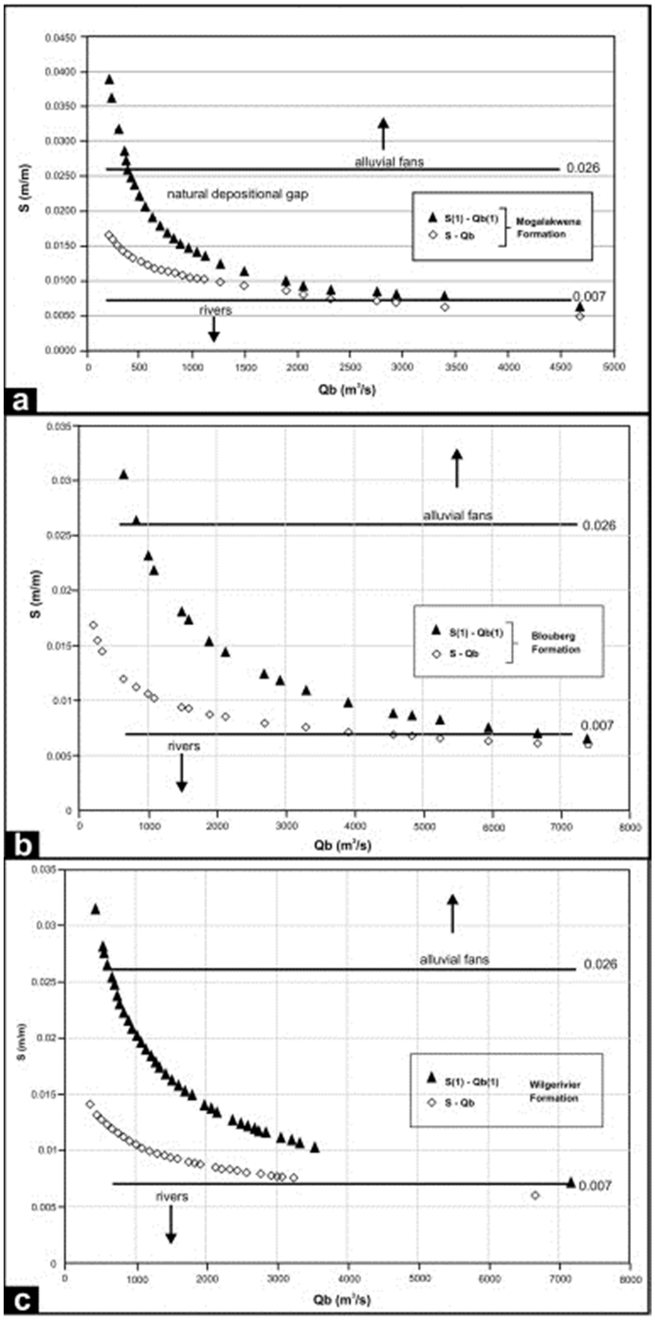


Figure 5 Binary plot of palaeohydrological values (methodology and the various parameters determined for the Waterberg rocks are discussed in detail in [Eriksson et al., 2008a] and [Eriksson et al., 2008b]) calculated for (a) the Mogalakwena Formation (b) the Blouberg Formation and (c) the Wilgerivier Formation of the Waterberg Group: palaeoslope (S) versus mean annual bankfull discharge values (Q_b). The two values of S (S and S[1]) reflect different equations within the standard palaeohydrological methodology. Note also the maximum gradient (S) for rivers (0.007 m/m) and the minimum gradient for alluvial fans (0.026 m/m) of Blair and McPherson (1994). Note similarity of palaeoslope values for these three formations.

Figure 6 Field photograph of steeply inclined (foresets are close to 30° to the bounding surfaces) planar cross-bedding in the Makgabeng Formation, Waterberg group; cross-bed foresets comprise inversely-graded aeolian laminae. View of photo is approximately at right angles to the dip direction (left to right). Makgabeng Plateau, Limpopo Province, South Africa; person in right foreground for scale.

Figure 7 Localized erosively-based massive sandstone (with cracked weathering pattern in outcrop) cutting through low angle inversely-graded aeolian laminae in palaeo-dune toesets, Makgabeng Formation, Waterberg Group, South Africa. The massive sandstones are interpreted as the product of massive rainfall events resulting in hyperconcentrated flows which passed down dune lee faces, eroding them in the toesets (Simpson et al., 2002). Centimetric scale; location on Makgabeng Plateau, Limpopo Province.

