

Facies Models 6. Diagenetic Facies of Iron Formation

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Facies Models 6. Diagenetic Facies of Iron Formation

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Introduction

Facies is the term used to denote the aspect of part of a sedimentary rock body, as contrasted with other parts. Primary facies are defined by features present at the time of deposition: primary lithology, sedimentary texture and sedimentary structures. Diagenetic facies are defined by diagenetic features. In iron formations, diagenetic facies are primarily defined by the nature of the iron-bearing minerals present.

Information on the distribution of iron minerals in iron formations can be distilled into a facies model in the same way as a model of sedimentary facies is distilled from information on primary lithology, sedimentary texture and structure. In fact, a model of diagenetic facies of iron formation is the oldest facies model available, the models of Borchert (1952) and James (1959). Of course, at that time, the mineralogy of iron formations was thought to be primary, in accordance with the general underestimation of diagenesis in the 1950s, and iron mineral facies of iron formations were therefore thought to be sedimentary facies. This fact cannot detract from the value of the James-Borchert model which I will have to paraphrase.

Just as a model for primary facies, the model of diagenetic facies must fulfill four purposes (Walker 1976).

- 1) It must act as a *norm*, for purposes of comparison;
- 2) It must act as a *framework* and *guide* for future observations;

3) It must act as a *predictor* in new geological situations;

4) It must act as a basis for physico-chemical interpretation of the diagenetic environment or system that it represents.

These four purposes are literally taken from Walker's (1976) introduction to the concept of facies models, except for two words. A model of *diagenetic* facies cannot serve as basis for *hydrodynamic* interpretation of the environment, and because *diagenetic* processes are determined by the *physico-chemistry* of the sedimentary system *after* deposition, a model of *diagenetic* facies must permit interpretation of these physico-chemical processes.

Many stages of diagenesis. Iron formations like other chemical sediments, underwent many stages of diagenesis. I define here as *early diagenetic* those processes that took place during the lithification and which are contemporaneous with the infilling of primary porosity, with compaction, and with the consolidation of the sediment. *Late diagenetic* processes took place after lithification. Of course, there is no clearcut limit between late diagenesis (taking place at low temperatures), load metamorphism (taking place at somewhat higher temperatures) and very low grade regional metamorphism (taking place during an orogeny, but not necessarily at higher temperatures than load metamorphism).

Figure 1 diagrammatically shows the most important mineral reactions that have been deduced from petrographic observation of iron formation of the Labrador trough. The transitions from limestone into iron formation have been very little studied. Such transitions exist in the Transvaal Supergroup (Button, 1976), in the Hamersley Group (Davy, 1974) and in the Archean of the Slave Province (Kimberley, oral commun., 1976). Therefore, nothing is known about the first step of diagenesis namely the conversion of calcareous sediment to ferriferous chert. However, oolites of iron formations more closely resemble aragonite oolites than calcite oolites; this fact suggests that conversion of calcareous sediment to iron formation occurred during very early diagenesis, and preceded the inversion of aragonite to calcite (Kimberley, 1974).

On the other hand, transformations of iron minerals shown in Figure 1 have

been deduced from petrographic observations like those illustrated by Figures 2 to 5. In Figure 2, concentric texture of oolites is beautifully preserved by sub-microscopic hematite crystallites. Figure 3 shows partial destruction of the concentric texture by coarse-grained magnetite. In Figure 4, nearly all of the concentric texture has disappeared save for a few relicts here and there. Clearly, magnetite formed here at some later stage than hematite. A similar observation is illustrated in Figure 5, where magnetite not only is younger than hematite dust, but where the growth of magnetite is accompanied by differential compaction; consequently, magnetite formed here during a comparatively early stage of diagenesis, before the complete lithification of the sediment. This simple example illustrates the observations that have been used to construct the scheme shown in Figure 1.

