ECONOMIC GEOLOGY RESEARCH UNIT

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IRON FORMATION AS AN END-MEMBER IN CARBONATE SEDIMENTARY CYCLES IN THE TRANSVAAL SUPERGROUP, SOUTH AFRICA

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by

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ABSTRACT

The deposition of iron-rich sediments in the dolomite of the Transvaal Supergroup some 2 200 million years ago was not a random process. Iron formation precursor can be shown to be an end-member in a series of well-defined chemical and sedimentary cycles. The stratigraphic record bears witness to a repeated chemical conditioning of the depositional environment, which resulted in the deposition of iron-rich carbonates, iron formation precursor, and of banded iron formation. Chemical conditioning can usually be related to carbonate depositional structures suggestive of the presence of shoals in the basin. It is envisaged that these banks or shoals of carbonate detritus caused a partial restriction between the sedimentary environment and the open ocean. The abnormal concentrations of iron, silicon, and other elements which developed in these restricted basins resulted in the precipitation of successively more iron-rich sediments and chert, and culminated in deposition of banded iron formation.

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CONTENTS

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INTRODUCTION

The origin of banded iron formation has been the subject of geological debate for many years. A uniformitarian approach to the problem has had only partial success, since recent analogues of banded iron formation are apparently unknown. Most investigators have examined the problem through detailed studies of the petrology, mineralogy, chemistry, and sedimentational features of the iron formation itself. This approach has resulted in the acquisition of a great deal of useful new information, but has not produced any widespread consensus regarding the origin of this peculiar sediment-type. In this paper, the results of a different approach to the problem are outlined. Iron-rich assemblages are viewed in their stratigraphic setting, and are seen to occupy specific niches in carbonate cycles. A study of the sedimentary structures developed within the carbonate rocks provides a basis for deciphering the origin of the contiguous iron formation.

IRON FORMATION IN THE TRANSVAAL SUPERGROUP : REGIONAL SETTING

The banded iron formation to be discussed here is a widely distributed unit within the Transvaal Basin. The basin is developed on the Kaapvaal block, one of the oldest stable continental nuclei of the African continent. This crustal unit has been tectonically and thermally stable for at least 2 500 million years. The Transvaal sediments were deposited on this continental block in environments which varied from fluvial, through deltaic, to marine. The carbonates deposited in the basin are abundantly stromatolitic over large areas, and show many of the characteristics of present-day shelf-carbonate deposits.

The Transvaal Supergroup has an age which falls between the limits of 2 100 and 2 300 m.y. (Davies and others, 1969; Van Niekerk and Burger, 1964). Recently, a basaltic andesite interbedded in the Pretoria Group (the uppermost broad unit within the Transvaal Supergroup) has been dated at 2224 ± 21 million years (D. Crampton, Bernard Price Institute for Geophysical Research, personal communication, 1972).

The area studied during the present investigation is situated at the eastern extremity of the Transvaal Basin, in the northeastern portion of the Transvaal Province, Republic of South Africa. In this region, the Transvaal sediments rest upon an Archaean basement and are intruded and overlain by the mafic phase of the Bushveld Complex (Figure 1).

The Transvaal Supergroup is subdivided (from base to top) into four major stratigraphic units, the Wolkberg Group (0-1 900 metres), the Black Reef Quartzite (0-500 metres), the Olifants River Group, (100-2 500 metres), and the Pretoria Group (up to 7 000 metres). The relationships to be described were encountered in the carbonates and iron formation of the Olifants River Group.

STRATIGRAPHY OF THE OLIFANTS RIVER GROUP

The Olifants River Group (the Dolomite Series of the South African literature) is comprised predominantly of chemical sediments. It is subdivided into the Malmani Dolomite (lower Dolomite Stage), the Penge Formation (Banded Ironstone Stage) and the Duitschland Formation (Upper Dolomite Stage). The Pretoria Group covers these units unconformably, and, towards the south, progressively eliminates the upper formations, coming to rest on lower units in this direction (see Figure 1).