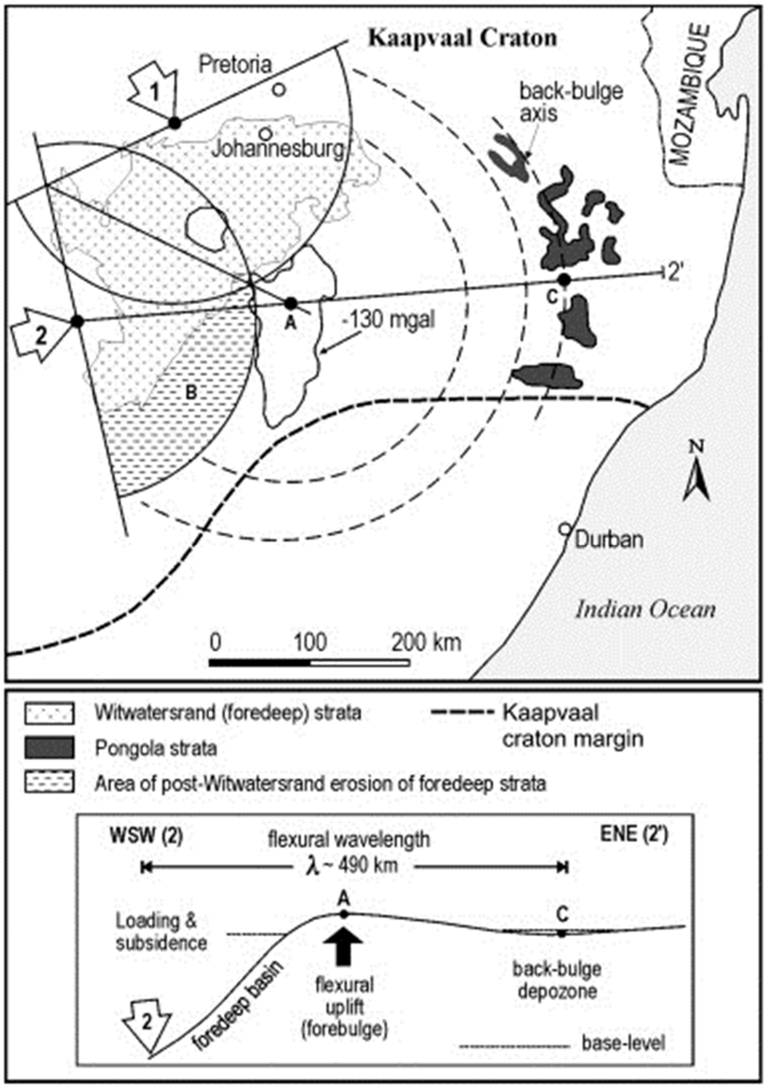


Figure 8 Sketch map (top) and schematic profile through inferred Witwatersrand foreland basin system (below). The preserved Witwatersrand basin equates to the foredeep depozone (foredeep sub-basin), with area "B" representing an area of subsequent erosion of these foredeep strata. The two solid line half-circles, centred on the areas of maximum loading (numbered "1" and "2" for accreting northern and western composite terranes, respectively), outline the approximate distribution of the foredeep depozone. The forebulge developed outside the area covered by these two half-circles, with its apex, point "A" (see in map and profile below) enclosed by the −130 mgal isoline of the gravity field. The three dashed circles suggest contour lines of the foreland system centred around the forebulge apex (at A), with the outermost such circle marking the position of the backbulge sub-basin axis (which equates with the depo-axis of the Pongola Supergroup basin), as suggested in theoretical flexural profile models (cf. Catuneanu, 2001 and references therein). In the case of the greater Witwatersrand basin, the forebulge remained emergent, separating discrete foredeep (filled by the Witwatersrand Supergroup) and backbulge (filled by the Pongola Supergroup) sub-basins. The cross-sectional profile 2–2′ on the map is shown below in the profile. Modified after Catuneanu (2001).

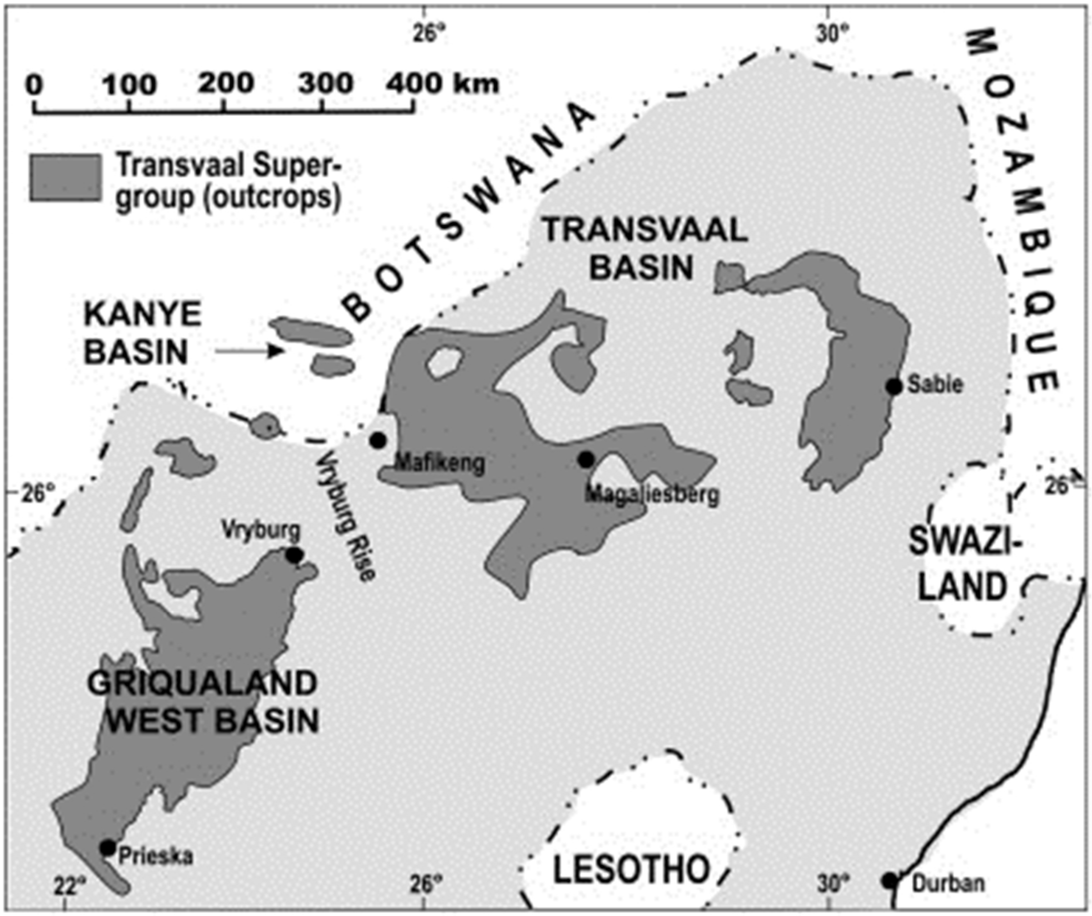


Figure 9 Sketch map showing the three Transvaal (Supergroup) sub-basins: Transvaal itself and Griqualand West, separated by the Vryburg Rise, a palaeohigh, with the Kanye to the north of this feature.

