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Deep-water manganese deposits in the mid- to late Proterozoic Penganga Group of the Pranhita–Godavari Valley, South India

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Abstract: Facies analysis of the unmetamorphosed sediments enclosing the stratiform manganese oxide deposits of the mid- to late Proterozoic Penganga Group identifies the base of slope of a distally steepened deep-water ramp as their site of accumulation. The interpretation is based on their close association with a variety of mass flow deposits ranging from limestone conglomerates to calcarenites and deep-water, plane-bedded micritic limestone devoid of current- or wave-generated structures as well as detritus coarser than fine silt. These deposits occur within a major transgressive succession. The base of slope origin of stratiform manganese deposits is uncommon in the rock record and their origin is to be constrained against the background of base of slope depositional setting.

Sedimentary manganese ores are known to accumulate in highly variable depositional milieu from lacustrine to oceanic settings (Roy 1981, 1988). They are found in both shallow- and deep-water settings at present, whereas shelf-related depositional regime appears to represent the most favoured site of manganese deposition in the rock record (Beukes 1983; Cannon & Force 1983; Frakes & Bolton 1984; Bolton & Frakes 1985; Force & Cannon 1988). Occurrences in shallow depositional settings is well in accord with the oxygenating conditions required for deposition of manganese minerals in their high valency state. In contrast, the dearth of land-based deep-water manganese deposit remains an enigma in sedimentary manganese metallogenesis (Roy 1988) when their common occurrence in modern oceanic basins is considered.

In recent years Bühn *et al.* (1992, 1993) suggested an outer shelf origin of manganese deposits associated with metamorphosed Damara Sequence of Namibia based mainly on lithofacies and tectonic setting. Beukes (1989), of course, suggested that the manganese in the Transvaal Supergroup was derived from deep marine setting with a small hydrothermal component deduced on the basis of geochemical considerations.

The unmetamorphosed middle to late Proterozoic manganese oxide deposits (Roy *et al.* 1990) of the Penganga Group provides a unique opportunity for detailed facies analysis of the enclosing sediments. These manganese deposits are hosted within deep-water micritic limestones (Roy 1988; Chaudhuri *et al.* 1989; Roy *et al.* 1990). In this paper we propose to constrain the depositional setting of manganese deposits through facies analysis of the host carbonates.

Geologic setting

The Proterozoic of the Pranhita–Godavari Valley

The Penganga Group constitutes a part of the middle to late Proterozoic Godavari Supergroup (Chaudhuri & Howard 1985; Chaudhuri & Chanda 1991) of the Pranhita–Godavari Valley basin in South India. The basin (Pranhita–Godavari Valley basin), commonly cited as a major Proterozoic continental rift basin (Naqvi & Rogers 1987), served as a repository of sedimentary sequences formed in a variable spectrum of environments ranging from continental to deep marine. The Supergroup has been subdivided into several groups on the basis of genetically related lithologic assemblages. The age relationship between different groups is still uncertain. However, the lower and middle parts of the Supergroup comprise carbonate-dominated assemblages, both shallow and deep marine whereas the upper part is represented by continental siliciclastics (Chaudhuri & Chanda 1991). The Penganga Group, occupying the northwestern part of the outcrop belt (Fig. 1), is dominated by a deep water limestone–shale assemblage with a subordinate basal unit of shallow marine siliciclastics. Until now it is not known to have any depositional contact with other carbonate-dominated groups, but it

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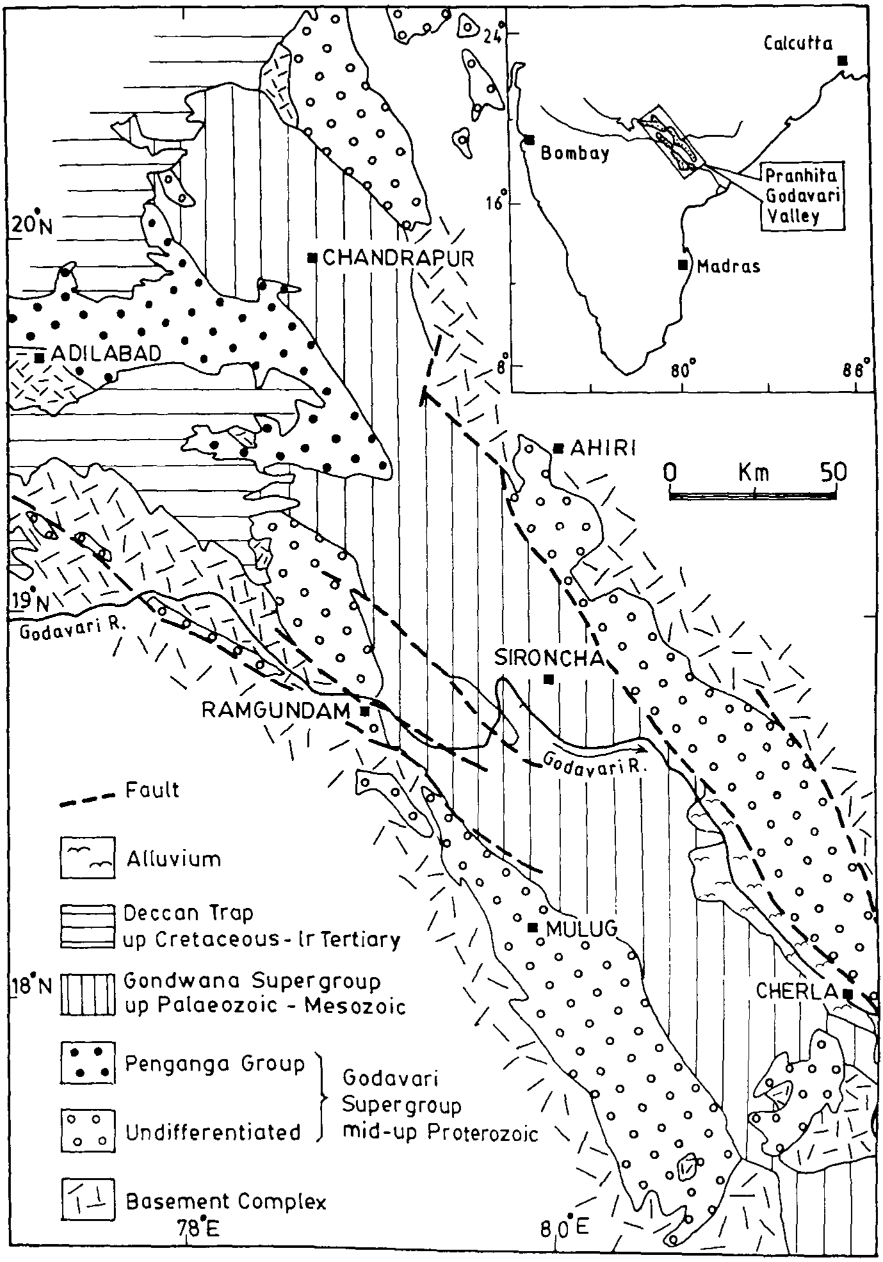