In the eastern and northeastern portions of the Transvaal Basin, the Malmani Dolomite has been subdivided into five regionally-persistant gross-lithologic units (Button, 1973). The lowest

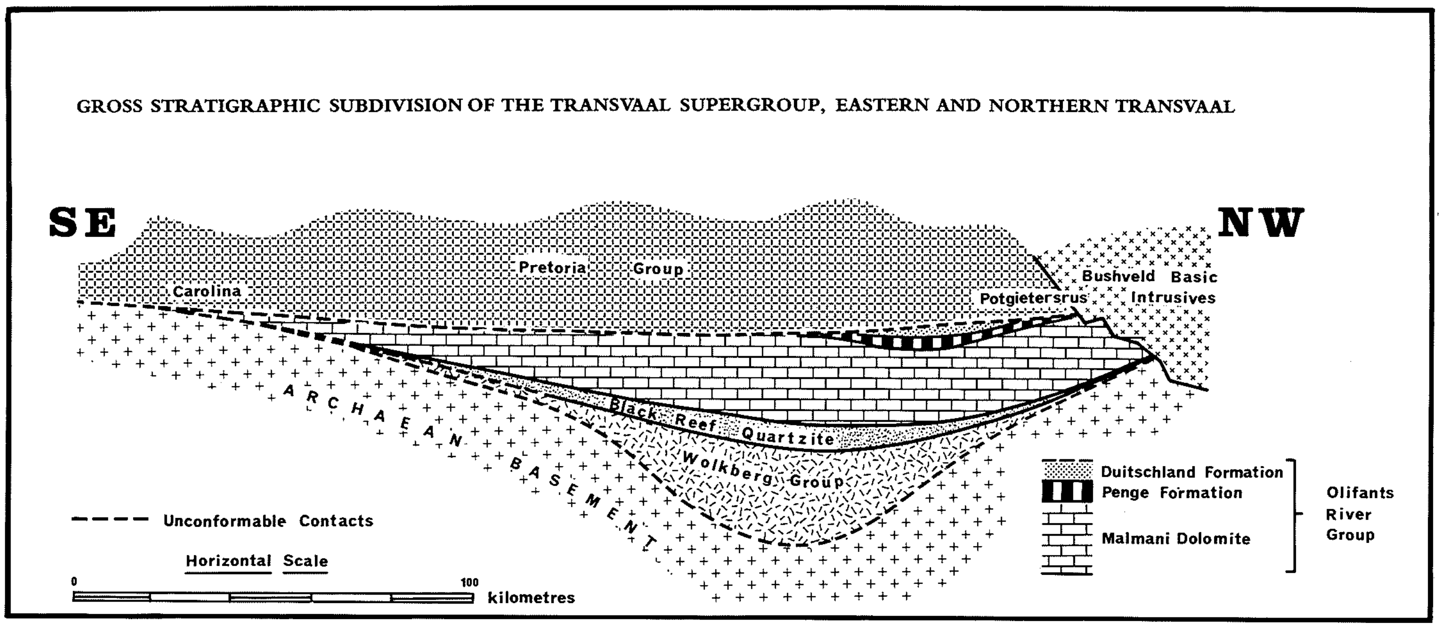

Figure 1 : Regional stratigraphic subdivision of the Transvaal Supergroup in the eastern Transvaal.

unit is the transition zone which grades downwards to the underlying Black Reef Quartzite. It is composed of iron- and manganese-rich dolomite, is poor in chert, and is associated with abundant beds of quartzite and carbonaceous mudstone (Figure 2). The transition zone, which attains thicknesses of up to 200 metres, grades up to the lower dolomite and chert zone, a unit consisting of light-coloured dolomite with abundant chert. This subdivision, which is up to 650 metres thick, is capped by a disconformity that is marked by a thin discontinuous layer of chert-in-shale breccia.

A new sedimentary cycle commences with the deposition of a laterally impersistent arenaceous unit (the Blyde River Quartzite), followed by layers of carbonaceous mudstone in iron- and manganese-rich, dark-coloured, chert-poor dolomite. A thin body of primary limestone is preserved sporadically near the base of this 120-metre-thick unit. The assemblage, in many ways similar to the basal transition zone, grades upwards to the 450-metre-thick upper dolomite and chert zone (lighter-coloured dolomite with abundant chert). Once again an intraformational disconformity terminates this unit. The final mega-cycle of the Malmani Dolomite starts with a thin chert-in-shale breccia, which is overlain by a layer of carbonaceous mudstone (up to 20 metres in thickness). Some 350 metres of carbonate sediments overly the mudstone, and constitute what has been termed the mixed zone of the Malmani Dolomite (Button, 1973). This unit and its transition to the iron formation of the overlying Penge Formation provide most of the geologic relationships on which this paper is based.

STRATIGRAPHIC RELATIONSHIPS IN THE MIXED ZONE AND THEIR BEARING ON IRON FORMATION GENESIS

Broadly, the mixed zone consists of a dolomite, often dark-coloured and relatively free of chert. Beds of carbonaceous mudstone are present, especially near the base of the unit. Other sediment types include bodies of primary limestone and layers of a rock here termed proto-iron formation or iron formation precursor. The latter is an assemblage of ferruginous carbonates with layers of chert and grunerite-rock or grunerite-bearing mudstone. The grunerite-bearing rock often includes layers rich in scattered crystals of magnetite or hematite. Lenses of amosite (the fibrous polymorph of grunerite) are also present in the assemblage.

