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BIF-hosted iron ore deposits—Hamersley style

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EXPLORATION MODEL

Examples

Metamorphosed supergene: Mt Tom Price, Mt Whaleback, Paraburdoo, Channar, Giles (Western Australia); Carajas (Brazil); Sishen (South Africa).
Non-metamorphosed supergene: (Dominant Hamersley resources.) Marandoo, BHP OB 29, Area C, Hope Downs, West Angela, Ophthalmias

Target

- New finds of outcropping ores are rare.
- Exploration now includes search for buried ores suitable for open pit mining.
- 5–300 Mt, depending on proximity to established infrastructure.
- >62% Fe < 0.08% P.

Mining and treatment

- Stratabound tabular bodies—mining is limited only by open-pit design needs.
- Grade within narrow limits, by on-site and port blending of lump (6–30 mm >64% Fe); fines (<6 mm >62% Fe).
- Typical export specification of ≤ 0.075% P.

Local geological criteria

- Simple hydrodynamic (sub)-artesian systems with BIF as the aquifer; exposed to the atmosphere along strike (kms), with exits for dissolved gangue.
- Surface-expressed faults or fractures needed to initiate the replacement process at depth (ores grow upward).
- Other pre-ore faulting and thrusting can add considerable structural complexity, but are not essential for enrichment.

Mineralisation features

- Ores are almost entirely *ferric* iron.
- Lamination and texture are preserved.
- Magnetite oxidises to martite (hematite).
- Pseudomorphs of goethite after gangue minerals (includes P, from BIF apatite) are characteristic of non-metamorphosed ores.
- BIF residues include chert, with kaolinite from silicate horizons.
- Burial metamorphism converts goethite to microplaty hematite, destroying internal texture, but bedding/lamination remains to high metamorphic grades.

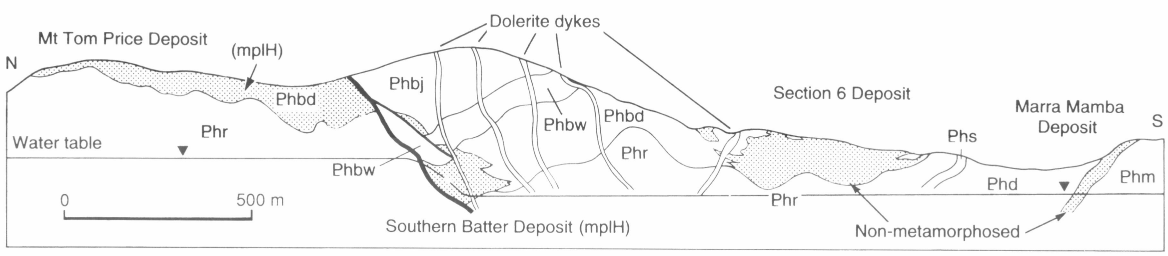

Figure 1. Generalised section at Mt Tom Price, showing metamorphosed iron ore deposits (*mpl H*) in the Brockman IF and non-metamorphosed deposits in Brockman and Marra Mamba IFs.

Regional geological criteria

- Synforms in BIF sequences of any age.
- Enrichment occurs at unconformities exposed to an oxic atmosphere (post-2000 Ma).
- Two major periods of ore forming worldwide (Precambrian and Mesozoic) with possible Paleozoic (North America).
- Burial metamorphism (goethite to hematite) is needed for premium ores.

Alteration

- Weathering of ore produces 3 tiers:

1. 1–2 m of hematite-dominated outcrop, texture/bedding/lamination preserved.
2. Deep hydrated zone (hematite forms Al-goethite) with destruction of texture and bedding, a type of lateritic ferricrete.
3. Ore zone, with post-enrichment leaching of goethite increasing friability.

