The Structural Setting and Hydrothermal Alteration of BIF-hosted High-Grade Iron Ore Deposits, the Hamersley Province, Western Australia

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All work in this thesis is original, and that of the candidate, unless otherwise acknowledged, specified or referenced

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4th February 2015

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ABSTRACT

The Hamersley province of northwest Western Australia is one of the world’s premier iron ore regions. The high-grade iron ore deposits are mostly hosted within banded iron formation (BIF) sequences of the Brockman and Marra Mamba Iron Formations of the Hamersley Group. The two main types of bedded iron deposits (BID) are: martite-microplaty hematite containing between 60 and 68 wt.% Fe, and martite-goethite containing between 56 and 63 wt.% Fe. Examples of martite-microplaty hematite deposits include Mount Whaleback, Mount Tom Price, Channar and Paraburdoo. The martite-microplaty hematite deposits are spatially associated with normal faults that formed during D2 deformation (~2200 Ma), and have later been affected by D3 deformation (~1650 Ma) and formed by the multistage interaction of hydrothermal fluids with the host BIF formation. Microplaty hematite within martite-microplaty hematite deposits display a variety of sizes, ranging from 20 to 300μm, and textures, ranging from platy to tabular. Microplaty hematite is commonly associated with supergene-modified hydrothermal deposits but can also form in the hydration zone of supergene deposits.

Analysis of the oxygen isotope compositions of magnetite and hematite from BIF, hydrothermal alteration assemblages, and martite-microplaty iron ore was conducted on samples from the Mount Tom Price, Paraburdoo, and Channar iron ore deposits. The results showed that the δ¹⁸O values of magnetite and hematite from hydrothermal alteration assemblages and high-grade iron ore range from –9.0 to –2.9 per mil, a depletion of 5 to 15 per mil relative to the host BIF. The δ¹⁸O values are spatially controlled by faults within the deposits, a response to higher fluid flux and larger influence on the isotopic compositions by the hydrothermal fluids. The oxygen isotope composition of hydrothermal fluids (δ¹⁸Ofluid) indicates that the decrease in the ¹⁸O content of iron oxides was due to the interaction of both basinal brines and meteoric fluids with the original BIF. Late-stage talc-bearing ore at the Mount Tom Price deposit formed in the presence of a pulse of ¹⁸O-enriched basinal brine, indicating that hydrothermal fluids may have repeatedly interacted with the BIFs during the Paleoproterozoic.

At the Paraburdoo deposit, high-grade iron ore (>63 wt. % Fe) of the 4EE orebody of the 4E deposit (>200Mt @ 63.5% Fe) occurs as a southerly-dipping sheet within banded iron formation (BIF) of the Early Proterozoic Dales Gorge and Joffre Members. Structural reconstruction of the 4E deposit established that reactivation of the 18E fault and the development of the NW-striking, steep SW dipping 4E and 4EE normal faults resulted in the preservation of the 4EE orebody below the 4E deposit, and 400m below the modern topographic surface.

Three hypogene alteration zones between unmineralised BIF and high-grade iron ore are observed: (1) distal magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite, (2) intermediate magnetite-dolomite-
hematite-chlorite-quartz-stilpnomelane, and (3) proximal hematite-dolomite-chlorite±pyrite±magnetite alteration zones. Hydrothermal alteration is temporally and spatially constrained by NW trending dolerite dikes that intruded the 4E and 4EE faults prior to hypogene alteration. Six vein types (V₁ to V₆) are recognized at the 4E deposit. The veins both crosscut and parallel the primary BIF layers and are emplaced contemporaneous with the hydrothermal alteration zones which record the transformation of unmineralised BIF to high-grade iron ore.

An integrated structural-hydrothermal alteration and fluid model proposes that during early Stage 1a hypogene fluid flow at the 4E orebody occurred during a period of continental extension and enhanced heat flow within sedimentary basins to the south of the Paraburdoo Range. The heated basinal brines were focussed by the NW-striking, steep SW-dipping 4E and 4EE normal faults and reacted with the BIF’s of the Dales Gorge and Joffre Members. The warm to hot (160° to 255°C) Ca-rich (26.6 to 31.9 eq. wt.% CaCl₂) basinal brine interacted with the magnetite-chert layers of the Dales Gorge and Joffre Members transforming them into magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite BIF. The Fe-rich brine (up to 2.8 wt.%) originated likely from evaporated seawater that had lost Mg and Na and gained Li and Ca through fluid-rock reactions with volcaniclastic rocks and carbonate successions within the Wittenoom Formation. The first incursion of deeply-circulating, low-salinity (5.8 to 9.5 eq.wt.% NaCl) heated (106° to 201°C) modified meteoric water is recorded in late stage 1A. This modified meteoric water had lost some of its Na through wallrock interaction with plagioclase and micas, possibly from interaction with dolerite from the Weeli Wooli Formation that directly overlies the Joffre Member and Dales Gorge Members.

Stage 1b involved continuing reactions between the hydrothermal fluids and the magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite BIF and produced both the intermediate magnetite-dolomite-hematite-chlorite-pyrite alteration zone and the proximal hematite-dolomite-magnetite-stilpnomelane alteration assemblages. Microplaty (10-80μm), platy (100-250μm), and anhedral hematite increasingly replace magnetite in the intermediate alteration zone in order to form the proximal alteration zones which consist of microplaty, platy, anhedral hematite and magnetite. The intermediate and proximal alteration zone represent the mixing of a hot (250° to 400°C), high-salinity Ca-rich (30-40 eq.wt.% CaCl₂), Sr-enriched rich basinal brine and low-temperature and low salinity (~5 eq.wt.% NaCl) modified meteoric water that was heated (~100°-200°C) during its descent into the upper crust. The heterogeneous mixing of the two end-member fluids resulted in the trapping of primary fluid inclusion assemblages containing a wide range of trapping temperatures (up to 200°C) and salinities (up to 25 eq. wt.% NaCl).

Stage 1c of the hypogene hydrothermal fluid is characterized by low-temperature (<110°C), low-salinity (~5 wt.% NaCl) meteoric water, that interacted with the proximal hematite-dolomite-magnetite-stilpnomelane altered BIF leaving a porous hematite-apatite high grade ore. Supergene alteration affected the orebody since the Cretaceous and is characterized by hematite-goethite alteration assemblage. It resulted in the destruction of hypogene alteration zones which are only preserved below the depth of modern weathering.
The martite-goethite bedded orebodies resulted from late Mesozoic supergene alteration of BIF. During this process magnetite was oxidized to martite, whereas silicates and carbonates were oxidized and hydrated to goethite or leached without replacement. The controls on the localization of supergene martite-goethite deposits, for example, the Hope Downs, Cloud Break, and Area C deposits include preexisting structures, such as faults, thrusts, and folds. These structures acted as fluid conduits that directed descending supergene fluids into the host BIF. Dolerite dikes and shale layers further focused and controlled fluid flow. High iron grades at the Area C and Hope Downs deposits are associated with synclinal structures where increased supergene fluid flow. The phosphorus (P) in martite-goethite and martite-microplaty hematite deposits was repeatedly remobilized by both hypogene and/or supergene fluids. The P distribution was controlled by several factors, such as fluid flux in fault zones, permeability of shale layers, and synclinal folds, which resulted in locally high concentrations (>0.10 wt %) of P in the deposits.
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This research was supported by an Australian Research Council Linkage grant, with RioTinto Exploration. The financial support provided resourcing for the work and allowed modern analytical techniques not commonly used in the study of iron ore deposits, to be applied. I would like to thank the Centre for Exploration Targeting (University of Western Australia) for providing a culture of learning that driver independent and creative thinking.

Most importantly I would like to thank my supervisors Steffen Hagemann and Hilke Dalstra for the undying support and unrelenting pursuit of scientific excellence. Their depth of knowledge and pragmatic approach to problem solving has been a great learning tool and has provided me with the tools and attitude to complete this work, as well as developing valuable life skills.
CHAPTER 1: INTRODUCTION


The three papers are presented in the chronological order that they were submitted to Economic Geology. Each successive paper builds on the previous paper’s finding in order to further test concepts and ideas. Additional information about reviewers, handling editors, type of publication (ie journal publication, book chapter) is also provided.

The key conclusions of the thesis, discussion of the results, together with recommendations for further work, are presented in Chapter 5.

The Appendix 1 contains all co-authored refereed journal articles, refereed extended abstracts and non-refereed abstract. A sample register, petrographic data, structural mapping, geological sections, maps, and geochemical data that were used as part of this thesis are presented electronically in Appendix 2.

The following publications resulted from the work completed during this PhD thesis:

Refereed, first authored journal articles:


3. Thorne, W., Hagemann, S., Dalstra, and Banks, D., 2014, Structural control, hydrothermal alteration zonation and fluid chemistry of the concealed, BIF-hosted high-grade 4EE iron orebody at the Paraburadoo 4E deposit, Hamersley Province, Western Australia, Economic Geology, 109: 1529-1562 (Chapter 4).

Refereed, co-authored journal articles:


Refereed extended abstracts:


Thorne, W.S., Hagemann, S.G., and Vennemann, T., 2007a, Oxygen Isotope Compositions of iron oxides from high-grade (>63 wt. % Fe) BIF-related iron deposits of the Pilbara province, Western Australia, in, Andrew et al., ed., Digging Deeper, Proceedings of the Ninth Biennial SGA Meeting, Dublin 2007, p. 1215-1218


Co-authored refereed extended abstracts:


Non-refereed abstract:

CHAPTER 2: BANDED IRON FORMATION-RELATED IRON ORE DEPOSITS OF THE HAMERSLEY PROVINCE, WESTERN AUSTRALIA.

Statement of candidature contribution:
As first author, I was solely responsible for writing of the paper, including designing and drafting of all figures. The role of co-authors was as follows:

- Steffen Hagemann contributed to the organisation, editorial, and scientific scrutiny of the paper.
- Adam Webb contributed to the drafting of Figs. 1, 3 and 8 as well assisting in the descriptions of BHPB’s Mt Whaleback and Area C deposits.
- John Clout contributed to the description of FMG’s Cloud Break deposit, the discussion of the mineralogy of martite-microplaty hematite ore and the zoning of supergene minerals, and the drafting of Figs. 9 and 11.

This chapter which is publication n°1 provides the most comprehensive summary of BIF-related iron ore deposits in the Hamersley province in the last 20 years. The paper contains previously unpublished descriptions of Rio Tinto Iron Ore’s (RTIO) Hope Downs deposit, Fortescue Metals Group’s (FMG) Cloud Break deposit and BHP Billiton’s (BHPB) Area C deposit. Revised models on the genesis of RTIO’s Mt Tom Price and Paraburdoo deposit as well as BHPB’s Mt Whaleback deposit are included, reflecting the large amount of research conducted at these deposits, primarily in the last decade.

This paper also provides a discussion of the most contentious scientific issues within these deposits including the formation of microplaty hematite, phosphorus distribution within the deposits, and the zoning of supergene minerals. The paper describes how microplaty hematite is present within many high grade iron ore bodies and that its presence cannot be used as diagnostic mineral for high-grade ore formed by hydrothermal processes. The common usage of the term, microplaty hematite, makes it difficult to differentiate the different ore types and their geochemical properties and, ultimately, the processes that lead to their formation. The distribution of phosphorus in high-grade iron ore deposits has always been of interest to iron ore companies due to being a deleterious element in the steel making process. The paper describes how phosphorus distribution is controlled by several factors, such as fluid flux in fault zones, the permeability of shale layers and synclinal folds, which resulted in locally high concentrations (>0.10 wt %) of phosphorus in the deposits. Supergene alteration within martite-goethite and martite-microplaty hematite deposits is characterised by a consistent vertical primary zonation patterns, including secondary goethite and replacement textures such as goethite infill and dehydration of goethite to hematite.

New genetic models for BIF-hosted iron ore deposits are presented and discussed at length. Based on these models the fluid source for the heated hypogene fluids and meteoric water responsible for the transformation of BIF to high-grade iron ore is suggested to be a rift-basin to the south of the Hamersley province.

This paper was submitted to Economic Geology, and reviewed by Drs Jay Barton and Bruce M. Simonson. Editorial work was completed by Dr Jens Gutzmer. The paper was published in Reviews in Economic Geology, v. 15, p. 197-221.
Chapter 8
Banded Iron Formation-Related Iron Ore Deposits of the Hamersley Province,
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Abstract

The Hamersley province of northwest Western Australia is one of the world’s premier iron ore regions. The high-grade iron ore deposits are mostly hosted within banded iron formation (BIF) sequences of the Brockman and Marra Mamba Iron Formations of the Hamersley Group and consist of two types: martite-microplaty hematite containing between 60 and 68 wt percent Fe, and martite-goethite containing between 56 and 63 wt percent Fe. Examples of martite-microplaty hematite include Mount Whaleback, Mount Tom Price, and Paraburdoo and examples of martite-goethite ore deposits include Mining Area C (Area C), Hope Downs, and the Chichester Range. The high-grade martite-microplaty hematite ores, which formed in the Paleoproterozoic, have a three-stage origin. Stage 1 involved the release, from the underlying sedimentary successions, of low (110ºC) to high (280ºC) temperature, highly saline (20–25.5 wt % NaCl-CaCl₂ equiv; Ca > Na > K) basinal brines that interacted with the underlying Wittenoom Formation and moved upward in normal faults, such as the Southern Batter fault at Mount Tom Price, the 4E fault at Paraburdoo, and the Central and Eastern Footwall faults at Mount Whaleback, into the host BIF. The hypogene fluids migrated laterally within large-scale folds with permeability controlled by shale layers and northwest-trending dolerite dike sets. The BIF was laterally and vertically altered into magnetite-siderite-stilpnomelane and hematite-ankerite ± magnetite assemblages at Mount Tom Price, a hematite-dolomite-chlorite-pyrite assemblage at Paraburdoo, and formed a dolomite-chlorite assemblage in the Mount McRae Shale at Mount Whaleback. Stage 2 involved deeply circulating, low-temperature (<110ºC), Na-rich meteoric waters that interacted with evaporites prior to their interaction with the BIF. The descending meteoric waters interacted with the carbonate-altered BIF to produce a martite-microplaty hematite-apatite assemblage prior to supergene alteration. Stage 3, the supergene stage during the Mesozoic to Tertiary, is the final stage in the transformation of BIF to high-grade ore. Shallow supergene fluids interacted with the martite-microplaty hematite-apatite assemblage to form a highly porous high-grade (>63 wt % Fe) martite-microplaty hematite ore. Supergene alteration is likely to have occurred for at least 80 m.y. and close to the present topographic surface. High-pressure (>0.10 wt %) martite-microplaty hematite assemblages can therefore form and may remain concealed beneath BIF, below Proterozoic erosion surfaces.

The martite-goethite bedded orebodies resulted from late Mesozoic supergene alteration of BIF. During this process magnetite was oxidized to martite, whereas silicates and carbonates were oxidized and hydrated to goethite or leached without replacement. The controls on the localization of supergene martite-goethite deposits, for example, the Hope Downs, Cloud Break, and Area C deposits include preexisting structures, such as faults, thrusts, and folds. These structures acted as fluid conduits that directed descending supergene fluids into the host BIF. Dolerite dikes and shale layers further focused and controlled fluid flow. High iron grades at the Area C and Hope Downs deposits are associated with synclinal structures where increased supergene fluid flow caused multiple phases of goethite leaching, precipitation, and cementation.

Microplaty hematite encompasses a variety of sizes, ranging from 20 to 300µm, and textures, ranging from platy to tabular. Microplaty hematite is commonly associated with supergene-modified hydrothermal deposits but can also form in the hydration zone of supergene deposits. The phosphorus (P) in supergene and supergene-modified hydrothermal deposits was repeatedly remobilized by both hypogene and/or supergene fluids. The P distribution was controlled by several factors, such as fluid flux in fault zones, permeability of shale layers, and synclinal folds, which resulted in locally high concentrations (>0.10 wt %) of P in the deposits.

It is unlikely that a single model for the formation of the martite-microplaty hematite ore deposits can explain all the structural, stratigraphic, hypogene alteration, and ore characteristics at the Mount Whaleback, Mount Tom Price, and Paraburdoo deposits. Continued collaborative research directed at elucidation of a single tectonic history of the Pilbara, based on collection of similar structural and geochemical data sets from these deposits, will advance genetic ore models and aid in exploration for concealed orebodies.

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Introduction

The Hamersley province of northwest Western Australia (Fig. 1), one of the world’s premier iron ore districts, is currently producing approximately 300 million metric tons (Mt) per annum of hematite ore that is equivalent to 40 percent of the world’s total iron ore seaborne trade (DOIR, 2003). The bedded iron ore deposits of the province are typically classified as either martite-microplaty hematite or martite-goethite and are predominantly hosted within banded iron formation (BIF) sequences of the Brockman and Marra Mamba Iron Formations of the Hamersley Group (Fig. 2). In total they host in excess of 25 billion metric tons (Bt) of iron ore (>55 wt % Fe) across the Hamersley province.

The martite-microplaty hematite ore is characterized by martite and ubiquitous microplates of hematite (i.e., microplaty hematite), variable hardness and porosity, and ranges in iron content between 60 and 68 wt percent Fe (Fig. 1; Clout and Simmonson, 2005). The deposits can be enormous in size (Mount Whaleback, >1,500 Mt) and extend to great depths (>400 m). High-grade martite-microplaty hematite orebodies are nearly all confined to the Brockman Iron Formation, although recent work (Clout, 2005) has identified smaller deposits in the Marra Mamba Iron Formation. The origin of these high-grade martite-microplaty hematite deposits has been controversial, although recent work has predominantly supported an origin linked to both superegene and hydrothermal processes (Li et al., 1993; Martin et al., 1998; Barley et al., 1999; Hagemann et al., 1999; Oliver and Dickens, 1999; Powell et al., 1999; Taylor et al., 2001; Webb et al., 2002; Brown et al., 2004; Dalstra and Guedes, 2004; Thorne et al., 2004, 2005, 2007a, b; Oliver, 2007).

The martite-goethite bedded orebodies resulted from late Mesozoic superegene alteration of precursor BIF (Morris, 1980, 1985; Clout, 2005). During this process magnetite was oxidized to martite, whereas silicates and carbonates were leached or replaced by goethite, or leached without replacement (Morris, 1985). These orebodies form deposits up to hundreds of millions of tons but rarely extend to depths >100 m (Harnsworth et al., 1990).

This paper aims to summarize results of recent research on the high-grade BIF-hosted iron ore deposits in the Hamersley province. Detailed descriptions, including the resource, stratigraphic, structural setting, ore mineralogy, hydrothermal alteration, and superegene fluid chemistry, and preliminary descriptive models of six characteristic high-grade iron ore deposits from the Hamersley province: Mount Whaleback, Mount Tom Price, Paraburdoo, Hope Downs, Mining Area C (Area C), and deposits from the Chichester Range, are presented in the first part (Fig. 1). The second part summarizes and discusses important similarities of all deposits, including the mineralogy of martite-microplaty hematite ore, the distribution of phosphorus within deposits, and the effect of superegene alteration. The third part of the paper provides a review of existing genetic models and discusses their similarities and shortcomings. Finally, open questions and exploration significance are discussed and further research work is suggested.

Mining history

Following the lifting of a Federal Government embargo on the export of iron ore in 1960, iron ore mining commenced in the Hamersley province at Mount Tom Price in 1966, followed by Mount Whaleback (1969) and Paraburdoo (1972). Together these three mines underpinned the majority of production in the region until the 1990s. The deposits are hosted within the Dales Gorge and Joffre Members of the Brockman Iron Formation (Fig. 2) and are characterized by high-grade martite-microplaty hematite ore (<63 wt % Fe) and low phosphorus (P <0.07 wt %). Continued mining of these world-class (>700 Mt, >63 wt % Fe) high-grade deposits for at least 40 years has slowly depleted the regions resources of the high Fe and low P martite-microplaty hematite ore.

Since the early 1990s numerous other deposits within the Brockman Iron Formation (Fig. 2) have commenced mining activities. Along the southern edge of the Hamersley province the Channar and Eastern Range deposits (Fig. 1) near Paraburdoo commenced mining in 1996 and 2004, respectively, and are included within the Greater Paraburdoo mining area. High-grade ore (>60 wt %) is mined predominantly within the Dales Gorge and Joffre Members and 18 Mt/yr (2004) was produced from the Greater Paraburdoo mining area. North of Mount Tom Price, the Brockman BS2 mine (Fig. 1), which was initially a high-grade detrital mining operation (exhausted in 1998), changed to mining high-grade martite-goethite iron ore (<63 wt %, P >0.07 wt %) within the Dales Gorge Member. After major upgrades the Brockman BS2 mine reopened in 2003 and has a production capacity of 8 Mt/yr. Importantly, all three deposits are characterized by both fault-controlled martite-microplaty hematite ore and martite-goethite ore types.

Until the last decade, exploitation within the Hamersley province focused on the high-grade martite-microplaty hematite, low P orebodies. Martite-goethite orebodies hosted within the Marra Mamba Iron Formation (Fig. 2) are typically of lower grade (55-65 wt % Fe) and have a low phosphorus content (<0.07 wt %); they were gradually developed as an alternative supply of low P iron ore (e.g., mining commenced at BHPB’s OB29 deposit in 1974; Fig. 1). As this ore type gained market acceptance other mines, including Marandoo (1994), Namundjili (1998), West Angeles (2002), and Mining Area C (2003) started production (Fig. 1). These ores are preferentially hosted in the upper Mount Newman Member of the Marra Mamba Iron Formation and are generally associated with synclinal folds. Annual production from these mines totals 60 Mt with the majority supplied from the West Angeles and Mining Area C mines.

Several other large Marra Mamba Iron Formation-hosted high-grade iron ore deposits are currently being developed. The Hope Downs deposit (Fig. 1; Hope North and Hope South) is located approximately 75 km northwest of Newman (Fig. 1). The deposit is hosted within the Mount Newman Member of the Marra Mamba Iron Formation with satellite deposits in the Brockman Iron Formation (Fig. 2). Mining is scheduled to start in 2008 with a production of 22 Mt/yr. The Chichester Range deposits (Fig. 1; Christmas Creek and Cloud Break) are located approximately 110 km northeast of Newman with the Cloud Break deposit scheduled to be mined in 2008 with an initial annual production of 45 Mt/yr. Interestingly, these deposits differ from other Marra Mamba deposits across the Hamersley province as they are hosted within the lower Nammuldi Member (Clout, 2005; Fig. 2).
FIG. 1. Location of selected high-grade iron ore deposits in the Hamersley province, Western Australia, described within this paper classified by generic model.
Channel iron deposits, including Mesa J, Yandicoogina, and Yandi (Fig. 1), consist of Miocene iron-rich detritus and provided around 41 percent (57–58 wt % Fe) of the total of 268 Mt of iron ore mined from the Hamersley province in 2006 (Clout and Simonson, 2005). These important deposits are not discussed within this paper but are comprehensively reviewed in Ramanaidou et al. (2003), MacPhail and Stone (2004), Heim et al. (2006), and Morris and Ramanaidou (2007).

Major BIF-Related Iron Deposits in the Hamersley Province

The Hamersley province hosts a number of significant iron ore deposits (Fig. 1) within the two most economically significant BIF units: the Marra Mamba and Brockman Iron Formations (Fig. 2). The following descriptions of the Mount Whaleback, Mount Tom Price, Paraburdoo, Hope Downs, Area C, and Chichester Range deposits (Fig. 1) are included in this review because of the significant amount of geological and geochemical data available on them and also because different end-member genetic models were proposed for their origin: namely supergene, supergene-modified syn-genetic, and supergene-modified hydrothermal.

