Mineralium Deposita (1999) 34: 784–789

LETTER

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Hydrothermal origin for the 2 billion year old Mount Tom Price giant iron ore deposit, Hamersley Province, Western Australia

Received: 16 February 1999 / Accepted: 14 May 1999

Abstract Giant iron-ore deposits, such as those in the Hamersley Province of northwestern Australia, may contain more than a billion tonnes of almost pure iron oxides and are the world's major source of iron. It is generally accepted that these deposits result from supergene oxidation of host banded iron formation (BIF), accompanied by leaching of silicate and carbonate minerals. New textural evidence however, shows that formation of iron ore at one of those deposits, Mount Tom Price, involved initial high temperature crystallisation of magnetite-siderite-iron silicate assemblages. This was followed by development of hematite- and ferroan dolomite-bearing assemblages with subsequent oxidation of magnetite, leaching of carbonates and silicates and crystallisation of further hematite. Preliminary fluid inclusion studies indicate both low and high salinity aqueous fluids as well as complex salt-rich inclusions with the range of fluid types most likely reflecting interaction of hydrothermal brines with descending meteoric fluids. Initial hematite crystallisation occurred at about 250 °C and high fluid pressures and continued as temperatures decreased. Although the largely hydrothermal origin for mineralisation at Mount Tom Price is in conflict with previously proposed supergene models, it remains consistent with interpretations that the biosphere contained significant oxygen at the time of mineralisation.

Introduction

The Hamersley Province in northwestern Australia contains some of the world's largest deposits of iron ore (Morris 1985; Harmsworth et al. 1990). These are hosted by Archaean to Palaeoproterozoic (2.6 to 2.4 Ga) BIFs of the Mount Bruce Megasequence Set, a succession of volcanic and sedimentary rocks that overlies the Pilbara Craton (Blake and Barley 1992). The BIFs are finely laminated, metamorphosed chemical sedimentary rocks, composed largely of microcrystalline quartz (chert), iron oxides, and iron silicate and carbonate minerals. Shales, carbonate sedimentary rocks, basalt-dolerite and rhyolite are interlayered with the BIFs (Blake and Barley 1992; Barley et al. 1997). The current outcrop patterns of the BIFs result from Palaeoproterozoic deformation. Initial deformation occurred in a foreland setting at ~2.4 Ga (Krapez 1999). This was followed by a second episode of north-south oriented compression (the Ophthalmia Orogeny) prior to development of the Wyloo rift-basin and ultimate development of an aulacogen (divergent stage of the Ashburton Province). Subsequent transpression, during oblique collision of the Pilbara and Yilgarn Cratons, closed the aulacogen. The ages of Ophthalmian compression and Wyloo extension are not well-constrained. However, it appears that Wyloo rifting occurred after 2.2 Ga, probably at about 2.0 Ga (Krapez 1999).

Editorial handling: R. Goldfarb

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Numerous BIF-hosted iron ore deposits occur in the southern part of the Hamersley Province (Fig. 1), in the area of most intense deformation and highest metamorphic grade. Much of the ore is contained in giant deposits of almost pure iron oxides. Although most ore conforms to the bedding of host BIF, its location is generally structurally controlled (Harmsworth et al. 1990). The BIF-hosted ores can be divided into two types (Morris 1980, 1985; Harmsworth et al. 1990): martite-goethite ores and martite hematite ores. Martite-goethite ores range from firmly indurated, brown goethite-rich material to friable yellow ochre, all with martite (hematite replacing original magnetite) and early formed prismatic hematite. These deposits range up to several hundred million tonnes. They show variable preservation of BIF features such as banding and pseudomorphing of gangue minerals by goethite. Because they are typically found in young weathering profiles, and contain abundant hydrous iron oxides, this group of ores is generally accepted to result from supergene enrichment of BIF during Mesozoic to Tertiary lateritic weathering (Morris et al. 1980; Morris 1985).