The Facies Model

Diagenetic facies of iron formation are basically the mineral facies introduced by James (1954). James defined facies by the *predominating* iron mineral and distinguished oxide (hematite, magnetite), silicate, carbonate, and sulphide facies. From this starting point, the mineral facies concept has evolved in two directions. First, Dimroth and Chauvel (1973) defined mineral facies by the relicts of very early diagenetic, iron minerals. These are the very fine grained, submicroscopic, hematite, and fine grained siderite and greenalite. Thus, hematite, siderite, and silicate facies are defined, to which a sulphide facies (mainly pyrite-rich shales) may be added. This scheme fails in formations that do not contain relicts of the very early diagenetic minerals. These have to be defined less precisely as "recrystallized iron formations".

Second, Zajac (1973) and Klein and Fink (1976) defined facies by the *assemblage of iron minerals* now present. This concept can be applied to any iron formation regardless of its diagenetic or metamorphic grade. Generally, it is assumed that all iron minerals present are stably co-existing; however, I will demonstrate below an example where this very likely is not the case.

Obviously, Dimroth and Chauvel's facies represent a relatively early stage of diagenesis and, therefore, show more

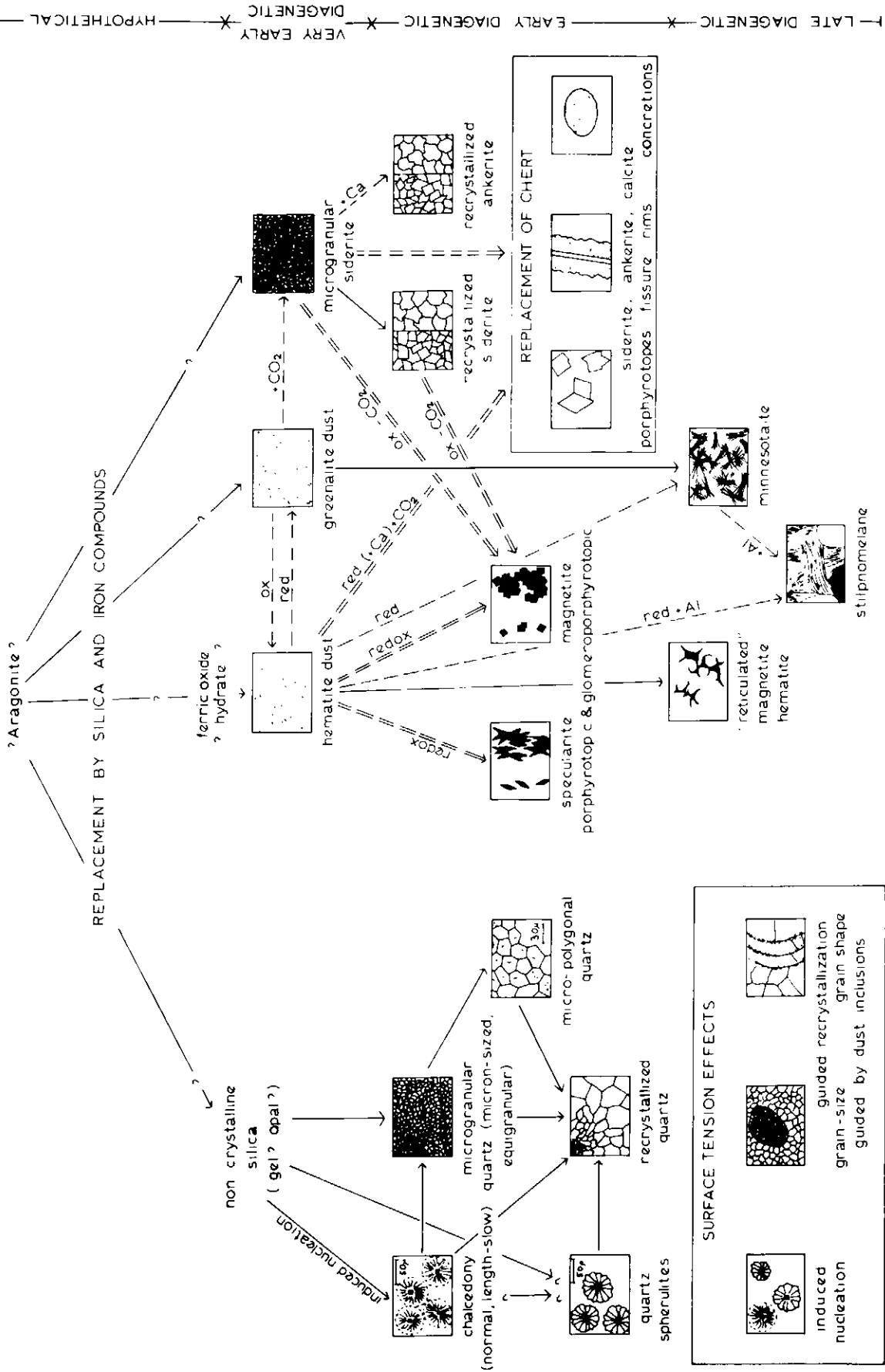


Figure 1.

Mineralogical evolution of the Sokoman Iron Formation, Labrador trough (after Dimroth and Chauvel, 1973) transformations hematite = greenalite - siderite is from the Gunflint Formation (Goodwin, 1956 and

author's observations, Single lines indicate mineral-by-mineral replacement and recrystallization, double lines indicate transformations interpreted essentially as solution-precipitation

reactions Transformations involving redox reactions shown by broken lines (ox = oxidation; red = reduction) ox = $\text{Fe}^{2+} \rightarrow \text{Fe}^{3+}$ solution and reprecipitation red = another place by oxidation

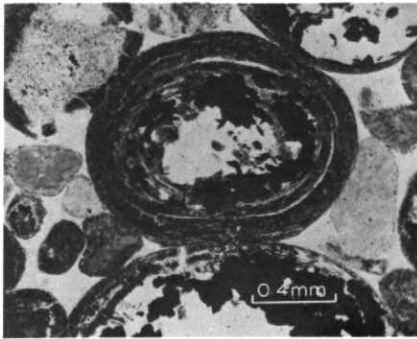


Figure 2. Hematite oolite. Concentric texture is well preserved by submicroscopic crystallites of hematite. Some recrystallizations of hematite in centre of oolite. Section B13-15, Labrador trough.

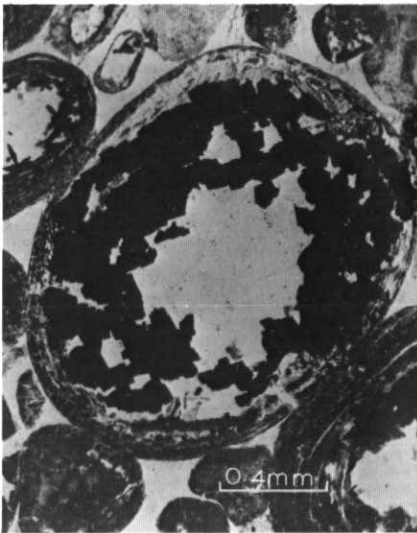


Figure 3. Same section, about one mm from Figure 2. Most of the concentric texture of the oolite has been destroyed by growth of magnetite.

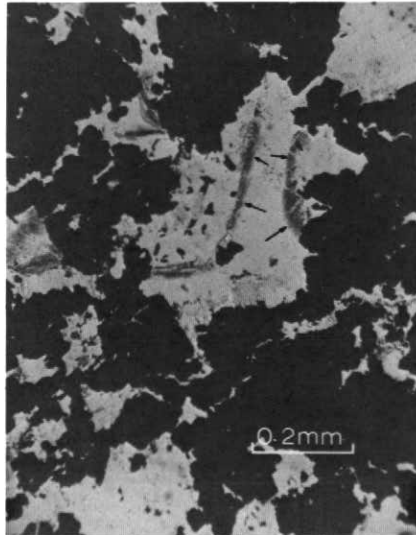


Figure 4. Same thin section, about 2 mm from Figure 3: All sedimentary textures have been destroyed by growth of magnetite porphyroblasts. A few remnants of oolite skins remain (arrows) and are preserved by submicroscopic crystallites of hematite.