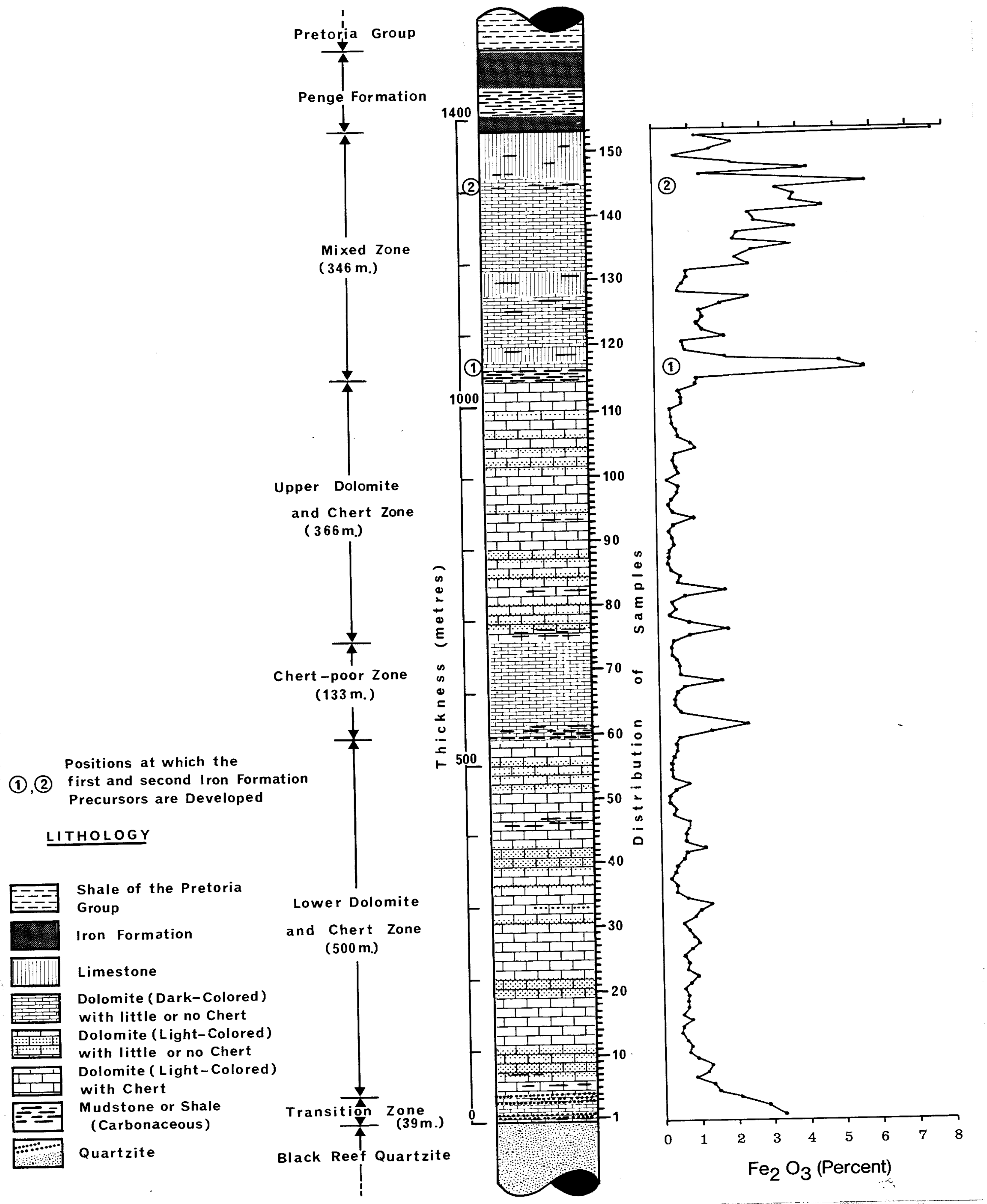

Figure 2 : The stratigraphy of the Malmani Dolomite and the Penge Formation, with a profile showing the distribution of iron through the former.

Iron formation precursor has been seen to grade into true iron formation, and is, in places, interlayered with it. Field evidence indicates that the precursor sediment was deposited in a setting closely allied to that in which iron formation formed. Conclusions regarding the origin of iron formation precursor can be applied (with modifications) to the principal iron formation of the Transvaal Supergroup.