Fig. 1. Geological map of the Pranhita–Godavari Valley showing distribution of different subdivisions of the Godavari Supergroup.

unconformably underlies the continental siliciclastics.

Felsic volcanics and pyroclastics have been lately reported from the deep as well as the shallow-water sediments in some groups of the Godavari Supergroup (Chaudhuri & Chanda 1991). No evidence of volcanic activity has so far been detected in the Penganga sediments around Adilabad, but has been reported from a succession in the central part of the Pranhita–Godavari Valley, which may be homotaxial to the Penganga succession.

The Penganga Group

The Penganga Group around Adilabad (Fig. 2) unconformably overlies a granitic/gneissic basement, and is divided into three formations, which in the ascending order are the Pranhita Sandstone, the Chanda Limestone and the Sat Nala Shale (Chaudhuri *et al.* 1989). The Pranhita Sandstone is a cross-bedded, subarkosic sandstone of shallow shelf origin. It passes upward to the Chanda Limestone which consists essentially of plane-bedded micritic limestones with thin marl intercalations. The limestone is punctuated by lime–clast conglomerates of mass flow origin and turbidites at various stratigraphic levels. Current-/wave-generated structures or detritus coarser than fine silt is conspicuously absent except for in a few turbidite beds. Shallow-water calcareous constituents like ooids and stromatolites are also absent.

The Chanda Limestone can be subdivided into a number of stratigraphically controlled, colour-defined units (Fig. 3a). The basal unit is brown, which successively changes upward to pink, steel grey, black, steel grey and lastly to brown at the top, before grading into the brown Sat Nala Shale. The same order of superposition of the colour-defined units has been noted everywhere in the relatively shoreward part of the basin. Towards the interior, however, the lower brown, the pink and part of the lower steel grey limestone grade into a highly siliceous grey limestone (Fig. 3a & b). The succession in the interior part of the basin is characterized by abundant intraformational mass flow deposits, whereas these are only locally developed in the shoreward part. In both cases, however, the mass flow deposits are invariably interbedded with micritic limestone. Almost all the limestone intervals contain early diagenetic pyrite nodules, and the highest concentration of the pyrite has been recorded in the siliceous grey interval.

Manganese deposits of the Penganga Group

The manganese ore deposits are confined entirely to the siliceous grey limestone, and occur as thin, persistent stratiform bodies enclosed within the limestone with sharp upper and lower contacts. The ore-bearing intervals occur at least in two different stratigraphic levels, and the thicker horizon has an average