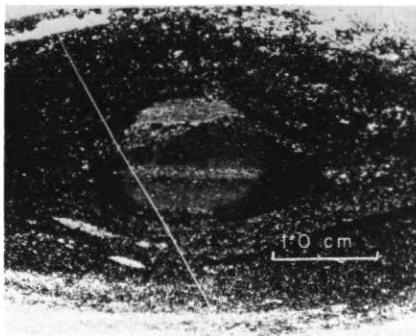


Figure 5. The spherical grey spot in the center is a nodule of red jasper, stained by submicroscopic hematite crystallites. Lamination and sedimentary textures (oids, peloids) are beautifully preserved by hematite dust. The dark material surrounding the jasper spot is composed mainly of coarse-grained magnetite and quartz. All sedimentary texture has been destroyed in this material, but some of laminae are continuous from the jasper into the surrounding magnetite-quartz fabric and indicate differential compaction around the jasper nodule. Section C2-S2, Labrador trough.

clearcut relations to sedimentary facies than the assemblage facies of Klein and Zajac which represent a relatively late stage mineralogy. On the other hand, mineral assemblage facies are very important to the economic geologist since the beneficiating properties of taconite iron ore depend largely on the late diagenetic mineral assemblage.

Overprinting mineral facies. Several redox reactions can be shown to have taken place in iron formations (Fig. 1), particularly oxidation of siderite to magnetite and/or hematite and reduction of hematite to magnetite, iron silicate or carbonate. An example of such overprinting is shown in Figures 6-8 hematite oolites and pisolites (Fig. 6), have been partly replaced by iron carbonate, probably siderite (Fig. 7, 8). It should be noted that siderite has never been observed by the writer in contact with the submicroscopic crystallites of hematite staining jasper, suggesting that both minerals do not stably co-exist. On the other hand, grain contacts between siderite and coarsely crystalline hematite are not uncommon; this association of minerals very likely is metastable, metastable preservation of specular hematite being due to its extremely low solubility in natural solutions.

Physico-chemical factors. Which iron mineral is stable during diagenesis depends mainly on Eh, pH, pCO_2 and pS'' of the pore solutions. Traditionally, the stability fields of various iron minerals are shown on Eh-pH diagrams at fixed values of pCO_2 and pS'' (Garrels and Christ, 1965; James, 1966). However, in marine sediments, pH varies little, and is largely determined by pCO_2 . Therefore, I follow Berner (1971) in representing the stability fields of the iron minerals in an Eh- pCO_2 diagram at very low pS'' , and in an Eh- pS'' diagram (Fig. 9). The stability fields of ankerite and of iron silicate are unknown. They have been stippled in Figure 9 on the basis of empirical evidence.

Eh depends on the concentration, in the pore solution, of oxygen, or of organic decay products; of course, both are incompatible. pS'' depends on the activity of sulphate reducing bacteria, which increases with the proportion of organic matter in the sediment and is limited by the availability of dissolved sulphate. pCO_2 is a measure of the CO_2 -

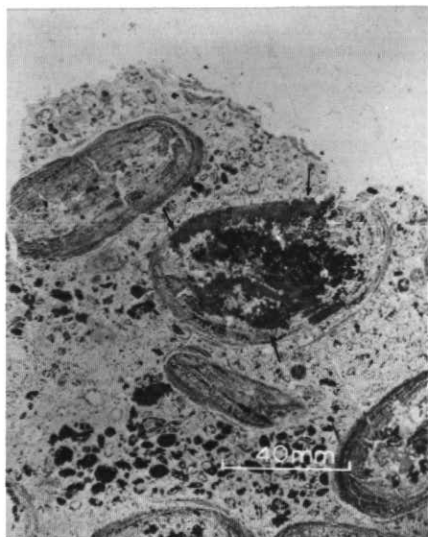


Figure 6.
Oolitic and pisolitic iron formation. Concentric texture of pisolites is beautifully preserved by submicroscopic crystallites of hematite. Carbonate porphyroblasts (probably siderite, dark grey shown by arrows) destroy the concentric texture. Thus, this is a hematite facies rock, overprinted by younger carbonate facies. Section B13-10, Labrador trough.

