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Facies Models 5. Models of Physical Sedimentation of Iron Formations

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Facies Models 5. Models of Physical Sedimentation of Iron Formations

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Introduction

A single model of physical sedimentation is sufficient to describe terrigenous sedimentary rocks. Iron formations, on the other hand, must be viewed in terms of two models: 1) A model of physical sedimentation relating the sedimentary structures and textures of the rock to hydrodynamic processes during deposition, and 2) a chemical model that relates the present mineralogy of the rock to diagenetic (and metamorphic) processes. This paper will discuss models of the physical sedimentation of iron formations. Chemical models of the diagenesis of these rocks will be introduced in a later paper.

Any sedimentary rock containing more than 15 per cent iron, except heavy mineral placers, may be called an iron formation. No single facies model can apply to so heterogeneous a rock class. However, we can work with two groups of facies models, since there are two main groups of iron formation (Table I): With few exceptions iron-rich sediments are: 1) detrital chemical sediments analogous to limestones, or 2) iron-rich shales. Models of the physical facies of the first group of iron-rich sediments are based on a comparison of the sedimentary textures and structures of iron formation and limestone. I will discuss here a facies model of shelf iron formations based on the analogy with limestones and will briefly sketch the rudiments of a facies model of pelagic iron formation. Facies models of terrigenous sediments, some of which have been discussed in No. 2 to 4 of this series of articles, can be applied to the iron-rich shales.

Relatively little is known on the sedimentology of iron formations and most previous work (e.g., Borchert, 1952; James, 1954, 1966) has been centered on the mineralogy of these rocks. Therefore, I will have to distill a local model of shelf iron formations from the example of the Sokoman Formation in the Labrador trough, Quebec, Canada. This model has been obtained mainly by lateral correlation of stratigraphic sections by observation of lateral facies changes, and by comparison with analogous recent limestones of the Persian Gulf. I will use some Archean iron formations of the Superior Province to sketch the rudiments of a model of pelagic iron formations. Much work will have to be done to develop and generalize both models.

Table I
Classes of iron-rich sedimentary rocks

I. Detrital chemical sediments

1. *Cherty iron formation* ("Banded iron formation").
Textures: analogous to limestone textures.
Composition: iron rich chert containing hematite, magnetite, siderite, ankerite, or (predominantly alumina-poor) silicates as predominating iron minerals. Relatively poor in Al and P.

2. *Aluminous iron formations* ("Minette-type ironstone").
Textures: analogous to limestone textures.
Composition: Aluminous iron silicates (chamosite, chlorite, stilpnomelane), iron oxides and carbonates. Relatively rich in Al and P.

II. Iron-rich shales

3. *Pyritic shales* ("Sulfide iron formation"). Bituminous shales containing nodules or laminae of pyrite. Grade into massive pyrite bodies by coalescence of pyrite laminae and nodules.

4. *Siderite-rich shales* ("Clay ironstones, Coal ironstones"). Bituminous shales with siderite concretions. Grade into massive siderite bodies by coalescence of concretions.

III. Other types

5. *Iron-rich laterites.*

6. *Bog iron ores.*

7. *Manganese nodules* and oceanic iron crusts.

8. *Iron-rich muds* precipitated from hydrothermal brines, Lahn-Dill type iron oxide ores and stratiform, volcanogenic sulfide deposits.

9. *Placers* of magnetite, hematite or ilmenite sand.

Sedimentary facies are influenced, to a degree, by biologic processes. Therefore ecologic factors have to be considered, where facies models are based on a comparison of ancient sedimentary rocks and recent sediments. Consideration of ecologic factors is particularly important if Precambrian rocks are compared with recent sediments, as I will do in this paper.

Precambrian sedimentary rocks differ from recent sediments mainly in the following properties: 1) Skeletal and framework building organisms did not exist in the Precambrian. Therefore, Precambrian sedimentary rocks lack skeletal debris, reefs built by framework-builders, and reef debris. 2) Sea-grasses and higher algae that form baffles at the bottom of the present shallow sea did not exist in the Precambrian. However, blue-green algae appparently fulfilled that function in the Precambrian at least to a degree. Blue-green algae also acted as sediment-binders and constructed reef-like mounds. 3) Mud-eating and bottom stirring organisms did not exist in the Precambrian. Therefore delicate laminations are preserved in Precambrian sedimentary rocks whereas they commonly have been destroyed in the Phanerozoic. Thin beds of differently textured sediment survived in the Precambrian, whereas they commonly have been mixed in the Phanerozoic.