(a) *Primary Nature of Limestone of the Mixed Zone*

For the purpose of this paper, it is necessary to stress that the carbonates of the mixed zone were probably all originally limestones. Field evidence for this conclusion is abundant. Firstly, many of the major limestone bodies occur at different stratigraphic levels in the mixed zone. Secondly, limestone bodies are invariably surrounded by an envelope of partially dolomitized carbonate (Figure 3). In these envelopes, dolomitization can be seen to have proceded along and across bedding planes. In some cases, limestone remnants are found as isolated lenses spread out at intervals in a bed of dolomite. Elsewhere, partial dolomitization accentuates sedimentary structures. In one instance, for example, small stromatolite columns are dolomitized, while the inter-column fill consists of a primary limestone detritus.

Further evidence for the primary nature of limestone lies in the degree of preservation of sedimentary structures. In the limestone, especially where not thermally metamorphosed by the Bushveld Complex, delicate sedimentary structures are perfectly preserved. Fossilized algal threads have recently been found in this primary carbonate. Zones containing well-defined structures in limestone, when followed along a plane of stratification, are seen to pass into dolomite which displays, at best, weak preservation of primary sedimentation structures. Elsewhere, the dolomite is often devoid of sedimentary structures.

(b) *Simultaneous Dolomitization and Iron-Enrichment of Primary Limestone*

The primary limestone contains relatively small proportions of iron. Analysis of 14 specimens revealed an average total iron content of the limestone of 1.4 percent (as Fe_2O_3); half the samples contained less than 1.0 percent of the oxide. Conversely, the replacing dolomite contains, on the average, 3.2 percent Fe_2O_3 (average of 24 analyses) and, in places, carries as much as 13 percent of this oxide. This relation is highlighted by the weathering process. Grey-weathering limestone remnants are found in dolomite which weathers chocolate-brown, russet, mustard, or vivid yellow. The conclusion that much of the iron in the dolomite was introduced during dolomitization of an iron-poor limestone is strongly put forward, and is the second critical step in the main argument.

The dolomitization process is known, from a number of modern cases, to proceed pene-contemporaneously with primary calcium carbonate deposition (Deffeyes and others, 1965; Illing and others, 1965). It follows that the dolomitizing solution (presumably the waters immediately above the sediment-water interface) built up significant concentrations of iron and magnesium. It is believed that gravity-induced fluxing of unconsolidated calcium carbonate sediments by these relatively high specific gravity brines provides the most suitable explanation for the origin of the iron-rich dolomite. Beneath the principal iron formation, field evidence suggests that these brines were of a different composition, the supply of Mg cations having been exhausted. Here, primary limestone was replaced by iron-rich carbonates, a process of partial sideritization.

(c) *Cyclical Distribution of Iron and Iron Formation Precursor Through the Malmani Dolomite*

It is instructive to examine the iron distribution pattern through the Malmani Dolomite, as a whole, and through the mixed zone, in particular. Grab samples of carbonate, spaced nine metres apart, were collected from the Malmani Dolomite in a single traverse in the northeastern Transvaal. The iron distribution curve (percentage as Fe_2O_3) is plotted alongside the stratigraphy in Figure 2.

One diffuse and four well-defined iron peaks are present. The cyclical alternation of iron-poor and iron-rich sediment is strikingly apparent. The lowest peak is associated with the dark-coloured dolomite of the transition zone. The multiple small peaks near the centre of the column are associated with the dark dolomites of the chert-poor unit. The three prominent peaks reflect carbonate in the mixed zone near the top of the succession. Using the arguments developed