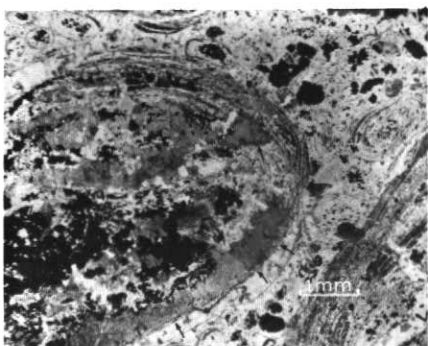


Figure 7.
Detail from Figure 6. Concentric texture of pisolite, preserved by hematite crystallites has been destroyed during subsequent growth of (dark grey) iron carbonate (arrows).

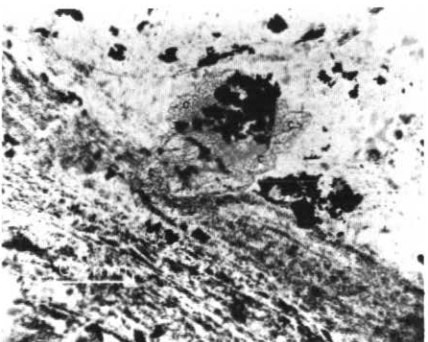


Figure 8.
Detail from Figure 6. A carbonate porphyroblast (grey, c) replaced part of a pisolite and the cementing chert. Thus carbonate facies is late diagenetic.

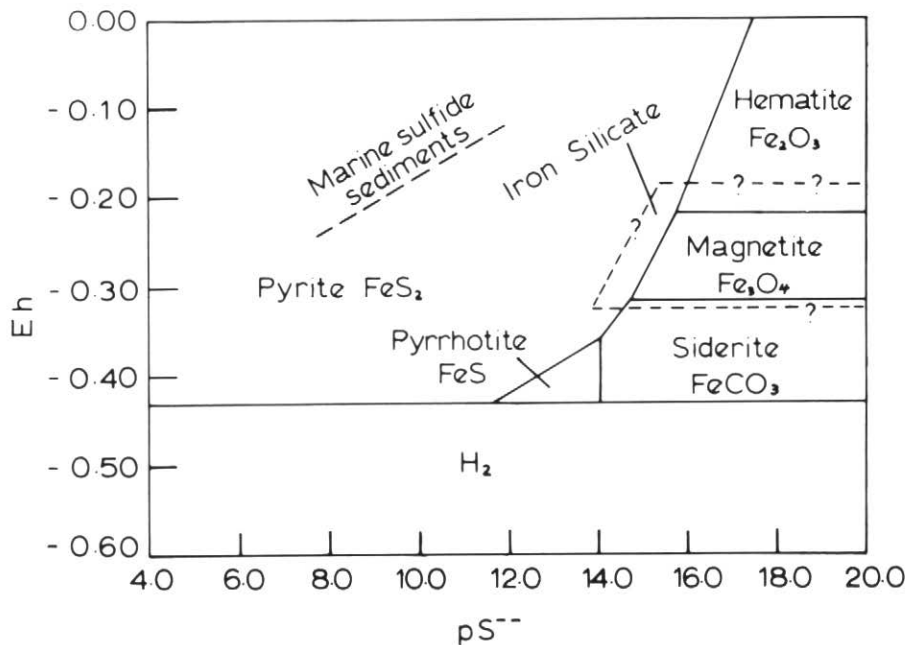
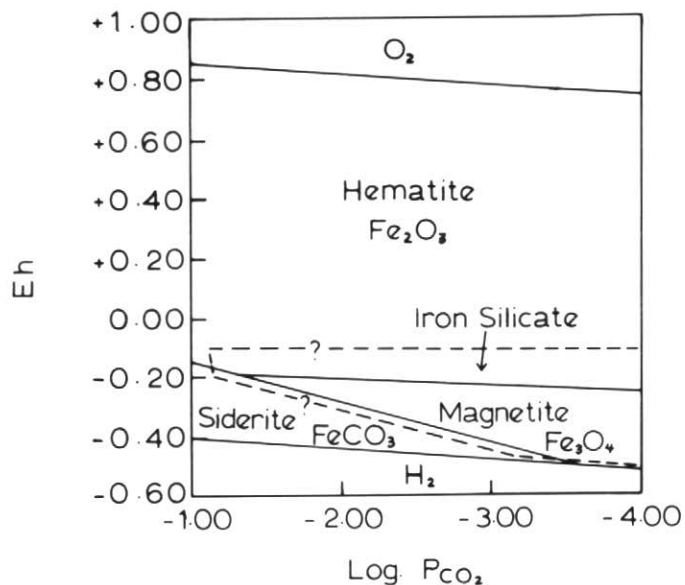