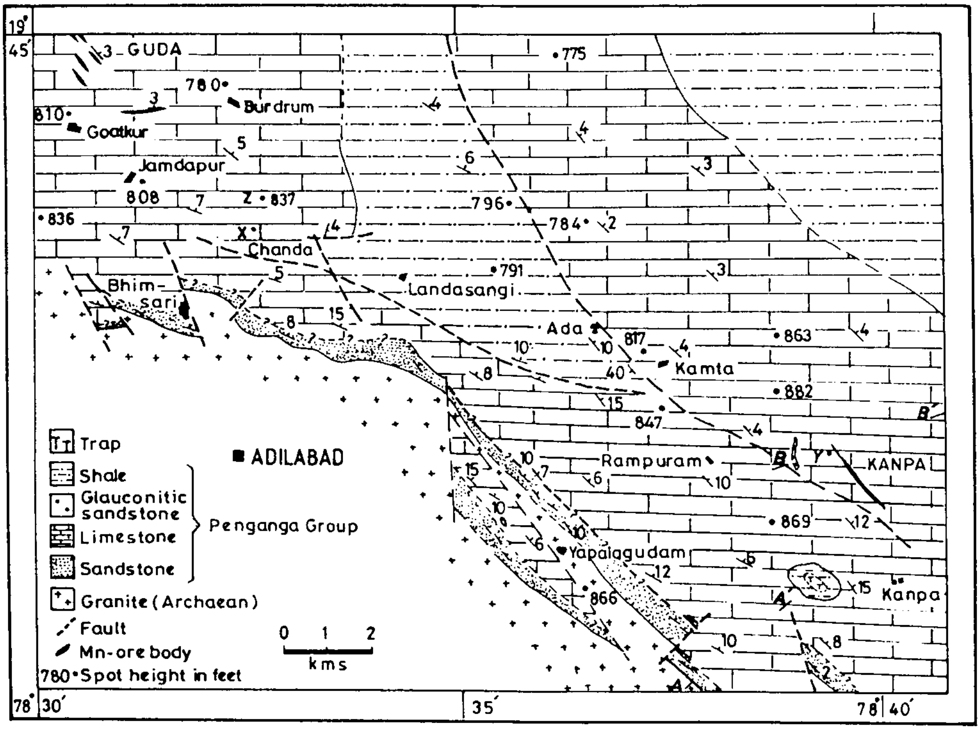

Fig. 2. Geological map of the Penganga Group around Adilabad, Andhra Pradesh, India (modified after Chaudhuri *et al.* 1989).

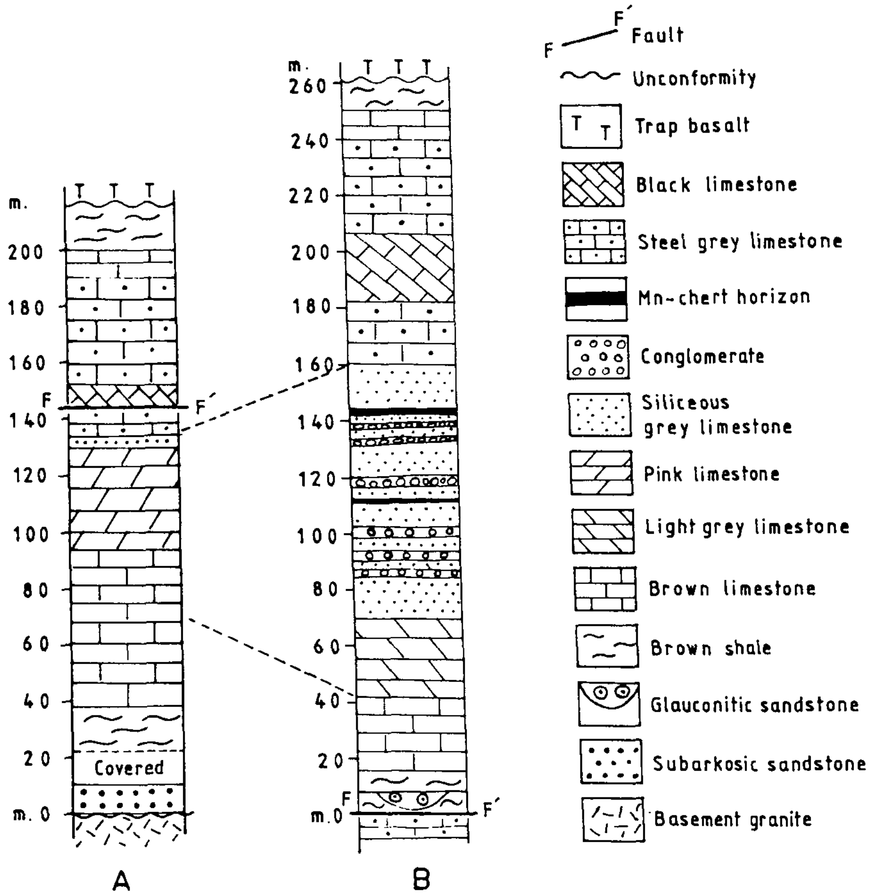

Fig. 3. Stratigraphic sections showing colour-defined intervals of the Chanda Limestone. (**a**) along AA and (**b**) along BB in Fig. 2.

total thickness of 70 cm. The ore-bearing horizons (Fig. 4) comprise thin beds and laminae of manganese oxide ores, ranging in thickness from 1 mm to approximately 5 cm, interstratified with comparably thick beds of chert–jasper and/or limestone.

Fig. 4. Manganese oxide ore beds interstratified with bedded chert.

Todorokite and birnessite are the major manganese oxide minerals with subordinate amounts of manganite, braunite, bixbyite, Psilomelane, pyrolusite and cryptomelane (see Roy *et al.* 1990 for details). Roy *et al.* (1990) interpreted the todorokite and birnessite as the pristine primary sedimentary minerals, which were subsequently diagenetically converted to other phases.

Siliceous grey limestones and their facies types

The siliceous grey limestone, the exclusive host of the manganese ore bodies, is a light bluish to light greenish micritic limestone containing about 20% silica. The rock is extremely homogeneous, both mineralogically and texturally throughout its outcrop, covering about 400 km^2 in the study area. The silica was identified as quartz in XRD analysis. It also occurs in the micrite size range and is very uniformly dispersed throughout the body of the limestone. The compositional homogeneity of the micritic beds is interrupted by beds slightly enriched in silica and clayey insoluble residues or by millimetres-thick pressure-solution seams. The siliceous limestone includes a large number of autoclastic mass flow deposits ranging from coarse debris flow conglomerates to calcarenites and possible calcisiltites. Precompactional authigenic pyrite is a common constituent of the limestone. It occurs as nodules varying in size from <1 cm to 10 cm and is found to be fairly uniformly distributed throughout the beds. Barite occurs locally as small nodules.

Facies types

Bedded micritic limestone facies. This is the most abundant of all the facies types and is characterised by five types of bedding style, labelled (a) to (e), described below. All the bed types are laterally very persistent, and may be traced for several tens to hundreds of metres, depending on exposure conditions.

Type a: thick, massive beds without any internal lamination (Fig. 5). The bed thickness normally ranges between 10 and 30 cm.