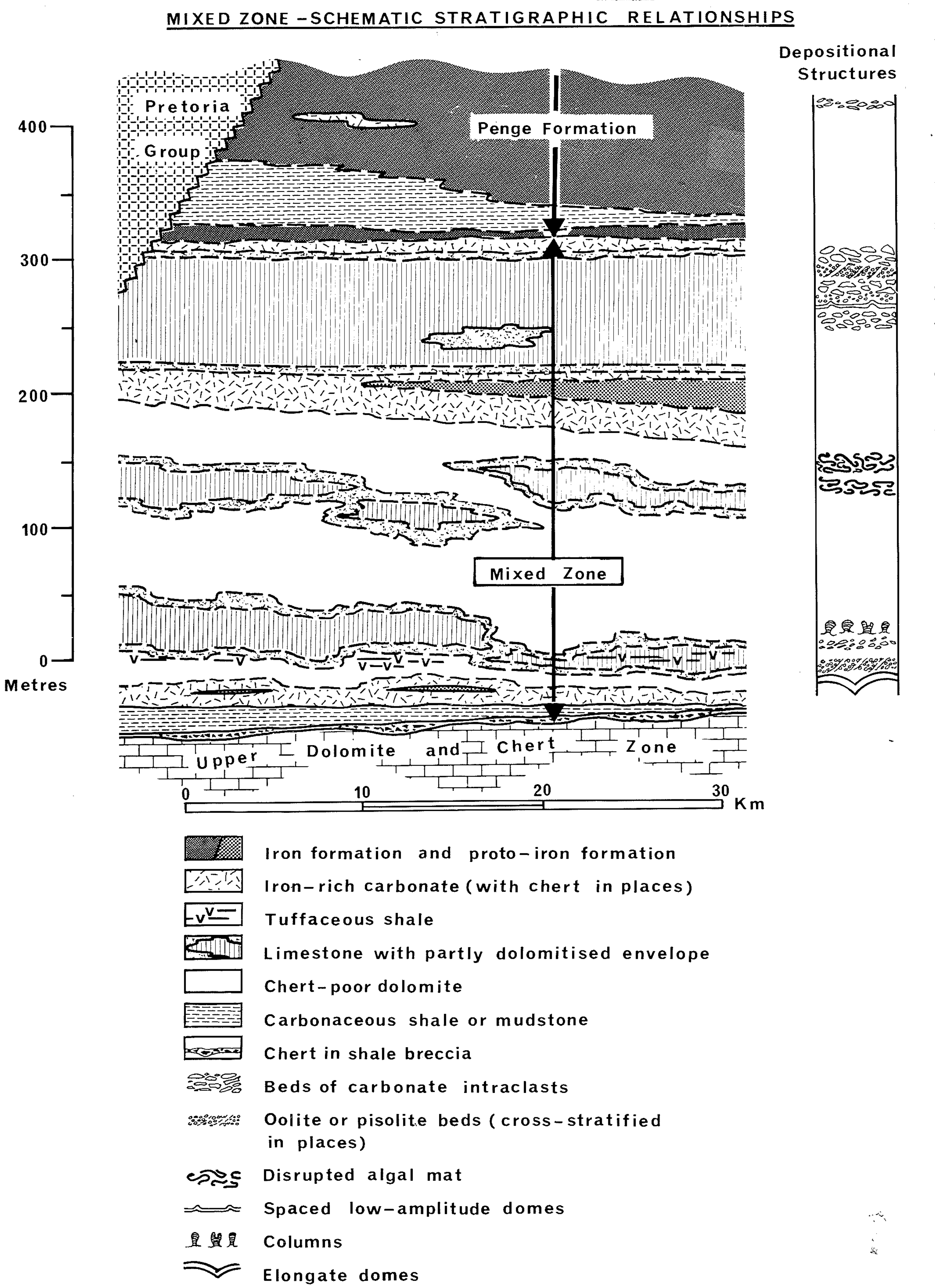

Figure 3 : Schematic stratigraphy of the mixed zone, Malmani Dolomite, in the eastern and northeastern Transvaal.

above, it can be stated that the waters of the Transvaal epeiric sea were cyclically enriched in iron. It should be noted that the iron peaks in the mixed zone show progressively higher values upwards in the unit.

The beds and lenses of iron formation precursor occupy well-defined positions in the mixed zone (Figure 3). It is no coincidence that the horizons at which proto-iron formation are developed coincide with the three iron peaks of the mixed zone shown in the iron-distribution diagram of Figure 2.

It is logical to conclude that iron formation precursor must have been deposited from waters significantly enriched in iron. The horizons of iron formation precursor are over- and underlain by iron-rich carbonates. It follows that the process of iron enrichment (with or without accompanying dolomitization) of the primary limestone must have occurred almost immediately after precipitation of the latter. In no other way can the juxtaposition of proto-iron formation and iron-rich carbonates be satisfactorily explained.

Reference to the schematic representation of mixed zone stratigraphy (Figure 3) indicates that iron formation precursor was deposited at two stratigraphic levels before the principal iron formation of the supergroup was deposited. The lower precursor layer is thin (generally less than one metre thick) and is laterally impersistent. It occurs in a thicker body (some 10 to 20 metres) of iron-rich dolomite which can be traced for nearly 200 kilometres along strike. The second iron formation precursor varies from 20 to 30 metres in thickness, and can be followed for about 90 km along strike. It occurs in a greater thickness of chert-bearing, iron-rich dolomite which is laterally more extensive than the proto-iron formation which it encloses. A third iron-rich unit (the basal iron formation unit in the Penge Formation) can be followed for about 50 km along strike. Its original extent is unknown, since, on its south-eastern extremity, it has been removed by pre-Pretoria Group erosion. This iron formation grades upwards to a thickness of carbonaceous mudstone, which passes, in turn, to the principal iron formation in the Penge Formation.