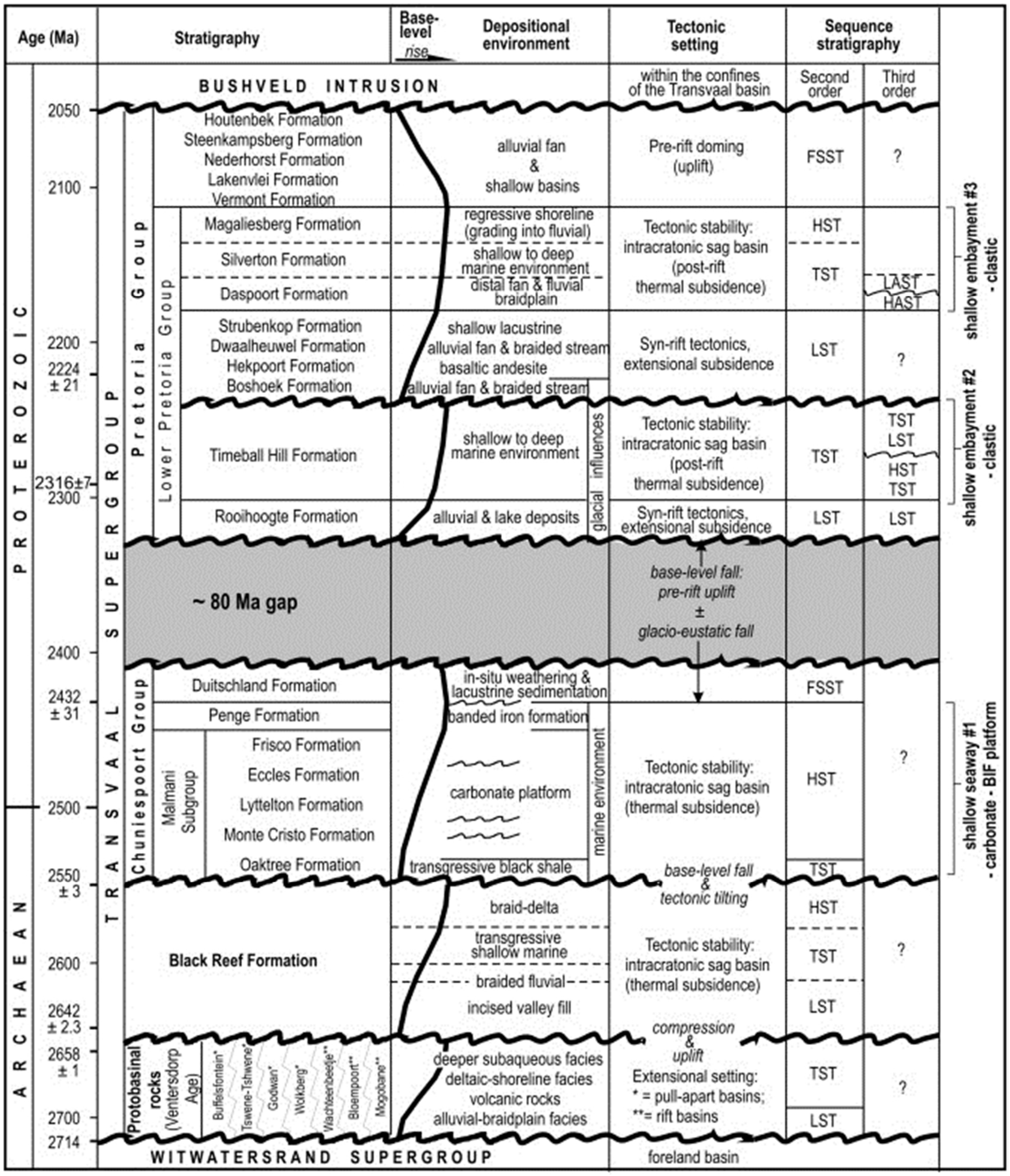
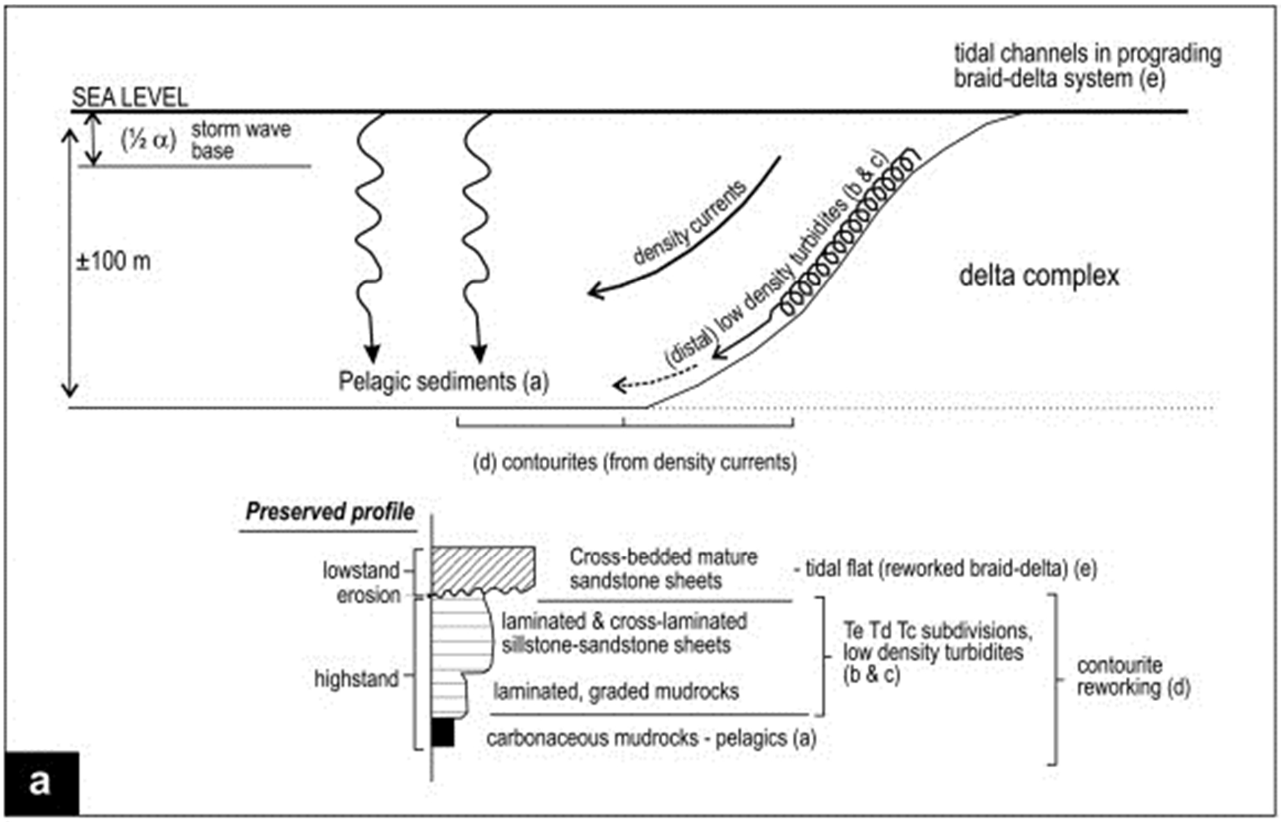