Figure 9.
Stability fields of some iron minerals in Eh-pCO₂ and Eh-pS'' diagrams (after Berner, 1971). The stability field of iron silicate is unknown.

content of the pore solutions; it depends largely on carbonate precipitation or dissolution during diagenesis and on addition of CO₂ to pore solutions due to decay of organic matter. Therefore, the organic content of the sediment and availability of oxygen during diagenesis are the main factors that determine which iron mineral will form in the sediment. Reduction of Fe³⁺ to Fe²⁺ is caused by the presence of organic matter in the sediment, oxidation of Fe²⁺ to Fe³⁺ by the presence of oxygen.

Sedimentological factors. The organic content of the sediment depends largely on the organic productivity of the environment and on the grain-size of the sediments. Because most organic detritus is plankton, organic matter is enriched in fine grained sediment. During *earliest diagenesis* oxygen diffuses into the sediment, diffusion rates increasing with increasing grain size. During later diagenesis oxygen is excluded from the sediment unless it is transported in descending solutions. On the other hand, ascending diagenetic solutions generally contain decay products of organic matter and, therefore, are reducing.

As an effect of these relations, earliest diagenesis generally takes place under oxidizing or slightly reducing conditions in well-washed sands; it takes place under strongly to moderately reducing conditions in muds, *unless these have been deposited in environments of low productivity and at low rates of sedimentation.* Conditions during later diagenesis depend largely on the direction of pore fluid exchange, pore fluids migrating mainly in well-washed sands.

Iron minerals in recent sediments.

Recent iron mineral formation apparently can be understood along the lines outlined above (Berner, 1971). Thus, ferric oxide or oxyhydrates form in the presence of free oxygen, and survive diagenesis in the absence of organic matter. Well washed sands are most likely to contain ferric oxides as are muds deposited in areas of low productivity and at low sedimentation rates (e.g., deep-sea clays). Chamosite forms in overall aerobic environments, in the presence of small amounts of organic matter (Porrenga, 1966). Siderite and sulphides are precipitated in muds relatively rich in organic matter

(Berner, 1970; Ho and Coleman, 1969); which is precipitated seems largely to depend on the availability of SO₄²⁻ in the pore solutions.

Iron minerals in ancient iron formations. James (1954, 1966) recognized a similar relationship between the texture, carbon content and mineralogy of iron formations as outlined above. In the Labrador trough, all well washed sands contain either hematite dust or iron silicate as very early diagenetic iron mineral. Most micrite- and intramicrite-type iron formations contain very early diagenetic siderite or iron silicate; however, some do contain hematite dust. As could be expected, relations between later diagenetic mineralogy and sedimentary texture are less clear-cut. A very large class of iron formation is derived from well-washed peloid or ooid sands. Traces of hematite dust prove that these were originally hematite facies rocks; however, this hematite has been replaced nearly completely by coarse grained magnetite, ankerite and siderite during later diagenesis. Thus, they are oxide facies rocks that have been reduced to siderite facies rocks later during their diagenesis.