The second group of deposits includes the low phosphorus martite-hematite (or microplaty hematite) ores at Mount Tom Price, Whaleback (Newman) and Paraburdoo (Fig. 1). This group is characterised by well-preserved primary lamination, but with variable loss of internal texture and the growth of abundant secondary prismatic and microplaty hematite (< 0.001 to > 0.25 mm). The origin of these deposits is more controversial, with a general view that they represent martite-goethite ore bodies that were metamorphosed (goethite replaced by hematite at > 100 °C) during Palaeoproterozoic burial (Morris 1985). The presence of pebbles of martite-hematite ore in ~2.0 Ga Wyloo conglomerates confirms the Palaeoproterozoic age of mineralisation (Morris 1985). Deposits of this type also occur in South Africa and Brazil (Morris 1985; Van Schalwkyk and Beukes 1986). The immense size of the biggest of these deposits (> 3 billion tonnes, N4E deposit, Carajas, Brazil) makes them the largest and most concentrated enrichment of any single metalliferous element in the Earth's crust. If these deposits formed by low-temperature, near surface supergene oxidation of BIF during the Palaeoproterozoic, they provide important evidence for deep lateritic-style weathering, between about about 2.2 and 2.0 Ga, a period when the oxygen content of the atmosphere is interpreted to have increased (Holland and Beukes 1990; Holland 1992).