Type b: thick, massive beds with poorly developed, laterally impersistent, plane to wavy internal laminae (Fig. 6a).

Type c: thin, alternate silica-rich and silica-poor plane beds varying in thickness from 2 to 5 cm. The individual beds do not show any internal lamination.

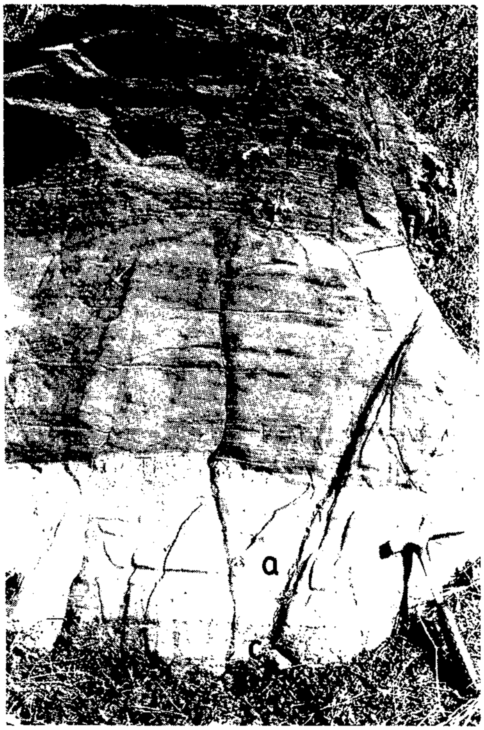

Fig. 6. Thick bed (**a**) of micritic siliceous limestone with poorly developed plane to wavy lamination (Type b) grades upward into thin plane-laminated (**b**) micritic limestone (Type d). Calcarenite bed (**c**) at the base.

Type d: plane to slightly undulatory thin laminae, ranging in thickness from 2 to 5 mm, closely associated with normally graded conglomerates, calcarenites and massive beds of type b (Fig. 6b).

Type e: thin beds, about 5 cm thick, with internal climbing ripple lamination (Fig. 7) characterized by well preserved stoss side and

Fig. 5. Thick massive even beds of micritic, siliceous grey limestone (Type a). Note insoluble residue-rich partings or thin beds (Type c) in between thick beds.

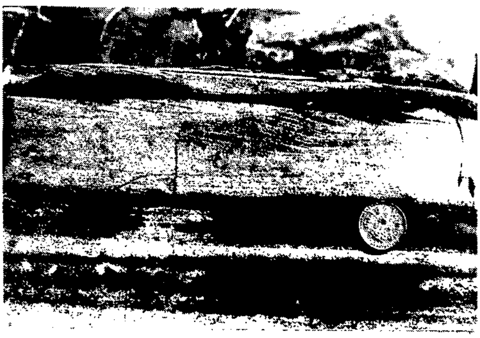

Fig. 7. Climbing ripple lamination within micritic siliceous grey limestone (Type e). Note low angle of climb.

low angle of climb (similar to type B of Jopling & Walker 1968).

Interpretation. The extremely homogeneous plane beds and laminae in micritic limestones without any shallow-water components closely resemble hemipelgic to pelagic slope carbonates. Lime-mudstones and calcisiltites with variable amounts of silt to clay-size insoluble residues are typical of ancient carbonate slope deposits (Wilson 1969; Cook & Enos 1977; Flügel 1982; Cook & Mullins 1983). In modern environments, fine-grained, undisturbed pelagic carbonate oozes normally are volumetrically most important on upper gullied part of the slope, rather than in the areas where they are diluted by sediment gravity flow deposits (Mullins & Neumann 1979). Thin, plane and climbing ripple lamination in the siliceous limestone strongly suggest emplacement by turbidity currents (Walker 1967; Mullins & Neumann 1979).

Lime-clast conglomerate facies. The lime-clast conglomerates comprise boulder to pebble size, platy, autoclastic lime-clasts set within lime-mud matrix. Clasts of bedded chert–jasper and interstratified chert and manganese occur in a few bodies. The conglomerate beds, in general, are laterally extensive sheets to tabular bodies with sharp lower and upper bounding surfaces. The lower bounding surfaces in several bodies are erosional whereas in the majority of the beds, they are non-erosive and planar. Thickness of the beds varies between 15 cm and 3.5 m. Individual beds may be clast-supported (Fig. 8), matrix-supported or are matrix-supported in the upper and clast-supported in the lower part. The matrix content, in general, is quite high, and may range up to about 25% by volume. In a majority of the conglomerate bodies, the clasts are highly disorganized (Figs 8 & 9) although these are oriented parallel to sub-parallel to the bounding surfaces in a few bodies. Most commonly, individual beds are either ungraded or normally graded (Fig. 10). About 15 cm thick inversely graded beds (Fig. 9) and thicker inverse to normally graded beds have been observed locally.

Fig. 9. Clast-supported, inversely graded conglomerate.

Interpretation. Complete absence of extrabasinal clasts or of clasts derived from shallow-water carbonates strongly suggests that the conglomerates were derived intrabasinally and originated within the slope environments. Poorly sorted, clast-supported to mud-supported fabrics, often with floating clasts in inverse, inverse to normally graded as well as ungraded massive beds strongly suggest that the conglomerates were emplaced by debris flows (Hampton 1975; Walker 1975; Middleton & Hampton 1976; Lowe 1982). The normally graded conglomerates with pebble to coarse sand-grade clasts and a few floating outsize clasts (Fig. 10) could be a product of cohesive turbulent debris flow or high, density turbidity flows (Lowe 1982). Inversely graded beds with appreciable amount of muddy matrix (Fig. 9)

Fig. 8. Clast-supported, ungraded lime-clast conglomerate of debris flow origin. Note boulder-size, disorganized clasts.