Recent studies of the Mount Tom Price, Mount Whaleback, and Paraburdoo deposits (Barley et al., 1999; Ridley, 1999, Taylor et al., 2001; Cochrane, 2003; Dalstra, 2005; Thorne et al., 2004, 2005; Sepe, 2007) agree that burial metamorphism and the formation of a magnetite-chert–rich BIF occurred prior to the formation of high-grade iron ore. Smith et al. (1982) used prehnite, pumpellyite, epidote, and actinolite in mafic rocks of the Fortescue Group to define four metamorphic zones within the Hamersley province that increase from north to south: prehnite-pumpellyite, prehnite-pumpellyite-epidote, prehnite-pumpellyite-epidote-actinolite, and (prehnite)-epidote-actinolite zones. Peak metamorphic temperatures during the Ophthalmian orogeny (D2; 2.14 ± 0.03 Ga; Alibert and McCulloch, 1993) reached at least 200°C (Smith et al., 1982).

Mount Whaleback

The Mount Whaleback hematite deposit is located in a faulted outlier of Hamersley Group rocks south of the
Ophthalmian Range (Harmsworth et al., 1990; Brown et al., 2004), approximately 4 km west of Newman (Fig. 1). It is the single largest metalliferous deposit in Australia, originally containing approximately 1.8 Bt tons of high-grade (~65 wt % Fe) martite-microplaty hematite ore with low contaminants (e.g., ~2.5 wt % SiO₂, ~0.05 wt % P, and ~1.5 wt % Al₂O₃). Total Brockman Iron Formation mineral resources at Mount Whaleback and its satellite deposits in June 2007 totaled 1.601 Mt containing approximately 62 wt percent Fe, with low contaminants (0.09 wt % P, 5.5 wt % SiO₂, and 2.4 wt % Al₂O₃; BHP Billiton, 2007).

The Mount Whaleback deposit is structurally complex with at least five deformation events (i.e. D₁ to D₅) having influenced the orebody (Brown et al., 2004). The main structure of the area is characterized by east-west–trending folds related to the major Ophthalmian orogeny (D₂; ca. 2.3–2.2 Ga.) that dominates the southeastern Hamersley province. The D₂-related parasitic folds locally broaden the width of the metamorphic event (Tyler and Thorne, 1990). Importantly, phic conditions and is superimposed onto the earlier burial east part of the Ophthalmia fold belt. Formation of this cleavages is to the west, although local plunge of the synclines is to the west, although local plunge reversals occur as a result of later deformation (i.e., D₅). An axial plane cleavage is associated with D₂ folds in the south-east part of the Ophthalmia fold belt. Formation of this cleavage had taken place under low- and very low grade metamorphic conditions and is superimposed onto the earlier burial metamorphic event (Tyler and Thorne, 1990). Importantly, D₂-related parasitic folds locally broaden the width of the Dales Gorge and Joffre Members of the Brockman Iron Formation that host the high-grade ore (Fig. 3).

Two low-angle normal faults, the East Footwall and Central faults (D₂; ca. 2.2 Ga; Brown et al., 2004), that significantly affect the orebody (Brown et al., 2004). The main structure of the orebody is characterized by a few hundred meters of horizontal displacement and dip very gently to the south-southwest (Fig. 3). High-grade martite-microplaty hematite ore at Mount Whaleback is spatially associated with these normal faults. The distinct displacement on stratigraphic markers within the orebody suggests either that two or more phases of movement on these faults pre- and postdated ore formation or that fault movement was broadly synchronous with ore formation. The steep, south-dipping Mount Whaleback fault forms the northern boundary of the orebody where it juxtaposes Hamersley Group rocks against older Fortescue Group rocks to the north (Fig. 4). This east-northeast–trending fault can be traced for tens of kilometers, has a throw estimated to be between 500 and 900 m (Kneeshaw, 1975), and its formation has been related to crustal-scale extension during deposition of the Bresnahan Group (D₂; ca. 1.5–1.6 Ga.; Tyler et al., 1991).

Geologic field relationships constrain the age of high-grade ore formation. A northwest-trending dolerite dike set, which spatially controls the extent of the orebody and is hydrothermally altered to a talc-chlorite assemblage, predates the ore. The dike crosscut the Cheela Springs Basalt, a unit that has been dated at 2300 ± 15 Ma (SHRIMP Zircon U-Pb; Martin et al., 1998). The Mount McGrath Formation contains clasts of high-grade ore and is older than the June Hill Volcanics at 1843± 2 Ma (zircon 207Pb/206Pb; Pidgeon and Horwitz, 1991).

High-grade iron ore (Fig. 5B) is hosted predominantly in the Dales Gorge Member of the Brockman Iron Formation (Fig. 3), with lesser amounts in the underlying Colonial Chert Member of the Mount McRae Shale and the overlying Joffre Member of the Brockman Iron Formation (Fig. 3). Strati-graphic thicknesses of all units within the deposit vary considerably due to intense folding and mass changes associated with the transformation of BIF to high-grade iron ore (Webb et al., 2004). The high-grade ore is composed almost entirely of subhedral martite grains and aggregates set in a porous (up to 30 vol %), randomly oriented network of microplaty hematite (Fig. 5B; Harmsworth et al., 1990; Webb et al., 2003).

Until recently, a “supergene-metamorphic” model was widely cited for the genesis of the Mount Whaleback deposit (e.g., Morris, 1980, 1985, 1998; Morris et al., 1980; Harmsworth et al., 1990). Several authors (Powell et al., 1999; Brown, 2004; Webb et al., 2004) suggested that hydothermal fluids are responsible for the transformation of BIF to high-grade martite-microplaty ore (Table 1). Webb et al. (2004) studied intervals of the Mount McRae Shale, located at depth within the deposit (Fig. 3) and adjacent to high-grade ore, which consist of fine- to medium-grained dolomite cut by numerous chlorite and ferroan-dolomite and/or ankerite veins (Webb et al., 2004). Spatially associated with many of these veins are large crystals of fine-grained ferroan-dolomite and ankerite (Fig. 5A). These authors suggested that the silica-bearing phases were replaced by carbonate phases and the fluids causing these reactions were likely basinal brines or deeply circulating meteoric fluids.

Powell et al. (1999) conducted both oxygen isotope and fluid inclusion investigations on veins from several outcrops in the Newman area where hematite ± quartz veins, broadly of syn-D₂ timing, are locally surrounded by microplaty hematite ore. Fluid inclusions trapped in quartz from veins with centimeter-scale microplaty hematite margins show evidence of two fluids. The first fluid is a 270° to 380°C water-rich fluid most likely consisting of modified meteoric water, and the second is a 380° to 420°C, highly saline FeCl₂-H₂O ± NaCl ± CaCl₂ basinal brine (Table 1). The low δ¹⁸O values of the hematite ores (~–3‰) relative to BIF (~5‰) are caused by the equilibration of hematite with meteoric or basinal brines. Powell et al. (1999) suggested that the quartz-hematite veins formed during the Ophthalmian orogeny (D₂), and because the δ¹⁸O hematite values (~–1.5‰) of the veins approach those of the high-grade ore (~–3‰), the two were linked hydrodynamically, and that ore formation, at moderate temperature, was triggered during the Ophthalmian orogeny. Similarly Brown et al. (2004) studied veins comprised of massive, milky white quartz with minor hematite selvages and infill that formed around the Mount Whaleback deposit. Their structural mapping and analysis of vein patterns indicate that they likely formed synchronous with martite-microplaty hematite ore formation. Fluid inclusion measurements indicate that the veins formed at temperatures between 203° and 231°C (Table 1; pressure corrected) with fluid salinities ranging between 11 and 11 wt percent NaCl equiv at the time of vein formation (Table 1). Proton-induced X-ray emission analyses of these fluid inclusions indicate that the trapped fluid contained ~14,000 ppm Cl, ~1,600 ppm K,
that dominates the southeastern Hamersley province. The D2 metamorphic event (Tyler and Thorne, 1990). Importantly, age had taken place under low- and very low grade metamorphosed east part of the Ophthalmia fold belt. Formation of this cleavage is associated with D2 folds in the south.

The Mount Whaleback deposit is structurally complex with several main orebodies and related features. The main structure affected by the major Ophthalmian orogeny (D2; ca. 2.3–2.2 Ga.) and the second is a 380° 420°C, highly saline FeCl2-H2O ± Ni, Cu, Zn, As, Br, Bi, Sr, and Y (Table 1; Brown et al., 2004). Oliver (2007) used stable isotope and geochemical modeling to produce potential fluid reaction pathways that result in the formation of high-grade martite-microplaty hematite iron ores from BIF. The fluid model suggests that the predominant isotopic and geochemical composition of the ores was produced by reaction of BIF with voluminous, oxidized, probably Paleoproterozoic meteoric water at about 200°C (Table 1).

**Mount Tom Price**

The Mount Tom Price deposit is located near the eastern closure of the Mount Turner syncline (Fig. 1) that trends approximately east-west and appears to be the earliest fold in the area (D2). The original resource for this deposit, which is 7.5 km long, up to 1.6 km wide (avg 0.6 km), and has a maximum depth of 400 m below surface, was estimated at 900 Mt, grading 63.9 wt percent Fe and 0.03 wt percent P (Fig. 4; Harmsworth et al. 1990). As of 2003, the deposit contained a reserve of 199 Mt at 64.5 wt percent Fe and a measured and indicated resource of 100 Mt at 63.7 wt percent Fe, low P, high-grade ore. More than 90 percent of the ore at Mount Tom Price is located in the Dales Gorge Member, with smaller amounts in the Joffre Member (Fig. 4) where it is in faulted contact with the Dales Gorge Member (Barley et al., 1999). The complete thickness of the Dales Gorge Member is mineralized at the eastern end of the deposit (e.g. South East Prong deposit; Fig. 1), with only the upper part of this unit mineralized farther west (e.g. North deposit; Fig. 1). Several major structures control the form of the deposit. The Southern Batter Fault is a northwest-striking, southwest-dipping normal fault with a throw of up to 300 m (Fig. 4). In the center of the deposit, the fault splays into two, or locally three, closely spaced subparallel faults, each with a smaller normal offset than the main structure. The Box Cut fault (Fig. 4) strikes east-west, has a steep southerly dip, and produces displacement of up to 300 m. The fault juxtaposes high-grade hematite ore of the North East Prong against dolomite from the Paraburdoo Member to the north (Fig. 4). The South...
FIG. 4. Plan geology and cross sections of the Mount Tom Price deposit. Geology plan displays the distribution of high-grade ore types within the deposit and the major structural features. Cross sections through the North (A-B) and Southern Ridge (C-D) deposits display the spatial distribution of hypogene alteration and high-grade ore (modified after Taylor et al., 2001, and Thorne et al., 2007b).
East Prong fault (Fig. 4) is a northerly dipping reverse fault which forms the steep northern edge of the South East Prong deposit and has a maximum displacement of 120 m. A significant structural discontinuity separates the more complexly folded younger rocks of the Brockman Iron Formation from the gently synclinal older Marra Mamba Iron Formation at the eastern end of the deposit where there is a substantial loss of carbonate section underneath. A second phase of west-northwest–northwest folds (D3) oblique to D2 folds, affects much of the mine area (Harmsworth et al., 1990). These folds are complex and noncylindrical with their axial planes fanning about a west-northwest axis (Taylor et al., 2001). A set of northwestern-trending dolerite dikes extends through the Southern Ridge and Syncline deposits and is unaffected by D2- and D3-related folds. The dolerite dikes are altered to talc-chlorite assemblages by hydrothermal fluids (Taylor et al., 2001).

High-grade hematite ore consists of martite and microplaty hematite (Fig. 5D) and the ore preserves the meso- and microlayering of the primary banded iron formation layers (Fig. 4). The ore is porous and has low concentrations of SiO₂, Al₂O₃, P, Na₂O, and K₂O. Martite-microplaty hematite ore consists of randomly oriented fine-grained platy hematite and martite (Fig. 5D). Individual microplates (0.001–0.25 mm) are bladed or cigar shaped with sharp terminations (Taylor et al., 2001) that form overgrowths on subhedral martite grain margins (Ridley, 1999; Taylor et al., 2001; Thorne et al., 2004,
The intervening shale layers are pale pink clay seams that constitute the remaining impurities in the ore. Petrological and geochemical studies at Mount Tom Price have identified three hypogene alteration zones between BIF and high-grade iron ore: (1) distal magnetite-siderite-stilpnomelane, (2) intermediate hematite-magnetite-ankerite-talc-chlorite (Fig. 5C), and (3) proximal martite-microplaty hematite-magnetite-apatite alteration zones (Barley et al., 1999; Taylor et al., 2001; Thorne et al., 2004, 2005).

The most current ore deposit models (Barley et al., 1999; Taylor et al., 2001; Cochrane, 2003; Thorne et al., 2004, 2005, 2007b) involve a hydrothermal-supergene origin for the formation of the high-grade Mount Tom Price deposit. Taylor et al. (2001) presented a three-stage hydrothermal model for the formation of the Mount Tom Price high-grade iron deposit that is summarized below. The first, hypogene, stage of ore formation transformed magnetite-chert-rich BIF into a magnetite-siderite alteration assemblage. In this stage of alteration, low-temperature, highly saline bicarbonate-saturated fluids from the underlying carbonate-shale Wittenoom Formation migrated upward along fault zones into the lower part of the Brockman Iron Formation. During the second, deep meteoric stage of ore formation the magnetite-siderite assemblage was oxidized to a hematite-ankerite-microplaty hematite assemblage, with magnetite converted to martite. The fluid responsible was a low-temperature (~100°C), low-salinity, meteoric fluid. A second stage of gangue removal followed this oxidation stage and stripped all carbonate from both magnetite and hematite zones, leaving highly porous and permeable iron ore layers with a high apatite content interbedded with magnesium-rich shale layers. The final, purely supergene, stage of upgrading is indistinguishable from modern weathering but penetrated deep below the present surface. Magnesium silicates were converted to a kaolinitic residue, greatly thinning the shale layers, apatite was destroyed, and both calcium and phosphorus were leached from the ore.

Thorne et al. (2004, 2005) invoked a two-stage hypogene-supergene model for the formation of the North deposit (Fig. 3). Stage 1 involved early hypogene alteration by upward movement of hydrothermal, CaCl2-rich brines (110°–280°C) that likely originated during regional deformation of the Hamersley province. The brine exchanged with the underlying Wittenoom Formation as it moved into the large-scale fault zone toward the periphery of the deposit. Late-stage hydrothermal alteration involved the interaction of low-temperature, low-salinity (<110°C; Thorne et al., 2007a; Table 1) modified meteoric waters, and a 380°–420°C, highly saline FeCl3-H2O x NaCl x CaCl2 basinal brine (Brown et al., 2004: 203–211°C, 1.1–11 wt % NaCl equiv: ~14,000 ppm Cl, ~1,600 ppm K, ~650 ppm Ca, ~675 ppm Ti, and varying concentrations of Mn, Fe, Ni, Co, Zn, As, Br, Rh, Sc, and Y). Late-stage meteoric water (~200°–250°C) (Webb et al. 2004) involved oxygen isotope compositions of iron oxides in one of the major ore-control structures to demonstrate that intense hypogene fluid flow and fluid/rock reactions occurred within the core zone of the normal faults but decreased systematically away from the fault zone toward the periphery of the deposit. Late-stage hydrothermal alteration involved the interaction of low-temperature, low-salinity (<110°C; Thorne et al., 2007a; Table 1) meteoric water, with the hematite-ankerite-magnetite assemblage leaving a porous martite-microplaty hematite-apatite assemblage. Stage 2 supergene alteration in the Tertiary (Vasconcelos, 1998) resulted in the removal of residual ankerite.
and apatite and the weathering of the shale layers to kaolinite-rich clay.

Hagemann et al. (1999) conducted a fluid chemistry study on quartz vein and hematite samples from the Southern Bar- 
ter fault zone within the Southern Ridge deposit at Mount Tom Price. The quartz veins displayed δ18O values from 16.1 
to 18.3 per mil and hematite from –3.1 to 0.8 per mil. Fluids in 
equilibrium with quartz are calculated to range from –0.1 
to 4.1 per mil and interpreted to originate from meteoric 
water and/or basinal brines that exchanged variably with the 
wall rocks. Analyses of fluid inclusion waters from the South-
ern Ridge yield δD values from ~35 to ~90 per mil and con-
sist largely of two populations, one with light δD values from 
~53 to ~66 per mil and another with heavy values from ~34 to 
~43 per mil. Hagemann et al. (1999) interpreted the data as 
compatible with a mixture of basinal brines and meteoric wa-
ters that reacted with the sedimentary wall rocks. Fluid in-
clusion analyses of quartz-hematite veins suggest that the ma-
jority of aqueous fluids trapped in fluid inclusions are related 
to complex basinal brines.

Recent work has constrained the timing of both the hy-
drothermal and supergene alteration within the Mount Tom 
Price deposit. Rasmussen et al. (2007) studied U-Pb 
chronometers, xenotime, and monazite, from high-grade 
hematite ore from the Mount Tom Price. Both phosphate 
minerals occur as inclusions within the hematite ore and as 
coarser crystals intergrown with martite and microplaty 
hematite, indicating that the xenotime and monazite precipi-
tated during the transformation of BIF to high-grade ore and 
grew during multiple discrete events. Rasmussen et al. (2007) 
suggested that ore genesis may have commenced as early as 
2.15 Ga, with subsequent hydrothermal remobilization at 
2.05, 1.84, 1.67, 1.59, 1.54, 1.48, and 0.85 Ga. Vaszconcelos 
(1998) and Cochrane (2003) obtained ages from 40Ar/39Ar dat-
ing on Mn oxides, indicating that supergene alteration at the 
Mount Tom Price deposit spanned the last 80 m.y.

Paraburdoo

The Paraburdoo deposit is located at the southern margin 
of the Hamersley province, approximately 65 km south of 
Mount Tom Price (Fig. 1), and is part of a series of deposits 
along the Paraburdoo Range (66W to Channar deposits; Dal-
stra, 2005). In 1972 the premining reserve was 300 Mt at 63 
wt percent Fe, with low impurities (0.00% P 3.5 wt % SiO2, 
2.1 wt % Al2O3). By 1975 the total reserves and indicated 
resources had increased to 700 Mt of >60 wt percent Fe. As of 
2003, the deposit contained a reserve of 23 Mt at 64.5 wt per-
cent Fe and a measured and indicated resource of 35 Mt at 
63.6 wt percent Fe, low P, high-grade ore.

The main Paraburdoo iron deposit consists of two orebod-
ies: 4 West and 4 East (Fig. 6; Harnsworth et al., 1990), and 
high-grade iron ore is equally hosted within the Dales Gorge 
and Joffre Member of the Brockman Iron Formation. A 
minor amount of high-grade iron ore is also associated with 
flat-lying faults in the Yandicoogina Shale and the lowermost 
section of the Weelt Wolli Formation (Fig. 2). Structural re-
construction of the Paraburdoo deposit suggests that the flat 
fault zones that now underlie the deposit were steeply dip-
ping normal faults, prior to late tilting (Dalstra, 2005). Con-
glomerates containing pebbles of iron ore in the Mount Mc-
Graham Formation (Figs. 2, 6) indicate the first erosion of high-
grade iron ore that developed from Hamersley BIFs (Thorne 
and Seymour, 1991). Northwest-trending dikes (ca. 2008 Ma; 
SHRIMP 207Pb/206Pb baddeleyite; Müller et al., 2005) that 
limit the areal extent of iron orebodies at Paraburdoo (Fig. 6) 
intrude extensional faults that were active and hydrothermally 
disturbed during ore formation (Barley et al., 1999; Dalstra et 
al., 2005) but do not intrude the Mount McGrah Formation. 
The dikes were emplaced after the Panhandle folding (D1NC; 
Müller et al., 2005), and dikes, country rock, and ore were 
completely eroded before the deposition of the Mount McGrah 
Formation (Taylor et al., 2001). Thus, the formation of high-
grade iron ore took place during or shortly after emplacement 
of these ca. 2008 Ma dikes.

Most of the Paraburdoo deposit is in the Tertiary supgene 
alteration profile and exhibits a strong goethite overprint of 
the primary martite-microplaty hematite ore (Fig 5F). How-
ever, below the 4 East deposit and the depth of modern 
weathering, carbonate-altered BIF (50–55 wt % Fe; 
hematite-dolomite-chlorite-pyrite; Fig. 5E) occurs trans-
itional between BIF (~35 wt % Fe) and high-grade (>65 wt 
% Fe) iron ore (Thorne et al., 2007b). Sepe (2007) and 
Thorne et al. (2007b) proposed a three-stage hydrothermal-
supergene model for the 4 East deposit at Paraburdoo (Fig. 
6). The first hypogene stage involved hypogene alteration by 
upward movement along fault zones, of 170°C, basinal brines 
(20.9 wt % CaCl2 equiv with Ca > Na > K; Table 1) into the 
Dales Gorge and Joffre Members of the Brockman Iron For-
mation. The low Cl/Br and Na/Br ratios, high Ca concentra-
tions, and elevated base metal content of the hydrothermal 
fluids suggest that they were modified basinal brines sourced 
from evaporated seawater from which halite had precipitated 
and, prior to being expelled from the Wittenoom Formation, 
had caused dolomitization of limestone succession(s) (cf. 
Lüders et al., 2005; Thorne et al. 2007b). Fluid rock reactions 
transformed magnetite-chert-rich BIF into the intermediate 
hematite-dolomite-chlorite-pyrite assemblage. The second 
stage involved deeply circulating, low-temperature (~<100°C) 
meteoric water that interacted with the hematite-dolomite-
cholorite-pyrite assemblage, resulting in the formation of the 
proximal martite-microplaty hematite-apatite assemblage. 
Carbonate was leached from the rock and chlorite and pyrite 
were replaced by anhedral hematite. The high Cl/Br ratio of 
the fluid is interpreted to have formed from fluids that had in-
teracted with evaporites prior to interaction with the BIF for-
mations. Stage 3 supergene alteration during the Tertiary en-
tirely overprinted the proximal alteration zone converting 
shale layers to kaolinite-rich clay, destroying apatite, and 
leaching calcium and phosphorus from the ore.

Hope Downs

The Hope Downs deposit consists of the Hope North and 
Hope South orebodies and is located approximately 75 km 
northwest of Newman (Fig. 1) in Western Australia. The 
Hope Downs deposit consists of high-grade (>60 wt % Fe) 
martite-goethite ore that is hosted by the Mount Newman 
Member of the Marra Mamba Iron Formation. The Hope 
North deposit has proven reserves of 37 Mt at 61.6 wt per-
cent Fe, with low impurities (0.06 wt % P; 2.0 wt % SiO2, and 
1.5 wt % Al2O3). The Hope South deposit has proven reserves

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The Hope Downs deposits are located at the eastern end of the east-west–trending Weeli Wolli anticline (D2; Fig. 7), which plunges to the east and the west forming an elongated dome with minor plunge reversals along the axis and numerous subsidiary folds, particularly on the limbs. The central part of the Weeli Wolli anticline is characterized by flat-lying to gently dipping strata that abruptly turn into steeply dipping synclines plunging to the east at Hope Downs (Fig. 7). Upright concentric folds formed during D2 (Martin et al., 1998) and appear to be related to the D2 Ophthalmian major dome and basin structure of the Hamersley province. Superimposed on this structure are a number of
Fig. 7. Plan geology and cross section of the Hope Downs deposit, showing distribution of high-grade martite-goethite and detritals. A–B displays a northeast-southwest-trending section across the Weeli Wolli anticline (modified after Lascelles, 2006a). C–D is a north-south section across the Hope Downs North deposit displaying structural and stratigraphic controls on ore formation (Courtesy of Rio Tinto Exploration).
Deformed limbs in moderately narrow, linear to arcuate belts deformation produced pervasive east-west–trending, upright chevron-style minor folds that appear to be related to D3 deformation (Paquay and Ness, 1998).