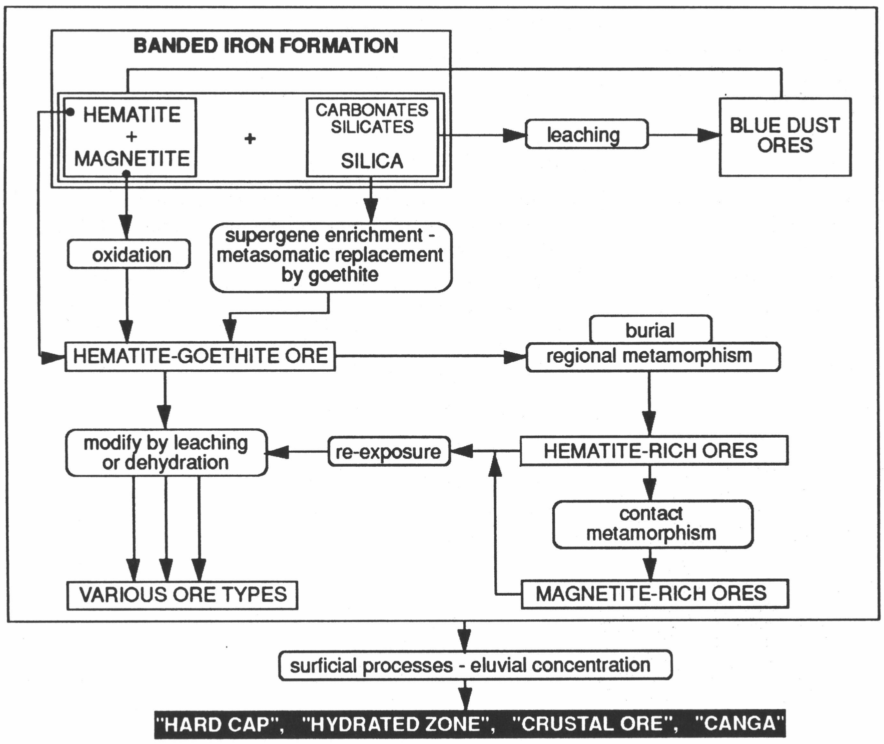

Figure 2. Schematic of genetic modelling for the major BIF-hosted iron ores of the world.