The pattern of development of iron formation precursor is one of thin, laterally impersistent units at the base of the mixed zone, giving way upwards to a thicker and laterally more persistent unit which culminates in the iron formation of the Penge unit. It is stressed that the main iron formation of the Transvaal basin did not represent a sudden event unrelated to the previous history of the basin. To the contrary, iron-rich carbonates were formed in early cycles, followed by iron-rich carbonates with proto-iron formation in later cycles, the cyclical pattern of iron enrichment culminating in the thick and extensive Penge Formation.

MECHANISM OF IRON CONCENTRATION

To understand the mechanism for the cyclical change in iron concentration, with time, in portions of the Transvaal epeiric sea, changes in iron content of the carbonates have been related to other cyclical changes in the dolomite stratigraphy. On a mega-cyclical scale, chert-poor, dark-coloured dolomite (rich in iron and manganese) is generally found at the base of transgressive cycles, which often cap intraformational disconformities. These iron-rich dolomites are found in association with higher-than-normal amounts of clastic sediment (mudstone or quartzite or both), and, in many cases, with remnant bodies of primary limestone. This mixed clastic and chemical assemblage was probably laid down at the interface of a delta, depositing quartzite and mudstone, and an epeiric sea, depositing carbonate sediments. Minor facies shifts or migrations of fluvial distributaries resulted in deposition of the pile of carbonate and clastic sediments which mark the bases of these transgressive cycles.

Eriksson (1971), in discussing the iron distribution in the middle and lower portions of the Malmani Dolomite in an area to the southwest of Johannesburg, stated that higher-than-normal iron and manganese concentrations were considered to be indicative of a coastal environment. He favoured the mechanism implicit in the findings of Wolf and others (1967) that, in arid regions, higher-than-normal concentrations of iron and manganese occur in near-shore carbonates due to evaporative concentration of these elements in sea water.

In the transgressive mega-cycles of the dolomite in the northeastern Transvaal, and elsewhere, field-studies have shown that iron-rich carbonates pass upwards into stratigraphic units in which signs of clastic carbonate sedimentation are abundant. The dark dolomites of the basal transition zone pass upwards into sediments in which beds of oölites are abundant (Eriksson, 1972; Button, 1973). Similarly, the iron-rich dolomite (with iron formation precursor) at the base of the mixed zone is overlain by dolomite and limestone in which oölite or pisolite beds are developed, together with layers of intraclastic breccia.

The above cases involve transgressive hemi-cycles; there are, however, two regressive situations in the mixed zone, where iron-poor carbonates grade upwards to iron-rich carbonates with proto-iron formation. In the first case, below the second iron formation precursor (Figure 3), layers of an intraclastic breccia are present, the clasts consisting of fragments of a once-pliable algal mat, now often contorted and overfolded. Davies (1970) described identical disrupted algal mats from recent sediments of Shark Bay, Western Australia. In the second case, below the lower of the Penge iron formation units, bed upon bed of intraclastic conglomerate and breccia are preserved, together with abundant oölite layers which are sometimes cross-bedded. Individual slabs of carbonate in the breccia are up to 50 cm long and 15 cm thick.

The critical relationship is that, in many cycles, both transgressive and regressive, carbonate detritus can be inferred to have been present in an off-shore sense with respect to the site of iron-rich carbonate deposition. The situation envisaged is shown in a schematic way in Figure 4 in which the profile of a hypothetical shoreline during mixed zone time is portrayed. At any one time, thin beds of mudstone were being deposited in a deltaic situation. At the delta front, intertonguing of carbonates with these allochthonous sediments occurred. At some distance off the delta front, a bank or shoal of carbonate detritus is inferred to have been present. This barrier, formed in the turbulent zone where oceanic waves bottom out, was responsible for impeding the free communication of oceanic waters with those in the lagoon or barred basin.