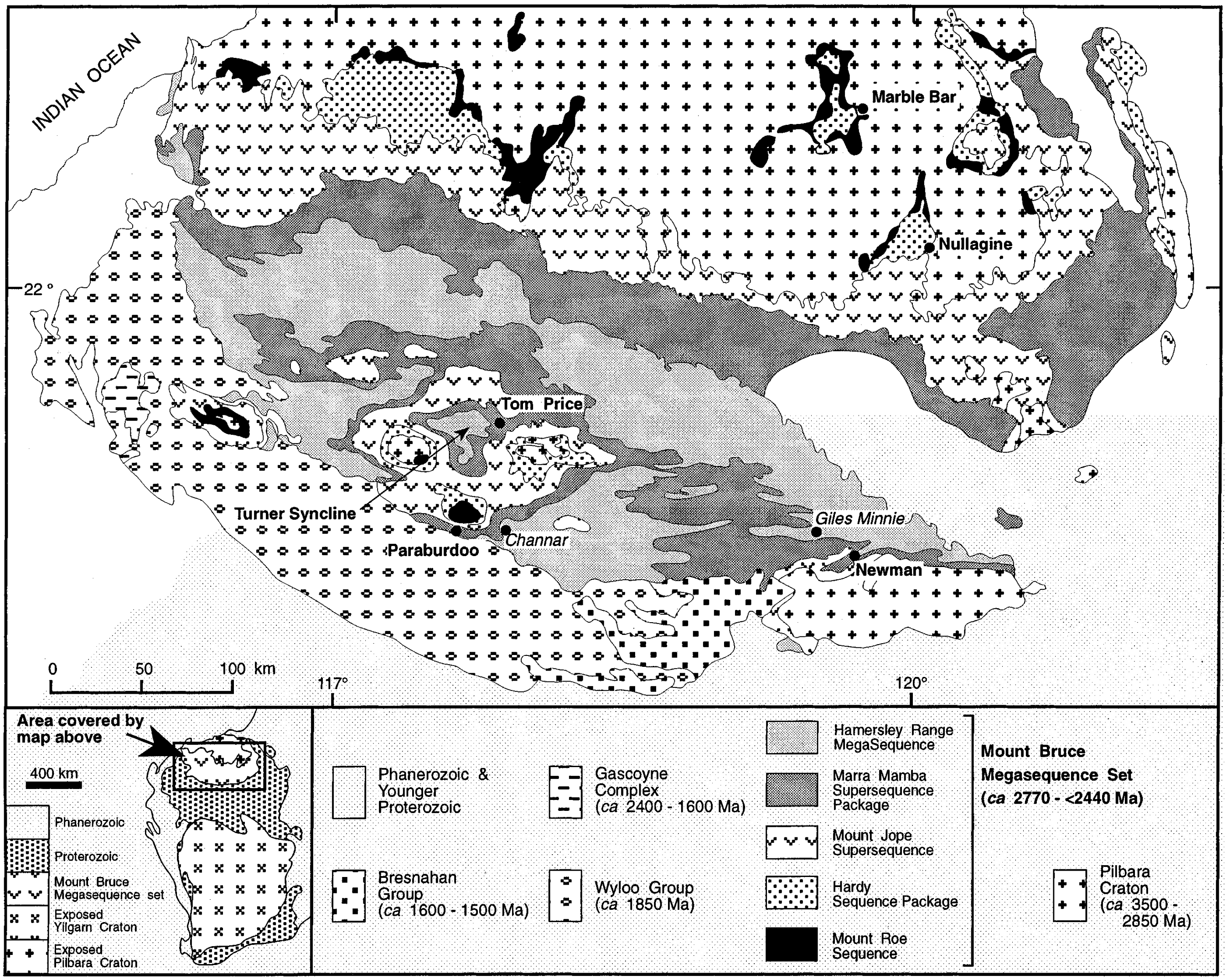


Fig. 1 Geological map of the Hamersley Province showing outcrop of the Mount Bruce Megasequence Set and location of major hematite deposits

However, it is difficult to explain the combination of: (1) the great depth of some ore (>400 m) accompanied by centimetre to metre scale transitions from ore with >60% Fe to BIF with <30% Fe (locally with unmineralised BIF above ore); (2) the clear structural control on the location of ore; and (3) the removal of huge amounts of silica from BIF, by supergene enrichment followed by metamorphism. Consequently, some authors (Li et al. 1993; Powell et al. 1999), following early models (Dorr 1965; Kneeshaw 1975), have suggested a hydrothermal origin for martite-hematite mineralisation. However, as most transitions from unmineralised BIF to ore are overprinted by Tertiary weathering, unambiguous evidence supporting a hydrothermal origin has yet to be presented for any deposit. Deep diamond core-drilling at the Mount Tom Price mine provides a unique opportunity to examine BIF to ore transitions where key paragenetic relationships have not been obscured by later weathering.

Mount Tom Price

Mount Tom Price is located near the eastern closure of the Mount Turner Syncline (Fig. 1). The original resource for this deposit, which is 7.5 km long, up to 1.6 km wide (average 0.6 km) and has a maximum depth of 400 m below surface, was estimated at 900 million tonnes of 63.9% iron and ~0.05% phosphorus (Harmsworth et al. 1990). The orebody occupies two east-trending, shallow-plunging synclines and is disrupted by a west-northwest trending normal fault (Fig. 2). This fault, the Southern Batter Fault, limits outcrop of the ore on the southern side of the main orebody, where ore extensions are down-thrown beneath the Southern Ridge. Mineralisation occurs in BIF of the Dales Gorge Member of the Brockman Iron Formation in the Hamersley Range Megasequence, with minor iron enrichment of adjacent units. The complete thickness of the Dales Gorge Member is enriched at the eastern end of the deposit, with only the upper part of this unit mineralised further west. Hematite mineralisation statically overprints folds and foliations related to the major (Ophthalmian) phase of regional folding (Dettbarn 1991) and cross cuts these structures on Southern Ridge. Martite-hematite ore also occurs at substantial depth beneath unmineralised BIF at two localities: under the Southern Ridge and at North deposit (Fig. 2).

BIF to ore transitions

The following discussion is mainly based on analysis of diamond core-drilling at the North Deposit that shows upwards and lateral transitions over a few metres from magnetite-chert BIF (25–35% Fe), to enriched magnetite(±hematite)-siderite(±iron silicate) BIF