Subhedral magnetite with grains averaging around 500 µm forms mesolayers between the chert and the magnetite layers. Dolomite, and calcite are typically abundant and commonly shows a texture of interlocking subhedral crystals. Iron-rich black shale layers and very rare magnetite. Siderite typically consists is typified by a coarser scale of layering than most of the other BIF of the Hamersley Group with mesolayers commonly between 2 to 10 cm and rarely up to 30 cm thick. Chert layers consist mainly of interlayered siderite and/or ankerite and chert mesolayers with minor stilpnomelane and carbonaceous amorphous silica was lost prior to lithification to form mesolayers (Fig. 5H). High-grade ore consists of hematite, brown and yellow goethite of earthy consistency, with minor manganese oxides and secondary quartz in near-surface rocks. The majority of the hematite is in the form of martite (Fig. 5G) but some cryptocrystalline to amorphous hematite (red ochre) near the surface is derived from dehydration of goethite and by direct precipitation from suspension during weathering. The ore in the deeper parts of the deposit consists mainly of highly porous to friable martite with minor interstitial goethite and approximately 15 vol percent of earthy goethite. Martite is abundant throughout the ore ranging from massive mesolayers of interlocking grains to thin microlayers and isolated euhedral grains.

Paquay and Ness (1998) proposed that the Hope Downs deposit formed as a result of intensive supergene alteration. Lascelles (2006a), in contrast, invokes a model for the origin of the host iron formation and the ore deposits, in which density currents transported reworked iron silicates and hydroxides in colloidal suspension onto an unstable sea floor. Hydrous amorphous silica was lost prior to lithification to form chert-free BIF. Supergene alteration of the chert-free BIF produced the present high-grade martite-goethite ore. Paquay (2001). South over north thrusts associated with this deformation preferentially developed on the stratigraphic contact between Mount Newman and West Angela Members and has an apparent movement of up to 200 m (Kepert, 2001). Associated with these thrusts are gently north-dipping splays that Kepert (2001) interpreted as low-angle extensional faults compatible with the D3 Riedel shear configuration. Subsequent D3 deformation produced gentle to open, east-west–trending symmetric folds with upright axial surfaces subparallel to D3 folds (Kepert, 2001).

Several key structures such as south-dipping thrust faults control the geometry of the orebody at the Area C deposit (Fig. 8). These are interpreted to have formed during D3 deformation and have later been affected by D4 deformation, resulting in the undulation of the main thrust and steepening of the deeper, low-angle fault (Bodycoat, 2007). South-dipping thrust faults (D3) developed along the contact between the Mount Newman Member and West Angela Shale (Fig. 8) and associated folds and thickened the main ore-hosting units in the Mount Newman Member (Fig. 8). Martite-goethite ore is widely developed in both the Marra Mamba and Brockman Iron Formations in the Mining Area C region (Fig. 9). The high-grade ore is spatially related to the Mesozoic–Tertiary paleosurface, indicating that it is most likely developed from near-surface to very deep (>200 m) supergene enrichment of BIF (Morris, 1985, Harnsworth et al., 1990).

In the Marra Mamba Iron Formation, high-grade ore is hosted in the upper Mount Newman and lower West Angela Members, although it can continue within the MacLeod Member closer to the surface (Kepert, 2001). Stratigraphically lower (i.e., Nammuldi Member; Fig. 2) high-grade orebodies are typically thin and very discontinuous along strike. Ore thicknesses vary between 20 to 80 m and are controlled by major structures.

**Chichester Range deposits**

The Chichester range deposits, Christmas Creek and Cloud Break, are located approximately 110 km north-east of Newman (Fig. 1) in Western Australia. The deposits are hosted within the Nammuldi Member at the base of the Marra Mamba Iron Formation (Figs. 2, 9). Production is due to commence in early 2008. The mineral resource for the Christmas Creek deposit (Fig. 9) is 1,415 Mt at 58.4 wt percent Fe with moderate levels of impurities (0.054 wt % P, 4.3 wt % SiO2, and 2.49 wt % Al2O3). The Cloud Break deposit has a resource of 816 Mt at 58.7 wt percent Fe with slightly lower impurities (0.056 wt % P, 4.1 wt % SiO2, and 2.4 wt % Al2O3). Folds of martite-microplaty ore consist of low Fe, massive-microplaty hematite with almost no goethite and 62 to 66 wt percent Fe2O3, 2 to 4 wt percent SiO2, 0.5 to 2.0 wt percent Al2O3, and 0.03 to 0.05 wt percent P.

Regionally, the Chichester range dips less than 5° to the south but at the Christmas Creek deposit the regional strike is offset by north-south– to northeast-southwest–trending faults and open folds with a south-southwesterly plunging fold axis. The latter are developed throughout the Chichester Range (Bagas, 2005) and in the Fortescue Group immediately to the north (Bagas et al., 2004; Bagas, 2005). The relationship of the folds in the Christmas Creek deposit to folding affecting the Hamersley province to the south (i.e., D2
to recumbent folds with intervening long, flat, relatively un-deformation produced pervasive east-west–trending, upright chevron-style minor folds that appear to be related to D 3 deformation.

Subhedral magnetite with grains averaging around 500 cmally shows a texture of interlocking subhedral crystals. Iron rich black shale layers and very rare magnetite. Siderite typi-
and chert mesolayers with minor stilpnomelane and carbon-

layers consist mainly of interlayered siderite and/or ankerite other BIF of the Hamersley Group with mesolayers com-

1998).

cline result in intense deformation of BIF (Paquay and Ness, 2005) and the northern margin of the Weeli Wolli anti-
to the south-southeast on the northern limbs of the main cen-

2001). Associated with these thrusts are gently north-dipping deformation preferentially developed on the stratigraphic

1-2001). South-dip-

resulting in the undulation of the main thrust and steepening

2001). South over north thrusts associated with this

subparallel to D2 folds (Kepert, 2001).
and D3) remains unclear. Folds may alternatively represent the waning phases of post-Fortescue Group gravity-driven tectonics (Kepert, 2001).

The Cloud Break and Christmas Creek deposits are located within the Nammuldi Member and consist of a widespread 5- to 20-m-thick supergene weathering blanket of near-surface martite-goethite ore (Fig. 5H) over a total strike length of over 80 km (Fig. 9) and dip extent of up to 4 km. High-grade ore crops out locally along the Chichester Range and dips at about 4º south under a thin (0–50 m) cover of Tertiary to Recent sediments (gravels and clays). High-grade ore formation is controlled by northeast-southwest–trending faults and folds and has been crosscut and exposed locally by erosion in creeks that follow the structures. There are also pods of fine- to medium-grained martite-microplaty hematite ore closely associated with northeast- to southwest-trending brittle shear zones. The microplaty hematite varies in shape and size from coarse (100–200 µm) euhedral crystals at the Christmas Creek and Cloud Break deposits, to very fine (10–60 µm) blades that are common at the Mount Nicholas deposit (Figs. 1, 5I). Disseminated fine-grained martite-microplaty hematite also occurs in the supergene ore, especially beneath the modern weathering zone.

Small synclines and faults have focused supergene fluids within the Cloud Break and Christmas Creek deposits and along the Chichester Range, resulting in the preferential development of supergene ore within these structures. The distribution of martite-microplaty hematite ore within the Chichester range is interpreted to be the result of an initial Proterozoic structurally controlled hypogene alteration within the Nammuldi Member of the Marra Mamba Iron Formation (Hannon et al., 2005). The earlier hydrothermal martite-microplaty ore was subsequently overprinted by supergene alteration.

**Characteristics of High-Grade Iron Ore**

The mineralogical characteristics of high-grade iron ore from deposits of hydrothermal, supergene or syngenetic origin are the result of interaction between fluids and the protolith BIF. In the last decade, the expansion of iron ore mining to deeper levels within existing deposits (e.g., Mount Tom Price, Paraburdoo, or Mount Whaleback), development of new mining districts (e.g., the Chichester Range), and the greater production from deposits within the Marra Mamba Iron Formation (e.g., Area C or Hope Downs), in combination with studies on the ore petrology, mineralogy, and geochemistry, have led to a better understanding of the processes that control the mineralogical characteristics.

**Mineralogy of martite-microplaty hematite ore**

Microplaty hematite (mpH) is an important constituent of both hard and soft high-grade iron ore (Figs. 5, 10). Hard and soft ores are an interlocking mass of randomly oriented plates of hematite and martite crystals, with somewhat greater porosity in the soft ore. High-grade microplaty hematite ore is best developed in the Brockman Iron Formation, although
Microplaty hematite in high-grade iron ore may have formed either in response to hypogene hydrothermal processes (Barley et al., 1999; Taylor et al., 2001; Webb et al., 2002; Thorne et al., 2004, 2005) or as part of weathering, i.e., the dehydration of goethite in the upper hardcap of deposits (Harnsworth et al., 1990). The size of microplaty hematite is variable and has been defined differently by many authors: 10 to 200 µm (Muir, 1985) and 20 to 250 µm or larger (Taylor et al., 2001) in high-grade ore at Mount Tom Price; 10 to 60 µm in hematite-ankerite magnetite alteration assemblage at the North Deposit, Mount Tom Price (Thorne et al., 2004); 100 to 200 µm as euhedral crystals at Christmas Creek and Cloud Break; and 10 to 60 µm as straight-wall blades at Mount Nicholas (Chichester Range; Clout, 2005).

Microplaty hematite is well described by Taylor et al. (2001, p. 547) as "individual microplates that are bladed with 'blunt' terminations, or cigar shaped with sharp terminations." Clout (2005) described microplaty hematite as coarse, euhedral crystals at the Christmas Creek and Cloud Break deposits and as very fine, straight-wall blades at the Mount Nicholas deposit (Fig. 5).

At the Mount Tom Price and Paraburadoo deposits detailed fluid chemistry studies on carbonate and quartz that form coeval with microplaty hematite have shown that the microplaty hematite was formed from hydrothermal fluids (Taylor et al., 2001; Thorne et al., 2004, 2005, 2007a, b). Thorne et al. (2004, 2005, 2007a) also described platy magnetite (100–300 µm) and hematite as replacement of platy magnetite, in ore zones within the Mount Tom Price and Channar deposits. Platy magnetite is locally abundant and forms from the replacement of siderite. This platy form of hematite is also classified as microplaty hematite.

Several other textural forms of hematite are significant within martite-microplaty hematite high-grade ore. Thorne et al. (2007a) noted that there is significant variation in the proportion of textural forms of hematite (e.g., anhedral, platy, bladed, microplaty) at various deposits within the Pilbara province (Fig. 10). The hard martite-microplaty hematite ores at Mount Tom Price are generally rich in anhedral, fine, mounded hematite, most likely derived by continuous compaction of the softer and more porous microplaty hematite (Taylor et al., 2001). Clout and Simonson (2005) described 50-nm-thick layers of interlocking mosaic textured hematite from the more intensively mineralized zones at Mount Tom Price. Thorne et al. (2007a) noted a higher proportion (up to 90 vol %) of anhedral hematite from the 4E and 2 Lens deposits at Paraburadoo, where iron ore is typically soft and friable and lacks the characteristic interlocking textures observed at Mount Tom Price. Thorne et al. (2007a) and Sepe (2007) suggested that the relative proportions of anhedral hematite and microplaty hematite and their ability to receive and/or martite within the carbonate-altered BIF may be a function of temperature, fluid flux, and chemistry of the hypogene fluids.

Clout (2005) described nanometer-sized plates of hematite, referred to as "Texture X" from high-grade ore from the Christmas Creek deposit. Texture X samples have a very high (65–68 wt %) iron content and are characteristic of many high-grade microplaty hematite ores. The nanometer-scale needles, tubes, and hematite plates mostly occur within the more porous martite-microplaty hematite ore samples and well below the depth of weathering or dehydration in the Christmas Creek and Cloud Break deposits. Texture X also occurs in dehydrated zones that form in perched water tables in the martite-goethite ore types (Clout, 2002).

The range of sizes of hematite within high-grade ore varies from the nanometer scale (60–600 nm), to microns (1–3 µm) and hundreds of microns (100–200 µm), which Clout (2005) interpreted to represent various stages of nucleation, crystallization, and progressive growth of hematite. At present, there is no clear definition of what constitutes microplaty hematite. Microplaty hematite is present within many high-grade iron deposits and in variable amounts, sizes, and textural forms but alone cannot be used as diagnostic mineral for high-grade ore formed by hydrothermal processes. Of equal importance are anhedral hematite and Texture X, which are formed by the same hydrothermal processes. The common usage of the term, both within the Hamersley province and worldwide (Gutzmer et al., 2006), reduces the ability to differentiate the ore types and geochemical properties of high-grade ore and ultimately the processes that lead to their formation. The ability of microplaty hematite to form interlocking networks producing a hard, porous ore is economically important, as the ore is sold at a premium. The P-T-X conditions during which microplaty hematite forms, therefore, become increasingly important when exploring for new orebodies.

**Phosphorus distribution**

There are both significant inter- and intradeposit variations of phosphorus content within the iron ore deposits of the Hamersley province. Phosphorus (P) is an important contaminant in steelmaking and the majority of high-grade ore production from the Hamersley province averages less than 0.08 wt percent P (Dukino et al., 2000). At present the bulk of production is a blend of low P (i.e., <0.07 wt % P) martite-microplaty hematite (e.g., Mount Whaleback, Mount Tom Price) and/or martite-goethite ores (e.g., Mining Area C) with the more common high P (i.e., >0.07 wt %) martite-goethite ores (e.g., Brockman B34 and parts of OB18). The low P martite-microplaty hematite ore is the major component of the blend, however resources of low P martite-microplaty ore are now becoming rapidly depleted (e.g. Mount Tom Price; Dukino et al., 2000).

The P concentrations in the ores reflect the original amounts of apatite in the BIF (Muir, 1985). Morris (1985) was able to demonstrate a relationship between phosphorus distribution and layering, where apatite that is interpreted to be authigenic forms random, semicontinuous layers and scattered, rounded grains ~20 µm in diameter. The Marra Mamba Iron Formation has inherently lower apatite content (<0.04 wt %) than the Brockman Iron Formation (~0.10 wt %; Morris, 1985). Supergene altered martite-goethite ores hosted by the Brockman Iron Formation have a high P content (>0.07 wt % P), whereas supergene altered hydrothermal martite-microplaty hematite ores in the Brockman Iron Formation...
Formation are typically marked by a lower P content (<0.07 wt % P).

Within the Mount Tom Price deposit three generations of phosphorus-bearing minerals have been identified (Cochrane, 2003): primary crystalline apatite layers, secondary hydrothermal apatite, as well as complex phosphate minerals and P-rich hydroxides. The primary apatite layers are similar to those described by Morris (1985). Hydrothermal apatite is included in talc-apatite veins that crosscut the hydrothermal martite-microplaty-apatite ore; it is preserved below the depth of modern supergene alteration. Fluid inclusion studies show that the veins were formed from Fe- and Si-rich, low-salinity (NaCl) fluids that range from 220° to 280°C (Cochrane, 2003). Apatite was precipitated along geochemical barriers, in particular shale layers and dolerite dikes. Minor quantities of P are within matrix minerals such as magnetite, hematite, monazite, and xenotime (Rasmussen et al., 2007). Supergene weathering of the martite-microplaty hematite-apatite ore resulted in P being leached by descending supergene fluids and reprecipitated at lithologic contacts or within synclinal structures.

Apatite is readily mobilized under supergene conditions, dissolving slowly in the pH range of 7 to 8 and more rapidly under acidic conditions (Clout, 2005). Hence, most of the original apatite in the BIF would be dissolved by ground water during the replacement of silicate and carbonate minerals with goethite in the BIF, releasing (PO₄)³⁻ ions into the water during the replacement of silicate and carbonate minerals. The primary apatite layers are dissolved slowly in the pH range of 7 to 8 and more rapidly under acidic conditions (Clout, 2005). Hence, most of the original apatite in the BIF would be dissolved by ground water during the replacement of silicate and carbonate minerals with goethite in the BIF, releasing (PO₄)³⁻ ions into the ground water. Phosphorus becomes associated with goethite through surface adsorption of ions or complexes onto the growing surfaces of gel-like precursors and is incorporated into the goethite structure (Dukino et al., 2000). Goethite can also be introduced into the iron ores by later ground-water movement as high P colloform material along joints.

Dehydration of goethite preserves the crystal outline of the earlier goethite. Dehydrated zones generally form immediately above secondary goethite or transported away in solution once it reached the water table. Beneath the hardcap is the hydrated zone where most microcracks and cavities are filled with colloform secondary goethite. Hematite is partly replaced by goethite especially along joint planes. The ore is denser with low-porosity and higher goethite content compared with the primary ore, especially for the microplaty hematite ores. Any yellow ochreous goethite is typically partly or completely replaced by secondary brown goethite. The hydrated zone in microplaty hematite deposits is typically thin (<10 m vertically) when compared to martite-goethite deposits (10–25 m vertically).

Within any given hydrated zone there can be one or more stacked near-horizontal dehydrated zones, which are best developed 8 to 30 m below surface. In the dehydrated zone colloform secondary goethite and primary brown or ochreous goethite are largely dehydrated to hematite. Hematite formed from dehydration preserves the crystal outline of the earlier goethite. Dehydrated zones generally form immediately above shale layers or faults. Circulation of ground water in the upper portion of a deposit may have been influenced by aquicludes such as shale layers. Iron-rich ground water is interpreted to have ponded above near-horizontal aquicludes, encouraging abundant secondary goethite to deposit in localized zones and subsequently, partly or completely, dehydrated to hematite.
thus causing the density of the host ore to increase. The dehydrated zone may also contain localized high levels of manganese oxides (e.g., pyrohite and minor cryptomelane).

The primary ore zone often represents the thickest (up to 400 m) part of the ore profile in microplaty hematite deposits and is dominated by porous martite-microplaty hematite ore. The boundary between the hydrated zone and primary ore in martite-goethite deposits is more subtle but usually coincides with the absence of goethite infill textures and a marked increase in ochreous goethite in defined stratigraphic units (e.g., zone C in the Newman Member).

The leached zone is interpreted to form below the interpreted paleowater table, where yellow goethite of earthy consistency predominates over brown goethite, especially for most martite-goethite deposits but also many microplaty hematite deposits (Fig. 11). Immediately above the water table, there is a region where ochreous goethite appears to have replaced secondary and primary brown goethite. In martite-goethite ores, the presence of ochreous goethite is associated with a marked increase in the amount of kaolinite, traces of pyrite and friable residual leached carbonates and silicates from the BIF host. The ochreous goethite-brown goethite boundary is often located at the current water table; this is not always the case and may also be located significantly above it (e.g., at the Marandoo deposit; Fig. 1).

Genetic Models for BIF-Related Hamersley Iron Ore Deposits

This section provides the most current genetic models for the formation of high-grade BIF-hosted iron ore deposits in the Hamersley province. In order to discuss these models the six deposits described above are subdivided into two categories: supergene (Fig. 1; Area C, Hope Downs, and Chichester Range); and supergene-modified hydrothermal ores (Fig. 1; Mount Whaleback, Mount Tom Price, Paraburdoo).

Supergene model

There is a general consensus that the martite-goethite ores of the Hamersley province formed as a result of supergene enrichment of BIF (Paguay and Ness, 1998; Kepert, 2001; Taylor et al., 2001; Clout, 2002, 2005; Dalstra and Guedes, 2004; Thorne et al., 2004; Bodycoat, 2007). In the supergene model, high-grade iron ore is interpreted to have been formed by ground water that replaced silicate and carbonate minerals in the BIF with goethite (Morris, 1985). Despite stratigraphic thinning of up to ~35 percent and multiple phases of postore leaching, all levels of the primary BIF layering are preserved (Morris, 1985). Protolith textures are represented by martite (=primary hematite) and goethite pseudomorphs after chert, carbonates, and silicates. They are typically associated with the paleosurface but some extend to >250 m in depth (Morris, 1985). All three martite-goethite deposits considered in this study, Hope Downs (Mount Newman Member; Fig. 7), Area C (Mount Newman Member; Fig. 8), and Cloud Break (Nammuldi Member; Fig. 9) are located within the Marra Mamba Iron Formation, although the same supergene processes are thought to have formed martite-goethite deposits in the Brockman Iron Formation (e.g., sec. 6, 7 at Mt Tom Price; Fig. 5; Taylor et al., 2001). The major controls on the localization of the martite-goethite deposits are structure (e.g., faults, thrust, and folds) and descending supergene fluids (Hodkiewicz et al., 2005).

At all three deposits the formation of supergene martite-goethite deposits is structurally controlled by thrust fault development related to D2 deformation (Kneeshaw and Kepert, 2002). Deformation by D3 has resulted in rotation of fault surfaces at Area C (Fig. 8) such as the reorientation of the main thrust from a south-to-north-dipping surface at the C deposit (Fig. 8; Bodycoat, 2007). The faults act as fluid pathways for supergene fluid flow into the BIF and allow supergene fluids to penetrate to depths of up to 250 m (Hodkiewicz et al., 2005). Where faults and thrusts allowed supergene fluid flow into tightly to overturned synclines, the synclines tend to contain iron ore with higher iron grades than flatlying or gently dipping strata. The subvertical orientation of iron grade within faulted synclines is controlled by the orientation of the hanging-wall and footwall contacts with the hanging-wall contact of the thrusts containing higher Fe grades (Hodkiewicz et al., 2005). The interplay of thrusts, faults, and folds has resulted in thrust stacks of ore-forming horizons, and hence an increased volume of mineralized stratigraphy at the orebodies.

The structural controls on the location of the martite-goethite deposits are also responsible for continued modification of the primary supergene martite-goethite ore. Synclinal structures focus greater volumes of ground water, resulting in the leaching of goethite from the martite-goethite orebody. Multiple phases of goethite precipitation and cementation can result in a less porous and denser ore of higher Fe grade. This secondary goethite commonly dehydrates to hematite (Clout, 2002).

The formation of the Area C, Hope Downs, and Cloud Break deposits show similar stratigraphic and structural controls. Textural features of ore and gangue mineralogy, such as replacement of chert layers by iron hydroxides, goethite pseudomorphs after hematite and magnetite crystals as well as the distribution of iron-forming minerals such as hematite and goethite within each deposit, supports a supergene model for the formation of the deposits (Morris, 1985). In sharp contrast to this genetic model, a supergene-modified syngenetic model has been proposed for the Hope Downs deposit (Lascelles, 2006a). This model proposes that during diagenesis the hydrous amorphous silica was lost prior to lithification to form chert-free BIF. Weathering of the chert-free BIF produced the high-grade hematite ore that is exposed today. Locally, Lascelles (2006a) documented that high-grade ore lacks any evidence of the prior presence of chert layers or the replacement of chert layers by carbonate. Synsedimentary chert-free BIF is also identified at Mount Gibson in the Yilgarn province (Lascelles, 2006b). The model is supported by the observations that the high-grade ore lacks chert layers and the apparent lack of evidence for hydrothermal alteration. The model proposes that the formation of the chert-free BIF ore may have controlled the structure of the deposit rather than the structure controlling the deposit location (cf. Fig. 11; Lascelles, 2006a).