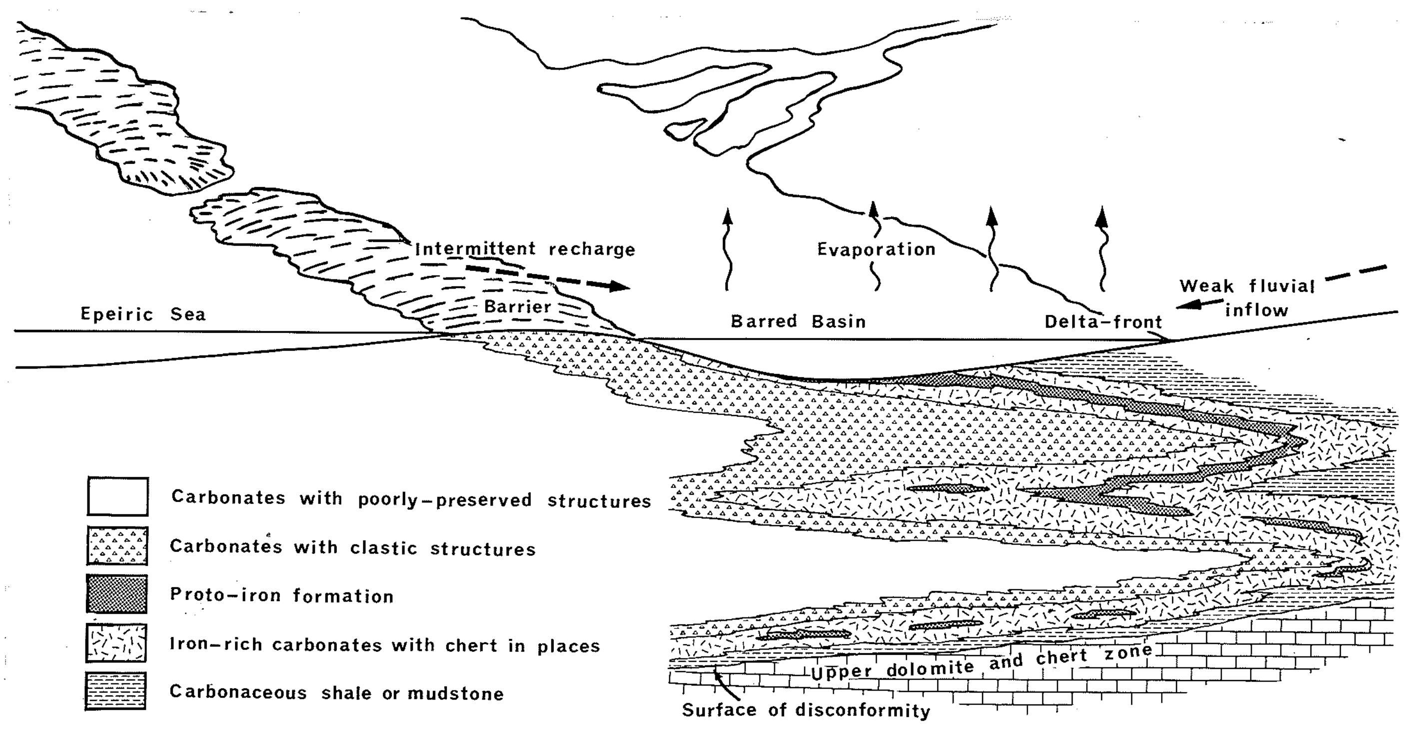

Figure 4 : Three-dimensional diagram showing the envisaged facies-belts during mixed zone times, and the cyclical lithology produced by transgressive and regressive facies-belt shifts.

The size of the evaporitic basin must have varied with time, but was certainly measurable in terms of hundreds of kilometres, as shown by field evidence in the area studied. In Figure 4, the vertical scale has been grossly exaggerated to show stratigraphic relationships. Bottom-slopes were probably measurable in terms of a few centimetres per kilometre. The shoal or bar probably rose only a few tens of centimetres above mean high-tide.

Within the restricted basin, a build-up in the concentration of silicon, iron, and magnesium took place. Such concentrations could have occurred through two processes similar to those called upon by Deffeyes and his co-workers (1965) to explain the high magnesium:calcium ratios in the present-day barred basins on Bonaire, Netherlands Antilles. In the first case, removal of calcium cations (through limestone precipitation) could have increased the relative proportions of magnesium and iron. Secondly, a repetitive process of evaporation in the barred basin, followed by re-filling with fluvial and oceanic waters (the latter by higher-than-normal tides, by flow through restricted passes, or by seepage), could have resulted in a brine with high concentrations of iron and magnesium. At intermediate concentrations, a gravity-induced fluxing of unconsolidated $CaCO_3$ sediments by the brine resulted in the production of iron-rich carbonates. At higher concentrations, direct precipitation of iron-rich minerals and chert occurred, these presently being seen as the hematite, magnetite, grunerite, and chert rocks comprising the iron formation.

The ultimate source of iron in the waters of the Transvaal Basin is, as in the case of other basins of iron formation deposition, an enigma. The only obvious signs of volcanicity in the mixed zone are beds of a tuffaceous shale found from about 20 to 70 metres up from the base of the unit. There is no direct evidence linking volcanicity with iron-enrichment in basinal waters. In places, the opposite can be inferred, tuffaceous beds being encountered in iron-poor primary limestone. In pre-mixed zone cycles, iron-rich carbonate phases are similarly devoid of obvious signs of volcanism.