Comparison of Iron-Formation and Limestone

The siliceous and aluminous iron formations, variously termed cherty (or banded) iron formations and Minette-type ironstones, are detrital chemical sediments. They have sedimentary textures and structures analogous to the sedimentary textures and structures of limestones. Analogues to all textural types of detrital limestones exist: micritic, intraclastic, peloidal and pelleted, oolitic, pisolitic, stromatolitic. These varieties are briefly described in Table II, but Figures 2 to 8 and 9 to 14 might be better documentation than any description in words would be.

Table II
Textural types of cherty iron formation

1. *Micrite-type:* Deposited as a mud whose particles are too fine grained to survive diagenesis. Only lamination and stratification are visible as depositional structures. Small-scale cross-beds here and there prove deposition as particulate, non-cohesive matter. (Fig. 8, 9).

2. *Pelleted:* Fine pellet textures of silt or very fine sand size. (Fig. 2).

3. *Intraclastic:* Containing gravel-size fragments (intraclasts) whose internal textures prove derivation from pene-contemporaneous sediment. Fragments may be embedded in a micrite-type matrix (Fig. 5, 10), or may be bound by a cement that has been introduced during diagenesis.

4. *Peloidal:* Contains sand-size fragments (peloids) without internal textures. Peloids may be embedded in a micrite-type matrix or may be bound by a clear chert cement that has been introduced during diagenesis (Fig. 3).

5. *Oolitic:* Contains concentrically laminated ooids, either set in a micrite-type matrix or, more commonly bound by a clear chert cement introduced during diagenesis (Fig. 10).

6. *Pisolitic:* Contains pisolites (concentrically laminated bodies similar to but larger than ooids, probably produced by blue-green algae) set either in a micrite-type matrix or cemented by clear chert. (Fig. 6).

7. *Stromatolitic:* Wavy, columnar or digitating stromatolites. (Fig. 7).

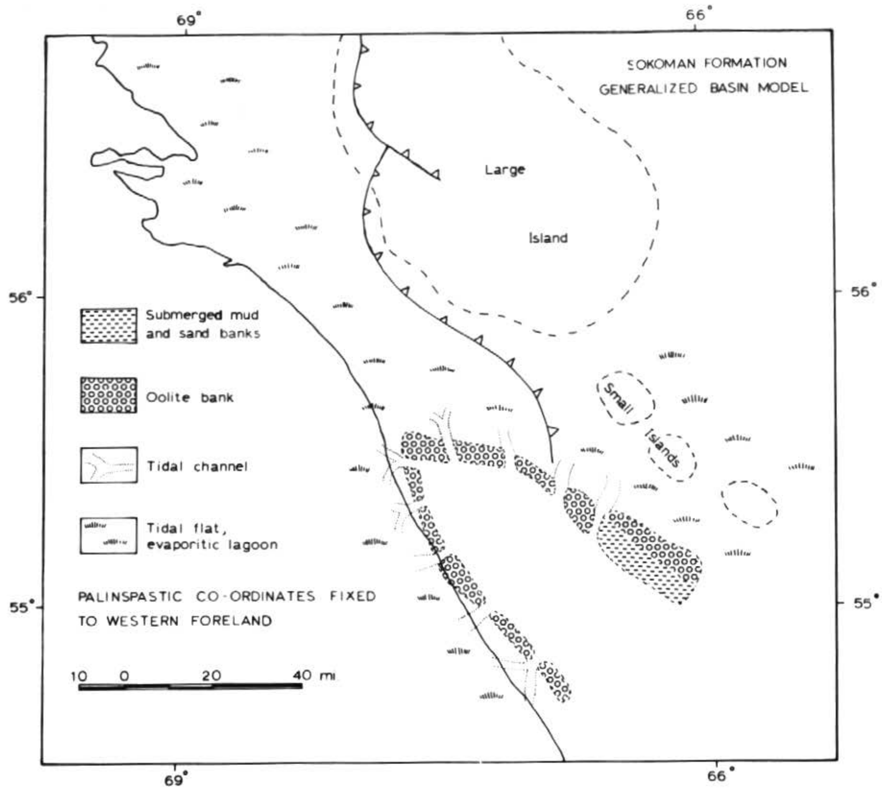

Figure 1
Paleogeography of the Labrador trough during deposition of the Sokoman Formation.