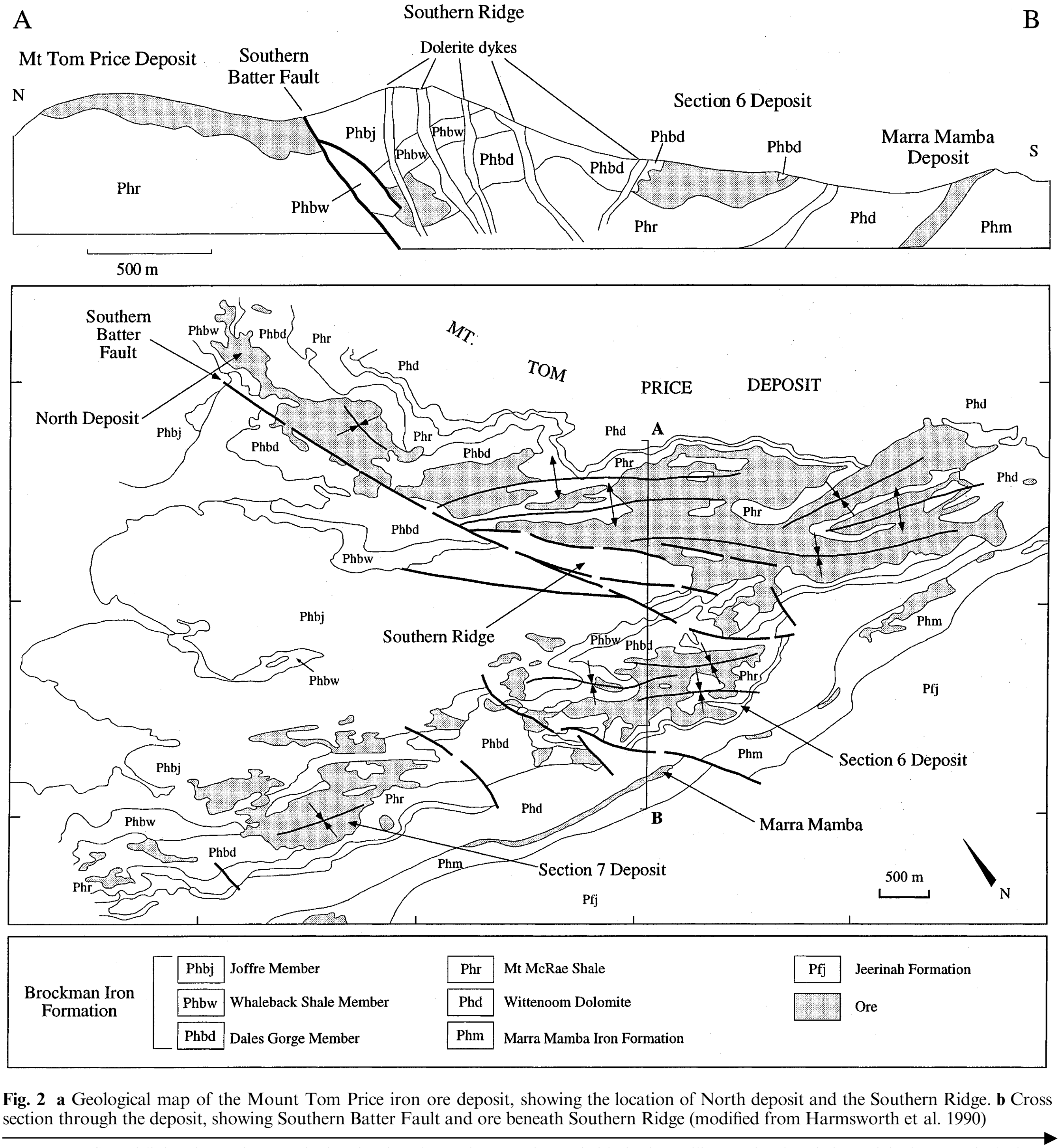