Supergene-modified hydrothermal model

Geologic evidence from the high-grade iron ore deposits of Mount Whaleback, Mount Tom Price, and Paraburdoo (Fig.
1) all support a hydrothermal process of transformation of BIF (~30–35 wt % Fe) to high-grade martite-microplaty hematite ore (>63 wt % Fe). The deposits exhibit similar stratigraphic and structural controls, ore mineralogy and textures, and hypogene alteration assemblages that define hydrothermal alteration zones in BIF and high-grade martite-microplaty ore. The interpretations provided here are based predominantly on detailed observations at the Mount Tom Price and Paraburdoo deposits. Components of the model proposed for the Mount Whaleback are speculative and discussed later. Two aspects of the supergene-modified hydrothermal model need to be elaborated on: timing of ore formation, and structural controls of ore formation.

1. Mafic sills dated ca. 2208 ± 10 Ma (SHRIMP 207Pb/206Pb baddeleyite; Müller et al., 2005) that were folded during the Ophthalmian orogeny and then cut by the unconformity at the base of the Lower Wyloo Group, constrain the Ophthalmian orogeny to the period between ca. 2208 and 2031 Ma, before Lower Wyloo Group extension, sedimentation, and flood-basalt volcanism. The maximum age of ore formation is constrained by the ca. 2008 ± 16 Ma age (SHRIMP 207Pb/206Pb baddeleyite; Müller et al., 2005) of the northwest-trending dolerite dike set, which predates the ore and limits the areal extent of iron orebodies at Paraburdoo and Tom Price (Figs. 4, 6). The dikes intruded extensional faults active during ore formation, and were hydrothermally altered during ore formation (Barley et al., 1999; Taylor et al., 2001; Dalstra, 2005) but do not intrude the 1843 ± 2 Ma (zircon 207Pb/206Pb; Pidgeon and Horwitz, 1991), which postdates the formation of high-grade ore. High-grade martite-microplaty hematite ore is likely to have formed during a ca. 2050 to 2000 Ma period of continental extension (Müller et al., 2005) that postdated the Ophthalmian orogeny rather than the 2400 to 2500 Ma timing proposed by Powell et al. (1999).

2. The three major martite-microplaty hematite deposits of the Hamersley province are spatially associated with faults (Fig. 12) that offset stratigraphy on the order of hundreds of meters (Taylor et al., 2001; Dalstra, 2005). At the time of ore formation, the Mount Whaleback (Central and Eastern Footwall faults; Fig. 3), the Mount Tom Price (Southern Batter fault; Fig. 4) and Paraburdoo (e.g. 4E Basal fault; Fig. 6) were all spatially associated with northwest- to west-trending normal fault systems, which were active during a basin-wide extensional event at the end of the deposition of the Lower Wyloo Group (Fig. 2; Taylor et al., 2001; Dalstra, 2005). At Mount Tom Price, fold hinges, fault splays, impervious shales, and dolerites which act as aquitards, control the fluid flow within the orebody (Thorne et al., 2005). The normal faults and associated splays that are associated with the orebodies
provide the critical pathways that allowed the movement of basinal brines and/or deep circulating meteoric water into the BIF and the transformation to high-grade martite-microplaty hematite ore (Hagemann et al., 1999; Taylor et al., 2001; Thorne et al., 2004, 2005, 2007a, b; Clout and Simmonson, 2005) suggested that major normal faults like the Southern Batter and the Mount Whaleback faults could represent back thumbs in the hanging wall of major regional thrusts, with ore fluids being preferentially focused through the back thumbs since they are most likely to form more dilatant zones.

Two successive stages of Paleoproterozoic hydrothermal fluid-induced alteration and one Tertiary supergene stage can be distinguished in the transformation from BIF to high-grade martite-microplaty ore, with each stage being the necessary precursor to the next (Fig. 12; Table 1; Barley et al., 1999; Hagemann et al., 1999; Ridley, 1999; Taylor et al., 2001; Cochrane, 2003; Thorne et al., 2004, 2005, Webb et al., 2004, Oliver, 2007; Thorne et al., 2007b).

Stage 1a: The initial hypogene stage is presently only observed at Mount Tom Price. Low-temperature (107°C–114°C), highly saline (25.5 wt % NaCl-CaCl₂ equiv) brines from the underlying Wittenoom Formation moved upward along the Southern Batter fault (Fig. 12) and were focused into silica-rich rocks of the Dales Gorge Member by the shafts of the underlying Mount McRae Shale Formation and overlying Whaleback Shale Formation, which acted as aquitards. Within the Dales Gorge Member, hypogene fluids migrated laterally within large-scale folds with permeability controlled by shale layers and the northwest-trending dolerite dike sets. The mafic dikes were altered to chlorite-talc with accessory leucoxene and pyrite. Magnetite-chert–rich BIF was transformed laterally and vertically into magnetite-siderite-stilpnomelane assemblages. In the North deposit at Mount Tom Price the siderite–iron silicate veins (V₂) crosscut and parallel BIF layers. The BIF wall rock adjacent to V₂ veins (cf. Thorne et al., 2004) is locally brecciated with clasts of magnetite-rich mesolayers set in a matrix of bladed magnetite, siderite, and iron silicates.

Stage 1b: The second hypogene stage is observed at all three deposits (Fig. 12). Basinal brines (21–25% wt % CaCl₂ equiv; Ca > Na > K) were focused along normal faults (i.e. Southern Batter fault, Mount Tom Price; 4E fault, Paraburdoo; Central fault, Mount Whaleback) into the host BIF formation. At the Paraburdoo deposit, the interaction of 150° to 190°C basinal brines with the BIF transformed BIF into hypogene hematite-dolomite-chlorite-pyrite assemblages. At Mount Tom Price the interaction of 220° to 280°C basinal brines with the hypogene magnetite-siderite-stilpnomelane alteration assemblage of stage 1a resulted in the formation of a distinct hypogene hematite-ankerite ± magnetite alteration assemblage, leaving only remnants of stage 1a. Simultaneously, within the core of the deposit, the initial downflow and rising of meteoric water would likely have started. Microplaty hematite crystallized as both individual blades and dense clusters that form overgrowths on magnetite and as individual plates within ankerite crystals. Secondary apatite is present as inclusions within magnetite and microplaty hematite and as anhedral crystals within ankerite crystals. Fault movement formed the matrix-supported wall-rock breccias that consist of angular and rotated clasts of altered wall rock within an ankerite-microplaty hematite matrix.

Stage 2: This stage involved deep circulating, low-temperature (~110°C; Table 1), Na-rich meteoric water. The descending meteoric water interacted with the carbonate-altered BIF to produce a martite-microplaty hematite-apatite ore (Fig. 12). The meteoric water was likely formed from the interaction with evaporite successions prior to interaction with the BIF formations.

Stage 3: The supergene stage, is the third and final stage in the transformation of BIF to high-grade ore (Fig. 12). Shallow supergene fluid interacted with the hydrothermal, martite-microplaty hematite-apatite alteration assemblage to form highly porous, high-grade martite-microplaty hematite ore. Supergene alteration is likely to have begun in the Cretaceous and occurred for at least 50 m.y. and is ongoing today (Vasconcelos, 1998; Cochrane, 2003). Weathering of shale layers resulted in considerable volume reduction and the formation of kaolinite-rich clay with elevated titanium contents. These clay minerals constitute the main impurity in the orebody.

Discussion

Genetic models for the formation of both supergene and supergene-modified hydrothermal deposits share similar structural controls, such as faults, thrusts, and synclines. The structures act as traps for either hydrothermal and/or supergene fluids that transformed BIF (~35 wt % Fe) to high-grade iron ore (56–69 wt % Fe). Hydrothermal alteration at Mount Tom Price and Paraburdoo form a well-defined vertical and lateral transition between BIF and high-grade ore. Detailed geochemistry, including microthermometry, laser-ablation of fluid inclusions, ion-chromatography of fluid inclusions trapped in carbonates, quartz, magnetite, and hematite, and C and O isotope composition of quartz, carbonates, and oxides confirm (Hagemann et al., 1999; Cochrane, 2003; Thorne et al., 2004, 2005, 2007a, b) that basinal brines and/or meteoric waters were responsible for the formation of carbonate-altered BIF and, ultimately, the formation of high-grade martite-microplaty ore. Despite these significant advances, several key features of the supergene-modified hydrothermal model remain poorly understood.

First, the relative volume of basinal brines and/or meteoric waters responsible for the crystallization of microplaty hematite within the deposits remains unknown. Thorne et al. (2004) described that at the North deposit (Fig. 4), basinal brines (24 wt % NaCl equiv) interacted with the magnetite-siderite-stilpnomelane assemblage, forming the hematite-ankerite-magnetite alteration assemblage, and the crystallization of microplaty hematite (Fig. 5C). In this reaction, iron liberated from siderite and iron silicates forms microplaty hematite. Taylor et al. (2001) proposed that at the Southern Ridge deposit (Fig. 4), deep circulation of low-salinity meteoric waters oxidized the siderite to microplaty hematite and magnetite to martite. Fluid inclusions in quartz veins from the Southern Batter fault in the Southern Ridge deposit (Hagemann et al., 1999) show evidence of mixing of basinal fluids and meteoric waters. Within the core of the Southern Ridge deposit, the Southern Batter fault and associated spals...
facilitate the downstream of meteoric water. Fluid mixing between descending meteoric water and ascending basinal brines caused the intermediate martite-ankerite-microplaty hematite alteration assemblage. At the North deposit, located at the periphery of the mineralizing system, ascending basinal brines were the dominant fluid source, forming the intermediate hematite-ankerite-magnetite alteration assemblage. The lack of detailed geochemistry and microthermometry data on the Mount Whaleback deposit makes it difficult to confirm the role of basinal brines and/or meteoric waters in the transformation of BIF to high-grade ore. Importantly, the formation of high-grade martite-microplaty hematite ore was dependent on both a period of silica loss and a period of oxidation during or soon after extensional deformation (Webb et al., 2003). Potentially, the influx of warm, silica-saturated fluids from underlying sedimentary successions into extensional faults triggered silica loss in the BIF by carbonate replacement of the chest layers. Focusing of oxidized, surface-derived fluids by faults, into the carbonate-rich BIF formed the high-grade martite-microplaty hematite orebody (Webb et al., 2004).

The source of iron and the exact mechanism for the crystallization of the different textural forms of hematite, such as anhedral, microplaty, or bladed, is not well understood. Quantitative LA-ICPMS microanalysis of primary fluid inclusions from dolomite in dolomite-chlorite veins from the 4E deposit, Paraburdoo (Thorne et al., 2007b) shows no evidence of significant iron enrichment within the fluids. The formation of hematite took place from the dissolution of the primary magnetite within the BIF and precipitation of microplaty anhedral hematite within a dolomite-chlorite matrix. Textural evidence showing microplaty hematite replacing martite (Webb et al., 2004) and iron silicates (Thorne et al., 2004) also indicate that the majority of iron is likely to be sourced locally from within the iron formation.

Textural and petrological data from hydrothermal alteration zones (Thorne et al., 2004) at the North deposit, Mount Tom Price, show that, where siderite is replaced by ankerite, iron is liberated and is precipitated as microplaty hematite. This reaction results in the formation of an interlocking network of microplaty hematite and martite crystals, resulting in a hard martite-microplaty ore. At Paraburdoo, the magnetite-siderite-stilpnomelane assemblage is not observed (Thorne et al., 2007a) and hematite is crystallized within dolomite as anhedral and microplaty hematite (Fig. 5E) but with limited interlocking of the hematite microplates resulting in a more friable and softer ore (Fig. 5F). The paragenetic sequence of alteration mineralogy and high-grade ore at Mount Tom Price and Paraburdoo suggest that in order to form hard, porous ores, the formation of an early magnetite-siderite-stilpnomelane alteration assemblage is significant.

Rivers (1998) conducted a detailed investigation into the petrology and geochemistry on dolomite from the Wittenoom Formation and suggested that dolomites were likely formed during diagenesis. Thorne et al. (2007b) suggested that the basinal brines, prior to being expelled from the Wittenoom Formation, were magnesium-rich brines, formed from evaporated seawater that had interacted with carbonate successions forming dolomite and producing calcium enrichment of the basinal brines. As the basinal brines were not responsible for the dolomitization of the Wittenoom Formation it is likely that the modification of the basinal brines occurred within sedimentary successions distal from the high-grade martite-microplaty hematite deposits prior to their movement into the Wittenoom Formation.

The existence of a rift-basin to the south of the Hamersley province at the time of ore formation provides the most likely heat and fluid source for hydrothermal fluids (Cawood and Tyler, 2004). Although not preserved within the stratigraphic record in the Ashburton basin, widespread quartz pseudomorphs after evaporitic minerals are interbedded with stromatolites in 2.20 Ga sedimentary rocks in the Yerrida rift basin of Western Australia (El Tabakh et al., 1999). The Yerrida basin, together with other Early Proterozoic basins of Western Australia, is inferred to have developed in a rifted passive margin between the Pilbara and Yilgarn cratons where extensive hydrothermal and volcanic activity took place (Pirajno et al., 1998). Hydrothermal fluids are postulated to have reacted with the sedimentary rocks and produced epigenetic lead carbonate deposits (Pirajno et al., 1998). The Yerrida basin is possibly a heat and fluid source for the basinal brines that are critical in the formation of the high-grade martite-microplaty hematite deposits. Xenotime and monazite in high-grade hematite orebodies from the Mount Tom Price deposit have recorded repeated episodes of hydrothermal fluid flow that commenced at 2.15 Ga and continued intermittently for more than one billion years (Rasmussen et al., 2007). Hydrothermal fluid flow responsible for the formation of high-grade martite-microplaty hematite ore may be a direct result of basin development and repeated tectonic activity to the south of the Hamersley province, in particular, during a ca. 2050 to 2000 Ma period of continental extension (Muller et al., 2005). Morris (2002) argued that there were not sufficient hydrothermal fluids within the Hamersley province to form the Mount Tom Price deposit, neither the larger Mount Whaleback deposit. Oliver (2007) calculated the fluid flux required for the formation of the Mount Whaleback deposit by meteoric waters. Using the silica solubility data of Fournier and Potter (1982), and assuming that silica was stripped from at least a 1-km path length (the open cut at Mount Whaleback is 5 km long), they calculated that the amount of silica dissolved from the deposits at 200°C and 1 kbar requires a fluid flux on the order of $6 \times 10^7$ moles H$_2$O cm$^{-2}$. A large rift basin, as suggested above, may have provided this necessary large volume of hydrothermal fluids that were subsequently focused through structures along the southern margin of the Pilbara craton during deformation in the bordering Capricorn orogen.

Taylor et al. (2001) and Muller et al. (2005) provided both structural and geochronological data that suggest that iron ore formation took place at approximately 2000 Ma, several hundreds of millions of years after the Ophthalmian orogeny (D$_2$). Evidence from quartz and quartz-hematite veins surrounding the Mount Whaleback deposit suggests that the peak of hydrothermal activity in the district was apparently synchronous with the main phase of microplaty hematite ore formation during the last stages of the Ophthalmian orogeny and the first stages of post-Ophthalmian extension. If the timing of ore formation proposed for the Mount Tom Price and Paraburdoo deposits is indeed similar to that of the Mount
Whitney deposit, then the quartz and quartz-hematite veins surrounding the Mount Whaleback deposit likely record regional metamorphism and fluid flow during the Ophthalmian orogeny and not the hypogene fluids and/or meteoric waters responsible for the formation of high-grade martite-microplaty hematite ore.

Dalstra (2005) and Kneeshaw and Kepe (2002) provided differing views on the orientation of faults within the Mount Whaleback deposit during the formation of the high-grade orebodies. The model proposed in this paper for the formation of the Mount Whaleback deposit is based on structural reconstructions by Dalstra et al. (2002) that attributed the subhorizontal orientation of the Central and East Footwall faults as the possible result of tilting associated with rotation on younger faults (e.g., Mount Whaleback fault system: Figs. 5, 12; Taylor et al., 2001). The structural reconstruction is based on published and unpublished data from the Mount Whaleback deposit together with structural data from the Giles Mina and Lido deposits, 50 and 10 km west of Mount Whaleback, respectively. Alternatively, Kneeshaw and Kepe (2002) argued for an originally shallow dipping orientation for the Central and East Footwall faults. The posture rotation at the Mount Whaleback deposit could explain why there are no hypogene carbonate alteration zones preserved at the deposit (Fig. 12). A consequence of the posture rotation is that the deepest part of the orebody prior to the rotation is now mined out, or close to the surface, where weathering is likely to have destroyed any evidence of hypogene alteration. Presently, the model proposed by Dalstra (2002) is preferred as it explains a number of features such as the lack of hypogene alteration, the orientation of faults, and strong spatial relationships between faults and high-grade ore.

Morris (1985) and Harnsworth et al. (1990) suggested that the iron ore clasts within the Mount McGraft Formation were sourced from the nearby Paraburdoo deposit, and therefore the high-grade ore must have formed and was eroded during the Paleoproterozoic. However, the stratigraphic origin of the mineralized clasts has never been determined. It is equally possible that the clasts are sourced from the Weeli Woold Formation, and that the current Paraburdoo deposit was not exposed to the surface until the late Mesozoic.

For the Hope Downs deposit, Lascelles (2006a) invoked a modified supergene syngenetic model whereby supergene alteration of a chert-free BIF produced the present high-grade martite-goethite ore based on the high-grade ore, lacking any evidence of the prior presence of chert layers or the replacement of chert layers by carbonate. Synsedimentary chert-free BIF was also identified at Mount Gibson in the Yilgarn province (Lascelles, 2006b). Paquay and Ness (1986) described the structural and stratigraphic controls of the Hope Downs deposit and proposed a supergene model based on the observed ore textures (e.g., goethite replacing chert bands) and mineralogy (e.g., martite-goethite-earthy goethite).

The genetic ore models described in this paper are end members of the processes that form the economic BIF-related iron deposits of the Hamersley province. Within the Channar and Eastern Range deposits, located to the east of the Paraburdoo deposit (Fig. 1), martite-goethite orebodies both overprint and surround martite-microplaty hematite orebodies. It is likely that the faults and fracture networks that focused the flow of hypogene fluid during the Paleoproterozoic provide the necessary structural architecture for descending supergene fluids.

Conclusions

In the past decade significant advances have been made in the understanding of the formation of high-grade iron ore deposits of the Hamersley province. Advances include the structural reconstruction of deposits, mapping and detailed geochemistry of hypogene alteration zones, and a greater understanding of the processes and controls of supergene alteration.

The high-grade martite-microplaty hematite deposits are located within the Dales Gorge and Joffre Member of the Brockman Iron Formation, and the Marra Mamba Iron Formation and are spatially associated with normal faults and thrust faults that formed during D2 deformation (~2200 Ma), and have later been affected by D3 deformation (~1650 Ma). Petrological and geochemical studies at Mount Whaleback, Mount Tom Price, and Paraburdoo have identified hypogene carbonate alteration zones between BIF and high-grade iron ore. The replacement of chert layers by carbonate(s) results in the removal of quartz of the BIF and the crystallization of microplaty hematite within the carbonate matrix. The high-grade ore is composed almost entirely of subhedral martite grains, which range in size from 100 to 200 μm, and aggregates set in a porous (up to 30% porosity), randomly oriented network of microplaty hematite that ranges in size from 20 to 200 μm. Variable amounts of fine (<10 μm) anhedral, and large (200–300 μm) platy hematite within the ore result in a softer, more friable ore.

The high-grade martite-goethite deposits of Area C, Hope Downs, Christmas Creek, and Cool Break are spatially related to Mesozoic-Tertiary supergene alteration and the ore is composed of martite and goethite. Increased supergene fluid flow within synclines has caused multiple phases of goethite leaching precipitation and cementation.

Presently the most accepted controls for the formation of BIF-related high-grade iron ore deposits is that they form in areas where faults, active during the Paleoproterozoic, structurally link the underlying sedimentary packages of the Witteetoom Formation with the silico-rich BIF of the Brockman and Marra Mamba Iron Formations. When large volumes of ascending basinal brines are focused into the magnetite-chert-rich BIF, the rocks are transformed into a hypogene carbonate-altered BIF. Interaction of descending meteoric waters with the carbonate-altered BIF produces a porous martite-microplaty hematiteapatite ore. The final stage of late Mesozoic supergene-alteration removed much of the apatite from the martite-microplaty hematite-apatite assemblage, forming martite-microplaty high-grade ore and converted shale layers to kaolinite-rich clay.

Exploration Significance

After 50 years of intensive exploration and mining across the Hamersley province the discovery of new outcropping deposits is unlikely, therefore, making the search for concealed high-grade orebodies the new frontier for iron ore companies (Hargrav et al., 2007). Recent research on high-grade orebodies has resulted in the development of genetic models that
can provide exploration criteria aimed at discriminating between barren and hydrothermally altered and iron-enriched BIFs within the Hamersley province.

The most fundamental aspect of the supergene-modified hydrothermal model is that the hydrothermal fluids can form high P martite-microplaty hematite assemblages below the depth of supergene alteration. Also, and contrary to Morris (1985) who proposed that the iron ore formation required exposure during the Paleoproterozoic, the formation of the high-grade martite-microplaty hematite deposits is not related to Proterozoic erosion surfaces. The identification of possible hydrothermal martite-microplaty hematite ore within the Chichester Range deposits significantly improves the potential of the Marra Mamba Iron Formation to host significant martite-microplaty hematite occurrences in similar structural setting as those hosted within the Brockman Iron Formation (Clout, 2005).

The awareness of hydrothermal alteration zones, surrounding high-grade BIF-related iron deposits, allows companies to place emphasis on the identification and the spatial distribution of hypogene carbonate alteration and replacement textures. These can then be used as vectors toward high-grade iron ore in exploration drilling. Hagemann et al. (2007) showed that the manganese content of carbonates increases toward high-grade martite-microplaty ore at Mount Tom Price, thus potentially permitting the delineation of a geochemical vector toward high-grade ore. The understanding that faults and thrusts are important controls in the localization of supergene martite-goethite and supergene-modified hydrothermal martite-microplaty hematite deposits provides an important tool for exploration beneath cover. Conventional structural and stratigraphic mapping and reconstructions of the structural history of entire iron districts might allow the identification of areas where fluids can be focused by faults into structural traps (i.e., synelines).

Open Questions and Future Research

Several key questions remain within the supergene-modified hypogene models proposed for the Mount Whaleback, Mount Tom Price, and Paraburdoo deposits. One of these is the relative roles and composition of ascending and/or descending hypogene fluids and meteoric waters, respectively. Further work to integrate and expand the fluid data sets, in particular, within fault zones where evidence of fluid mixing is likely to be preserved, is needed. The mechanism and P-T-X condition for the crystallization of microplaty hematite also requires further work. Several mechanisms are invoked, both redox and nonredox, based on the requirement for oxygen to transform magnetite to hematite. If oxygen is not required to convert magnetite to hematite (cf. Ohmoto, 2003) then the descending oxygenated hydrothermal fluids, as proposed by, for example, Taylor et al. (2001), are not required to transform magnetite to martite or form microplaty hematite at Mount Tom Price. This significantly changes the relationship(s) of the deposit to the Proterozoic weathering surface and increases the depth at which high-grade orebodies may be found.