Figure 10 Schematic summary figure of stratigraphy, inferred depositional palaeoenvironments, tectonic settings and interpreted sequence stratigraphy for the Transvaal Supergroup within the Transvaal preservational basin (modified after [Catuneanu and Eriksson, 1999] and [Eriksson et al., 2001b]; see latter for geochronological detail and references). Note four main subdivisions of the basin-fill: “protobasinal rocks” at the base, Black Reef Formation, Chuniespoort Group, and uppermost Pretoria Group; also major unconformity at base of Pretoria Group. Note also, three inferred epeiric settings: a shallow carbonate–BIF epeiric seaway interpreted for the Chuniespoort Group, and two postulated clastic epeiric embayments at the levels of the Rooihoogte-Timeball Hill Formations and the Daspoort–Silverton–Magaliesberg Formations (far right of figure). Wavy lines suggest unconformable contacts. Abbreviations: LST = lowstand systems tract; TST = transgressive systems tract; HST = highstand systems tract; FSST = falling stage systems tract; LAST = low accommodation systems tract; HAST = high accommodation systems tract.



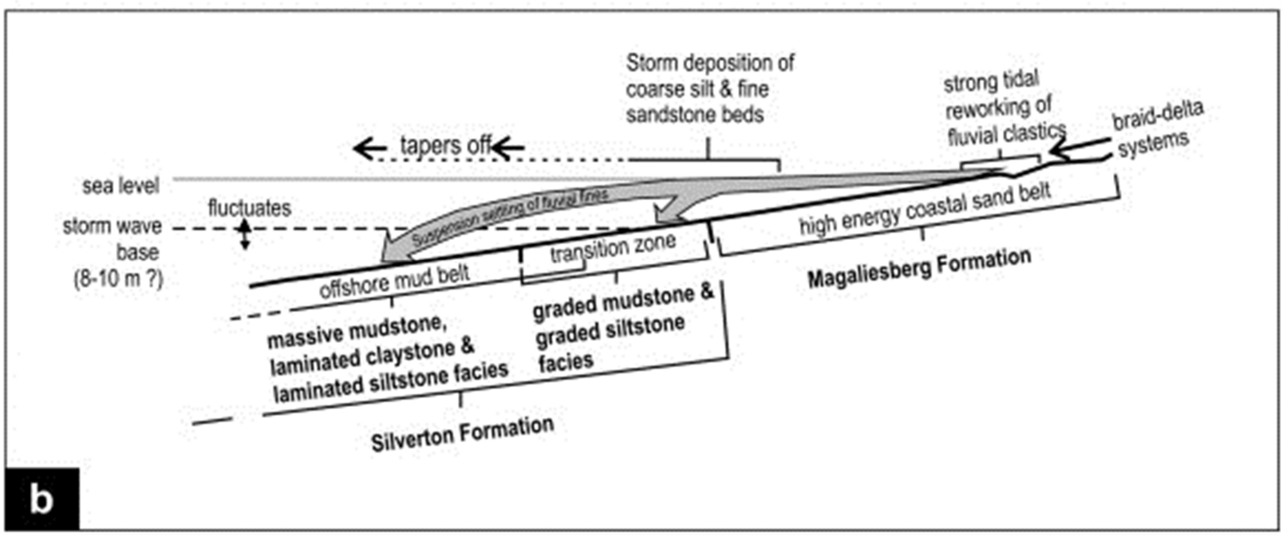


Figure 11 Clastic epeiric embayment models interpreted for: (a) the Timeball Hill Formation and (b) the Silverton–Magaliesberg Formations, both of the Palaeoproterozoic Pretoria Group, Transvaal Supergroup (modified after [Eriksson and Reczko, 1998], [Eriksson et al., 2002] and [Eriksson et al., 2004d]).