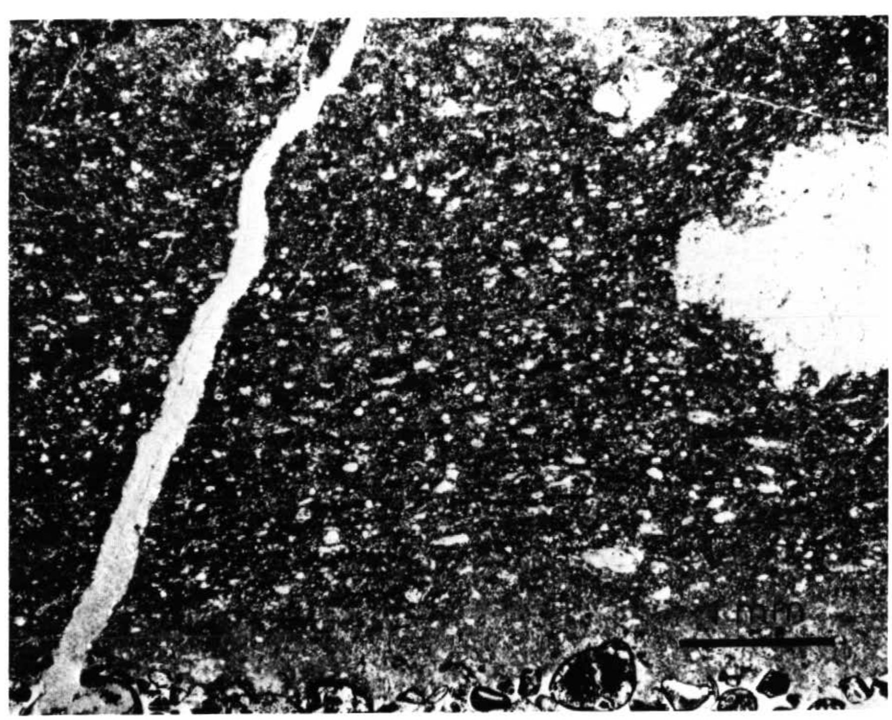

Figure 2
Pelleted iron formation. Relicts of silt-size particles (pellets) are visible. Hematite is the iron mineral.

These sedimentary textures and structures permit a number of important conclusions: 1) iron formations, like most limestones, are detrital sediments; they have been mechanically deposited as particulate matter. 2) particles came in various sizes: clay-size (Fig. 8), silt-size (Fig. 2), sand-size (Fig. 3), gravel-size (Fig. 5). 3) particles were relatively hard, and were non-cohesive. The fresh sediment had mechanical properties rather similar to those of lime muds, sands and gravels.