Establishing the source of the bauxite brines responsible for hypogene alteration and ultimately the formation of high-grade martite-microplaty hematite ore requires further work. The hydrothermal fluid source and the controls on fluid movement within the Hamersley province will have significant influence on the modeling of potential structurally controlled sites where hypogene alteration of the BIF may take place and, ultimately, the location of high-grade iron ore deposits.

The supergene origin of martite-goethite deposit has previously been widely accepted (Paquay and Ness, 1998). However, recent work by Lascelles (2006a) provided an alternative supergene-modified syngenetic model for the Hope Downs deposit. Stable isotope analyses of hematite, goethite, manganese, and carbonates can assist in elucidating the origin of fluids responsible for the transformation of BIF to high-grade martite-goethite ore at the deposit. Dating of alteration phases (e.g., hematite and goethite) may also help resolve the timing and ultimate formation of the Hope Downs deposit.

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We thank the Rio Tinto Exploration, Pilbara Iron Pty. Ltd., BHP Billiton and Fortescue Metals Group who allowed us access to current data on their deposits. Without their support in both logistics and geology many of the details and geological plans of their deposits would not have been possible. We also thank Des Lascelles, Jay Barton, Bruce M. Simonson, and Jens Gotzner for constructive criticism of an earlier version of this manuscript.

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Dudink, R. D., England, B. M., and Kneeshaw, M., 2000, Phosphorus distribution in martite-microplaty hematite ore requires further work. Tom Price. This significantly changes the relationship(s) of the redox and nonredox, based on the requirement for oxygen to be preserved, is needed. The mechanism and P-T-structural and stratigraphic mapping and reconstructions of an important tool for exploration beneath cover. Conventional chemical vector toward high-grade ore. The understanding of this manuscript.

Several key questions remain within the supergene-modified ore. Dating of alteration phases takes place and, ultimately, the location of high-grade iron ore enrichment: Unpublished B.Sc. thesis, Brisbane, University of Queensland, 20 p.


CHAPTER 3: OXYGEN ISOTOPE COMPOSITIONS OF IRON OXIDES FROM HIGH-GRADE BIF-HOSTED IRON ORE DEPOSITS OF THE CENTRAL HAMERSLEY PROVINCE, WESTERN AUSTRALIA: CONSTRAINTS ON THE EVOLUTION OF HYDROTHERMAL FLUIDS.

Statement of candidature contribution:
As first author, I conducted the isotope analyses of all samples (under the supervision of Prof. Torsten Vennemann), writing of the paper, and drafting of all figures. The role of co-authors was as follows:

- Steffen Hagemann contributed to the organisation, editorial, and scientific input and scrutiny of the paper.
- Torsten Vennemann assisted in the isotope analyses at the Isotope Laboratory of the University of Lausanne and provided valuable contributions to the interpretation of the results.
- Nick Oliver provided input into the regional significance of the data and discussion on hydrothermal processes.

This paper provides the most comprehensive work on the oxygen isotope composition of magnetite and hematite from BIF, hydrothermal alteration assemblages and high-grade iron ore from iron ore deposits in the western part of the Hamersley province. Importantly, the constraints on the temperature of the hydrothermal fluids through microthermometry (Chapter 4) have allowed the calculation of the oxygen isotopic composition of hydrothermal fluids.

Oxygen isotope analysis of magnetite and hematite from hydrothermal alteration assemblages and high-grade iron ore show a $^{18}\text{O}$ depletion of 5 to 15 per mil relative to the host BIF. The depletion in $\delta^{18}\text{O}$ values of hydrothermal alteration assemblages and high-grade iron ore is spatially controlled by normal faults, with the samples proximal to the faults exhibiting the greatest depletion in $^{18}\text{O}$ values. This is interpreted to be a response to higher fluid flux of hydrothermal fluids within fault zones. The structural control of hydrothermal fluids flow is further investigated in Chapter 6. Calculations of the oxygen isotope composition of hydrothermal fluids ($\delta^{18}\text{O}_{\text{HOT}}$) indicates that the decrease in the $^{18}\text{O}$ content of iron oxides is due to the interaction of both basinal brines and meteoric fluids with the original BIF.

The paper also describes how high-grade iron ore formed by hydrothermal fluid preserves its $\delta^{18}\text{O}$ composition, even when exposed to modern weathering processes. This characteristic allows the $\delta^{18}\text{O}$ composition of high-grade iron ore samples to be used as a diagnostic exploration tool for discrimination between iron ore formed from hypogene or supergene processes. The discovery that the stable isotope composition of hematite can produce meaningful insights into the formation of the ore bodies, lead to the investigation of crushed-leach analysis of hematite that is discussed in Chapter 4.

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Oxygen Isotope Compositions of Iron Oxides from High-Grade BIF-Hosted Iron Ore Deposits of the Central Hamersley Province, Western Australia: Constraints on the Evolution of Hydrothermal Fluids

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Abstract
The Hamersley province of northwest Western Australia is one of the world's premier iron ore regions with high-grade martite-microplaty hematite iron ore deposits mostly hosted within banded iron formation (BIF) sequences of the Brockman Iron Formations of the Hamersley Group. These high-grade iron ores contain between 60 and 68 wt percent Fe, and formed by the multistage interaction of hydrothermal fluids with the host BIF formation. The oxygen isotope compositions of magnetite and hematite from BIF, hydrothermal alteration assemblages, and high-grade iron ore were analyzed from the Mount Tom Price, Paraburdoo, and Channar iron ore deposits. The δ18O values of magnetite and hematite from hydrothermal alteration assemblages and high-grade iron ore range from −9.0 to −2.9 per mil, a depletion of 5 to 15 per mil relative to the host BIF. The δ18O values are spatially controlled by faults within the deposits, a response to higher fluid flux and larger influence on the isotopic compositions by the hydrothermal fluids. The oxygen isotope composition of hydrothermal fluids (δ18Ofluid) indicates that the decrease in the 18O content of iron oxides was due to the interaction of both basal brines and meteoric fluids with the original BIF. Late-stage talc-bearing ore at the Mount Tom Price deposit formed in the presence of a pulse of 18O-enriched basinal brine, indicating that hydrothermal fluids may have repeatedly interacted with the BIFs during the Paleoproterozoic.

Introduction
The recognition that hydrothermal fluids play an essential role in the formation of high-grade (≥63 wt % Fe) martite-microplaty hematite iron ore deposits of the Pilbara province has been recent (i.e., Barley et al., 1999; Hagemann et al., 1999; Powell et al., 1999). High-grade martite-microplaty hematite iron ore deposits, such as Mount Tom Price and Paraburdoo (Fig. 1), and Mount Whaleback farther east (Fig. 1; Powell et al., 1999), consist of almost pure hematite, limiting the types of geochemical analysis suitable to determine the roles of hydrothermal fluids in their formation. Because iron oxides are present in all stages during the transformation of banded iron formation (BIF; ≥35 wt % Fe) to high-grade martite-microplaty hematite ore, the study of their oxygen isotope compositions provides a suitable way for determining the interaction of fluids with the BIF. Numerous studies have documented the oxygen isotope compositions of iron oxides from BIFs (Becker and Clayton, 1976; Powell et al., 1999; Oliver et al., 2007) and high-grade martite-microplaty hematite ore deposits (Hagemann et al., 1999; Oliver and Dickens, 1999; Powell et al., 1999; Gutzmer et al., 2006; Oliver et al., 2007) from the Hamersley province. Becker and Clayton (1976) completed the most detailed oxygen isotope study of BIF to date from the Brockman Iron Formation (Fig. 1). Quartz-magnetite and quartz-siderite pairs from the Dales Gorge Member (Fig. 1) were found to have undergone isotopic exchange at a temperature estimated, on the basis of the isotopic fractionations, to be above 270°C and probably less than 310°C, during metamorphism (Table 1). Studies of the Mount Tom Price, Mount Whaleback, and Paraburdoo deposits (Ridley, 1999; Taylor et al., 2001; Dalstra, 2005; Thorne et al., 2004, 2005, 2007b; Sepe, 2007) have shown that hydrothermal alteration assemblages and high-grade iron ore crosscut the dominant magnetite-quartz-rich banding and the axial plane cleavage associated with F2 folds formed during the Ophthalmian orogeny (D2; 2.14 ± 0.03 Ga; Albert and McColloch, 1950). Peak metamorphic temperatures reached at least 200°C (Smith et al., 1982) during the Ophthalmian orogeny, and have been superimposed onto an earlier burial metamorphic event (Tyler and Thorne, 1990).

Powell et al. (1999) also studied the oxygen isotope compositions of quartz, hematite and magnetite from BIF layers of the Dales Gorge Member (Fig. 1). Stable isotope geothermometry (quartz-magnetite-hematite) indicated a temperature range of 270° to 405°C for magnetite, and 180° to 405°C for hematite similar to those obtained by Becker and Clayton (1976) for the Dales Gorge Member BIF for peak metamorphic temperatures (Table 1). Recent work on high-grade martite-microplaty hematite iron ore deposits (Hagemann et al., 1999; Oliver and Dickens, 1999; Powell et al., 1999; Gutzmer et al., 2006; Oliver et al., 2007) have focused on the isotopic compositions of quartz, magnetite, and hematite to determine the relative roles of ascending hydrothermal fluids and descending meteoric water in the transformation of BIF to high-grade martite-microplaty ore. Powell et al. (1999), Gutzmer et al. (2006),...
Mt Whaleback
martite-microplaty ore. Hagemann et al. (1999) studied the origin with the BIF, resulting in the formation of high-grade ore, and talc-bearing ore are used to determine the following:

1. controls on the spatial distribution of the oxygen isotope compositions of iron oxides within and between each deposit,
2. the relationship of oxygen isotope compositions to faults that may control hydrothermal fluid flow,
3. the cause of changes in oxygen isotope compositions of oxide minerals during the paragenetic steps from BIF to high-grade iron ore, and
4. the reasons for variations in oxygen isotope compositions of hematite from martite- and microplaty hematite-rich layers of high-grade ore samples.

Regional Geology

The tectonic history of the Hamersley province has been well documented by Tyler and Thorne (1990), Powell and Horwitz (1994), Krapez (1999), and Taylor et al. (2001). The stratigraphy of the Hamersley province previously has been described in detail, particularly by MacLeod (1966), Trendall and Blockley (1970), Harnsworth et al. (1990), and Taylor et al. (2001).

The southern Pilbara province (Fig. 1) has undergone several deformation events: the Ophthalmia orogeny (D2; 2140 ± 30 Ma, Sm-Nd riebeckite; Albright and McCulloch, 1993), the Panhandle event (~2020 Ma; Miller et al., 2005), and the Capricorn orogeny (D3; ~1780 Ma; Tyler and Thorne, 1990; Krapez, 1999).

Early deformation (D1), possibly associated with diageneric processes, produced some extensional fabrics at mesoscales but no major repetition or removal of stratigraphy (e.g., Powell et al., 1999; Brown et al., 2004). The Ophthalmian orogeny (D2) produced large scale folds with east-west-trending fold axes throughout the province, such as the Turner syncline. The Panhandle event produced D3 folds with northwest-trending fold axes that are best developed in the western part of the
Table 1. Microthermometry and Fluid Geochemistry of Hydrothermal Fluids Responsible for the Hydrothermal Alteration Assemblages at the Studied Deposits

<table>
<thead>
<tr>
<th>Location</th>
<th>Host rock Member</th>
<th>Minerals</th>
<th>Metamorphic temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cathedral Gorge</td>
<td>Dales Gorge</td>
<td>Quartz-siderite and quartz magnetite</td>
<td>270° to 310°C</td>
</tr>
</tbody>
</table>

Hydrothermal Alteration Assemblages and High-Grade Ore

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Host rock</th>
<th>Mineral assemblage</th>
<th>Th(HOM) (°C) (σ, n)</th>
<th>Salinity (wt % CaCl₂ equiv)</th>
<th>Th(HOM) (°C) (σ, n)</th>
<th>Salinity (wt % NaCl equiv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mount Brockman Iron Formation</td>
<td>Dales Gorge</td>
<td>Distal: magnetite-siderite-stilpnomelane</td>
<td>107°–142°C (120°±10°C, 1σ, n = 66)</td>
<td>24.0–26.0 wt % CaCl₂ equiv</td>
<td>157°–192°C (170°±10°C, 1σ, n = 66)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Intermediate: hematite-magnetite-siderite-stilpnomelane</td>
<td>153°–346°C (253°±60°C, 1σ, n = 120)</td>
<td>22.4–24.5 wt % CaCl₂ equiv</td>
<td>203°–360°C (353°±60°C, 1σ, n = 120)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Proximal: martite-microplaty</td>
<td>61°–112°C (100°±17°C, 1σ, n = 25)</td>
<td>5.5–9.5 wt % CaCl₂ equiv</td>
<td>111°–162°C (152°±17°C, 1σ, n = 25)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Talc-bearing ore:</td>
<td>118°–253°C (220°±60°C, 1σ, n = 31)</td>
<td>22.8–25.9 wt % CaCl₂ equiv</td>
<td>168°–303°C (270°±60°C, 1σ, n = 31)</td>
<td></td>
</tr>
<tr>
<td>Mount Tom Price</td>
<td>Brockman Iron Formation</td>
<td>Dolomite-chlorite-pyrite</td>
<td>98°–135°C (123°±12°C, 1σ, n = 34)</td>
<td>9.0–16.0 wt % NaCl equiv</td>
<td>138°–193°C (163°±12°C, 1σ, n = 34)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Quartz-hematite veins</td>
<td>98°–135°C (123°±12°C, 1σ, n = 34)</td>
<td>9.0–16.0 wt % NaCl equiv</td>
<td>138°–193°C (163°±12°C, 1σ, n = 34)</td>
<td></td>
</tr>
</tbody>
</table>

province, and interacted with the earlier east-west Ophthalmian folds (D1) to produce a dome-and-basin fold pattern. Following the Panhandle event, and prior to deposition of the Mount McGrath Formation, reactivation of earlier northwest-striking faults, as well as the development of northwest-striking, steep southwest-dipping normal faults with throws of up to 2100 m, formed horst and graben structures along the southern margin of the Hamersley province (Dalstra, 2005). The northwest-trending dolerite dikes (ca. 2008 Ma; SHRIMP 207Pb/206Pb baddeleyite; Müller et al., 2005) intruded into these normal faults and spatially control iron ore formation along the Paraburdoo range.

Two later phases of ductile to brittle-ductile deformation between ca. 1840 and 1770 Ma represent the Capricorn orogeny (Müller et al., 2005), which was the consequence of a collision between the Pilbara and Yilgarn cratons (Tyler and Thorne, 1990). The Capricorn orogeny postdates the deposition of conglomerates of the Upper Wyloo Group (Fig. 1) containing microplaty hematite ore clasts (Tyler and Thorne, 1990) and consequently is not considered important for the main phases of ore genesis (e.g., Powell et al., 1999; Müller et al., 2005), although it may have contributed to some radiogenic isotopic resetting (Rasmussen et al., 2005).

Rocks of the upper parts of the Hamersley Group (Boolegeeda, Woongarra, and Welli Wolli Formations) are locally eroded on two unconformities (Fig. 1; Taylor et al., 2001; Dalstra, 2005). The first unconformity (slightly discordant) rests the Beasley River Quartzite of the Lower Wyloo Group onto the Turee Creek and Hamersley Groups (Fig. 1) of the Mount Bruce Supergroup (Taylor et al., 2001; Dalstra, 2005). The second unconformity places the Mount McGrath Formation of the Upper Wyloo Group (Fig. 1) onto the rocks of the Hamersley Group and Lower Wyloo Groups at an angle of approximately 10° (Dalstra, 2005). The presence of microplaty hematite pebbles (Morris, 1980) within the McGrath Formation has been used to indirectly constrain the youngest age for iron ore formation within the Hamersley province (1843 ± 2 Ma; zircon 207Pb/206Pb TIMS; Pidgeon and Hovrutz, 1991). A later set of Mesozoic to Neoproterozoic northeast-trending mafic dikes and faults (e.g., Whaleback fault at Mount Whaleback; Powell et al., 1999) cuts across both the Ashburton trough and the Hamersley Group and postdates the Capricorn orogeny. The Channar dolerite (752 ± 10 Ma; U-Pb zircon, SHRIMP; Taylor et al., 2001) cuts through high-grade iron ore at the Channar deposit, producing contact aureoles of coarse-grained magnetite.

Regional metamorphic grade ascribed to the Ophthalmian orogeny is lowest in the northern Hamersley province (prehnite-pumpellyite facies) and increases southward to greenschist facies along the proximal martite-microplaty hematite-magnetite-apatite alteration assemblage formed late in the ore-forming process. High-grade martite-microplaty hematite ore preserves the meso- and microlayering of the host BIF and consists of randomly oriented fine-grained platy hematite (10–100 µm) and martite (20–250 µm), with a porosity from 10 to 50 vol percent (Fig. 4). Martite is subhedral to euhedral, and exhibits prominent overgrowth of microplaty hematite from the grain margins (Fig. 3E; Barley et al., 1999; Taylor et al., 2001; Thorne et al., 2004, 2008).

Important to note for this study is that the deposits studied have not been subject to a recognizable metamorphic overprint after the formation of the high-grade martite-microplaty hematite deposits (Smith et al., 1982; Rasmussen et al., 2005; Gutzmer et al., 2006), suggesting that the metamorphic grade associated with the postore Capricorn orogeny was very low, below the temperatures attained during the main phases of ore genesis (see below).

Local Geologic Setting

The following descriptions of the Mount Tom Price, Paraburdoo, and Channar deposits (Fig. 1) are included in order to define the structural and stratigraphic controls on high-grade iron ore formation.

Mount Tom Price deposit

The Mount Tom Price deposit (Fig. 2) is located within the south-central Hamersley province (Fig. 1). It contained an original resource of 900 Mt at 63.9 wt percent Fe and <0.05 wt percent P (Harnsworth et al., 1990). The orebody is oriented as a southerly dipping sheet within the eastern fold closure of the Turner syncline. The deposit extends for 7 km from the North orebody in the northwest to the South East Prong orebody in the southeast, is up to 1.6 km wide (avg 600 m), with a maximum depth of 250 m below the pre-mining topographic surface (Taylor et al., 2001). The majority of the high-grade ore is hosted within the Dales George Member with lesser ore within the upper McRae Shale, lower Whaleback Shale, and Joffre Member (Fig. 2).

Three hydrothermal alteration zones transitional between BIF (Fig. 3A) and high-grade iron ore (Fig. 3F; Barley et al., 1999; Hagemann et al., 1999; Taylor et al., 2001; Thorne et al., 2004, 2005) have been identified at the deposit: an outer distal magnetite-siderite-stilpnomelane (Fig. 3B), an intermediate hematite-magnetite-ankerite-chlorite (Fig. 3C), and a proximal martite-microplaty hematite-magnetite-apatite alteration zone (Fig. 3D; Thorne et al., 2005). Talc-bearing ore (Fig. 3E), formed by the replacement of quartz by talc, locally forms pods of high-grade iron ore in the hanging wall of the Southern Batter fault. Locally, talc-bearing ore veins crosscut the proximal martite-microplaty hematite-apatite alteration assemblage (Fig. 3E), indicating that the alteration assemblage formed later in the ore-forming process.

The most current ore deposit models (Barley et al., 1999; Taylor et al., 2001; Cochrane, 2003; Thorne et al., 2004, 2005, 2008) invoke a multistage hydrothermal origin for the transformation of BIF to high-grade martite-microplaty ore at the Mount Tom Price deposit, involving both hydrothermal and supergene processes. Stage 1 involved early (~2000 Ma; Müller et al., 2005) hypogene alteration by upward movement of hydrothermal, CaCl2-rich brines from the underlying Wittenoom Formation into the large-scale folds of the Dales...
OXYGEN ISOTOPE COMPOSITIONS OF IRON ORE DEPOSITS OF THE CENTRAL HAMERSLEY PROVINCE, W.A.

Figure 2. Cross sections of the Mount Tom Price deposit. Cross sections through the North and Southern Ridge orebodies display the stratigraphic units and major normal faults within the orebodies, the spatial distribution of hydrothermal alteration assemblages and high-grade ore (modified after Thorne et al., 2007b), and the δ18O values (‰) of hematite and magnetite samples. Samples obtained from open-pit grab samples and diamond drill core. Note the low δ18O values of samples proximal to fault zones. Section shown facing west.
Gorge Member. Reactions with these hypogene basinal brines (107°C to 142°C, ~25 wt % equiv; NaCl-CaCl2-rich; Table 1) transformed magnetite-quartz–rich BIF to a magnetite-siderite-iron silicate assemblage. Continued hypogene alteration is characterized by an apparent increase in temperature (fluid inclusion homogenization from 153°C to 346°C, 23 wt % NaCl equiv, CaCl2-rich; Table 1), and an increase in $f_{O_2}$ recorded by the formation of a hematite-ankerite-magnetite assemblage and the crystallization of microplaty hematite. The latest hydrothermal alteration event involved the interaction of low-temperature, low-salinity (65°–110°C; Table 1) meteoric water, with the hematite-ankerite-magnetite assemblage being replaced by a porous martite-microplaty hematite-apatite assemblage. Supergene alteration is likely to have begun in the Cretaceous and continued for at least 80 m.y. and is ongoing today (Vasconcelos, 1998), resulting in the removal of residual ankerite and apatite and the weathering of the shale layers to kaolinite-rich clay (Taylor et al., 2001).

Paraburdoo deposit

The Paraburdoo deposit is located 65 km south of the Mount Tom Price deposit on the southern margin of the central Hamersley province (Fig. 1). The deposit contains, prior to mining, an ore reserve of more than 300 Mt at 64 wt percent Fe and 0.08 wt percent P (Taylor et al., 2001). High-grade ore is spatially controlled by flat faults that now underlie both the 4W and 4E orebodies (Dalstra, 2005), and northwest-trending dolomite dikes. High-grade ore is equally distributed between the Dales Gorge and Joffre Members with minor amounts in the Mount McRae Shale, Yandicoogina Shale, and Weeli Wolli Formation (Fig. 5). Most of the Paraburdoo deposit is in the Tertiary supergene alteration profile and exhibits a strong goethite overprint of the primary martite-microplaty hematite ore. However, below the 4 East and 2 Lens orebodies and the depth of modern weathering, carbonate-altered BIF (50–55 wt % Fe; hematite-dolomite-chlorite-pyrite; Figs. 4, 5) occurs transitional between BIF (~35 wt % Fe) and high-grade (> 65 wt % Fe) iron ore (Thorne et al., 2007b). Thorne et al. (2008) proposed a supergene-modified hydrothermal model for the 4 East and 2 Lens orebodies at Paraburdoo (Fig. 5). The first hypogene stage involved alteration by upward movement along fault zones, of 170ºC (Table 1), basinal brines (20.9 wt % CaCl2 equiv with Ca > Na > K; Table 1) into the Dales Gorge and Joffre Members of the Brockman Iron Formation. The low Cl/Br and Na/Br ratios, high Ca concentrations, and elevated base metal content of the hydrothermal fluids suggest that they were modified basinal brines sourced from evaporated seawater from which halite had precipitated prior to being expelled from the Wittenoom Formation. Fluid-rock reactions transformed magnetite-quartz–rich BIF into magnetite-calcite-stilpnomelane-anhedral hematite-pyrite and hematite-dolomite-chlorite-pyrite.
assemblages. Stage 2 hydrothermal alteration involved the interaction of low-temperature, low-salinity (65º–110ºC; Table 1) meteoric water, with the hematite-dolomite-chlorite-pyrite assemblage, resulting in the formation of the proximal martite-microplaty hematite-apatite assemblage. Stage 3 supergene alteration during the Tertiary entirely overprinted the proximal alteration zone converting, shale layers to kaolinite-rich clay, destroying apatite, and leaching calcium and phosphorus from the ore. High-grade ore at the Paraburdoo deposit characteristically contains abundant anhedral hematite (Fig. 4).