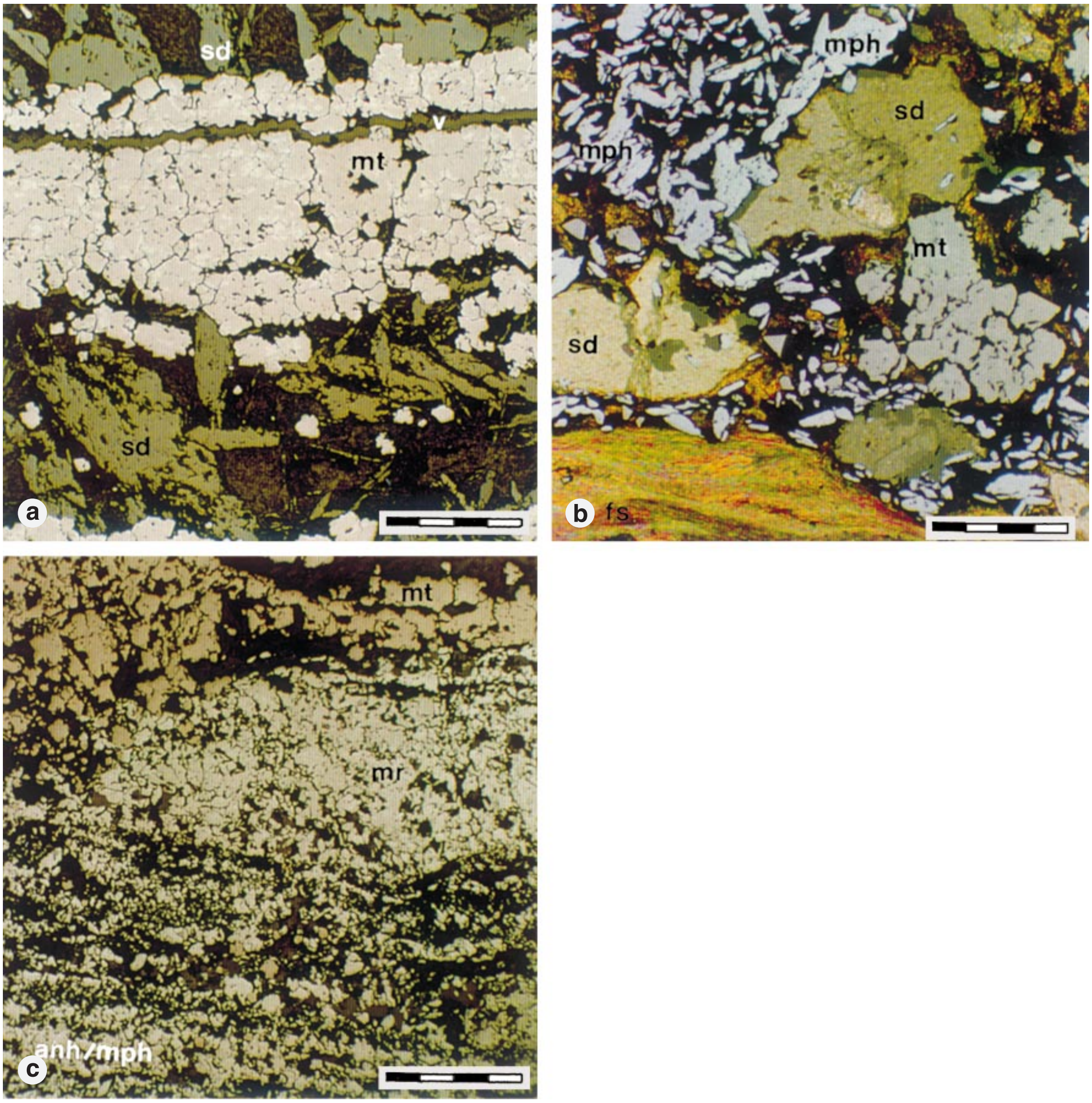