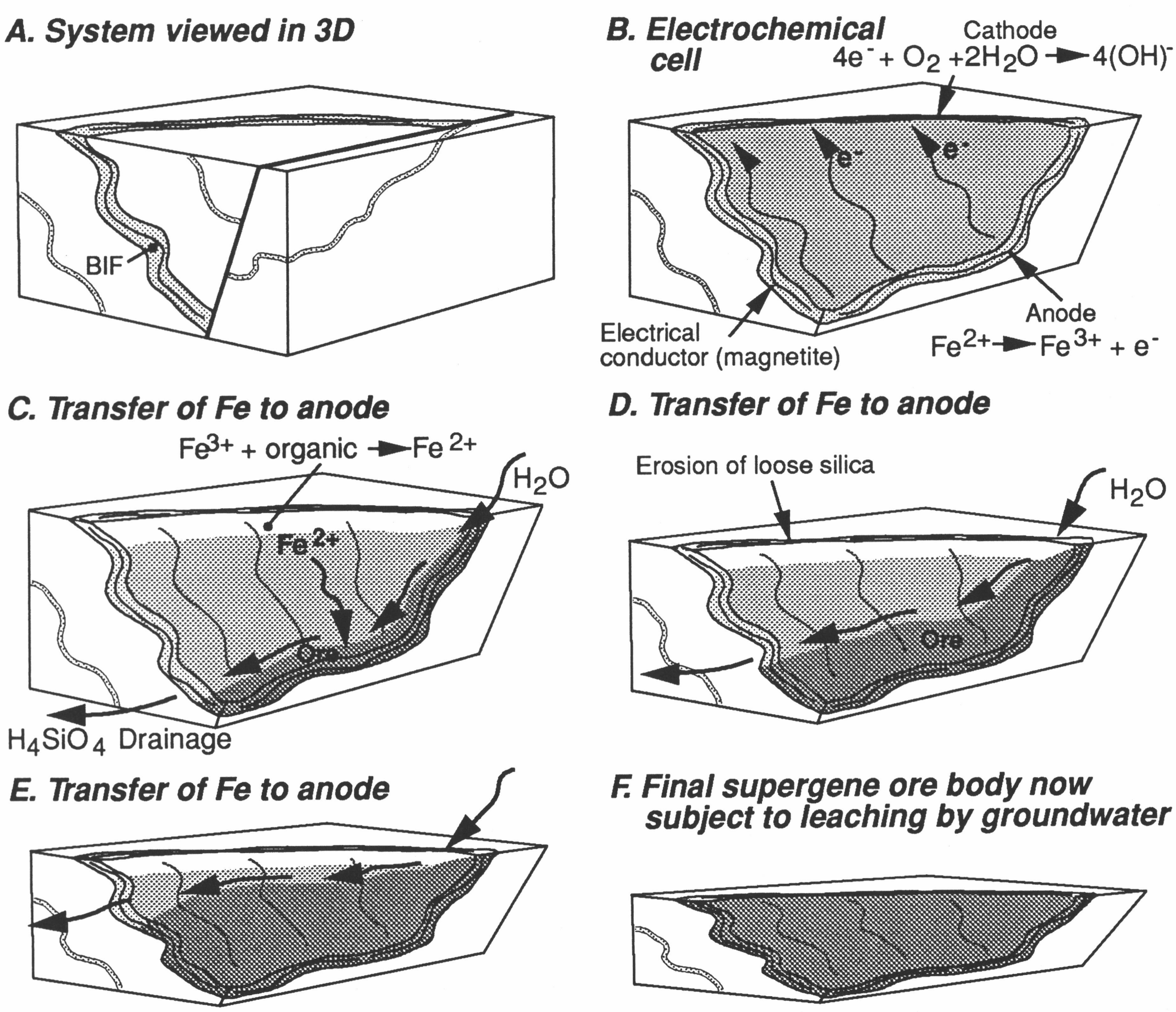

Figure 3. Model of supergene iron ore growth from below, as erosion removes the surface.

Introduction

Much of the following text is summarised from Morris (1980, 1985) and Harmsworth et al. (1990), which should be consulted for detailed citation. Though based mainly on research in the Hamersley Province ('Hamersleys'), experience has shown that the basic concepts apply to BIF-hosted ores world-wide.

Major ore types

Although the dominant world resources of iron ore are associated with the major late Archean–Proterozoic BIF depositories, ores are known from the whole gamut of these chemical sediments. Because export specifications for Australian bedded ores typically call for $\leq 0.075\%$ P, iron ores in Australia tend to be considered in terms of '*low phos or high phos*'.

BIF-hosted iron ores can be divided broadly into three groups:

1. *Leached BIF*: i.e. residual concentrates of Fe oxides ('blue dust ores'). Very low P.
2. *Martite–goethite ores*: moderate–high P, typically 0.07–0.17% P.
3. *Martite–microplaty hematite ores (± residual goethite)*: Low P, < 0.07%.

Deposits of group 2 and 3 ores are shown in Figure 1.

Weathering affects all three ore types in various ways, particularly in tropical areas, and early exploration was typically confined to the resulting complex profiles. This, compounded by limited communication between rival groups, led to the conflicting genetic models of the past.

The mature three-tiered weathering profile of ore in the Hamersleys includes:

- *an outcropping zone* of 1–2 m of hematite-rich ore, the 'carapace', with preserved BIF features resulting from dehydration of the goethite-replaced matrix, underlain by:
- a highly modified *hydrated zone*, up to 70 m, dominated by aluminous goethite, and marked by destruction of parental features.

These two tiers, comprising the 'hardcap', overlie

- the *ore proper*, which retains the original BIF features, but is typically affected by moderate–severe post-enrichment groundwater leaching of goethite and, less often, of hematite.