The Ratty Springs prospect lies approximately 2 km west of the Paraburdoo deposit (Fig. 1). Quartz-hematite veins are hosted within the layers of the Mount Sylvia Formation, and are up to 50 cm in width and have a strike length of several meters. The veins lie both oblique and parallel to the bedding in the wall rock, and are surrounded by alteration zones of large (up to 5 mm), elongate, tabular hematite and fine-grained hematite (>100 µm). Microthermometry on primary fluid inclusions trapped with quartz within quartz-hematite veins indicates that a low-temperature (100°–150°C), moderate salinity (9 to 16 wt % NaCl equiv) fluid, possibly formed from a mixture of heated meteoric fluids and basinal brines, was responsible for formation of the veins (Table 1). Similar veins have been identified within fault zones (Mount Tom Price; Hagemann et al., 1999) and surrounding deposits (Mount Whaleback; Powell et al., 1999; Brown et al., 2004), suggesting that the quartz-hematite veins may represent the periphery of the mineralizing system (Dalstra, 2005).

### Channar E3 deposit

The group of five deposits known collectively as the Channar mining area is approximately 20 km southeast of the town of Paraburdoo, on the southern limb of the Bellary anticline (Harmsworth et al., Fig. 1). The total resource at Channar mining area is 290 Mt at 63.0 percent iron and 0.09 wt percent phosphorus (Harmsworth et al., 1990). High-grade ore at the Channar East 3 deposit (Fig. 6), the Dales Gorge Member is cut by two northwest-trending dolerite dikes, of which the southernmost is intruded in a normal fault zone (Fig. 6). The eastern edge of the long dolerite dike is at a depth of 150 m below the present topographic surface. At the Channar East 3 deposit (Fig. 6), the Dales Gorge Member is cut by two northwest-trending dolerite dikes, of which the southernmost is intruded in a normal fault zone (Fig. 6). The eastern edge of the

<table>
<thead>
<tr>
<th><strong>Hydrothermal Alteration Assemblages</strong></th>
<th>Euhedral Magnetite (Mg)</th>
<th>Platy Magnetite (pMg)</th>
<th>Bladed Hematite (pHm)</th>
<th>Microplaty Hematite (mpH)</th>
<th>Martite (Mr)</th>
<th>Anhedral hematite (aHm)</th>
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<tbody>
<tr>
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<td></td>
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<tr>
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<td></td>
</tr>
<tr>
<td>Proximal (martite-dolomite-chlorite-pyrite)</td>
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<tr>
<td>Ratty Springs (quartz-hematite veins)</td>
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<td>Channar</td>
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<tr>
<td>Proximal (martite-dolomite-microplaty hematite-rhodochrosite-chlorite)</td>
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<tr>
<td>High-grade iron ore</td>
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<tr>
<td>Mount Tom Price</td>
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<tr>
<td>Paraburdoo</td>
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<table>
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<th><strong>Data</strong></th>
<th><strong>Banded Iron Formation</strong></th>
<th><strong>Hydrothermal Alteration Assemblages</strong></th>
<th><strong>High-grade iron ore</strong></th>
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<td></td>
<td>Dales Gorge Member</td>
<td>Mount Tom Price</td>
<td>Mount Tom Price</td>
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<td></td>
<td>Joffre Member</td>
<td></td>
<td>Channar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Distal (magnetite-siderite-stilpnomelane)</td>
<td>Proximal (martite)</td>
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<td></td>
<td></td>
<td>Intermediate (hematite-magnetite-stilpnomelane)</td>
<td>Dolomite (Mr)</td>
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<td></td>
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<td>Proximal (martite-microplaty hematite-apatite)</td>
<td>Microplaty (pHm)</td>
</tr>
<tr>
<td></td>
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<td>Talc-bearing ore (talc ± martite ± apatite)</td>
<td>Martite (Mr)</td>
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<td></td>
<td></td>
<td>Paraburdoo</td>
<td>High-grade iron ore</td>
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<tr>
<td></td>
<td></td>
<td>Proximal (hematite-dolomite-chlorite-pyrite)</td>
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<td>Ratty Springs (quartz-hematite veins)</td>
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<td>Channar</td>
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<td></td>
<td></td>
<td>Proximal (martite-dolomite-microplaty hematite-rhodochrosite-chlorite)</td>
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</tbody>
</table>
FIG. 5. Cross section of the 4E orebody in the Paraburdoo deposit displaying the major structural features within the deposit, the spatial distribution of hydrothermal alteration assemblages, and high-grade ore (modified after Taylor et al., 2001) and the δ18O values (‰) of hematite and magnetite samples. Section shown facing west.

FIG. 6. Cross section through the Channar E3 orebody displaying the major structural features within the deposit, the spatial distribution of hydrothermal alteration assemblages and the δ18O values (‰) of hematite samples. Section shown facing west (FWZ = footwall zone, DG1, DG2, DG3 = Mining subdivision of Dales Gorge Member, Courtesy of Rio Tinto Exploration).
orebody is defined by a northeast-trending dolerite dike that
crosscuts and postdates the formation of the high-grade iron
ore (Taylor et al., 2001).

One hydrothermal alteration zone has been identified as
transitional between BIF and high-grade martite-microplaty
ore. The martite-dolomite-microplaty hematite-rodhemchrorite-
chlorite assemblage was identified in the hanging wall of a
northwest-trending normal fault (Fig. 6). Similar structural
and stratigraphic controls (Dalstra, 2005), paragenetic se-
quence of alteration assemblages (Thorne et al., 2007a), and
fluid geochemistry (Table 1) of the Channar E3 orebody to
the Paraburdoo deposit suggest a shared genetic history for
the transformation of BIF to high-grade martite-microplaty
hematite between the deposits.

Petrology and Classification of BIF, Hydrothermal
Alteration Zones, and High-Grade Iron Ore

Sampling

A total of 59 samples were collected from diamond drill
core and hand samples from the Mount Tom Price (Fig. 2; n
= 35), Paraburdoo (Fig. 5; n = 20) and Channar (Fig. 6; n = 3)
deposits and Ratty Springs prospect (n = 1). Samples were
taken from BIF, hydrothermal alteration assemblages, and
high-grade iron ore hosted within the Dales Gorge, Joffre,
and Whaleback Shale members, and the Mount Sylvia For-
mation (Table 2).

Oxygen isotope analyses for this study were performed on
mineral separates or hand-picked grains from BIFs, veins
containing combinations of quartz, carbonate, magnetite,
hematite, and ores containing hematite and locally, goethite.
Hematite, magnetite and quartz were separated from BIF by
line crushing, hand magnet, magnetic separation, and use of
heavy liquids (sodium polytungstate) in a centrifuge, and
were also hand-picked from coarse vein samples. Although
future work will undoubtedly be able to refine the microana-
ytical techniques to deal with the problems of complex, fine-
grained mineral intergrowths in BIFs and ores, we have gone
to considerable lengths to ensure sample purity. The oxide-
silicate separation techniques were only concluded for an in-
dividual sample when petrographic examination of powders
indicated less than 1 percent contamination, adding a poten-
tial error of up to 0.2 per mil to δ18O values for oxides, com-
parable to the analytical errors.

Analytical methods

The oxygen isotope compositions (16O, 18O) of the samples
were measured at the Institute of Mineralogy and Geochem-
istry, University of Lausanne, Switzerland, using a method
similar to that described by Sharp (1990) and Rumble and
Hoering (1994), and that is described in more detail in Kase-
mann et al. (2001).

Between 0.5 to 2 mg of sample was loaded onto a small Pt-
sample holder and pumped out to a vacuum of about 10–6 mbar.

Table 2. The δ18O values (‰) of BIF, Hydrothermal Alteration Assemblages and High-Grade Iron from the Selected Deposits

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Sample type</th>
<th>Hole ID/ open pit location</th>
<th>Depth/L (m)</th>
<th>Mineralogy</th>
<th>Mass (mg)</th>
<th>δ18O VSMOW (‰)</th>
</tr>
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<tbody>
<tr>
<td>G831 329.5</td>
<td>1 DC</td>
<td>G831</td>
<td>329.5</td>
<td>Magnetite</td>
<td>2.28</td>
<td>4.1</td>
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<tr>
<td>G1186 142</td>
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<td>G1186</td>
<td>142.0</td>
<td>Magnetite</td>
<td>3.52</td>
<td>6.0</td>
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<tr>
<td>PA002 3</td>
<td>3 DC</td>
<td>GTO54E001</td>
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<td>8.8</td>
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<tr>
<td>4EST DDH44 304m</td>
<td>4 DC</td>
<td>4EST DDH44</td>
<td>304.0</td>
<td>Magnetite</td>
<td>3.82</td>
<td>13.0</td>
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</tbody>
</table>
| Distal (magnetite-siderite-
stilpnomelane) |
| GT03NTD001 46.45m | 5 DC | GT03NTD001               | 146.4       | Magnetite  | 2.27      | -7.7          |
| NTD0900 6 DC | NTD0900     | 111.1                    | Magnetite   | 2.27      | -9.0       |
| 317 7 DC   | G1186       | 106.5                    | Magnetite   | 2.41      | -8.6       |
| Intermediate (hematite-
magnetite-ankerite-
clinochlore) |
| NTD122 8 DC | GTO3NTD001   | 142.5                    | Martite-microplaty hematite | 2.69 | -5.3 |
| 261 9 DC   | G381-77     | 422.2                    | Martite-microplaty hematite | 3.46 | -6.0 |
| 517 10 DC  | G906-77     | 329.7                    | Martite-microplaty hematite | 2.74 | -5.5 |
| 518 11 DC  | G906-77     | 335.0                    | Martite-microplaty hematite | 2.13 | -4.6 |
| 77779 12 DC | ND77779     | 110.9                    | Martite-microplaty hematite | 2.05 | -6.2 |
| 510a 13 DC | G906-77     | 341.5                    | Martite-microplaty hematite | 1.49 | -3.5 |
| 514 14 DC  | G906-77     | 342.0                    | Martite-microplaty hematite | 2.62 | -3.9 |
| 520 15 DC  | G906-77     | 354.0                    | Martite-microplaty hematite | 2.27 | -4.3 |
| Conformal (martite-
microplaty hematite-
magnetite-spilitic) |
| L4970 17 DC | L4-77       | 293.1                    | Martite-microplaty hematite | 2.55 | -9.5 |
| 514 18 DC  | G906-77     | 284.0                    | Martite-microplaty hematite | 2.50 | -4.1 |
| 523 19 DC  | G906-77     | 375.6                    | Martite-microplaty hematite | 1.83 | -4.4 |
| 525 20 DC  | G906-77     | 385.1                    | Martite-microplaty hematite | 3.61 | -5.2 |
After preflourination of the sample chamber overnight, the samples were heated with a CO₂-laser in 50 mbars of pure F₂. Excess F₂ is separated from the O₂ produced by conversion to Cl₂ using KCl held at 150°C. The extracted O₂ is collected on a molecular sieve (5A) and subsequently expanded into the inlet of a Finnigan MAT 253 isotope ratio mass spectrometer. Oxygen isotope compositions are given in the standard δ notation, expressed relative to VSMOW in per mil. Replicate

<table>
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<tr>
<th>Sample ID</th>
<th>ID</th>
<th>Sample type</th>
<th>Hole ID/depth (m)</th>
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<th>Mass (mg)</th>
<th>δ¹⁸O Sample/‰</th>
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<td>DC</td>
<td>GTO34E901 55.5</td>
<td>Martite-microplaty hematite</td>
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<td>4500W 400</td>
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<td>Channar E3 410</td>
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<td>-2.9</td>
</tr>
<tr>
<td>STR011</td>
<td>38</td>
<td>HS</td>
<td>Southern Ridge 800</td>
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<td>2.27</td>
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<td>HS</td>
<td>L4-77 218.0</td>
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<td>-3.2</td>
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<tr>
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<td>41</td>
<td>DC</td>
<td>L4-77 250.0</td>
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<td>42</td>
<td>HS</td>
<td>4 East Main Pit 285</td>
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<td>43</td>
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<tr>
<td>11W-1</td>
<td>44</td>
<td>HS</td>
<td>11West 385</td>
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<tr>
<td>11W-1</td>
<td>45</td>
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<td>11W-2</td>
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<td>-5.4</td>
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<td>4WES-2</td>
<td>47</td>
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<td>Martite-microplaty hematite</td>
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<tr>
<td>866</td>
<td>48</td>
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<td>DD086EMP059 247.8</td>
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<tr>
<td>897</td>
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<td>DD086EMP059 266.0</td>
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<td>-4.8</td>
</tr>
<tr>
<td>898</td>
<td>50</td>
<td>DC</td>
<td>DD086EMP059 303.8</td>
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<td>1.77</td>
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</tr>
<tr>
<td>901</td>
<td>51</td>
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<td>Martite-microplaty hematite</td>
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<tr>
<td>906</td>
<td>52</td>
<td>DC</td>
<td>DD086EMP059 229.4</td>
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<td>-5.0</td>
</tr>
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<td>53</td>
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<tr>
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<td>-4.1</td>
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<td>55</td>
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<td>917</td>
<td>56</td>
<td>DC</td>
<td>DD086EEXP0000 229.9</td>
<td>Martite-microplaty hematite</td>
<td>1.77</td>
<td>-5.4</td>
</tr>
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<td>57</td>
<td>DC</td>
<td>DD086EMPO13 170.1</td>
<td>Martite-microplaty hematite</td>
<td>1.84</td>
<td>-4.6</td>
</tr>
<tr>
<td>942</td>
<td>58</td>
<td>DC</td>
<td>DD086EMPO13 185.2</td>
<td>Martite-microplaty hematite</td>
<td>2.84</td>
<td>-4.4</td>
</tr>
<tr>
<td>CHE3-388-1</td>
<td>59</td>
<td>HS</td>
<td>CHE-3 396</td>
<td>Martite</td>
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<td>-5.9</td>
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<tr>
<td>CHE3-388-2</td>
<td>60</td>
<td>HS</td>
<td>CHE-3 396</td>
<td>Microplaty hematite</td>
<td>2.96</td>
<td>-7.3</td>
</tr>
</tbody>
</table>

Paraburdoo (martite-microplaty hematite)

Intermediate (hematite-dolomite-chlorite-pyrite)

Paraburdoo (martite-microplaty hematite)

Channar (martite-microplaty hematite)

Paraburdoo (martite-microplaty hematite)

High-Grade Ore

Mt Tom Price (martite-microplaty hematite)

Talc-bearing ore (talc-apatite-martite)

Batty Springs (quartz-hematite veins)

Channar Proximal (martite-dolomite-microplaty hematite-rhodochrosite-chlorite)

Channar (martite-microplaty hematite)

Table 2. (Cont.)
oxygen isotope analyses of the standards used (NBS-28 quartz, n = 13, and UWG-2 garnet, n = 7; Valley et al., 1995) indicated an average reproducibility of ± 0.1 per mil for δ18O. The accuracy of δ18O values was commonly better than 0.2 per mil compared to accepted δ18O values for NBS-28 of 9.64 per mil and UWG-2 of 5.8 per mil.

Results

The results of the oxygen isotope analysis from the Mount Tom Price, Paraburdoo, and Channar deposits are presented in Tables 2 and 3, and Figure 7.

BIF: Magnetite from magnetite-quartz-rich BIF of the Dales Gorge Member and Joffre Member has positive δ18O values (4.1 and 6.0‰) with magnetite from the Joffre Member BIF having higher values (8.8 and 13.0‰). The δ18O values from this study are similar to previous studies by Becker and Clayton (1976), Oliver and Dickens (1999), and Powell et al. (1999), who reported δ18O values of magnetite from the Dales Gorge Member ranging from 1 to 9 per mil (Fig. 7).

Hydrothermal alteration assemblages

Magnetite and hematite from distal, intermediate, and proximal hydrothermal alteration assemblages from the Mount Tom Price, Paraburdoo, and Channar deposits are depleted in 18O when compared to magnetite from unaltered BIF (Tables 2, 3; Fig. 7).

At the Mount Tom Price deposit, magnetite from distal magnetite-siderite-stilpnomelane alteration is depleted by up to 13 per mil, and hematite from the intermediate, proximal alteration zone and high-grade ore samples are depleted by 8 to 10 per mil compared to the δ18O values for (metamorphic) magnetite from BIFs. At the Paraburdoo and Channar East 3 deposits, hematite from proximal hematite-dolomite-chlorite-pyrite and martite-dolomite-microplaty hematite-rhodochrosite-chlorite alteration zones are 18O depleted, on average, by 15 and 9 per mil, respectively, relative to δ18O values for BIF magnetite (Table 2; Fig. 8). Hematite from late-stage talc-bearing ore at the Mount Tom Price deposit has a δ18O value of 4 per mil less than BIF magnetite (Tables 2, 3; Fig. 7).

Quartz-hematite veins: Hematite from the quartz-hematite veins at the Ratty Springs prospect is depleted in 18O when compared to the BIF but has a similar value to those of

### Table 3. The δ18O values (‰), oxide-water fractionation (△18O, ‰) and calculated δ18Ofluid (‰) of BIF, hydrothermal alteration assemblages and high-grade iron from the selected deposits.

<table>
<thead>
<tr>
<th></th>
<th>δ18O of oxides</th>
<th>Fractionation oxide-water △18O</th>
<th>δ18Ofluid</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n</td>
<td>Min</td>
<td>Max</td>
</tr>
<tr>
<td>Banded iron formation</td>
<td>Dales Gorge Member</td>
<td>2</td>
<td>4.1</td>
</tr>
<tr>
<td></td>
<td>Joffre Member</td>
<td>2</td>
<td>8.8</td>
</tr>
<tr>
<td>Hydrothermal alteration assemblages</td>
<td>Mt Tom Price</td>
<td>Distal (magnetite-siderite-stilpnomelane)</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Intermediate (hematite-magnetite-ankerite-chlorite)</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Proximal (martite-microplaty hematite-magnetite-apatite)</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Talc-bearing ore (talc-apatite-martite)</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>Paraburdoo</td>
<td>168–303</td>
<td>–3.5 to –2.8</td>
</tr>
<tr>
<td></td>
<td>Intermediate (hematite-dolomite-chlorite-pyrite)</td>
<td>1</td>
<td>–5.6</td>
</tr>
<tr>
<td></td>
<td>Ratty Springs (platy hematite)</td>
<td>98–151</td>
<td>–5.8 to –2.3</td>
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<tr>
<td></td>
<td>Channar</td>
<td>Proximal (martite-dolomite-microplaty hematite-rhodochrosite-chlorite)</td>
<td>1</td>
</tr>
<tr>
<td>High-grade ore</td>
<td>Mt Tom Price</td>
<td>Distal (martite-microplaty hematite)</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>Paraburdoo</td>
<td>17</td>
<td>–7.3</td>
</tr>
<tr>
<td></td>
<td>Channar</td>
<td>2</td>
<td>–7.7</td>
</tr>
</tbody>
</table>
The δ¹⁸O values are similar to the δ¹⁸O values of hematite from quartz-hematite veins at the Paraburdoo deposits (–4.1 ‰; Table 3; Fig. 7) and from the Mount Whaleback deposit (–3.4 ‰; Powell et al., 1999). Compositional variations in mesobands A comparison of the δ¹⁸O values of microplaty-hematite (Fig. 3C) and martite (Fig. 3C) dominant bands within the high-grade ore samples at Channar shows that the δ¹⁸O values of microplaty hematite-dominant layers are more ¹⁸O depleted (–7.7‰) than martite-dominant layers (–7.0‰). In the martite and microplaty hematite samples from the Paraburdoo deposit, the microplaty hematite layers (–6.1‰) are more depleted in ¹⁸O than the martite layers (–5.8‰), whereas at Mount Tom Price the martite-dominant layers are more ¹⁸O-depleted (–5.5‰) than the microplaty hematite layers (–3.9‰). Hence the given variability is not consistent.

Fluids in equilibrium with the oxides (δ¹⁸Ofluid) During hydrothermal alteration, the change in the δ¹⁸O values of the iron oxides regarded as a mixture between the oxygen present in the iron oxides and the oxygen introduced by the fluid (Gregory, 1989). When either basinal brines or meteoric fluids interact with the BIF, the lowest δ¹⁸O values measured for iron oxides in a specific alteration assemblage are assumed to be the most likely to be in isotopic equilibrium with the fluid (e.g., Gutzmer, 2005). Otherwise stated, in such cases the fluid/rock ratio is likely to be the highest and values calculated for the fluids may approach those of the originally infiltrating fluid. For iron oxides with higher δ¹⁸O values, the original BIF values are likely to have been retained either because the fluid/rock ratio was low or because the fluid had a high δ¹⁸O value that may even have been in equilibrium with the original iron oxides. To account for these variables, we have calculated the fluid (δ¹⁸Ofluid) to be in equilibrium, based on the lowest δ¹⁸O values measured for iron oxides (Table 2) within a given alteration assemblage and over the range of trapping temperatures (Table 3). The δ¹⁸Ofluid in equilibrium with iron oxides was calculated (Fig. 7) using the isotope fractionations of Becker and Clayton (1976) for magnetite and of Yapp (1990; Table 3) for hematite. The former allows a direct comparison to the study on BIFs by Becker and Clayton (1976) and because the fractionations are similar to those determined by Cole et al. (2004) at 300°C. At low temperatures, the Yapp (1990) fractionation is considered reliable and correlates well with known δ¹⁸O values of modern and Miocene waters (i.e., Seidel et al., 2005) and at higher temperatures, the Yapp (1990) fractionation correlates well with several hematite-quartz, quartz-water, hematite-carbonate, and carbonate-water fractionations. The δ¹⁸Ofluid in equilibrium with magnetite and hematite was calculated at temperatures based on the fluid inclusion homogenization temperatures from BIF (Table 1), hydrothermal alteration assemblages, and high-grade iron ores. Trapping temperatures were pressure-corrected using stratigraphic reconstructions of the Mount

### Table: Compositional variations in mesobands

<table>
<thead>
<tr>
<th>Banded Iron Formation</th>
<th>Hydrothermal Alteration Assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dales Gorge Member</td>
<td>Mount Tom Price</td>
</tr>
<tr>
<td>Joffre Member</td>
<td>Distal (magnetite siderite-stiplnomelane)</td>
</tr>
<tr>
<td></td>
<td>Intermediate (hematite-magnetite-stiplnomelane)</td>
</tr>
<tr>
<td></td>
<td>Proximal (martite-microplaty hematite-apatite)</td>
</tr>
<tr>
<td></td>
<td>Talc-altered ore (talc ± martite ± apatite)</td>
</tr>
<tr>
<td></td>
<td>Paraburdoo</td>
</tr>
<tr>
<td></td>
<td>Proximal (hematite-dolomite-chlorite-pyrite)</td>
</tr>
<tr>
<td></td>
<td>Ratty Springs (quartz-hematite veins)</td>
</tr>
<tr>
<td></td>
<td>Channar</td>
</tr>
<tr>
<td></td>
<td>Proximal (martite dolomite-microplaty hematite-rhodochrosite-chlorite)</td>
</tr>
</tbody>
</table>

| High-grade iron ore   | Mount Tom Price                   |
|                       | Paraburdoo                        |
|                       | Channar                           |

### Fig. 7. The δ¹⁸O values (‰) of BIF, hydrothermal alteration assemblages, and high-grade iron from the studied deposits.