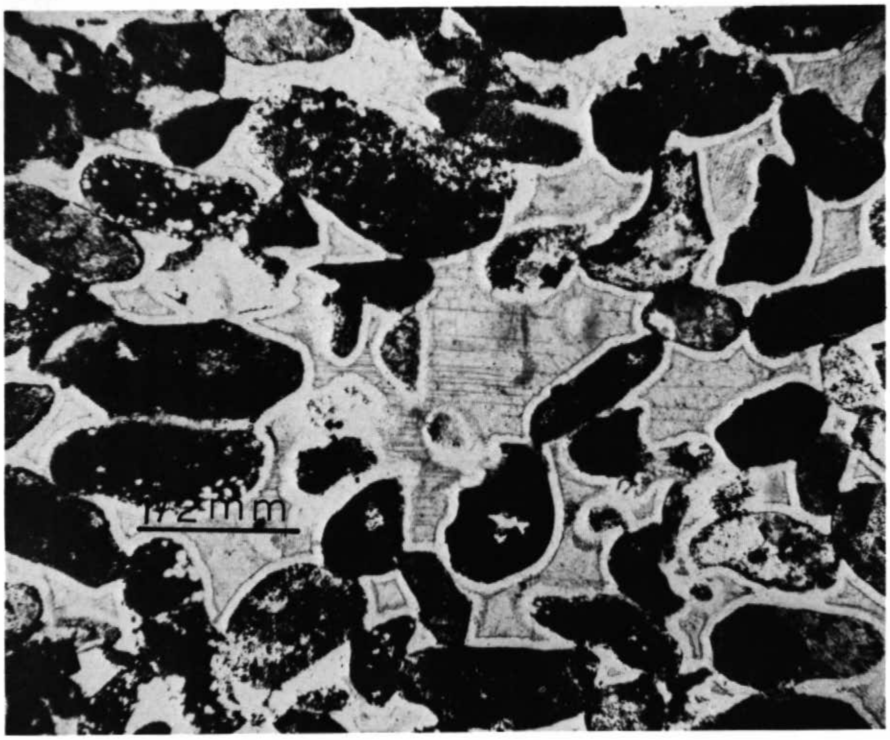

Figure 3
Peloidal iron formation. Structureless particles of sandsize (peloids) are well preserved. Hematite as iron mineral.

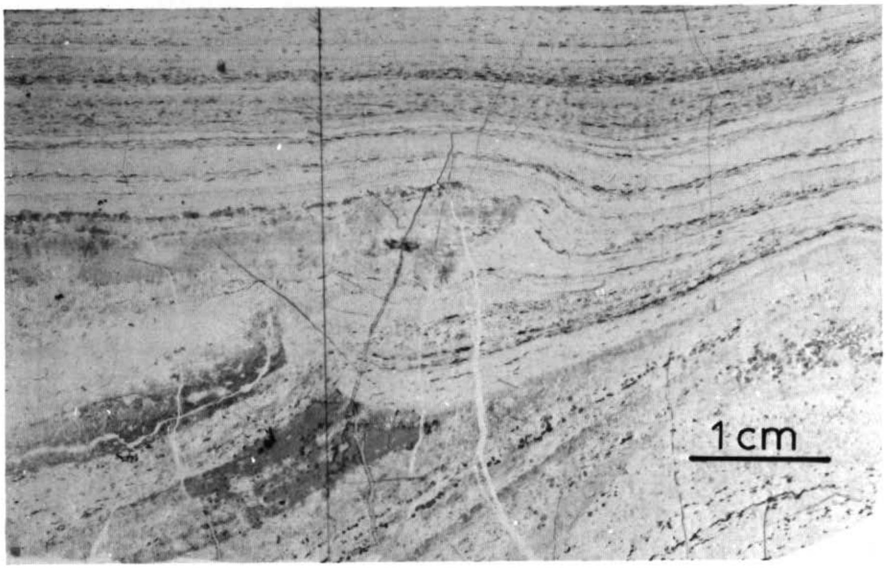

Figure 4
Peloidal iron formation, strongly recrystallized, with crossbedding.

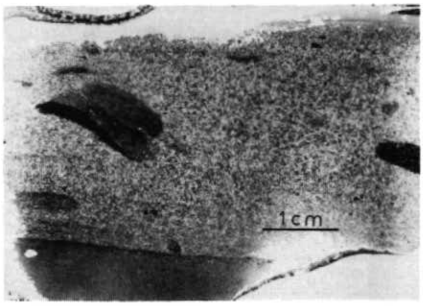

Figure 5
A micrite-type iron formation is overlain by an intraclastic bed. Coarse fragments show internal lamination, thus, are derived from penecontemporaneous sediment. The intraclasts are loosely set in a micrite-type matrix. Minnesotaite is the iron mineral.

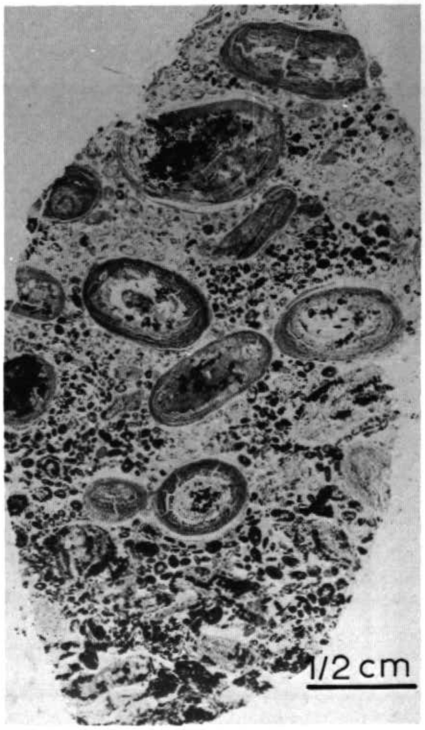

Figure 6
Pisolitic iron formation. Pisolites (P) and ooids (O) cemented by clear chert. Hematite and magnetite predominate.