Fig. 10. Clast-supported, normally graded pebbly conglomerate. Note floating outsize clasts and plane laminations developed at the top.

are identified as deposits of modified grain flows which would require a high gradient of 9–14° to maintain its mobility (Lowe 1976). The debris flow comprising only of pelagic carbonates further suggests that the flows originated by remolding of submarine slides (Cook 1979*a*, *b*). Tabular bed morphology without any appreciable basal erosion suggests that majority of beds were emplaced on the lower slope (Cook & Mullins 1983) and a few could also extend onto the adjacent basin floor (Crevello & Schlager 1980).

Calcarenite facies. The calcarenites occur as thin sheets (Fig. 11) interbedded with the micritic limestones and consist exclusively of small sand-size clasts of siliceous micritic limestone within a matrix of same composition. The sheets normally range in thickness from 5 to 15 cm, but may vary between 1 and 25 cm. The calcarenite sheets typically have sharp lower bounding surfaces, and are often gradationally overlain by plane-laminated (bed type d) or thick-bedded micrites with poorly developed internal lamination (bed type b). The coarsest of the calcarenites are pebbly and matrix-rich, and may be ungraded or normally graded. Relatively thinner beds are poor in matrix, and show well-developed normal grading.

Interpretation. The lithological similarity between the clast and the matrix in the calcarenites with those of the conglomerates strongly suggests that the calcarenites represent the finer grained products of the slides that generated the conglomerates. Massive, indistinctly graded to ungraded, matrix-rich calcarenites with or without muddy capping are likely to be a product of muddy turbidity flows (Ghibaudo 1992; Mullins & Cook 1986). Thin, normally graded calcarenites with plane-laminated micrite at the top (Fig. 11) represent top-truncated Bouma sequences T_A to T_{AB} (see Cook & Mullins 1983). Although turbidites may be deposited at different parts of the slope, the top-truncated turbidites are commonly found in the lower slope environments (Cook & Mullins 1983).

Fig. 11. Normally graded calcarenite capped by thin plane laminae.

Facies sequences and associations

The gravity flow deposits do not show any organization that resemble submarine fan sequences. Rather, they are more randomly distributed with two end member types, the thin-bedded turbidites and thick-bedded debris flow conglomerates. Random association of coarse and fine mass flow deposits militates against a point source canyon-fan depositional system. The sheet-like geometry of the mass flow units with slightly erosional to non-erosional bases indicates that the debris were emplaced by sheet flows. The sequence resembles the slope apron deposits, that are characteristic of carbonate platform sequences world wide (Cook *et al.* 1972; Schlager & Chermak 1979; Crevello & Schlager 1980; Mullins *et al.* 1984; Mullins & Cook 1986).

Despite the fact that debris flow conglomerates occur randomly in deep-water carbonate slope environments, and that huge debris flows are capable of travelling down the slope into the distal basin plain (Crevello & Schlager 1980), the study of a number of sections led to the identification of three preferred facies sequences (*sensu* Walker & Mutty 1973) that can be interpreted in terms of a carbonate slope apron model (Mullins & Cook 1986).

Sequence 1. The sequence (Fig. 12) is characterised by thick conglomerates interbedded with thick-bedded micritic limestones. The conglomerates characteristically show erosional lower bounding surfaces and are normally highly muddy often with matrix-supported clasts ranging up to boulder size. Most of the conglomerates are ungraded or normally graded near the top. Inversely graded conglomerates, though rare, are typical of this sequence. Only thin units (about 7 cm) of interstratified bedded chert and manganese ore may occur in this sequence.

Sequence 2. The sequence (Fig. 12) is dominated by conglomerates and calcarenites interbedded with different types of beds of the micritic limestone, namely types a c and e. The conglomerates are extensive parallel-sided sheets without any evidence of erosion at their base. They are normally clast-supported and usually ungraded. Conglomerates normally grading