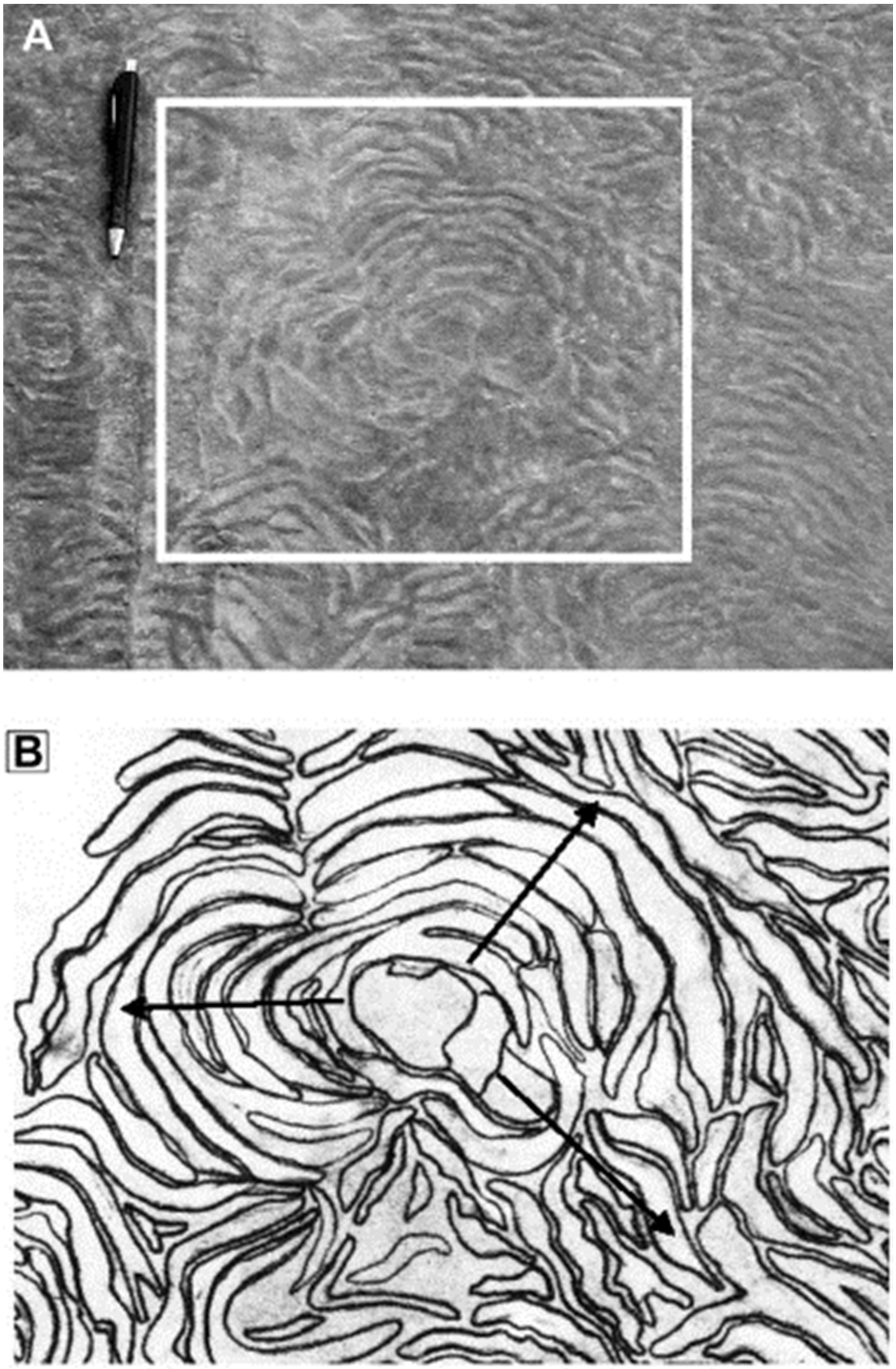


Figure 12 Mat ground, preserved intact, displaying a radial arrangement of laterally linked curved spindles (pen length 14 cm) (A). Sketch made from A, detailing stromatolite-like lateral accretion (B).

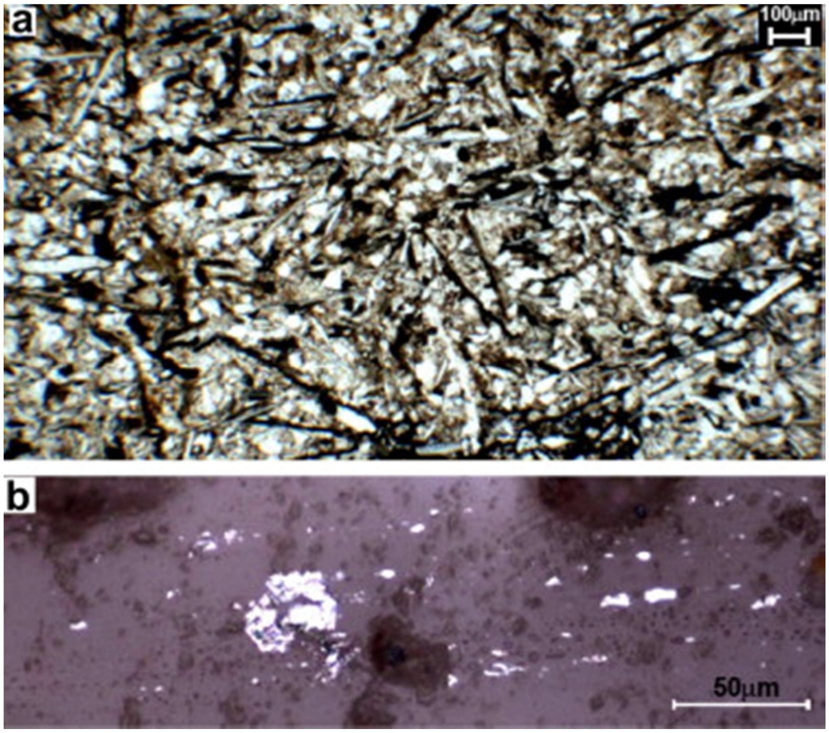


Figure 13 Hapazardly oriented platy minerals in fine-grained veneer on coastal marine bed surfaces in the Sonia Sandstone, depicting "fly-paper" effect (a); early diagenetic pyrite in clustured and also in disseminated form within the marine interval of the Sonia Sandstone (b).