The Sokoman Formation: Local Model of Shelf Iron Formations

Flat, prograding coastlines away from estuaries, like the south shore of the Persian Gulf (Purser, 1973), generally show a subdivision in four oceanographic domains: 1) a shallow marine shelf sea, 2) a narrow strip of sand bars, tidal deltas, reefs, or island, 3) a lagoonal platform, 4) a tidal flat. The Sokoman iron formation of the Central Labrador trough is very similar to the limestones that are presently being deposited at the south shore of the

Figure 7
Stromatolitic iron formation. Hematite predominates.

Persian Gulf, and a similar oceanographic subdivision can be recognized (Fig. 1). Each oceanographic domain has its own, characteristic facies. The following facies model is based mainly on a comparison of the Labrador trough iron formation with limestones in the Persian Gulf area, but takes in account certain features of limestones of the Bahama platform and the Florida coast.

Shallow marine shelf: Most parts of the shallow marine shelf, below storm wave base, are covered with micritic muds. Micrite-type shelf iron formation (Fig. 9) of the Labrador trough show delicate parallel laminae; such lamination is absent from shelf muds of the Persian Gulf due to the action of sediment-eating organisms.

Where marine currents are strong, very fine sand ($2\text{-}3\phi = 0.1 - 0.2$ mm) has been winnowed from coarse foreshore sediment, and has been carried far on to the shelf. Such shelf sands in the Persian Gulf are poorly sorted, due to admixture of terrigenous mud and of shells of foraminifera. In the Labrador trough, shelf sand has not been diluted by other components and is well sorted. Parallel stratification is characteristic.

Foreshore zone. Beds of coarse-grained oolite or peloid sand are intercalated between the shelf muds or sands in the forshore zone. Storm layers formed at the wind-exposed margin of mud shoals that reached above storm wave base. Such storm layers (Fig. 10) generally are composed of very coarse-grained muddy sand or of muddy gravel. They are poorly sorted and, not uncommonly, show graded bedding. A mud-matrix is absent where the area eroded by the storm lacked muddy sediment.

Figure 10
Iron formation deposited as muddy gravel. Intraclasts embedded in a micrite-type matrix. This is a storm layer deposited on top of a mud bank in the foreshore of the shelf. Hematite and magnetite predominate.

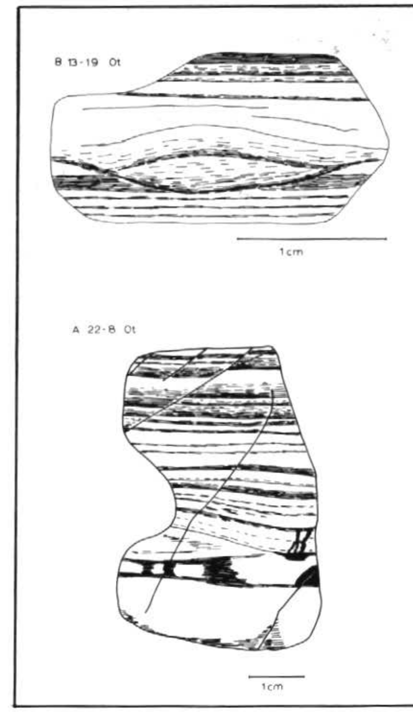

Figure 8
Micrite-type iron formation. Cross-lamination proves that the rock has been deposited as non-cohesive particulate matter, but no remnants of the particles survived diagenesis. Iron minerals are hematite and minnesotaite.

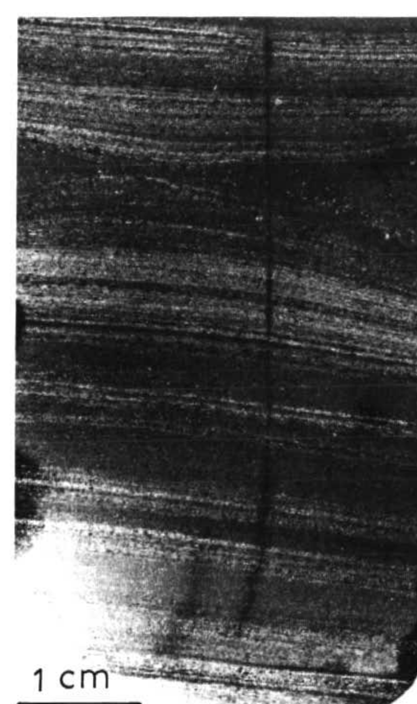

Figure 9
Micrite-type iron formation deposited on the shelf. Minnesotaite predominates.