Leached BIF

The leached BIF group is derived from BIF by direct dissolution of the gangue minerals by groundwater to give residual concentrates of Fe oxides. The ore quality ranges from high-grade residues of martite (oxidised magnetite = hematite) ± primary hematite (so-called 'blue dust ores'), to highly siliceous but very friable BIF, readily upgraded by magnetic or density methods. Ores of this derivation, often associated with enrichment ores, are found in high-rainfall areas such as Brazil and India, but not in the arid Hamersleys. A small deposit of this type is known from the wetter southern coastal area of Western Australia (South Downs deposit).

Martite–goethite

The martite–goethite group includes a wide range of ores, in deposits up to hundreds of millions of tonnes, which show residual BIF features, such as banding, as well as goethite-pseudomorphed microtextures. These goethite-rich ores range from firmly indurated brown material to very friable yellow ochre, all with martite and, often, primary hematite. Deleterious phosphorus levels range from around a moderate 0.07% to a very high 0.17% P, depending on the apatite content of the parent BIF and post-enrichment leaching of goethite, the major carrier of P. Because they lack evidence of metamorphism, are typically related to present erosion surfaces, and contain abundant hydrous Fe oxides, these hematite–goethite ores are generally accepted as resulting from supergene enrichment.

Martite–microplaty hematite

The martite–microplaty hematite group includes the premium, low-P hematite-rich ores (often inaccurately called *specularite* ores), the origin of which has attracted much debate and which are here attributed to metamorphism of post-2000 Ma supergene ores. The group is characterised by the presence of well-preserved primary banding, but with loss of the internal texture found in the non-metamorphosed ores, as a result of the growth of abundant secondary microplaty hematite (*mpl H*) from the goethite-pseudomorphed gangue. With increasing metamorphism they range to granular or 'micaceous' hematite. The immense size of some of these deposits (Mt Whaleback, Western Australia, >1500 Mt; N4E deposit, Carajas, Brazil, >3000 Mt) makes them the largest and most concentrated secondary accumulations of any single metalliferous element in the Earth's crust, approached in scale only by some aluminium orebodies.

A subgroup comprises small deposits of magnetite-rich ore, rarely exceeding a few million tonnes and generally associated with *mpl H* ores. These magnetite segregations can be attributed to thermal metamorphism of normal ores by basic intrusives.

Ore genesis

There are five features of the BIF-hosted Fe-enrichment orebodies that merit special attention when considering their origin.

Scale. Individual deposits range from a few hundred to over 3 billion tonnes of >64% Fe. They may extend along strike for five or more kilometres and more than two kilometres down structure. Typically, they grade abruptly from ore grade (>55% Fe) to altered BIF (<30% Fe) over distances that are negligible compared with the size of the deposits—sometimes less than a metre.

Purity of the ores. The ores are dominated by ferric Fe oxides and only extremely rarely contain exotic components that cannot be attributed to residues from the parent rocks.

Oxidation state of the ores. A few relatively small deposits are mainly magnetite; others contain magnetite and hematite, but, quantitatively, it is the largely oxidised deposits, comprising hematite or hematite–goethite ores, that dominate the world scene. Despite their oxidised character, these extend to depths well below the normal influence of the atmosphere, but there are no significant mineralogical differences between shallow (below hardcap) and deep ore.

Wall-rock alteration. It is generally agreed, even by advocates of hypogene models, that none of the major iron deposits contains evidence for the upward passage of fluids through the host rocks or associated faults. This is of considerable importance in view of the immense amount of solution necessary for enrichment, whatever mechanism is advocated.

Stratigraphic situation. Most iron ores are strata-bound and there is a total absence of genetically related 'mineralisation' in the associated rocks.

Syngenetic models

These hypotheses have received minimal endorsement. Modelling was based originally on a clastic origin for BIF, and on the strata-bound character of the ore zones, which were thus considered to be a type of placer deposit. With better access as the result of mining, and with the acceptance of the concept of a chemical origin for BIF, the models lost favour. Some unpublished support for syngenetic models is still expressed in the Hamersleys.