A correlation between mineralogy and carbon content has also been recognized (James, 1966). Iron formations containing hematite or iron silicate as earliest diagenetic iron mineral are free of carbon, rocks containing fine-grained siderite commonly contain a few tenths of a per cent carbon, and most pyrite-rich rocks are highly bituminous shales.

Applications of the Model

We now will investigate how the model fulfills the four functions defined above

1) *The model as framework and guide to future observation.* A few simple observations permit ready classification of iron formations into mineral facies. If the reader prefers my definition of mineral facies, he should inspect thin sections for the presence of hematite dust, or microcrystalline iron silicate (greenalite, minnesotaite) or siderite. If he prefers assemblage facies he must determine all iron minerals, paying particular attention to carbonate mineralogy; ankerite, dolomite and calcite are more common than is generally assumed.

2) *The model as norm.* Basically, this model has served as a norm since it was first proposed by James (1954). Of course, deviations from the norm exist: e.g., very rarely, sandstones sandwiched between black shale may contain considerable pyrite. Based on the model, we can try to give an answer to the question why is this so, as I will attempt to do below.

3) *The model as basis of physico-chemical interpretation.* The Gunflint Formation contains beds, 5 to 10 cm thick, of hematite iron formations intercalated between organic-rich pyritiferous shales (Figs. 10-12). Fine-grained hematite has been replaced at the margins of these beds, first by iron silicate (greenalite), then by carbonate (siderite). Clearly, replacement of earlier hematite by siderite is indicated since the sedimentary textures of the rock are successively destroyed during growth of carbonate.

My tentative interpretation is that solutions containing decay products of organic matter diffused into the bed of hematite iron formation from the adjoining shale layers, and established a reducing diagenetic environment.

Similarly, the sand beds containing pyrite discussed above are sandwiched between organic-rich shales in which sulphate-reducing bacteria were active. Thus, the abnormal precipitation of pyrite in a sandstone probably is due to infiltration of H₂S bearing diagenetic solutions from the adjoining shale into the sandstone bed.

4) *The model as a predictor.* Extremely reducing and extremely oxidizing environments never should be in direct contact. Thus, when we find two outcrops of iron formation, one containing hematite dust as earliest iron mineral, the other with microcrystalline siderite, we can be fairly certain that the two rocks do not belong to stratigraphic units that are directly in contact. However, we are quite unable to predict the thickness of any transitional zone, which may be quite thin.

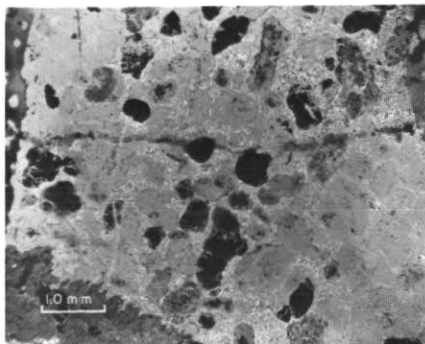


Figure 10.

Part of a jasper layer intercalated between black shale, replaced at the margin by iron silicate (minnesotaite) and carbonate. Here peloids preserved by submicroscopic hematite crystallites are shown. Section G7B, Gunflint Formation.

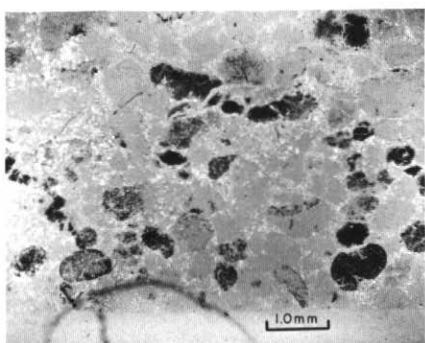


Figure 11.

Same section as Figure 10. Peloids have been replaced by stiplnomelane.

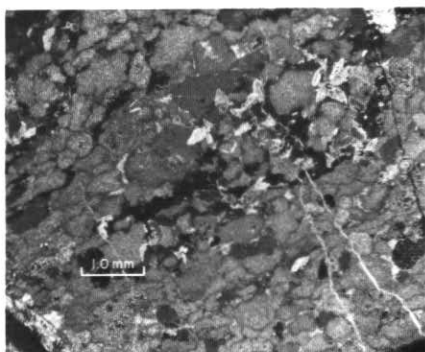


Figure 12.