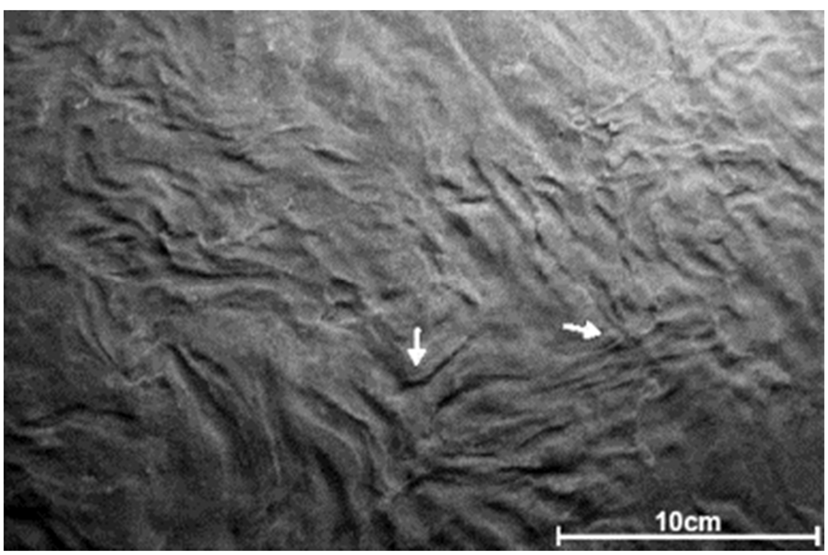


Figure 14 Crumpling feature on a sandstone bed surface within the coastal marine interval of the Sonia Sandstone. Note the minor drag folds in diverse directions (arrows).

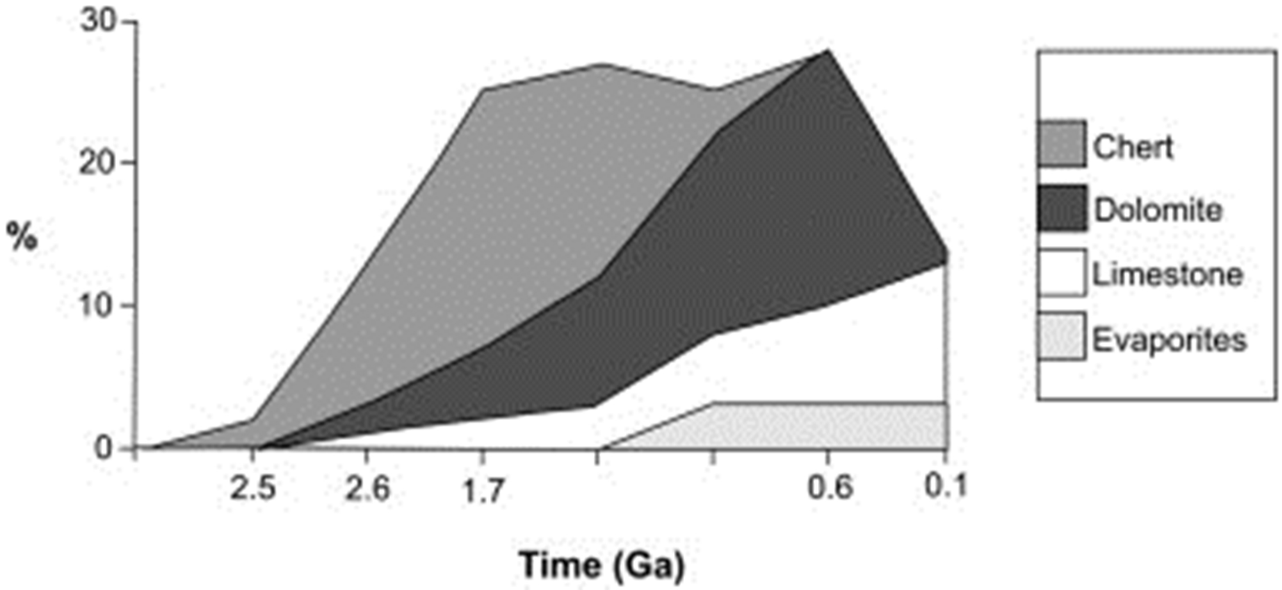


Figure 15 The relative abundance of non-siliciclastic sediments through time, showing dolomites to be much more abundant than limestones in the Proterozoic, when microbial ecosystems dominated the biosphere. Modified after Ronov (1965).

Figure 16 Cross-bedding and ripples in silicified carbonate rocks of the c. 3.5 Ga Dresser Formation, Warrawoona Group in the North Pole Dome area of the Pilbara craton. Pen for scale.

Figure 17 A number of stromatolitic structures are found in the c. 3.5 Ga Dresser Formation (Pilbara craton) – here, partially silicified wrinkly stromatolites pass laterally into domical forms. Coin for scale.

Figure 18 Steep-sided coniform stromatolites of the Strelley Pool Chert at the Trendall locality, Shaw River, Pilbara, Western Australia, showing both persistent and limited vertical continuity between laterally flat- and undulatory-laminated areas. The complex changes in form suggest competitive response to minor variabilities in biochemical conditions (e.g., chemical gradients, nutrient supply, light availability).