The range of calculated δ¹⁸Ofluid values (‰) is shown over the range of temperatures calculated by microthermometry.
Tom Price deposit (30°, Taylor et al., 2001, Thorne et al., 2001) and the Paraburdoo (40°) and Channar deposits (30°; Dalstra, 2005) during the likely time of hydrothermal alteration in the Paleoproterozoic.

**BIF:** The δ¹⁸Ofluid calculated to be in equilibrium with magnetite from both the Dales Gorge and Joffre members have similar values (Fig. 7) that range from 12.7 to 21.7 per mil at temperatures of 270° to 310°C.

**Hydrothermal alteration assemblages:** At Mount Tom Price the average δ¹⁸Ofluid calculated to be in equilibrium with magnetite from distal alteration has a range from -0.7 to -0.4 per mil at temperatures of 157° to 192°C (Table 3).

Hematite from the intermediate alteration assemblage at Mount Tom Price has a range of δ¹⁸O values from -3.5 to 1.2 per mil at temperatures of 153° to 346°C (Table 3; Fig. 7). The hematite from proximal alteration and high-grade martite-microplaty hematite ore has calculated δ¹⁸Ofluid values that range from -8.3 to -5.8 per mil and -6.7 to -4.3 per mil, respectively (Table 3; Fig. 7) at temperatures of 61° to 112°C.

**OXYGEN ISOTOPE COMPOSITIONS OF IRON ORE DEPOSITS OF THE CENTRAL HAMERSLEY PROVINCE, WA**

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Similar trends are seen at the Paraburdoo and Channar deposits. The δ¹⁸Ofluid values calculated to be in equilibrium with hematite from proximal alteration assemblages have more positive values than those calculated to be in equilibrium with hematite from high-grade martite-microplaty ore. Hematite from proximal alteration assemblages have δ¹⁸Ofluid of -0.9 to
–0.6 per mil at temperatures of 150° to 190°C at the Paraburdoo deposit and –1.5 to 3.2 per mil at temperatures of 139° to 325°C for the Channar E3 deposit (Table 3; Fig. 7). Hematite from high-grade ore has δ18Omin values of –9.6 to –6.0 per mil at Paraburdoo and –10.0 to –6.4 per mil at Channar E3 deposit at temperatures of 61° to 112°C (Table 3; Fig. 7).

Hematite from quartz-hematite veins at the Ratty Springs prospect has calculated δ18Omin values that range from –5.8 to –2.3 per mil at temperatures of 98° to 153°C, similar to high-grade martite-microplaty hematite ore from the nearby Paraburdoo deposit (Table 3, Fig. 7).

**Discussion**

**Spatial distribution of δ18O values**

At Mount Tom Price there is a clear spatial relationship between the depletion in δ18O of hematite and martite relative to the BIF and the Southern Batter fault zone (Fig. 2). Numerous studies (Hagemann et al., 1999; Taylor et al., 2001) have shown that the Southern Batter fault (Fig. 2) provides the likely fluid for ascending and descending hydrothermal fluids. The δ18O values of samples proximal to the Southern Batter fault are expected to be the most fluid dominated and, for a given temperature, their oxygen isotope compositions are likely to relate closely to those of the hydrothermal fluid. Farther from the fault zones, and at the periphery of the hydrothermal system, the δ18O values of the iron oxides represent oxygen sourced both from the BIF and, to a lesser extent compared to the proximal fault zones, from the hydrothermal fluids. Cope et al. (2008) studied the isotopic composition of hematite from high-grade ore at the Pic de Fon deposit in Guinea, and also found that the most δ18O-depleted samples were located proximal to shear zones. It is likely that the high-fluid flux within the fault-shear zones dominates the oxygen budget between the fluids and the rocks.

**Inheritance of isotopic compositions**

Samples from martite-microplaty ore at Mount Tom Price contain martite layers that are more depleted in δ18O than the microplaty hematite, a relationship that is not observed at either the Paraburdoo or Channar deposits. This is interpreted to be the result of the martite inheriting the highly δ18O-depleted composition of martite (δ18O = –8.4 ‰) from the martite-siderite-stilpnomelane alteration zone. Gutzmer (2006) observed a similar relationship at the Noamundi deposit in South Africa (Fig. 9), where the martite-textured hematite is more depleted in δ18O than microplaty hematite from the same deposit.

The replacement of martite by hematite took place at low to moderate temperatures of hydrothermal alteration (up to 300°C). At these low temperatures, geologically reasonable rates of isotopic exchange in mineral-fluid systems that are initially far from chemical equilibrium are controlled largely by chemical reactions such as crystallization from a fluid phase (Cole et al., 1983; Cole and Öhmoto, 1986; Stof-fregen, 1996) or transformation to a new phase (Cole and Chakraborthy, 1991). Isotopic exchange controlled by chemical reactions is typically several orders of magnitude faster than rates influenced by diffusion. Regardless of whether the reactants simply recrystallize in the presence of a fluid phase or become replaced by a new crystalline phase, it is clear that isotopic exchange will accompany the dissolution and precipitation leading to new crystal growth (Cole and Chakraborthy, 2001). During the recrystallization of martite to hematite, Fe-O bonds are broken and reformed and if this occurs in the presence of a fluid, the oxygen now newly incorporated into the martite may be derived not only from the magnetite but also from the hydrothermal fluid. Hence, the change in the δ18O values of the iron oxide will be a function of the fluid/rock ratio, and if this is low or if the fluid is already in isotopic equilibrium with the iron oxides, the newly formed phase may practically inherit the δ18O values of the original oxide.

Zheng (1995) studied the discrepancy in the quartz-magnetite oxygen isotope fractionation between previous theoretical calculations of Zheng (1991) and the experimental determinations of Chiba et al. (1989). The author suggested that the discrepancy may be caused by magnetite formed during the inheriting of the oxygen isotope composition of the precursor mineral hematite. By analogy, the hematite from the samples studied in this paper may well have inherited the δ18O-depleted oxygen isotope composition from its magnetite precursor.

**Fluid sources**

The oxygen isotope compositions of hydrothermal alteration assemblages and high-grade iron ore from the studied deposits are clearly depleted in δ18O by about 5 to 10 per mil when compared to BIF. This depletion clearly underlines the role of hydrothermal fluids in the transformation of BIF to high-grade ore evolved in hydrothermal models described by Powell et al. (1999), Taylor et al. (2001), and Thorne et al. (2004, 2008).

The oxygen isotope compositions of the hydrothermal fluids, while having a wide range for each deposit, indicate two fluid sources (Table 3; Fig. 5). The formation of the magnetite-carbonate and hematite-carbonate alteration assemblages in the presence of low- to moderate-temperature (157°–346°C), saline (15.5–26.0 wt % CaCl2 equiv) fluids, with a range in δ18O values (–3.5 to 3.2‰) suggestive of a meteoric water component to a basinal brine. A mixture of meteoric water with basinal brines is also supported by the results of Taylor et al. (2001) and Thorne et al. (2008) for the Mount Tom Price deposit, where fluid mixing has been suggested to result in the crystallization of microplaty hematite. The involvement of basinal brines is furthermore supported by Thorne et al. (2007a) who studied dolomite from hypogene dolomite-hematite-chlorite-pyrite veins at the Paraburdoo deposit. The dolomite was formed at moderate temperature (150°–190°C) with high Cl/Br, low Na/Br, and moderate-salinity, evolved brines (19.9–22.0 wt % CaCl2 equiv), having an elevated base metal content (Grandia et al., 2003).

The proximal martite-microplaty hematite and high-grade martite-microplaty hematite assemblages also formed by the interaction of δ18O-depleted, meteoric fluids at low temperatures (Table 2; Fig. 5) with the hypogene hematite-carbonate alteration assemblages. This supports the studies of Taylor et al. (2001) and Thorne et al. (2008), which evoke a stage of deep meteoric fluid circulation in the formation of high-grade iron ore.
At the Mount Tom Price deposit, low- to moderate-temperature (~115°–253°C) Ca-Mg-rich brines (22.8–25.9 wt % CaCl₂ equiv that interacted with BIF, forming the talc-bearing ore have δ¹⁸O values of 1.8 to 9.0 per mil, suggesting the involvement of basinal brines with a seawater origin but not excluding a possible contribution of meteoric water. The data supports the notion that the formation of talc-bearing ore is unrelated to the main ore-forming event and that hydrothermal fluid flow within the deposits was multistage and protracted. Rasmussen et al. (2007) suggested that ore genesis may have commenced as early as 2.15 Ga., with subsequent hydrothermal remobilization at 2.05, 1.84, 1.67, 1.59, 1.54, 1.48, and 0.85 Ga, a result of basin development and repeated tectonic activity to the south of the Hamersley province. The formation of talc-bearing ore may reflect one of these hydrothermal pulses.

Significance of quartz-hematite veins

Both the δ¹⁸O and the calculated δ¹⁸Ofluid values of hematite from quartz-hematite veins at the Ratty Springs prospect are similar to hematite from high-grade ore at the nearby Paraburdoo deposit (Fig. 8, Table 3). Sepe (2007) studied similar textured quartz-hematite veins as those at the Ratty Springs prospect that postdate hypogene alteration at the 2 Lens orebody of the Paraburdoo deposit. Microthermometry and ion chromatography on primary fluid inclusions within quartz from the quartz-hematite veins indicate that the veins likely formed from mixed high-salinity basinal brines and low-salinity meteoric fluids. Similar relationships were observed by Oliver and Dickens (1999), Powell et al. (1999), and Brown et al. (2004) for patchy hematite-quartz mineralization and associated veins near the Mount Whaleback deposit.

All the studies show a clear spatial relationship of quartz-hematite veins to high-grade iron ore deposits and similar fluid geochemistry to the hydrothermal fluids responsible for the transformation of BIF to high-grade iron ore. The timing of the veins relative to iron ore formation indicates that the quartz-hematite veins are related to iron ore formation during the extensional collapse of the Ophthalmian orogeny (~2200 Ma). Recent work by Müller et al. (2005) suggest that the iron ore formation, and therefore related quartz-hematite veins in the Hamersley province, is unrelated to the Ophthalmian orogeny and is likely to have formed during a 2.050 to 2.000 m.y. period of continental extension. Dalstra (2005) indicates that the quartz-hematite veins that postdate high-grade iron ore formation at the Paraburdoo deposit, quartz-hematite veins the Ratty Springs deposit are likely to have formed during reactivation of normal faults during the Capricorn orogeny (~1780 Ma; Tyler and Thorne, 1990).

Xenotime and monazite in high-grade hematite orebodies from the Mount Tom Price deposit have recorded repeated episodes of hydrothermal fluid flow that commenced at 2.15 Ga and continued intermittently for more than 1 Gyr (Rasmussen et al., 2007). It is likely that the quartz-hematite veins were formed during a period of hydrothermal fluid flow and focused within similar structural settings as the hydrothermal fluids that resulted in the transformation of BIF to high-grade iron ore.

Conclusions

The study of the oxygen isotope geochemistry of magnetite and hematite during the transition from BIF to high-grade ore in the central Hamersley province provides support for the hydrothermal formation of the deposit-like anachronism of both basinal brines and meteoric waters in their formation. This study has shown that the δ¹⁸O depletion in high-grade ore formed during the early hydrothermal alteration through the interaction of low- to moderate-temperature, saline brines mixing with the most negative δ¹⁸O values proximal to fault zones, and increasingly more positive δ¹⁸O values at the edge of the orebody. The values of δ¹⁸Ofluid support hydrothermal models that evoke both basinal brines and heated meteoric fluids for the transformation of BIF to high-grade iron ore. The results of this study are similar to those of other studies elsewhere on the oxygen isotope geochemistry of hematite from high-grade iron ore deposits (Gutzmer et al., 2006), suggesting that processes similar to those proposed in these studies are of general importance to the genesis of this ore type.

Acknowledgments

We would like to thank Rio Tinto Exploration and Rio Tinto Iron Ore, which allowed access and provided logistical support. Warren Thorn would particularly like to acknowledge Hilke Dalstra, Raul Bitencourt, Matt Ransdlen, Mike Stone, Pat MacKenzie, and Wayne Zarb for their invaluable assistance during field work and for provision of detailed geologic maps and sections without which the spatial interpretations made here would not have been possible. This research was supported by an Australian Research Council Linkage grant, Rio Tinto Exploration, and the University of Western Australia.

REFERENCES


OXYGEN ISOTOPE COMPOSITIONS OF IRON ORE DEPOSITS OF THE CENTRAL HAMERSLEY PROVINCE, W.A.


CHAPTER 4: STRUCTURAL CONTROL, HYDROTHERMAL ALTERATION ZONATION AND FLUID CHEMISTRY OF CONCEALED, BIF-HOSTED HIGH-GRADE 4EE IRON ORE BODY AT THE PARABURDOO 4E DEPOSIT, HAMERSLEY PROVINCE, WESTERN AUSTRALIA.

Statement of candidature contribution:

As first author, I conducted all fluid inclusion and laser ICP-Ms analyses, writing of the paper, and drafting of all figures. The role of co-authors is as follows:

- Steffen Hagemann contributed to the organisation, editorial, and scientific input and scrutiny of the paper. He provided scientific guidance on the fluid inclusion work.
- Hilke Dalstra provided the structural reconstruction of the Paraburdoo ranges and Figures 2, 4 and 5.
- David Banks assisted in the LA-ICPMS and crush-leach analysis at the Fluid Inclusion Laser Ablation ICP-MS Laboratory at the University of Leeds. He also contributed to the interpretation of results.

This paper provides the most comprehensive study of an iron ore deposit in the Pilbara published to date and builds significantly on previous research on the Paraburdoo deposit (Chapter 4). Based on thorough geological reconstruction of the 4E deposit, the paper introduces several new analytical techniques such as crushed-leach and LA-ICPMS analysis to provide meaningful insights into the ore-forming processes, and builds on previously used techniques (Chapter 5).

Structural reconstruction of the 4E deposit, together with structural mapping, core logging and advance geochemical analysis such as fluid inclusion microthermometry, crushed-leach ion chromatography and LA-ICPMS analysis of fluid inclusions, allowed the development of an integrated structural and hydrothermal model for the deposit.

The paper also provides new insights into the source of hypogene fluids and meteoric fluids responsible for the transformation of BIF to high-grade iron ore. Together with results from the research of oxygen isotopic composition of iron oxides at the Paraburdoo deposit (Chapter 5), the results from this paper indicate that the 4E deposit ultimately formed from a dual hydrothermal system courtesy of ascending basinal brines, possibly from sedimentary basins to the south of the Hamersley province, and descending, cold Paleopreterozoic meteoric water.

This paper was submitted to Economic Geology, and reviewed by Drs Francisci Javier Rios and Joydip Mukhopadhyay. Editorial work was completed by John Slack, Associate editor of Economic Geology. The paper was published in Economic Geology, v.109, p. 1529-1562.
Hamersley Province has become one of the world’s preeminent mining districts since initial mining of the iron ore resources in the 1960s. Deep drilling and mining below the depth of modern weathering at the Mt. Tom Price, Mt. Whaleback, and Paraburdoo deposits resulted in the recognition of carbonate-altered BIF transitional between low-grade BIF and high-grade iron ore. Early studies of the carbonate-altered BIF resulted in supergene-metamorphic models (Morris, 1980, al., 2008).
Fig. 1. Location of selected high-grade iron ore deposits in the Hamersley province, Western Australia, classified by generic model according to Thorne et al. (2008).
1985, 1988; Harmsworth et al., 1990) for the formation of high-grade, martite-microplaty hematite iron deposits, which have been superseded by more recent hypogene-supergene models (Barley et al., 1997, 1999; Martin et al., 1998; Hagemann et al., 1999; Powell et al., 1999; Ridley, 1999; Taylor et al., 2001; Cochrane, 2003; Dastrla and Guedes, 2004; Thorne et al., 2004, 2005, 2005a, b; 2008; Webb et al., 2004; Dalstra, 2005, 2007, 2009; Oliver, 2007).

Research at the Mt. Tom Price deposit (Fig. 1; Barley et al., 1999; Ridley, 1999; Taylor et al., 2001; Cochrane, 2003; Thorne et al., 2004, 2005, 2007b) has described the formation of high-grade, martite-microplaty hematite ore by the multi-stage interaction of hydrothermal fluids with the host BIF. Key to the development of the hydrothermal model was the discovery of hydrothermal alteration assemblages below the depth of modern weathering through deep drilling, primarily at the Southern Ridge and North deposits (Thorne et al., 2005).

At the Mt. Whaleback deposit (Fig. 1), deep drilling at the western end of the main pit into the McRae Shale, below the martite-microplaty ore and along the hanging wall of the Mount Whaleback fault, intersected carbonated-altered shale (Webb et al., 2004). Textural and geochemical evidence from the ferroan-dolomite and ankerite alteration assemblages, and crossing chlorite and carbonate veins within the McRae Shale support an early hypogene stage for the formation of the deposit (Webb et al., 2004).

Deep drilling at the Paraburdoon deposit (Fig. 1) intersected hydrothermally altered BIF, transitional between BIF and high-grade, martite-microplaty ore, both below and lateral to the 4E orebody. Thorne et al. (2007b, 2008) proposed a preliminary multistage hydrothermal-supergene model for the 4E deposit. The first hypogene stage involved alteration by upward movement, along fault zones, of the 170 °C basinal brines into the Dales upward movement, along fault zones, of 170 °C basinal brines (Webb et al., 2004).

Shale support an early hypogene stage for the formation of crosscutting chlorite and carbonate veins within the McRae Shale. Carbonate-altered shale of the Mt. Whaleback fault, intersected carbonate-altered shale the martite-microplaty ore and along the hanging wall of the 4E deposit. The first hypogene stage involved alteration by upward movement, along fault zones, of the 170 °C basinal brines into the Dales upward movement, along fault zones, of 170 °C basinal brines (Webb et al., 2004).

The stratigraphy and tectonic history of the Hamersley province has been well documented (MacLeod, 1966; Trendall and Blockley, 1970; Trendall, 1983; Harmsworth et al., 1990; Tyler and Thorne, 1990; Powell and Horwitz, 1994; Krapež, 1997, 1999; Taylor et al., 2001). Detailed mapping of the southern Pilbara margin by Müller et al. (2005) and Dalstra (2005), together with new age dating of mafic dikes and sills (Müller et al., 2005), has provided significant contributions to the tectonic history of the southern Hamersley province (Fig. 2). Müller et al. (2005) conducted in situ Pb/Pb dating of baddeleyite by sensitive high-resolution ion microprobe (SHRIMP), resolving the ages of two key suites of mafic intrusions and constraining for the first time the tectonic evolution of the Ashburton province and the age and setting of iron ore formation. Mafic sills that were folded during the Ophthalmian orogeny and then cut by the unconformity at the base of the Lower Wyloo Group have ages of ca. 2208 Ma (Fig. 2). A mafic dike swarm that intrudes the Lower Wyloo Group is hydrothermally altered at the Paraburdoon deposit is dated ca. 2008 Ma. These mafic dikes are slightly younger than a new syneruptive 2031 ± 6 Ma SHRIMP U-Pb zircon age for the Lower Wyloo Group. The ages constrain the Ophthalmian orogeny to the period between ca. 2208 and 2031 Ma; the ca. 2008 Ma age for the dike presents a new maximum age for iron ore formation and deposition of the Upper Wyloo Group, thereby linking ore genesis to a period of continental extension at ca. 2050 to 2000 Ma.

The southern Pilbara terrane has undergone three major tectonic events (Fig. 2); (1) the Ophthalmian orogeny (D2, Fig. 2; 2140 ± 50 Ma; Allibert and McCulloch, 1993), (2) the Panhandle orogeny (D3, Fig. 2; ~2000 Ma; Taylor et al., 2001; Müller et al., 2005), and (3) the Capricorn orogeny (D4, Fig. 2; ~2170 Ma; Tyler and Thorne, 1990; Krapež, 1999). Prior to this deformation, recumbent isoclinal folds and pods (D1; Dalstra, 2005) were developed within all BIFs, but are most pronounced in the Marra Mamba Iron Formation. The Ophthalmian orogeny (D2) produced large-scale, E-W-trending folds (F2; Taylor et al., 2001) throughout the province, with large E-W-trending dome and basin folds (e.g., Turner syncline) developed in the west. NW- to N-striking, steeply NE dipping normal faults (E2; Dastrla, 2005) that predated the deposition of the Beasley Quartzite (Fig. 2) are interpreted to have formed the early structural architecture for later fault-focused fluid flow (Southern Bather fault, Mt. Tom Price; Taylor et al., 2001). Dalstra (2005) provided a detailed reconstruction of the southern Hamersley province, demonstrating that extensional normal faults are uniformly associated spatially with high-grade
hematite deposits and, in particular, with faults that link the underlying dolomites of the Wittenoom Formation to the overlying iron formation, thus providing a pathway for mineralizing fluids.

The Panhandle orogeny (D3) produced NW-trending folds that are best developed in the western part of the province and predate intrusion of the numerous NW-trending dolerite dikes (Fig. 2; ca. 2008 Ma; Müller et al., 2005). The F3 folds interact with the earlier E-W-trending Ophthalman folds (F2) to produce a characteristic dome-and-ridge structural pattern (Taylor et al., 2001). Following the Panhandle orogeny and prior to deposition of the Mount McGrath Formation (Fig. 2), reactivation of earlier NW-striking faults, as well as the development of NW-striking, steeply dipping normal faults with throws of up to 2,100 m, formed horst and graben structures along the southern margin of the Hamersley province (Dalstra, 2005). The NW-trending dolerite dikes intruded into these faults and control ore distribution along the Paraburadoo Range.

The Capricorn orogeny (D4) took place before ca. 1770 Ma (Müller et al., 2005) and records the collision between the Pilbara and Yilgarn cratons (Tyler and Thorne, 1990). Deformation is coaxial to the Panhandle orogeny and rotated the southern margin of the Pilbara province south or north over 30° to 60°. Strike-slip reactivation of steep faults (after tilting) occurred during the later stages of the Capricorn orogeny.

The entire Hamersley Group (Fig. 2) is locally eroded on two unconformities (Dalstra, 2005). The first unconformity defines the base of the Lower Wyloo Group (up to 10° discordant) where the Beasley River Quartzite rests on the Turee Creek and Hamersley groups of the Mount Bruce Supergroup (Figs. 2, 3; Taylor et al., 2001; Dalstra, 2005). The second unconformity places the Mount McGrath Formation of the Upper Wyloo Group on rocks of the Fortescue to Lower Wyloo groups at an angle of approximately 10° to 20° (Dalstra, 2005). The presence of mica platy hematite pebbles within the McGrath Formation has been used to indirectly constrain the youngest age for high-grade iron ore formation within the Hamersley province (Morris, 1980; Taylor et al., 2001).