Fig. 2 **a** Geological map of the Mount Tom Price iron ore deposit, showing the location of North deposit and the Southern Ridge. **b** Cross section through the deposit, showing Southern Batter Fault and ore beneath Southern Ridge (modified from Harmsworth et al. 1990)

Fig. 3 **a** Reflected light photomicrograph showing interlayered magnetite and siderite-iron silicate microbands in enriched magnetite(± hematite)-siderite(± iron silicate) BIF from North deposit. Magnetite (*mt*; *brownish-grey*) is euhedral with minor hematite (*h*; *white*) inclusions, and siderite (*sd*) is bladed (*v* = void). Each block on the scale bar represents 100 μm. **b** Photomicrograph of the matrix of a hydrothermal breccia in enriched magnetite(± hematite)-siderite(± iron silicate) BIF from North deposit. The photomicrograph was taken using both transmitted and reflected light to highlight both oxide and non-oxide phases. Clasts of magnetite (*mt*), siderite (*sd*) and iron silicate (*fs*) occur in a matrix comprising prismatic to microplaty hematite (*mph*), ferroan dolomite and iron silicate. Each block on the scale bar represents 50 μm. **c** Photomicrograph of adjacent microbands of magnetite (*mt*), locally replaced by hematite (*martite*, *mr*) and anhedral and microplaty hematite (*anh*/*mph*) from magnetite-hematite mineralisation at North deposit. The iron oxides are separated by either voids (~10%) or relict carbonate, iron silicate and quartz. Each block on the scale bar represents 100 μm

Fig. 4 **a** Photomicrograph of porous martite-hematite ore from North deposit. Magnetite is completely replaced by hematite in a network of anhedral and microplaty hematite (*mph*). The iron oxides are separated by either voids (~30%) or relict recrystallised quartz. Each block on the scale bar represents 100 μm. **b** Detail of **a**, showing skeletal martite (*mr*) in a network of anhedral and microplaty hematite (*anh*/*mph*). The iron oxides are separated by voids (~30%) with relict recrystallised quartz (*q*). Each block on the scale bar represents 25 μm



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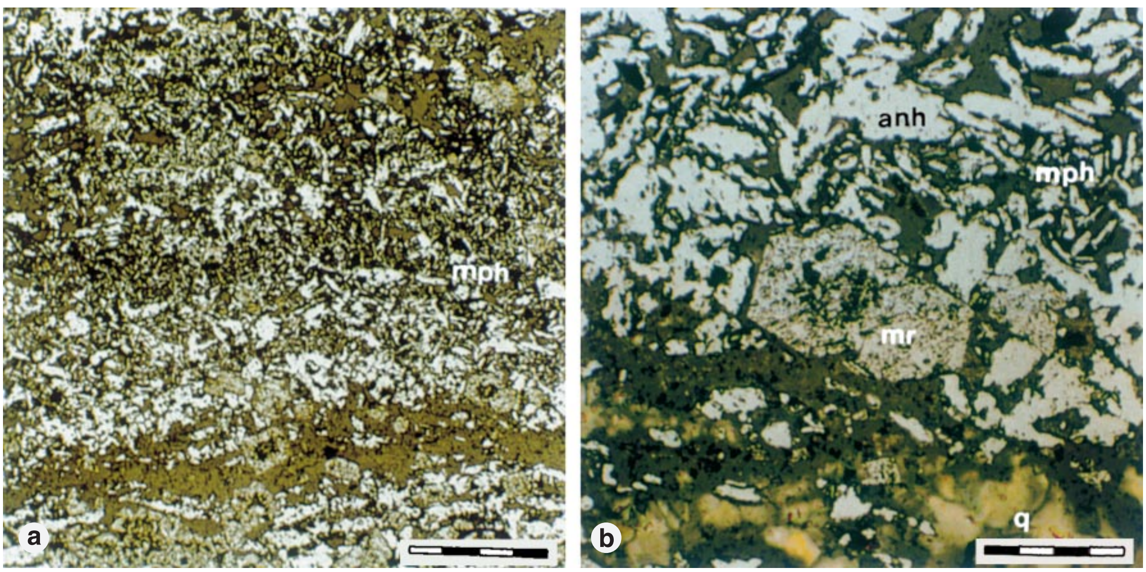

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(50–60% Fe), to magnetite- hematite mineralistion (~60% Fe), to martite-hematite(±goethite) ore (65% Fe). Unmineralised BIF is characterised by millimetre- to centimetre-scale alternating bands (termed microbands and mesobands, respectively) of chert (grain size <20 μm) and euhedral to subhedral magnetite (grain size 20 to 100 μm) with less abundant iron-rich silicate- and siderite-rich bands.