Supergene models

Support for supergene models has been strong in North America since the 1960s. However, Morey (1983) probably reflected the North American attitude to the variety of local supergene and hydrothermal models when he stated that none had proven to be entirely satisfactory. Ore formation related to 'weathering crusts' of various ages and to metamorphism of early concentrations apparently had considerable support in the former USSR (Sokolov & Grigor'ev 1977). A general supergene and supergene–metamorphic model (Morris 1980, 1985, Harmsworth et al. 1990) is widely accepted in the Hamersleys, largely as a result of detailed petrographic evidence.

Hypogene models

There are three quite different ore types for which hydrothermal origins are proposed. The first of these, the small magnetite deposits, such as those found with *mpl H* ores in the Marquette Range (USA), are possibly acceptable as hypogene, particularly with the evidence of massive cross-cutting magnetite veins and hydrothermal effects. More convincing are those, such as the specular, coarse, platy, micaceous or bladed hematite concentrations with magnetite, in vertical pipe-like folds affecting supergene ore at Koolyanobbing (Yilgarn Block, Western Australia). These pegmatite-like aggregates are associated with hydrothermal quartz and include hematite crystals up to 30 cm. These 'hypogene' ore types are in reality supergene concentrations modified locally by hydrothermal solutions.

Hypogene models have been used for the *mpl H* ores, but are difficult to justify in view of, among other points, 1) the absence of conduits or wall-rock alteration below the orebodies; 2) the presence of low-temperature phases, such as goethite, in deposits that have not been entirely leached or strongly metamorphosed; and 3) the absence of hydrothermal minerals.

To explain the absence of obvious conduits it is usually suggested (e.g. Dorr 1965) that diffusing solutions selectively dissolved undetectable amounts of iron from the surrounding BIF, and transported this to the ore sites. Here the fluids changed character to dissolve and remove all the non-Fe components, precipitating hematite in their place, again without evidence of their passage. Since at least greenschist conditions are stated or implied, this process must have occurred under as much as 10 km of cover. Ferric iron, in the absence of complexing agents, is virtually insoluble at pH >3, even at high temperatures. Data from oceanic ridge systems and contemporary volcanic systems attest to the efficiency of iron transport in the ferrous form (~1% Fe^{2+}) in supercritical aqueous fluid phases. But, equally, these fluids are not discriminating solvents. They invariably leave evidence of their passage in the form of strongly leached residues, a feature not associated with the typical BIF ores.

Assuming that the absence of conduits can be explained, an even greater problem arises in the redox situation. Detailed mechanisms for replacement processes are rare, because, as stated by Dorr (1965), a primary advocate of hypogene modelling, 'it is difficult to imagine possible conditions under which large scale and continuing oxidation of ferrous iron at the site of deposition might take place…no large-scale and continuing source of oxygen is easily found'.

Supergene enrichment and metamorphic upgrading—modelling

Nearly two decades of CSIRO–AMIRA research into iron ore has shown that supergene enrichment of BIF, followed in some cases by burial metamorphism, is the only mechanism that, currently, can reasonably explain the known data from the iron ore deposits, without the use of 'geological theology rather than scientific knowledge', as Dorr (1967) put it so well.

Figure 2 is a general schematic diagram, modified from Morris (1985), outlining the broad model. Essentially, oxidation of magnetite and metasomatic replacement of gangue minerals by goethite produces hematite–goethite ore. This ore may be modified by continued groundwater leaching or, by dehydration due to surface exposure, to produce a range of porous to extremely dense ore types. These *non-metamorphosed* ores of the Mesozoic comprise over 90% of the resources of bedded ore in the Hamersleys, but are little known elsewhere.