Shelf-marginal bank: Thick-bedded or massive oolite or peloid sand builds the sand banks, tidal deltas and islands at the shelf margin. The sand (Fig. 11) is coarse grained and very well sorted. Large-scale cross-bedding is common. Oolites formed where sand has been agitated more or less constantly at the sediment surface by alternating tidal currents. Peloids predominate where the sediment was rarely agitated; at present, this is the case where sediment has been stabilized by micro-organic slimes, blue-green algae, algae, or sea-grasses, and where tidal currents are weak.

Lagoonal platform. Sediments of the lagoonal platform are the most variable; they also show relatively great differences between otherwise analogous Precambrian and recent facies. At present, mud, muddy sand, or well sorted sand are deposited on lagoonal platforms, depending on local agitation by tidal currents and waves. Grain size of the sediment increases with increasing intensity of tidal action or wave agitation. Mud deposition is by no means restricted to lagoonal basins; on the contrary, it commonly occurs in water shallower than one metre, on mud banks and tidal flats. Strong storms may carry sand far into areas of mud-deposition, and may produce intraclast gravels. Superficial oolites are common and form where peloids are suspended for brief periods during storms. In the present, mud-eating and bottom-stirring organisms are extremely active and would destroy any small-scale bedding that existed.

In the Labrador trough, sediments deposited on the lagoonal platform are characterized by thin-bedded alternation of micrite-type sediment with oolite or peloid sand (Fig. 12). Surficial ooids are more common than fully developed ooids except in proximity to the basin-marginal oolite banks. Storm layers (Fig. 13) of sandy or muddy gravel are common. Upward-coarsening cycles, 5 to 15 metres thick, have been observed, and probably formed by the gradual infilling of lagoonal basins. Cross-bedding, erosion channels, scours, and flaser bedding are common.

Applications of the Model

Now that we have a local model of shelf iron formations we may ask how this model performs: 1) as a framework and guide to future observations, 2) as a norm with which other shelf iron formations may be compared, 3) as a basis for hydrodynamic interpretation, and 4) as a predictor in new geological situations.

(1) *The model as framework and guide to future observations.* The classification of textures (Table II) provides a simple framework for the description of iron formations. Furthermore, all other textures and structures common in limestones are to be expected in iron formations, for example, fenestral textures and solution porosity (Dimroth and Kimberley, 1976, Fig. 14). They have to be recorded carefully as basis for further interpretation.

(2) *The model as norm.* Facies of the Brockman Formation (Hamersley Range, Western Australia) and of the Penge and Kuruman Formation (Transvaal basin, South Africa)

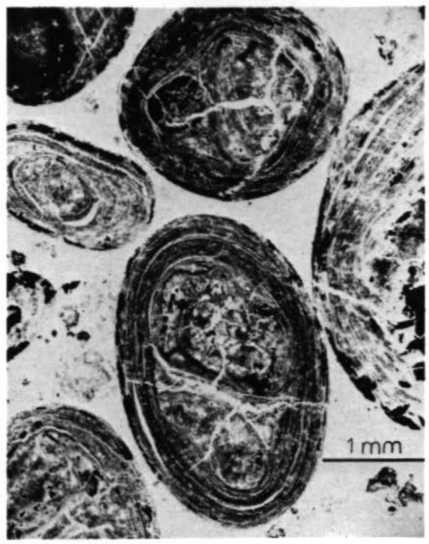

Figure 11
Iron formation deposited as well sorted, coarse-grained ooid sand, of the sand bank at the shelf margin. Hematite and magnetite predominate.