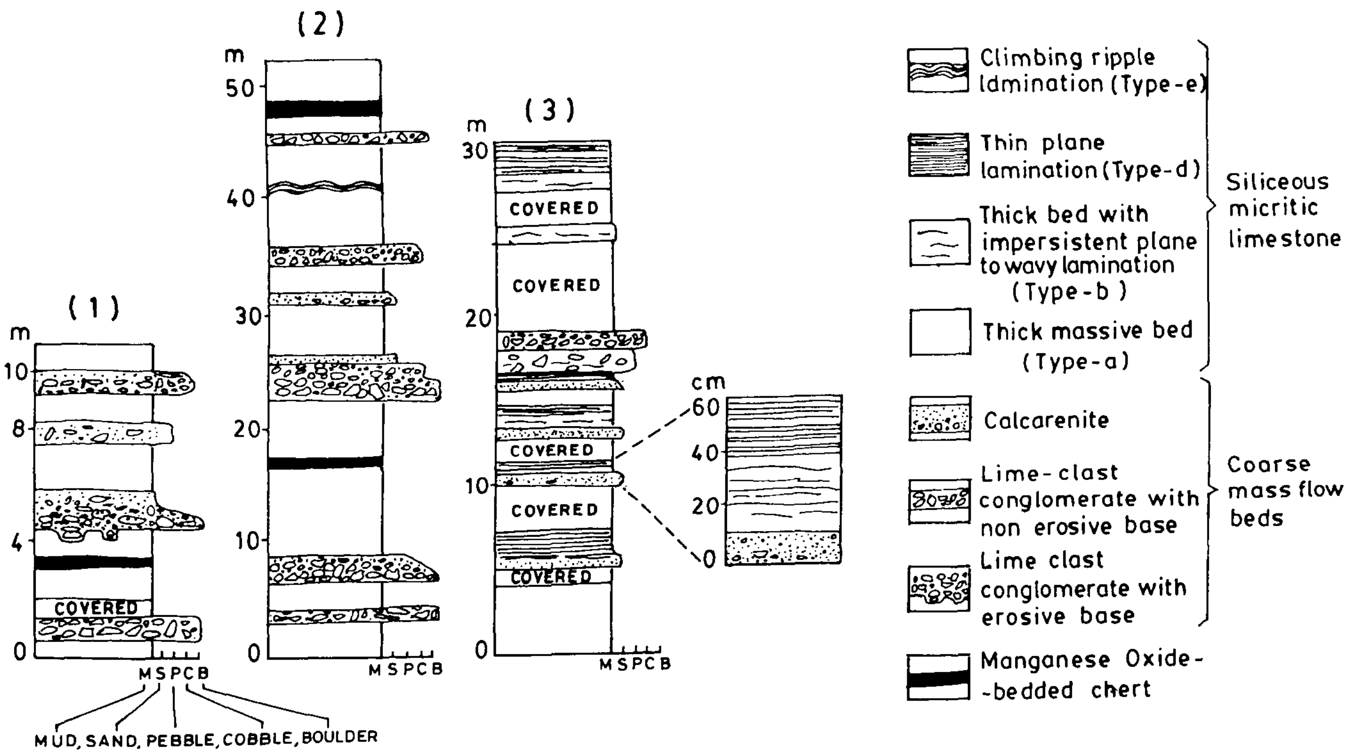

Fig. 12. Lithologs showing facies sequences: (**a**) Sheet-like lime-clast conglomerates with erosive base alternating with thick, massive micritic siliceous limestone. (Location X in Fig. 2). (**b**) Succession dominated by lime-clast conglomerate with non-erosive planar base. Conglomerates alternate with thick-bedded, massive micritic limestone and a few beds of calcarenite and climbing ripple laminated beds. (Location Y in Fig. 2). (**c**) Succession dominated by graded calcarenite and micritic limestone with thin lamination as well as thick beds with faint lamination. Lime-clast conglomerate beds are rare. (Location Z in Fig. 2).

upwards into micrites are, however, quite common. Clasts are generally smaller than those in the conglomerates of sequence 1. Calcarenites occur quite frequently, and are mostly top-truncated Bouma sequences, namely, T_A and T_{AB}. The manganese bearing horizons are almost exclusive to this sequence.

Sequence 3. The sequence (Fig. 12) is dominated by calcarenites and different types of micritic limestones. Coarse, chaotic conglomerates are rare. Thin sheets of normally graded or ungraded calcarenites interbedded with thick micritic beds with a poorly developed internal lamination or with units of very well developed thin plane laminae (types b and d) are typical of this sequence.

Facies association and depositional environment of the Penganga manganese

Monotonous, persistent plane-bedding and absence of any wave- or current-generated structures in the micritic limestones suggest a setting below storm wave base. Absence of any coarse clastics other than intraformational mass flow deposits is also consistent with this interpretation. Overall similarity of colour-defined intervals of the Chanda Limestone in terms of composition, structure and organisation suggests that all the intervals were deposited within a narrow bathymetric range and the succession evolved within a deep-water ramp setting. The relative abundance of mass flow deposits in the inner part, in comparison with shoreward part of the basin, suggests distal steepening of a gently sloping, deep carbonate platform. Interestingly, the siliceous grey limestone, the exclusive host of the manganese ore beds, is restricted to the slope developed at the distally steepened end of the platform.

Variations in the nature of the lower bounding surfaces of the slope-related conglomerates in the siliceous grey limestone, the clast size, the matrix content and the nature of the clast contacts, as well as the frequency of turbidites as opposed to debris flow beds collectively suggest that the facies sequences represent well-defined facies associations, and can be assigned to specific depositional environments. An overall off-slope fining trend with increasing abundance of calcarenite and calcisiltite characterize many carbonate slope environments (Cook & Mullins 1983; see also McIlreath & James 1978).

The basal erosional surfaces common to the conglomerates of sequence 1 suggest deposition within the cannibalistic zone of the slope. Turbidites and coarse debris flow conglomerates with broad shallow channelization are typical of the inner apron of the base of a slope setting (Mullins & Cook 1986). In contrast with sequence 1 sequence 3 is dominated by fine calcarenites and thin, plane-laminated micrites representing the more distal environments. They closely resemble the base of the slope to basin plain palaeoenvironment. Sequence 2 is

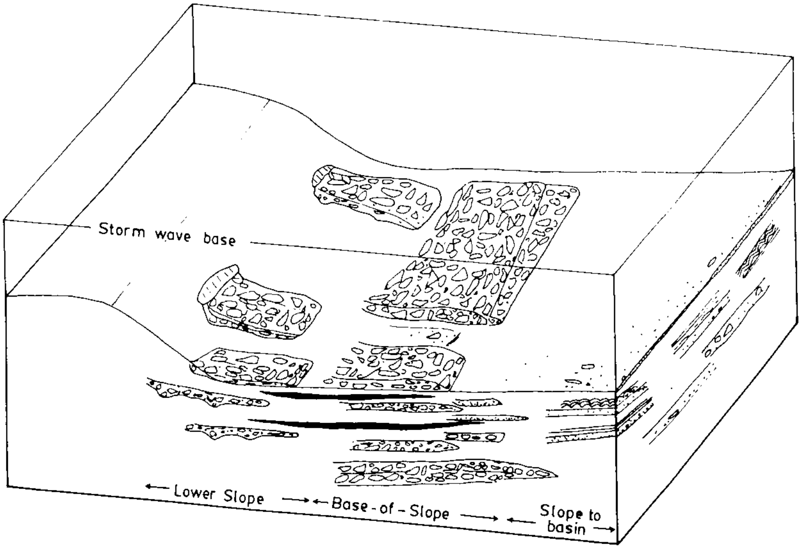

Fig. 13. Schematic model of palaeoenvironmental setting for manganese mineralization. Mn-ore horizons are restricted within lower slope to base of slope setting.