However, the earliest period of enrichment along exposed margins of the McGrath Trough probably resulted from the change to an oxic atmosphere at about 2000 Ma. Later, burial to about 4 km resulted in low-grade 'metamorphism' (ca 100°C) of the deposits, converting goethite to form hematite-rich ores with some residual goethite. Subsequent erosion in the Mesozoic re-exposed the ancient ores to groundwater leaching, which removed the bulk of the residual goethite, leaving low-P, hematite ore. These are the high-grade *mpl H* ores of the Tom Price–Whaleback type. The Paraburdoo type is less leached, containing remnant goethite and about 0.07% P. It is these *mpl H* types, with various metamorphic grades, that comprise the major high-grade ore resources outside Australia.

Six essential conditions for the genesis of supergene iron ore (Figure 3).

1. An Fe-rich aquifer (i.e. BIF) both as a source of iron and a target for enrichment.
2. Impervious layers, such as shales, above and below as aquicludes.
3. Suitable open structures, such as plunging synclines, together with deep-water access to initiate the system at depth, e.g. fault-engendered fracture zones or fold-axis fractures.
4. Suitable geochemical conditions, including exposure to the atmosphere and exits for dissolved components.
5. Suitable electrochemical conditions.
6. Tectonic stability for long periods.

To form high-grade *mpl H* ore of the Tom Price–Whaleback type requires a further step:

7. Burial metamorphism of supergene ore with goethite recrystallising to *mpl H*, followed by later exhumation of ore, and leaching of the remnant goethite with its included phosphorus.

The creation of ore in BIF requires the transfer of iron from near surface, by biogenically aided dissolution, into the reacting zone at depth. If, as in some models, the ore bodies grew from the surface downward, then the ore genesis must have been much faster than surface erosion in order to generate the great depths of ore typically observed. A more realistic appraisal indicates growth upward as the surface BIF erodes (Fig. 3).

Most deep-seated BIF-derived ferric ores have formed at depths beyond the likely reach of oxygenated water. Ferric iron is insoluble under most natural conditions. Thus, to have concentrated in its present position by pseudomorphous replacement of the BIF matrix, it must, presumably, have been transported in the soluble ferrous form, followed by oxidation and precipitation as ferric iron at depth. Because magnetite is a good conductor of electrons, atmospheric effects in the sub-outcrop can drive massive electrochemical cells in BIF (Morris et al. 1980). Solution of iron near the surface leaves a friable residue of silica, colloquially known as 'denatured BIF'. This readily erodes, exposing further material for reaction. Iron is transferred in groundwater through pores and fractures to the reacting zone at depth, replacing gangue minerals. Thus, the ore body grows upward, as erosion removes the surface, which forms a topographic low. With the end of the enrichment phase, weathering of the ore begins, while at depth, groundwater movement leaches goethite to produce porous ore. The typical ridges of the deposits today reflect the resistance to erosion of the hardcap.

Exploration

The known Hamersley resources of premium low-phosphorus *mpl H* ores are likely to be exhausted by the middle of the next century and, though the search continues for this ore type, increasing attention is being focussed on the ancillary ores for blending or development as discrete products.

With possible rare exceptions, all significant enrichment deposits have been found by their generally extensive outcrop, which largely explains why there have been no important premium ore discoveries in the past two decades of intensive exploration in the known BIF fields. Nevertheless, there is still potential for economically viable buried bodies, and the search continues. The obvious targets, using supergene–metamorphic modelling, are Precambrian unconformities, but if the ores have not been sufficiently metamorphosed or have not undergone a leaching stage, they may retain a significant goethite(+P) matrix. Though it could be expected that some phosphorus might have been exhaled with water from goethite during metamorphism, the ore may not reach the hoped for quality of the best exposed ores.

Systematic regional mapping is important for this exploration, and recognition of features equivalent to the Precambrian McGrath Trough and their relationship to post-2000 Ma exposure and later metamorphism will help to focus the search areas for the buried *mpl H* ores. Following this, the choice for more detailed exploration should be centred on structural situations that can be related to suitable pre-ore sub-artesian systems. Geophysical techniques have been used more as an adjunct to evaluation than for exploration in the past, but the development of airborne gravity methods could add a significant tool to the search for buried deposits.