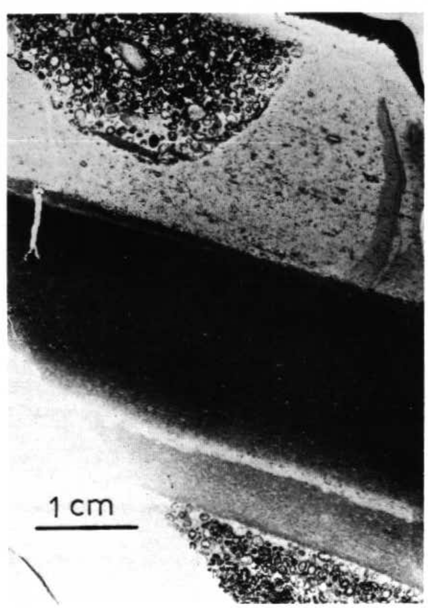

Figure 12
Alternating beds of oolitic and micrite-type iron formations, characteristic of the lagoonal platform. Hematite and magnetite predominate. Note the erosion channel.

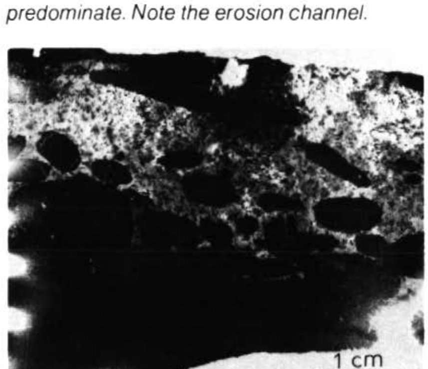

Figure 14
Laminated Algoma-type iron formation.

Figure 13
Intramicrite-type iron formation overlying laminated micrite. Note the graded bedding in the intramicrite, the erosional base of the bed, and loading of the underlying micrite by intraclasts. This is a storm layer deposited on a shallow mud-bank in the lagoonal domain. The iron mineral is siderite.

manifestly differ from facies of the Sokoman Formation (see Trendall, 1968). The two former units are exclusively or predominantly composed of laminated micrite-type iron formations, whereas micrite-type lithologies do not make up more than about 30 per cent of the Sokoman Formation. Based on the model we now ask why this is so.

(3) *The model as basis for hydrodynamic interpretations.* As noted above, the Kuruman formation is mainly composed of laminated, micrite-type iron formations (Beukes, 1973; Button, 1976). However, the formation does contain thin intercalated beds of oolite sand, of cross-bedded peloid sand, of intraclast conglomerate and of edgewise conglomerate. Some of the laminated rocks documented by Beukes (1973, Fig. 20B, C, 22C) more closely resemble peritidal lime muds than deep basinal lime muds.

These features suggest that facies differences beween the Sokoman and Kuruman Formations might be due to differences in basin hydrodynamics rather than to differences in basin depth: The Sokoman formation has been deposited in a basin exposed to fairly vigorous tidal action, as the present Persian Gulf; the Kuruman Formation possibly has been deposited in a similar basin from which, however, tidal currents were more or less excluded. Of course, this hypothesis has to be verified by much further work. Button (1976) has proposed a basin model for the Penge and Kuruman Formations that very closely resembles the basin model proposed for the Sokoman Formation.

The case of the Brockman Formation is quite different: Micrite-type rocks of that unit show *only* parallel lamination, quite similar to the varve-like laminations of evaporitic limestones (Trendall and Blockley, 1976). It appears that the laminated facies grades into peloidal rocks toward the extreme margin of the presently preserved part of the basin (A. D. T. Goode, pers. commun., 1976). These features suggest that the Brockman Formation has been deposited in a fairly deep basin, permanently below storm wave base. The Permian Marathon basin of the US may be a suitable Phanerozoic analogue; I know of no present-day equivalent to such a basin.

(4) *The model as predictor.* The tentative interpretations outlined above permit us to make some predictions: if our interpretations are correct, it is to be expected that micrite-type iron formations with desiccation cracks and with fenestral textures, associated with flat-pebble conglomerates should be found in the Penge and Kuruman Formations; such structures are not to be expected, and have not been found, in the Brockman Formation.

Pelagic Iron Formations

Our knowledge of Archean pelagic iron formations is far too limited to permit the distillation of a facies model that would fulfill the four functions required by Walker (1976). Yet, at least a sketch of a facies model may be presented. Algoma-type iron formations generally are thin lenses, rarely more than one or a few metres thick and rarely continous for more than a few kilometres. They have micrite type textures. Parallel lamination and, rarely, erosion surfaces (Gross, 1972; Schegelski, 1975) are the only primary sedimentary structures. Algoma-type iron formations do not grade into iron formation with different sedimentary textures and structures either laterally, or in vertical section.