Unmineralised BIF grades into enriched magnetite(±hematite)-siderite(±iron silicate) BIF adjacent to iron ore and shale bands. This BIF is characterised by siderite and very fine-grained iron-rich silicate minerals (such as minnesotaite, stiplnomelane and riebeckite) in the place of original chert microbands and mesobands. It is also enriched in iron (50 to 60% Fe), magnesium (up to 5% MgO), calcium (up to 10% CaO) and carbonate (loss on ignition, LOI, >10%) and depleted in silica (<20% SiO_2) relative to unmineralised BIF (typically ~30% Fe, <1% MgO, CaO and LOI, >40% SiO_2). Where preserved, former chert micro- and mesobands have generally been recrystallised to coarser-grained quartz (grain size 20 to 100 μm). Carbonate minerals are either equant or bladed, and intergrown with equant, subhedral magnetite (Fig. 3a) and anhedral, prismatic or platy hematite.

Hydrothermal breccias and quartz- and carbonate-bearing veins are also locally developed in these zones. Figure 3b shows intergrown magnetite, siderite, iron silicate and hematite in the matrix of a hydrothermal breccia. The presence of hematite in hydrothermal breccias and both within and in the selvedges of quartz and carbonate viens idicates that it crystallised at elevated fluid pressures (relative to supergene processes). Preliminary fluid inclusion analyses on material from North Deposit and Southern Ridge (Hagemann et al. 1999) indicate that quartz in contact with hematite contains primary low- and high-salinity H_2O-NaCl inclusions trapped between 150 and 250 °C, as well as complex salt-rich inclusions (18–25 equivalent wt% $CaCl_2$), and rare CO_2-rich inclusions. These clearly indicate that the quartz-carbonate veins as well as adjacent hematite formed at temperatures greater than 150 °C and up to 250 °C. These temperatures are similar to those estimated for hematite crystallisation in both BIF and ore during regional oxygen isotope studies by Becker and Clayton (1976) and Powell et al. (1999).

As mineralisation is approached hematite contents increase (to more than 50 vol% of iron oxides present) and ferroan dolomite becomes more abundant than siderite, reflecting increased oxidation. The transition from enriched magnetite(±hematite)-siderite(±iron silicate) BIF to magnetite- hematite mineralisation is abrupt and reflects the increased abundance of hematite in former carbonate and iron-silicate micro- and mesobands. Figure 3c shows a magnetite-rich mesoband (only partly oxidised to martite) adjacent to a mesoband of anhedral and platy hematite. Crystallisation of hematite prior to the complete oxidation of magnetite is characteristic of this style of mineralisation at Mount Tom Price and is at odds with the existing interpretation that hematite formed by metamorphism of pre-existing martite-goethite ore (Morris 1980, 1985). The final transition from magnetite-hematite mineralisation to martite-hematite ore occurs over several metres and reflects the complete oxidation of magnetite and the almost complete removal of non-oxide minerals producing a porous network composed of varying proportions of martite, and anhedral, prismatic and microplaty hematite (Fig. 4). Note the presence of fluid inclusions in relict recrystallised quartz in the lower part of Fig. 4b.

In the near-surface portion of the main ore body (Fig. 2) the Dales Gorge Member is almost entirely enriched to massive martite-hematite ore with minimal relict magnetite. However, a number of lines of evidence indicate that the processes described were responsible for formation of the deposit as a whole. The textures of near surface samples from the main ore body (Ayres 1971) are similar to those from North deposit shown in Figs. 3c and 4. The abundance of martite and presence of rare magnetite inclusions in ore were also cited by Ayres (1971) as evidence that ore was derived from magnetite-rich BIF. Limited analyses of BIF adjacent to the main ore body, obtained from early drilling (Ewers and Morris 1981), show it to be unusually magnetite- and carbonate-rich (with higher iron, magnesium, calcium and carbonate but lower silica contents than typical BIF). These samples were interpreted by Ewers and Morris (1981) as metasomatised BIF and unrelated to ore formation. However, comparison with North deposit indicates that they most likely represent iron-enriched, magnetite-siderite BIF formed during the first stage of development of the main ore body.