The original cover over *mpl H* bodies might perhaps be expected to show some evidence of an aureole of hydration, possibly with some increase in trace phosphorus, from the exhalation of water from goethite. However, since the cover sediments themselves were originally saturated with water before this low level of metamorphism (ca. 100°C), this is arguable. Some increase in quartz/hematite veining might be a more realistic expectation.

The best hope for viable buried *mpl H* deposits in the Hamersleys is those eroded to base level and concealed beneath alluvium in the intermontane areas. Clues to these lie in the identification of *mpl H* texture in alluvials around the margins of the McGrath Trough(s).

An alternative exploration model could be based on hypogene modelling. As discussed already, there are major difficulties in applying this concept to any known ores in the Hamersleys, but that does not invalidate the concept for a new ore type so far (1997) not seen. However, care should be taken to avoid mistaking post-ore metasomatic events in current ore types, such as the secondary magnetite, apatite, and pyrite of the Southern Ridge deposit at Mt Tom Price, for hydrothermal processes genetically related to iron ore formation.

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References

Dorr, J. van N., 1965. Nature and origin of high-grade hematite ores of Minas Gerais, Brazil. Economic Geology, 60, 1–46.

Dorr, J. van N., 1967. The Fort Gouraud, Mauritania, iron ore deposits—discussion. Economic Geology, 62, 567–572.

Gair, J.E., 1975. Bedrock geology and ore deposits of the Palmer Quadrangle, Marquette County, Michigan. U.S. Geological Survey, Professional Paper 769, 159 pp.

Gross, G.A., 1965. Geology of iron deposits in Canada: Volume I, General geology and evaluation of iron deposits. Geological Survey of Canada, Economic Geology Report 22.

Harmsworth, R.A., Kneeshaw, M., Morris, R.C., Robinson, C.J. & Shrivastava, P.K., 1990. BIF-derived iron ores of the Hamersley Province. In: F.E. Hughes (editor), Geology of the mineral deposits of Australia and Papua New Guinea. Australasian Institute of Mining and Metallurgy, Monograph 14, 617–642.

Kneeshaw, M., 1984. Pilbara iron ore classification—a proposal for a common classification for BIF-derived supergene ores. Australasian Institute of Mining and Metallurgy, Proceedings, 289, 157–162.

MacLeod, W.N.,1966. The geology and iron deposits of the Hamersley Range area, Western Australia. Geological Survey of Western Australia, Bulletin 117, 170 pp.

Morey, G.B., 1983. Animikie Basin, Lake Superior Region, U.S.A. In: Trendall, A.F. & Morris, R.C. (editors), Banded iron-formation: facts and problems. 13–68. Elsevier, Amsterdam.

Morris, R.C., 1980. A textural and mineralogical study of the relationship of iron ore to banded iron-formation in the Hamersley Iron Province of Western Australia. Economic Geology, 75, 184–209.

Morris, R.C., 1985. Genesis of iron ore in banded iron-formation by supergene and supergene–metamorphic process—a conceptual model. In: Wolf, K.H. (editor), Handbook of strata-bound and stratiform ore deposits. 13, 73–235. Elsevier, Amsterdam.

Morris, R.C., Thornber, M.R. & Ewers, W.E., 1980. Deep-seated iron ores from banded iron-formation. Nature, 288, 250–252.

Sokolov, G.A. & Grigor'ev, V.M., 1977. Deposits of iron. In: V.I. Smirnov (editor), Ore deposits of the USSR. 1, 7–113. Pitman, London,

Trendall, A.F., 1983. The Hamersley Basin. In: Trendall, A.F. & Morris, R.C. (editors), Iron-formation: facts and problems. 69–129. Elsevier, Amsterdam.

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