Algoma-type iron formations here and there are intercalated between volcanic flows. Much more common is intercalation between volcaniclastic (epiclastic and autoclastic) sediments. The iron formations are closely associated with grey or black (graphitic) cherts, with black shales, and with argillites all of which show parallel lamination as their only sedimentary structure, and with greywakes. Associated greywackes may be very fine grained and thin bedded (5 cm or less), or medium to coarse-grained and thick-bedded (30cm and more). Both are turbidites; the thin bedded greywackes probably are distal turbidites of Walker's (1967) type a→e, whereas the thick-bedded greywackes are proximal turbidites.

In certain cases, a characteristic vertical facies change has been observed: Proximal or distal turbidites are followed upwards by turbidites that have thin (one to several mm) laminae of magnetite-bearing chert at the top of the Bouma cycle (described in Walker, 1976 b). The proportion of chert gradually increases upward toward the bed, one to several metres thick, of magnetite chert or jasper. Upward, the proportion of iron formation decreases again. Such a sequence has been observed in Canada and in South Africa (compare Beukes, 1972, Fig. 7), but is neither universal nor is it the only sequence described.

Clearly, these iron formations are pelagic sediments since they occupy the uppermost interval of the Bouma cycle that is normally occupied by pelagic shale. They are intercalated indiscriminately between proximal or distal turbidites. Thus, apparently their deposition may have taken place on any part of the turbidite fan. It appears, therefore, that they were deposited in any environment from which volcaniclastic detritus was temporarily excluded. It is quite unknown whether the non-deposition of contemporaneous volcaniclastic material is due to tectonic causes (subsidence of the source area) or simply reflects a change in the pattern of the distributary channels.

This sketch does not fulfill the four functions of a facies model. It describes a common case of Algoma-type iron formation but is not necessarily a norm. It may serve as guide to future observations mainly in so far as it is important to search for, and to describe, cases where Algoma-type iron formations occur in a different context. The model has no predictive value, and conclusions on hydrodynamic conditions of deposition are trivial: obviously these iron formations settled from very dilute suspensions, in tranquil water, at a depth that was permanently below storm wave base.

Origin of Iron Formation

The facies model outlined above is largely independant of speculations on the origin of iron formation. Basically, there are two genetic hypothese: Cayeux (1922, 1911) suggested that the cherty iron formations formed by silicification of aluminous iron formations and the latter by penecontemporaneous remplacement of limestone. On the other hand, Cloud (1973), Eugster and Chou (1973) and Lepp and Goldich (1964) believed that iron formations were precipitated from a hypothetical "primordial" ocean; that primordial ocean is thought to have been devoid of oxygen and saturated with dissolved iron carbonate. Dimroth and Kimberley (1976) pointed out that there is strong evidence against the existence of such a "primordial" ocean even in the Archean.

Iron formations were deposited in the form of non-cohesive particulate matter; this may be interpreted as evidence against direct precipitation of silica and iron (in forms other than iron carbonate) since silica and iron oxide and silicate generally precipitate as gels. Of course, precipitation of iron carbonate is excluded, because of the evidence for deposition of iron formation from an oxygenic environment (Dimroth and Chauvel, 1973; Dimroth and Kimberley, 1976). The limestone-type textures of iron formations point to an origin of iron formation by limestone replacement. Thus, the author believes that iron formations most likely were precipitated as $CaCO_3$, probably as aragonite, organically or inorganically. The original aragonite sediment has been replaced by silica and iron compounds during early diagensis, but the details of the process, and the source of iron and silica (terrigenous clay? tuff?) are still unknown.

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This list is very brief on purpose. It is intended as a basic reader for those who wish to become familiar with the models outlined here and with other ideas and concepts on the origin of iron formations.

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Most hypotheses on the origin of Precambrian iron formations suffer from the defect that they attempt to link deposition of cherty iron formations to a hypothetical, oxygen-free ocean. This is in conflict not only with evidence for the presence of free oxygen in atmosphere and ocean since early Precambrian time, but also with the existence of very large cherty iron formations of late Precambrian age (compare Dimroth and Kimberley, 1976).

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