dominated by clast-supported conglomerates without any basal erosion conforming to the outer apron palaeoenvironment (cf. Mullins & Cook 1986), located intermediate between the inner apron and the zone of slope to basin transition. In all measured sections limestone conglomerates are intercalated with hemipelagic to pelagic micritic limestones. The manganese ore beds interstratified with bedded chert occur in the hemipelagic to pelagic intervals of mainly outer apron and only rarely in the inner apron environments of the base of slope setting (Fig. 13).

Discussion on the constraints of the origin of Penganga manganese ores

The Penganga deposits stand apart from most other known manganese deposits in terms of their mode of occurrence and depositional setting. Neoproterozoic manganese deposits are often associated with banded iron formations. Precambrian interlayered chert–manganese oxide deposits, on the other hand, are uncommon. Besides the occurrence around Adilabad (India), chert–manganese deposits are known from the little studied Archaean Chitradurga Group of South India (Roy 1981). Roy *et al.* (1990) proposed that Penganga manganese oxide deposits are of sedimentary origin and, based on Chaudhuri *et al.* (1989), concluded that the manganese formed below wave base. Analysis of trace and rare earth elements led Beukes (1989) to suggest that the manganese deposits in the Proterozoic Transvaal Supergroup were derived from deep marine water with a small hydrothermal component, without specifying the bathymetry of the site of accumulation. Bühn *et al.* (1992) interpreted that the manganese deposits in the Neoproterozoic Damara Sequence of Namibia formed under pelagic conditions in the outer shelf environments and related the development of shelf pelagic condition to a major transgressive event. The geological setting of the Penganga manganese deposit in broad terms is not far from that of the Namibian manganese deposits. However, the Penganga deposit is associated with the base of a slope of a distally steepened, deep carbonate platform, whereas the Namibian deposit formed in a siliciclastic, outer shelf environment. Again, unlike the Damara deposit, the Penganga deposit is free from any association with iron formation.

Sedimentary manganese deposits in shelf settings have often been linked with oceanic upwelling during transgressive–regressive cycles within an overall transgressive phase (Cannon & Force 1983; Frakes & Bolton 1984; Bolton & Frakes 1985; Force & Cannon 1988). Deep-water manganese deposits of the Damara Sequence have also been related to transgressive phase with attendant upwelling (Bühn *et al.* 1992, 1993). The carbonates overlying shallow-water siliciclastics in the Penganga succession indeed indicate a major transgression. Applicability of the hypothesis of upwelling or any inference on the origin for the Penganga manganese deposit, however, must take into consideration their base of slope depositional setting and lack of association with iron formation.

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References

BEUKES, N. J. 1983. Paleoenvironmental setting of iron-formations in the depositional basin of the Transvaal Supergroup, South Africa. *In*: TRENDALL, A. F. & MORRIS, R. C. (eds) *Iron-formations: Facts and Problems*. Amsterdam, Elsevier, 139–209.

—— 1989. Sedimentological and geochemical relationships between carbonate, iron-formation and manganese deposits in the Early Proterozoic Transvaal Supergroup, Griqualand West, South Africa. *Abstract volume, 28th International Geological Congress, Washington, DC, USA*, 143.

BOLTON, B. R. & FRAKES, L. A. 1985. Geology and genesis of manganese oolite, Chiatura, Georgia, USSR. *Geological Soceity of America Bulletin*, **96**, 1398–1406.

BÜHN, B., STANISTREET, I. G. & OKRUSCH, M. 1992. Late Proterozoic outer shelf manganese and iron deposits at Otjosondu (Namibia) related to the Damara Oceanic opening. *Economic Geology*, **87**, 1393–1411.

——, —— & —— 1993. Presevation of sedimentary features in Late Proterozoic manganese and iron-formations (Namibia) through upper amphibolite facies metamorphism: protoliths, paleoenvironment and ore genesis. *Resource Geology, Special Issue*, **17**, 12–26.

CANNON, W. F. & Force, E. R. 1983. Potential for high-grade shallow marine manganese deposits in North America. *In*: SHANKS, W. C. (ed.) *Unconventional mineral deposits*. New York, American Institute of Mining Metallurgy and Petroleum Engineers, 175–190.

CHAUDHURI, A. & CHANDA, S. K. 1991. The Proterozoic Basin of the Pranhita–Godavari Valley: An overview. *In*: TANDON, S. K., PANT, C. C. & CASSHYAP, S. M. (eds) Sedimentary Basins of India. Gyanodaya Prakashani, Nainital, 13–29.

—— & HOWARD, J. D. 1985. Ramagundam Sandstone: Middle Proterozoic shoal bar sequence. *Journal of Sedimentary Petrology*, **55**, 392–397.

——, DASGUPTA, S., BANDYOPADHYAY, G., SARKAR, S., BANDYOPADHYAY, P. C. & GOPALAN, K. 1989. Stratigraphy of the Penganga Group around Adilabad, Andhra Pradesh. *Journal of Geological Society of India*, **34**, 291–302.