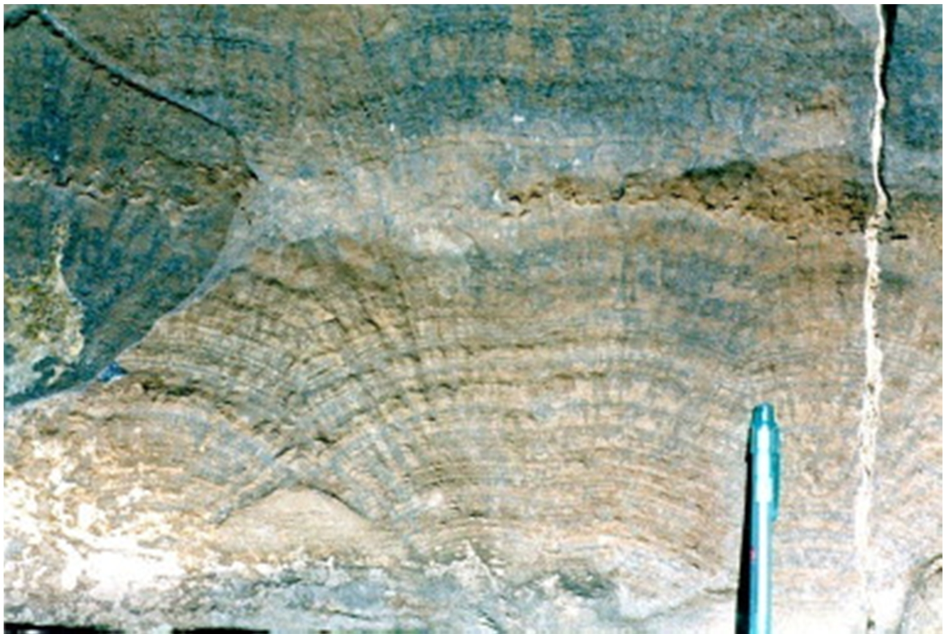


Figure 19 ‘Crystal fans’ from the Neoarchaean Reveilo Formation, Campbellrand Subgroup, South Africa. Originally interpreted as carbonate replacive of gypsum, they were later considered to be replacive after aragonite (Sumner and Grotinger, 1996). Hardie (2003) and Gandin et al. (2005) have since argued that the fans were in fact originally gypsum. Pen for scale.

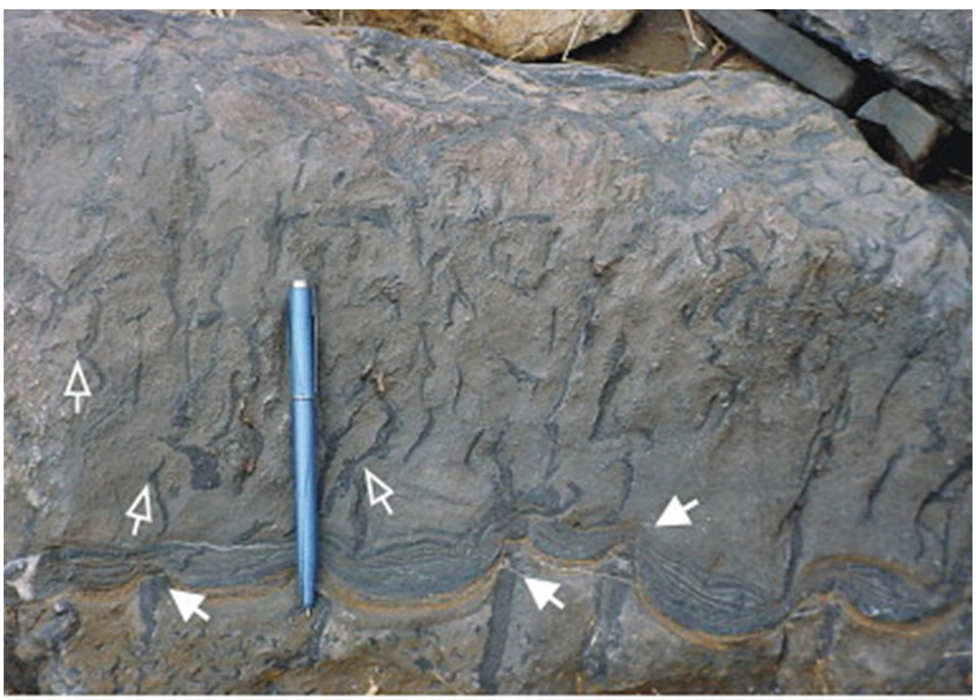


Figure 20 Swarms of molar tooth structures (oblique view) from the Mangurda Limestone, Pranhita-Godavari valley. Note ptygmatic folding within some of them (open arrows) and evidence of differential compaction against some of the vertically oriented structures (solid arrows).

Figure 21 Thinly interlaminated marl-limestone facies of the Bhander Limestone, India, with dark-coloured molar tooth structure in its upper, “shredded” part (arrow).

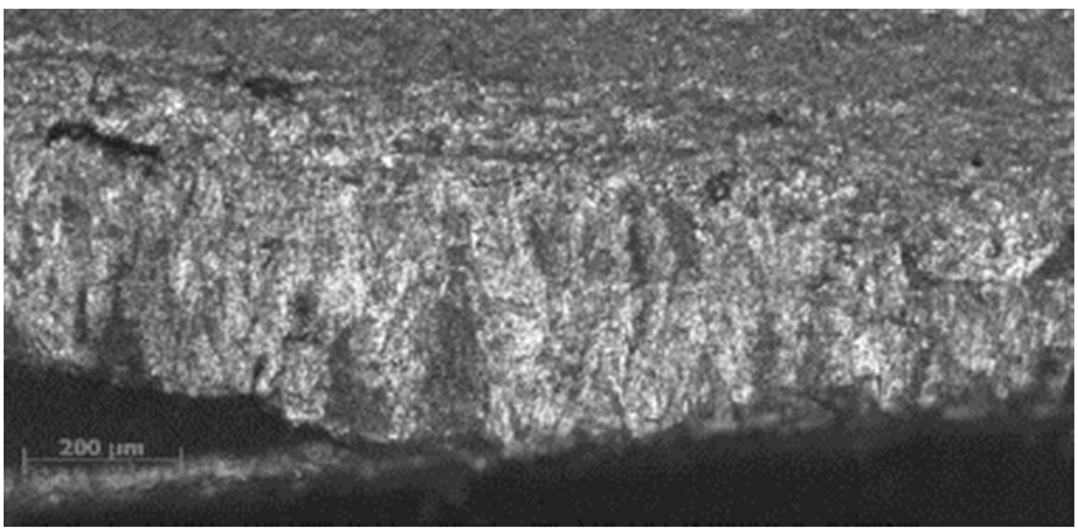


Figure 22 Swallow-tail structure in pseudomorphs after gypsum, found in close association with molar tooth structure within the Bhander Limestone, India.