Discussion

Together, the observations presented are consistent with the following origin for martite-hematite ore at Mount Tom Price. The initial stage of mineralisation involved crystallisation of magnetite(±hematite)-siderite(±iron silicate) assemblages at temperatures above 150 °C (and possibly above 250 °C) and high fluid pressures (i.e. hydrothermal conditions). This stage involved loss of silica (plus recrystallisation of chert to coarser-grained quartz) and addition of magnesium, calcium and carbonate. The high carbonate contents of these rocks coupled with the presence of rare CO_2-bearing fluid inclusions suggest that this stage of mineralisation was most likely due to interaction of BIF with high-temperature probably carbonate-rich fluid. Increasing oxidation (also at elevated temperatures) then resulted in crystallisation of prismatic hematite and ferroan dolomite. Continued oxidation converted residual magnetite to martite, with further crystallisation of hematite as prismatic and microplaty forms, as carbonate and remaining silicate minerals were leached. As evidenced by the fluid inclusion data, this stage most likely resulted from interaction of a highly saline, hydrothermal brine, initially at high temperatures, with a major external oxygen reservoir that was most likely meteoric water. As fluid mixing continued, temperatures decreased, possibly as the deposit was uplifted.

Powell et al. (1999) have recently interpreted the Hamersley iron ore deposits in terms of fluid flow related to uplift, or collapse, of the Ophthalmian fold-thrust belt. That study was based on a regional synthesis of stratigraphic, structural, fluid inclusion, and stable isotope data from BIF and ore, and the interpretation that prismatic hematite adjacent to regional, high temperature (>280 °C) Opthalmian quartz veins is genetically related to hematite in the ore bodies. Detailed textural studies by us and by Dettbarn (1991) show that mineralisation at Mount Tom Price clearly overprints and thus post-dates the main period of Opthalmian folding and metamorphism. Also, Powell et al. (1999) interpret, what we consider to be, Wyloo-aged extensional basins (see arguments in Krapez 1997a, b, 1999) that contain ore-grade hematite clasts, as belonging to the Ophthalmian foreland basin.

We consider that the weight of evidence is consistent with formation of the Mount Tom Price ore body as a response to heat- and fluid-flux during Wyloo rifting at about 2.0 Ga, rather than earlier dominantly compressive Ophthalmian deformation. In this regard, it is also worth noting that Barton and Johnson (1996) have recently implicated highly saline fluids and extensional tectonic settings in the genesis of another class of giant iron-oxide accumulations, the igneous-related iron-oxide (REE-Cu-Au-U) deposits.

It is difficult to reconcile the observations presented here with the interpretation that martite-hematite ore at Mount Tom Price was produced by burial metamorphism of original supergene martite-goethite mineralisation formed during a period of Palaeoproterozoic deep, lateritic-style weathering (Morris 1980, 1985). Consequently, apart from requiring revision of ore genesis models, a largely hydrothermal origin for Mount Tom Price ore has important implications for models of the evolution of atmospheric oxygen and early continental weathering. Because it requires a major input of oxygen from an external source, such as meteoric water, the hydrothermal model for Mount Tom Price is consistent with existing interpretations (Holland 1992) that the biosphere contained significant oxygen at the time of mineralisation. However, it requires that interpretations that suggest Palaeoproterozoic hematite deposits in the Hamersley Province, and elsewhere in the world, are simply the result of ancient deep lateritic-style weathering (±metamorphism), should be re-evaluated.

Acknowledgements We thank numerous Hamersley Iron and Rio Tinto geologists for their help during the period that this work was undertaken. In particular, we are most grateful to Andy Connor, Mahendra Pal, Tony Harding, John Phillips, Graeme Broadbent, Hilko Dahlsta, Dan Flynn and Dennis Taylor. David Groves is thanked for helping ALP with interpretations of mineral textures. He and Bryan Krapez are thanked for reading an early draft of this manuscript.

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