COOK, H. E. 1979*a*. Ancient continental slope sequences and their value in understanding modern slope development. *In*: DOYLE, L. S. & PILKEY, O. H. (eds) *Geology of continental slopes*. Society of Economic Paleontologists and Mineralogists, Special Publications, **27**, 287–305.

—— 1979b. Generation of debris flows and turbidity current flows from submarine slides. *American Association of Petroleum Geologists Bulletin*, **63**, 435.

—— & ENOS, P. 1977. *Deep-water carbonate environments*. Society of Economic Paleontologists and Mineralogists, Special Publication, **25**.

—— & MULLINS, H. T. 1983. Basin margin environment. *In*: SCHOLLE, P. A., BEBOUT, D. G. & MOORE, C. H. (eds) *Carbonate depositional environments*. American Association of Petroleum Geologists Memoir, **33**, 540–616.

——, MCDAVIED, P. N., MOUNTJOY, E. W. & PRAY, L. C. 1972. Allochthonous carbonate debris flows at Devonian bank ('reef') margins. Alberta, Canada. *Bulletin of the Canadian Petroleum Geologists*, **20**, 439–497.

CREVELLO, P. D. & SCHLAGER, W. 1980. Carbonate debris sheets and turbidites, Euxoma Sound, Bahamas. *Journal of Sedimentary Petrology*, **50**, 1121–1147.

FORCE, E. R. & CANNON, W. F. 1988. Depositional model of shallow marine manganese deposits around black shale basins. *Economic Geology*, **83**, 93–117.

FLÜGEL, E. 1982. *Microfacies analysis of limestones*, Springer Verlag, Berlin, 633pp.

FRAKES, L. A. & BOLTON, B. R. 1984. Origin of manganese giants: sea level changes and anoxic-oxic history. *Geology*, **12**, 83–86.

GHIBAUDO, G. 1992. Subaqueous sediment gravity flow deposits: practical criteria for their field description and classification. *Sedimentology*, **39**, 423–454.

HAMPTON, M. A. 1975. Competence of fine-grained debris flows. *Journal of Sedimentary Petrology*, **45**, 834–844.

JOPLING, A. V. & WALKER, R. G. 1968. Morphology and origin of ripple-drift cross lamination, with examples from the Pleistocene of Massachusettes. *Journal of Sedimentary Petrology*, **38**, 971–989.

LOWE, D. R. 1976. Grain flow and grain flow deposits. *Journal of Sedimentary Petrology*, **46**, 188–199.

—— 1982. Sediment gravity flows II: Depositional models with special reference to deposits of high density turbidity currents. *Journal of Sedimentary Petrology*, **52**, 279–298.

MCILREATH, I. A. & James, N. P. 1978. Facies models-13. Carbonate slopes. *Geoscience Canada*, **5**, 189–199.

MIDDLETON, G. V. & HAMPTON, M. A. 1976. Subaqueous sediment transport and deposition by sediment gravity flows. *In*: STANLEY, D. J. & SWIFT, D. J. P. (eds) *Marine sediment transport and environmental management*. New York, John Wiley and Sons, 197–218.

MULLINS, H. T. & Cook, H. E. 1986. Carbonate apron models: alternatives to the submarine fan model for paleoenvironmental analysis and hydrocarbon exploration. *Sedimentary Geology*, **48**, 37–79.

—— & NEUMANN, A. C. 1979. Deep-carbonate bank margin structure and sedimentation in the Northern Bahamas. *In*: DOYLE, L. S. & PILKEY, O. H. (eds) *Geology of continental slopes*. Society of Economic Paleontologists and Mineralogists, Special Publication, **27**, 165–192.

——, HEATH, K. C., BAREN, H. M. & NEWTON, C. R. 1984. Anatomy of modern open ocean carbonate slope: Northern Little Bahama Bank. *Sedimentology*, **31**, 141–168.

NAQVI, S. M. & ROGERS, J. J. W. 1987. *Precambrian Geology Formations of India*. Oxford University Press, Oxford.

ROY, S. 1981. *Manganese Deposits*. Academic Press, London, 458pp.

—— 1988. Manganese metallogenesis: A review. *Ore Geology Reviews*, **4**, 155–170.

——, BANDYOPADHYAY, P. C., PERSEIL, E. A. & FUKUOKA, M. 1990. Late diagenetic changes in manganese ores of the Upper Proterozoic Penganga Group, India. *Ore Geology Reviews*, **5**, 341–357.

SCHLAGER, W. & CHERMAK, A. 1979. Sediment facies of platform–basin transition, Tongue of the Ocean, Bahamas. *In*: DOYLE, L. S. & PILKEY, O. H. (eds) *Geology of continental slopes*. Society of Economic Paleontologists and Mineralogists, Special Publications, **27**, 193–208.

WALKER, R. G. 1967. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. *Journal of Sedimentary Petrology*, **37**, 25–43.

—— 1975. Generalised facies models for resedimented conglomerates of turbidite association. *Geological Society of America Bulletin*, **86**, 737–748.

—— & Mutti, E. 1973. Turbidite facies and facies associations. *In*: MIDDLETON, G. V. & BOUMA, A. H. (eds) *Turbidites and deep water sedimentation*. Anaheim, California. Society of Economic Paleontologists and Mineralogists, Pacific Section Short Course, 119–158.

WILSON, J. L. 1969. Microfacies and sedimentary structures in "deeper water" lime-mudstone. *In*: FRIEDMAN, G. M. (ed.) *Depositional environments in carbonate rocks*. Society of Economic Paleontologists and Mineralogists, Special Publications, **14**, 4–19.