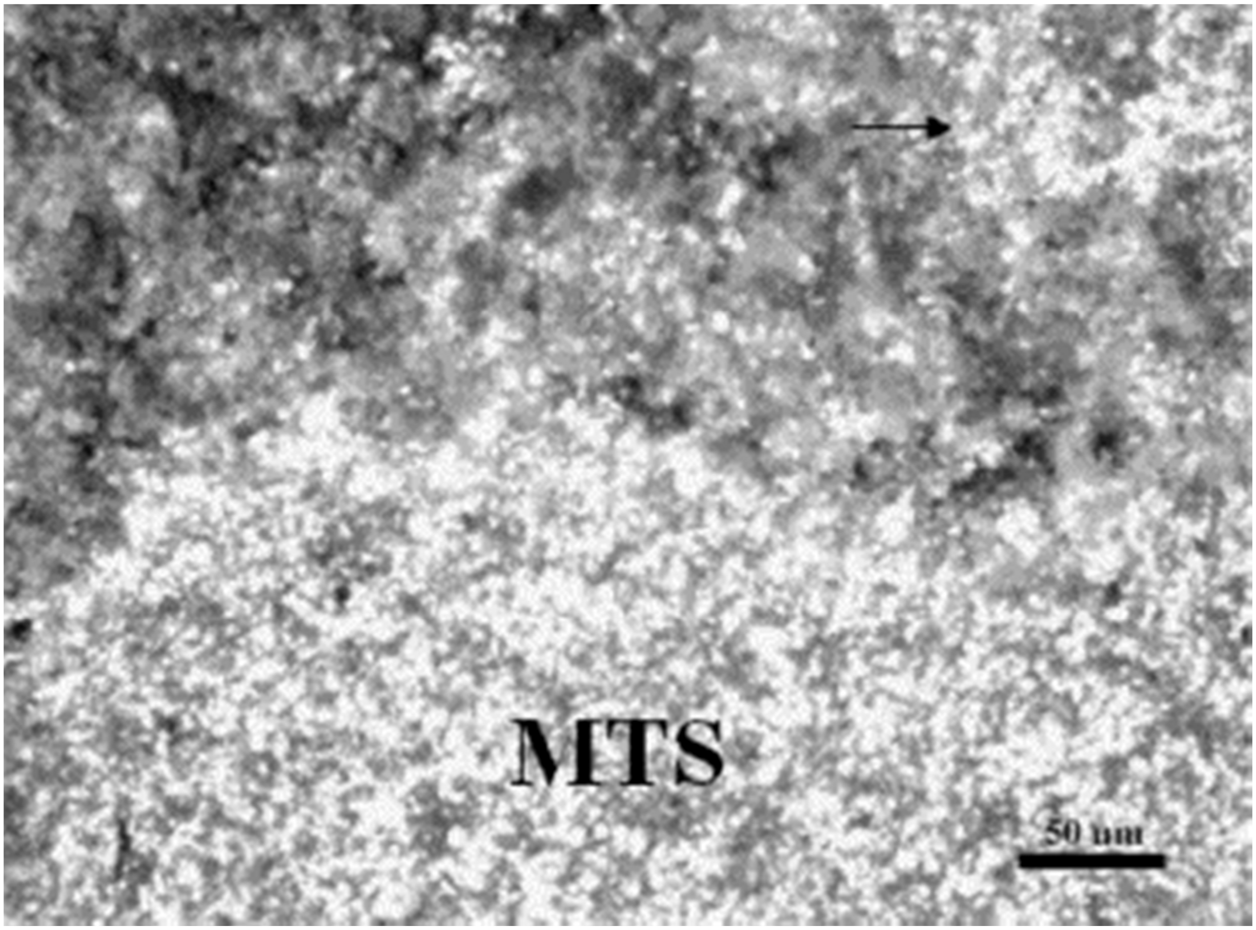


Figure 23 Under cathode luminescence, molar tooth structure (MTS) displays brighter illumination than the groundmass within the Mangurda Limestone, India. However, bright MTS crystals are sparsely distributed (arrow) within the groundmass as well.

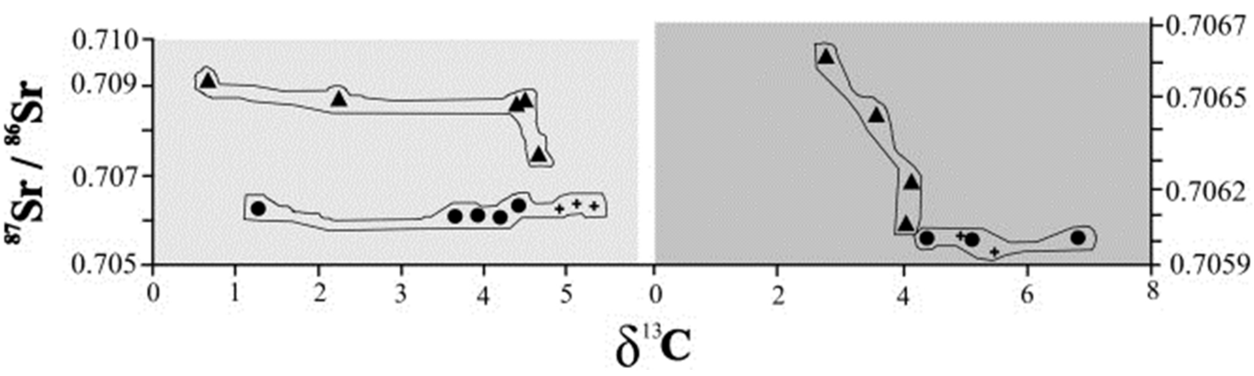


Figure 24 $^{87}Sr/^{86}Sr$ – $\delta^{13}C$ relation for 1st generation molar tooth structure (MTS) (cross symbol), 2nd generation MTS (circles) and the sediment that hosts them (triangles), in the Bhander Limestone, at left, and in the Mangurda Limestone, at right. Note significant difference between the MTS population and the host sediment for both the limestone formations.