In the area studied, there is much to commend the deep-weathering and leaching hypothesis for the derivation of iron and silica, especially considering the setting of the iron formation. It is found lying on up to 1 700 metres of carbonate sediment in which terriginous sedimentation is totally subordinate. The carbonates were deposited in an epeiric sea which was partly surrounded by an extremely stable and low-lying hinterland. This landmass, incapable of shedding significant volumes of clastic sediment by mechanical erosion, would have been an ideal site for deep chemical weathering. The association of iron-rich carbonates and iron formation with the small thicknesses of terriginous sediments developed in the Olifants River Group is surely significant, and could be explained in the following way. The flat, low-lying, and deeply weathered hinterland to the Transvaal epeiric sea could have contained large volumes of groundwater. A tectonic pulse which caused uplift of the basin-margin could have flushed out the groundwater, causing it to flow basinwards, carrying with it a suspended clay fraction and dissolved or colloidally-dispersed iron and silica. On entering the saline waters of the barred basin, the clay fraction probably flocculated. Evaporation of the waters in the basin would have produced the concentrated iron and silica brines, which ferruginised earlier-formed limestones and, at peak concentrations, precipitated iron formation.

DISCUSSION

It is a remarkable feature of the well-preserved Malmani Dolomite that signs of evaporitic beds are rare, if present at all. Using present-day models, the probability of a carbonate pile up to 1 700 metres thick preserving some significant evaporite beds would be high. It would be logical to suppose that extensive beds of evaporite, now represented by layers of carbonate-fragment breccia (Stanton, 1966), should be present. A systematic study of a large sector of the Transvaal Basin has indicated no such beds. The only regionally-persistent breccias present can be related to disconformities, and were formed by residual concentration of chert on a surface of erosion.

To argue that no evaporite environment existed over the long time span and large area represented by the dolomite of the Transvaal Supergroup is to stretch credibility. Evaporites could be expected to have formed either in a supratidal (dispersed nodules) or in a barred environment (bedded evaporites), where evaporitive removal of water could have caused high concentrations of dissolved cations. In the latter case, it would be necessary to envisage a barrier which separated the evaporite basin from the open ocean. No coral barriers could have existed as far back in time as the Transvaal depositional era, so that any barriers which existed must have been either stromatolitic or clastic. That clastic carbonate fragments can produce barriers in the modern setting has been demonstrated in the Persian Gulf, where Wood and Wolfe (1969) documented a shoal of algal grainstone boundstone separating the open ocean from a lagoon. They described a number of cycles produced by transgressive and regressive migrations of the shoal, lagoon, and supratidal depositional belts.

Using this Persian Gulf model, it can be supposed that the clastic carbonates in the Transvaal dolomites might have formed shoals or banks which separated a marginal facies of the basin from the open ocean. The examination of transgressive and regressive relationships in the Malmani Dolomite suggests that, in many cases, iron-rich carbonates, with chert and iron formation precursor, are found landward of shoal- or bank-forming rocks. The barriers inferred to have been present could have been responsible for the concentration and, ultimately, the deposition of iron-rich minerals.

The postulate of a restrictive barrier for iron formation deposition is not an innovation, having been suggested by Woolnough (1941), James (1954), and Goodwin (1956). One of the strongest areas of agreement in iron formation literature is that deposition of this sediment-type did not take place in an open ocean. The nature of the restrictive process must have varied from one deposit to the next. Some researchers have proposed inland lacustrine conditions (Govett, 1966; Eugster, 1969), others have envisaged basins barred from the ocean by anticlinal warps in the basin-floor (James, 1954). The arguments developed here are thus in agreement with those propounded by previous workers, who have approached the problem from all points of the geological compass. Such fundamental agreement should, at least, remove this aspect of iron formation deposition from the forum of geological debate. One can accept that, whatever the ultimate source of iron, deposition of iron formation occurred in a basin which was not in free communication with the open ocean.

One must view the carbonate-cyclical model of iron formation deposition as one model, directly applicable to the Transvaal Dolomite and the associated iron formation. The idea of restricted environment and the role of evaporation are not new. What the model does contribute is an explanation for the cyclical development of iron-rich sediments in a carbonate pile and the common association of iron formation with carbonate sedimentation. The argument suggests that signs of evaporite deposition are absent or rare because evaporite formation, as typified by the Phanerozoic, was not a process of any consequence prior to 2 000 m.y. ago. It is thought that the Proterozoic equivalents of this class of sediments are represented by iron formation. Strong field evidence exists for the presence of iron-rich brines at the site of deposition. It is put forward that the brines were formed by evaporitive concentration in a barred basin, which was separated from the open epeiric sea by banks or shoals of carbonate detritus.

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