Same section as Figure 10. Siderite porphyroblasts replaced peloids and part of the cements. Growth of siderite porphyroblasts is related to carbonate-filled fractures, to stylonites, and is most intense at the margin of the jasper bed.

Bibliography

This list of references has purposely been kept very brief.

1) Iron Mineral Facies

Borchert, H., 1960, Genesis of marine sedimentary iron ores: *Instit. Mining Metallurgy Bull.*, No. 640, p. 261-279.

James, H. L., 1954, Sedimentary facies of iron formation: *Econ. Geol.*, v. 49, p. 236-293.

James, H. L., 1966, Chemistry of the iron-rich sedimentary rocks: *U.S. Geol. Survey Prof. Paper*, 440W, 61 p.

Klein, C., Jr. and R. P. Fink, 1976, Petrology of the Sokomon in The Howells River area at the westernmost edge of the Labrador trough: *Econ. Geol.*, v. 71, p. 453-487.

Zajac, J. S., 1974, The Stratigraphy and Mineralogy of the Sokomon Formation in the Kuob Lake area, Quebec and Newfoundland: *Geol. Survey Canada Bull.* 220, 159 p.

In all these papers iron mineral facies is considered a primary, sedimentary facies. Conclusions on the sedimentary environment are drawn that cannot be maintained at the present state of knowledge.

2) Diagenesis of Iron Minerals in Recent Sediments

Berner, R. A., 1970, Sedimentary pyrite formation: *Amer. Jour. Sci.*, v. 268, p. 1-23.

Berner, R. A., 1971, *Principles of Chemical Sedimentology*: New York, McGraw Hill, 240 p.

Ho, C. and J. M. Coleman, 1969, Consolidation and cementation of recent sediments in the Atchafalaya Basin: *Geol. Soc. Amer. Bull.*, v. 80, p. 183-192.

Porrenga, D. H., 1966, Glauconite and Chamosite as depth indicators in the marine environment: *Marine Geol.*, v. 5, p. 495-501.

Berner's book should be familiar to any student of iron formations.

3) Physico-Chemistry of Iron

Garrels, R. M. and C. L. Christ, 1965, *Solutions Minerals and Equilibria*: Harper and Row, New York, 450 p.

4) Description of Iron Formations in Canada

Goodwin, A. M., 1956, Facies relations in the Gunflint iron formation: *Econ. Geol.*, v. 51, p. 565-591.

Goodwin, A. M., 1962, Structure stratigraphy and origin of iron formation, Michipicoten area, Algoma District, Ontario: *Geol. Soc. Amer. Bull.*, v. 73, p. 561-586.

Gross, G. A., 1968, Geology of iron deposits in Canada, v. 3, Iron ranges of the Labrador Geosyncline: *Geol. Survey Canada Econ. Geol. Dept.* 22, v. 3, 179 p.

5) Description of Other Iron Formations

Bayley, R. W. and H. L. James, 1973, Precambrian iron formations of the United States: *Econ. Geol.*, v. 68, p. 934-959.

Beukes, N.J., 1973, Precambrian iron formations of South Africa: *Econ. Geol.*, v. 68, p. 960-1004.

6) Other References Cited

Button, A., 1976, Iron formation as end-member in carbonate sedimentary cycles in the Transvaal Supergroup South Africa: *Econ. Geol.*, v. 71, p. 193-201.

Davy, R., 1974, A Geochemical study of a dolomite-BIF transition in the lower part of the Hamersley Group: *Geol. Survey Western Australia, Annual Rept.* 1974, p. 81-100.

Kimberley, M. M., Origin of iron ore by diagenetic replacement of calcareous oolite: *Nature*, v. 250, p. 319-320.

Walker, R. G., 1976, Facies models 1. General Introduction: *Geosci. Canada*, v. 3, p. 21-24.

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