A later set of wide (up to 100 m thick), NE-trending dikes cuts both the Ashburton trough and the Hamersley Group and postdates the Capricorn orogeny. The Channar dike (752 ± 10 Ma; Taylor et al., 2001) cuts through and locally metamorphoses the ore at the Channar deposit.

**Structural Reconstruction of the Paraburadoo Mining Area**

The Paraburadoo Range contains a number of hematite and hematite-goethite deposits, from the 66West deposit in the west to the Channar deposit in the east (Fig. 3). The largest and highest grade iron ore accumulations are the 4West and 4East orebodies of the Paraburadoo and Channar deposits (Fig. 3). Iron ores are hosted mainly by the Brockman Iron Formation, which is tilted 40° to 65° south in the west and less than 10° in the east near Channar.

The Lower Wyloo Group unconformity at the base of the Beasley Quartzite (BQ unconformity) is preserved in fault-bounded blocks along the western part of the Paraburadoo Range and is concordant to slightly discordant with respect to the older iron ore units, e.g., the Joffre Member and Weeli Wooli Formation. The Upper Wyloo Group (Mt. McGrath and Duck Creek formations) unconformably overlies the Hamersley and Lower Wyloo Group rocks at angles of about 10° to 20°. The Mt. McGrath conglomerate south of the Paraburadoo 4West deposit typically contains coarse clasts of low-grade Weeli Wooli BIF and quartzite at the base, grading upward.
into smaller pebbles consisting of microplaty hematite ore. Conglomerates at the base of the Beasley Quartzite are overwhelmingly devoid of ore pebbles but contain abundant clasts of low-grade BIF (Martin et al., 1998; Taylor et al., 2001).

Folds on the scale of meters to less than 1 m generally are the only compressional structures visible in outcrops. At Paraburdoo, the predominant folds trend northwest to west-northwest and, because of the tilting of the Wyloo Group unconformities, are most likely Capricorn (D4) in age. In contrast, faulting is pervasive in the area, with four major sets of faults recognized (Fig. 3; Table 1). Gently N-dipping faults with normal offsets up to several hundred meters are observed throughout the Paraburdoo Range and are largely parallel to the general strike of the iron formations (Table 1; range-parallel flat faults). Individual faults can be traced along strike for up to 12 km. Fault planes commonly are spoon shaped, with terminations at high angles to the bedding of BIFs, and gently west or east dipping, e.g., the 32East fault. Other large faults, such as the 4West, 4East, and 23East basal faults, are oblique to the strike of the range (Table 1; oblique faults), dip gently east or northeast, and also have normal displacements. However, because of the spoon-shaped nature of many faults, it is sometimes difficult to determine if they form a separate set. The third major fault set trends northwest with subvertical to steep northeast dips and a reverse sense of movement (Table 1; steep reverse faults). Examples are the Ratty Springs and 18East faults. Only at Channar are there known NW-trending steeply dipping normal faults (Table 1; steep normal faults).

Late dolerite dikes intruded preferentially along all fault systems in the Hamersley and Lower Wyloo Group, but nowhere intrude into the overlying Upper Wyloo Group.

**Timing**

The 4West and 4East basal faults (oblique faults) are among the oldest structures in the Paraburdoo area and originated prior to deposition of the Lower Wyloo Group (Figs. 2, 3; E2, Dalstra, 2005). They were reactivated during deposition of the Upper Wyloo Group (Fig. 2; E3), but have been inactive since then. Range-parallel flat faults, such as the 32East fault, lack offsets in the Mt. McGrath Formation, but the presence of similar offsets as oblique faults (Table 2) in the Hamersley and Lower Wyloo groups suggests that they are syn-E3 (Fig. 2). These faults were also inactive after this extensional event. A similar timing was proposed (Dalstra, 2005) for the reverse 18East and Ratty Springs faults (steep reverse faults).

NW-trending dikes that limit the areal extent of iron orebodies at Paraburdoo and intrude the 4East and 4West basal faults were active and hydrothermally altered during ore formation (Barley et al., 1999; Dalstra, 2005) but do not intrude the Mt. McGrath Formation. The dikes were emplaced after the Panhandle folding event (Fig. 2; D1 cap; Müller et al., 2005), and dikes, country rock, and ore were deeply eroded before deposition of the Mt. McGrath Formation (Taylor et al., 2001). Thus, the formation of high-grade iron ore took place during or shortly after the emplacement of these ca. 2008 Ma dikes.
In the Channar area, the main ore-controlling faults (Fig. 3: Howies Hole, Channar East and West faults) strike northwest and have moderate to steep northeast or southwest dips and normal displacement. Historic mapping shows that at Channar, subhorizontal to gently S dipping hematite-clast conglomerates of the Mt. McGrath Formation covered the hematite ore. The faults (steep normal faults) have offsets of up to 600 m in Hamersley Group strata, but smaller (dextral) displacement in the Upper Wyloo Group. Their timing is, therefore, constrained to E3, possibly as a reactivation of older E2 structures, with moderate strike-slip reactivation occurring during the Capricorn orogeny (Dalstra, 2005).

### Structural reconstruction of the southern Hamersley province

Back rotation of the hematite-clast conglomerate beds at the basin of the Mt. McGrath Formation toward their assumed near-horizontal depositional setting is required to reconstruct the setting of ore formation at Paraburdoo (e.g., Dalstra, 2005). In the west, near the 4West and 4East deposits, the conglomerate beds dip 40° to 50° south-southeast, indicating that the mineralized geometry was significantly modified during the Capricorn orogeny. The north-northeast rotation of the oldest (E2, E3) ore-controlling, oblique flat faults at 4East and 4West, over a horizontal axis, resulted in a northwest to north-northwest trend with moderately steep (45°–72°) dips to the northeast to east-northeast (Fig. 4, Table 1). Weakly mineralized (only E3) range-parallel faults originally had west-northwest trends and moderate north-northeast dips (Table 2). Back rotation also shows that the steeply NE dipping reverse faults, such as the 18East fault and the Ratty Springs fault, originated as moderately steep, SW-dipping normal faults. At the Channar deposit, the hematite-clast conglomerate beds of the Upper Wyloo Group are subhorizontal, and therefore the required rotation is minimal; the present fault geometry with steeply dipping normal faults (steep normal faults) reflects the setting at the time of mineralization.

Reconstructed long sections through the Paraburdoo Ranges prior to the Capricorn orogeny and to syn-Upper Wyloo Group and Lower Wyloo Group sedimentation are shown in Figure 4A and B, respectively. The depositional setting for the Lower Wyloo Group at Paraburdoo is a series of NW- to NWW-trending half grabens (E2) that formed by NE- to ENE-dipping normal faults (oblique faults and steep normal faults) progressively stepping down to the east-northeast (Fig. 4B). During deposition of the Upper Wyloo Group, this system of half grabens was reactivated, together with the formation of new normal faults having a more westerly trend and

### Table 1. Fault Systems in the Paraburdoo Ranges and Their Details

<table>
<thead>
<tr>
<th>Fault system</th>
<th>Representatives</th>
<th>Present orientation</th>
<th>Timing</th>
<th>Original orientation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Range-parallel flat faults</td>
<td>60West fault, 44West fault, 42East fault, 24East fault</td>
<td>Gently N to NNE dipping</td>
<td>Pre-Mt. McGrath (E3)</td>
<td>Moderately NNE dipping</td>
</tr>
<tr>
<td>Oblique flat faults¹</td>
<td>4West basal fault, 4East basal fault, 23East basal fault, 24East fault</td>
<td>Gently NE to E dipping</td>
<td>Pre-BQQ, reactivated pre-Mt. McGrath (E2-E3)</td>
<td>Moderately NE to ENE dipping</td>
</tr>
<tr>
<td>Steep reverse faults</td>
<td>Ratty Springs fault, 18East fault, 23East fault</td>
<td>Steeply NE dipping</td>
<td>Post-BQQ, pre-Mt. McGrath, late dextral reactivation (E3-F4)</td>
<td>Moderately to steeply SW dipping</td>
</tr>
<tr>
<td>Steep normal faults²</td>
<td>64East fault, Howies Hole fault, Channar East fault</td>
<td>Moderately NE dipping</td>
<td>Pre-BQQ², reactivated pre-Mt. McGrath and late dextral (E2–F4)</td>
<td>Moderately NE dipping, rare SW dips</td>
</tr>
</tbody>
</table>

BRQ = Beasley River Quartzite

¹Control mineralization at 4West, 4East, and 23/24East

²Control mineralization at Channar

### Table 2. Vein Types Identified at the 4E Deposit

<table>
<thead>
<tr>
<th>Vein type</th>
<th>Thickness (mm)</th>
<th>Equilibrium assemblage</th>
<th>Textures</th>
<th>Hydrothermal zonation</th>
</tr>
</thead>
<tbody>
<tr>
<td>V₁</td>
<td>0.50–6</td>
<td>Quartz</td>
<td>Typically massive with locally internal comb textures; common</td>
<td>Prior to hydrothermal alteration and iron ore formation</td>
</tr>
<tr>
<td>V₂</td>
<td>5–60</td>
<td>Dolomite-quartz ± pyrite</td>
<td>Commonly forms matrix-supported breccias that consist of angular and rotated clasts (5–50 mm) of altered wall rock within a dolomite-quartz-pyrite matrix</td>
<td>Distal</td>
</tr>
<tr>
<td>V₃</td>
<td>1–15</td>
<td>Dolomite ± pyrite</td>
<td>Associated with intermediate magnetite-dolomite-hematite-chlorite-pyrite alteration within BIF</td>
<td>Intermediate</td>
</tr>
<tr>
<td>V₄</td>
<td>1–50</td>
<td>Dolomite-chlorite ± pyrite</td>
<td>Associated with intermediate magnetite-dolomite-hematite-chlorite-pyrite alteration</td>
<td>Intermediate</td>
</tr>
<tr>
<td>V₅</td>
<td>20–2,000</td>
<td>Dolomite-hematite ± pyrite</td>
<td>Associated with proximal hematite-dolomite-magnetite-stilpnomelane alteration; pervasive within fault zones</td>
<td>Proximal</td>
</tr>
<tr>
<td>V₆</td>
<td>5–60</td>
<td>Quartz-hematite</td>
<td>Rare late-stage veins within BIF and dolerite</td>
<td>Late stage</td>
</tr>
</tbody>
</table>

In the Channar area, the main ore-controlling faults (Fig. 3: Howies Hole, Channar East and West faults) strike northwest and have moderate to steep northeast or southwest dips and normal displacement. Historic mapping shows that at Channar, subhorizontal to gently S dipping hematite-clast conglomerates of the Mt. McGrath Formation covered the hematite ore. The faults (steep normal faults) have offsets of up to 600 m in Hamersley Group strata, but smaller (dextral) displacement in the Upper Wyloo Group. Their timing is, therefore, constrained to E3, possibly as a reactivation of older E2 structures, with moderate strike-slip reactivation occurring during the Capricorn orogeny (Dalstra, 2005).
Fig. 4. A. Reconstructed long section through the Paraburdoo Ranges, syn-upper Wyloo Group deposition, showing the mineralized fault geometry and interpreted position of the orebodies (from Dalstra, 2005). B. Reconstructed long section through the Paraburdoo Ranges, syn-lower Wyloo Group deposition, pre-iron ore mineralization (from Dalstra, 2005). C. Stereonet (lower-hemisphere, equal-area) reconstruction of fault systems from the Paraburdoo Ranges. Gray dot (50) marks the pole to bedding in the upper Wyloo Group. Back rotation of the pole to the horizontal position and equal rotation of poles to major fault systems (OF, RPF, and SRF) define the primary orientation of these faults (OF, RPF, and SRF). Normal faults (SNF) in the Channar area occur in subhorizontal stratigraphy and do not require any back-rotation to define their primary orientations. This figure illustrates that all fault systems at Paraburdoo originated as moderately to steeply dipping normal faults and were later rotated to their presently flat or steep reverse orientations.
dips either steeply to the south-southwest or north-northeast (e.g., Table 1; range-parallel faults, steep normal faults, steep reverse faults; E3). The resulting geometry was a series of WNW- to NNW-trending grabens or half grabens (Fig. 4A). Hematite orebodies formed (Fig. 4A) early during the E3 event and were actively eroding during deposition of the Upper Wyloo Group, forming the hematite-clast conglomerates. Orebodies at the 4West, 4East, 23East, and 64East deposits formed in downthrown blocks at least several hundred meters below the surface, and therefore survived erosion prior to deposition of the Mt. McGrath Formation. Only the 11West and Channar orebodies formed on horsts, and were nearly completely stripped by erosion during deposition of the Mt. McGrath Formation, as evidenced by the presence of hematite-clast conglomerate near or on top of the ore. Other orebodies may have formed on the Ratty Springs and 18East horst, but these were eroded, their ores having been partly deposited in the hematite-clast conglomerates that flank these horsts.

**Geology of the 4EE Orebody**

The 4EE orebody has an inferred resource of >200 Mt at 63.5 wt % Fe of high-grade (>63 wt % Fe) iron ore that lies below the 4E open pit (Fig. 5). This orebody is concealed from the present topographic surface by 150 to 300 m of barren BIF, dolerite, and shale of the Yandicoogina Shale Member and Weeli Wooli Formation, and is located completely below the regional water table. Mapping within the 4East Main pit (4EMP) and 4East North pit (4E North), logging of diamond drill cores, and previous geologic interpretations were used to create cross sections (Fig. 6A, B) through the 4EE orebody. The 4EE orebody (>63 wt % Fe) forms a southerly (30°–40°) dipping sheet, parallel to the dip of strata of the Joffre and Dales Gorge members that host the orebody (Fig. 6). The deposit is bounded at depth by the 4East basal fault and the 4East basal fault dolerite, which intrudes the fault (Fig. 6A, B).

**Hypogene Hydrothermal Alteration Assemblages**

The transition from BIF (30–35 wt % Fe) to high-grade (>63 wt % Fe) iron ore in the 4EE orebody was investigated in 62 petrographic samples taken from 13 diamond drill holes (Fig. 5). Selected samples were examined using the scanning electron microscope at the Centre for Microscopy and Microanalysis (University of Western Australia) to determine semi-quantitative carbonate composition and document textures of oxide, carbonate, silicate, and sulfide minerals. Samples taken from below the depth of weathering (Fig. 6A, B) that contain hydrothermal alteration minerals such as iron silicates, ankerite, and microplaty hematite are interpreted to have formed through hypogene processes (cf. Barley et al., 1999; Taylor
Fig. 6. Section interpretations (A–B and C–D, Fig. 5) of the 4E orebody displaying diamond holes, structure, and stratigraphy and hypogene alteration assemblages. Inset of key core sample from diamond drill hole DD064EMP030 showing transition from low-grade BIF to high-grade iron ore.
et al., 2001; Thorne et al., 2004, 2008). Samples taken from within the weathering profile containing hydrous iron oxides such as goethite and limonite are considered supergene in nature (Morris, 1980, 1985; Barley et al., 1999; Thorne et al., 2008).

Hydrothermal alteration produced a laterally extensive zone below, lateral to, and above the high-grade iron ore (Fig. 6; DD064EMP030). The alteration zone comprises three subzones (1) distal magnetite-quartz-dolomite-stilpnomelane-hematite ± pyrite, (2) intermediate magnetite-dolomite-hematite-chlorite-stilpnomelane, and (3) proximal hematite-dolomite-chlorite ± pyrite ± magnetite (Figs. 7, 8). Hydrothermal alteration is restricted to BIF layers; shale layers preserve original mineralogy and textures. Hydrothermal alteration of dolerite within the 4East basal fault (Fig. 6) produced two zones: (1) intermediate chlorite-tremolite and (2) proximal chlorite-dolomite-talc-tremolite, with the intensity of alteration increasing toward high-grade iron ore. Regardless of the alteration intensity, the progressive changes in alteration mineralogy, described below, form a consistent pattern.

**Host rocks**

The BIF at the 4EE orebody is laterally comparable to that of the 4E orebody (Thorne et al., 2008) and is characterized by alternating magnetite and chert layers with subordinate carbonate, iron silicates, and pyrite (Figs. 8, 9A). Subhedral to euhedral magnetite (50–250 μm) commonly has anhedral hematite inclusions (1–15 μm). Chert-rich layers primarily contain microcrystalline quartz (1–20 μm; Fig. 9A) and lesser coarse polygonal (50–100 μm) or fibrous quartz intergrown with magnetite. Stilpnomelane forms thin layers (20–70 μm) of finely bladed (20–70 μm) crystals and an intergrown groundmass of pale green to brown crystals. Rare pyrite forms subhedral to euhedral crystals (30–150 μm) along high-grade hematite ore. Regardless of the alteration intensity, the progressive changes in alteration mineralogy, described below, form a consistent pattern.

### Table 1: Paragenetic Alteration Sequence in BIF

<table>
<thead>
<tr>
<th>Zone</th>
<th>Hypogene</th>
<th>Supergene</th>
</tr>
</thead>
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<tr>
<td><strong>IRON OXIDES</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Magnetite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>- euhedral</td>
<td></td>
<td></td>
</tr>
<tr>
<td>- anhedral</td>
<td></td>
<td></td>
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<tr>
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<tr>
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<tr>
<td>- microplaty</td>
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<td></td>
</tr>
<tr>
<td>- goethite</td>
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<td></td>
</tr>
<tr>
<td><strong>CARBONATES</strong></td>
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</tr>
<tr>
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<td></td>
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<tr>
<td>- siderite</td>
<td></td>
<td></td>
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<tr>
<td>- dolomite</td>
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<td></td>
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<tr>
<td><strong>SILICATES</strong></td>
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<td></td>
</tr>
<tr>
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<td></td>
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<tr>
<td>- stilpnomelane</td>
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<td></td>
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<tr>
<td>- chlorite</td>
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<tr>
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<tr>
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<tr>
<td><strong>OTHER</strong></td>
<td></td>
<td></td>
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<tr>
<td>- quartz</td>
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</tbody>
</table>

**Vein types**

- V1 0–5%
- V2 >5–10%
- V3 >10–20%
- V4 >20%

Fig. 7. The paragenetic alteration sequence in BIF at the 4E orebody. Proximity to mineralization increases to the right. Zone widths shown here bear no resemblance to actual widths observed in the field.
layers. The wall rock adjacent to V2 veins is locally brecciated with clasts (5–60 mm) of magnetite-rich mesobands in a matrix of anhedral magnetite, dolomite, stilpnomelane, and hematite.

**Intermediate alteration zone**

The intermediate alteration zone is typically several meters wide but may extend up to 30 m in width, particularly below the 4East basal fault in the footwall of the deposit (Fig. 6A, B), where alteration is spatially associated with flat-lying normal faults. This alteration zone is characterized by the mineral assemblage magnetite-dolomite-hematite-chlorite-pyrite that replaces mainly quartz and magnetite (Fig. 7). Euhedral magnetite shows minor replacement by martite along crystal boundaries, with tabular and anhedral hematite (50–250 μm) displaying limited replacement by magnetite (Fig. 9D). Microplaty hematite (10–50 μm) and anhedral hematite (5–20 μm) crystallized both as individual blades and dense aggregates within the dolomite matrix, and as overgrowths on magnetite. Pyrite occurs as large (100–800 μm) crystals within the dolomite matrix and forms along the contact of dolomite and magnetite layers. Apatite is present as inclusions within magnetite and microplaty hematite, and as anhedral crystals within dolomite crystals. Chlorite primarily forms masses intergrown with corroded magnetite layers and as thin (<200 μm) discontinuous layers in the dolomite matrix (Fig. 9E).

Two vein sets are recognized in the intermediate alteration zone: dolomite ± pyrite (V6; Table 2, Fig. 9C) and dolomite-chlorite ± pyrite (V6; Table 2, Fig. 9C). The V3 and V4 veins are abundant and cut BIF and shale layers, and locally cut the 4E basal fault dolomite. Breccias within the fault zones consist of angular and rotated clasts (5–60 mm) of altered wall rock within a dolomite-chlorite-hematite-pyrite matrix.

**Proximal alteration zone**

The proximal alteration zone is typically several meters wide but may extend up to 15 m in width, particularly below
Fig. 9. Microphotographs showing hypogene alteration mineralogy. A. Unmineralized magnetite (mg) and chert (qtz) microbands. Note microcrystalline dolomite (dol) within chert microband. B. Replacement of quartz mesoband (qtz) and dolomite (dol). C. Dolomite (dol) and stilpnomelane (stp) replacing quartz (qtz) within chert-rich bands. D. Euhedral magnetite (mg) displaying minor replacement of hematite (hem) along crystal boundaries, with anhedral magnetite crystallized as aggregates within the dolomite (dol) matrix. E. Chlorite (chl) forming masses intergrown with magnetite within intermediate alteration zone. F. Euhedral magnetite (mg) is replaced by martite, with microplaty hematite and anhedral hematite (hem) forming dense overgrowths within the dolomite matrix. G. High-grade ore dominated by microplaty hematite (hem), platy hematite, and martite. Goethite (goe) is present as fine-grained groundmass surrounding and locally rimming microplaty hematite. H. Platy hematite (hem) displaying radial patterns within quartz (qtz) vein within dolerite (dor).
the 4EE basal dolerite in the footwall of the deposit (Fig. 6A, B). A hematite-dolomite-magnetite-stilpnomelane alteration assemblage is characteristic (Fig. 7) and reflects the nearly complete replacement of quartz in the host BIF by dolomite, hematite, stilpnomelane, and pyrite. Iron oxides present include microplaty hematite (10–80 μm), platy hematite (100–250 μm), anhedral hematite, and magnetite (Fig. 9F). Epidote that constitutes a friable ore type and forms dense overgrowths on martite and disseminated crystals within the dolomite matrix (Fig. 8). The predominant veins (V5; Table 2; Fig. 8D) in the proximal alteration zone are filled with dolomite-hematite ± pyrite that cuts and parallels BIF layers. Wall rock adjacent to the V5 veins is locally brecciated with clasts (5–60 mm) of hematite-rich layers in a matrix of hematite, dolomite, and stilpnomelane (Fig. 8D).

Quartz-hematite veins (V6, Table 2; Fig. 8E) contain large (100–2,000 μm) platy and tabular hematite and quartz that typically displays comb textures. The veins have sharp contacts, lack visible wall-rock alteration, and cut primary layers within BIF and hydrothermally altered dolerite, thus indicating that their formation postdates hydrothermal alteration of BIF and dolerite that produced the high-grade iron ore.

Supergene high-grade hematite ore zone in the 4EE orebody

The high-grade hematite ore zone is up to several hundred meters wide and has a hematite ± goethite assemblage. Ore is dominated by microplaty hematite (10–80 μm), platy hematite (50–250 μm), and anhedral hematite (5–50 μm), together with lesser amounts of martite (50–200 μm; Figs. 7, SF, 9G). The hematite ± goethite assemblage reflects the removal of chlorite, dolomite, and most apatite from the proximal alteration zone. Goethite in the orebody is spatially related to the modern topographic surface and decreases modally with depth. Goethite also is more prevalent along faults and fractures that cut and parallel BIF layers. Wall rock adjacent to the V5 veins is locally brecciated with clasts (5–60 mm) of hematite-rich layers in a matrix of hematite, dolomite, and stilpnomelane (Fig. 8D).

Intermediate alteration zone

The intermediate alteration zone is typically several meters wide but may extend up to 10 m (Fig. 9H). The mineral assemblage chlorite-tremolite with traces of leucoxene and epidote is characteristic. Relict primary igneous minerals of the dolerite remain, including orthopyroxene, clinopyroxene, quartz, and plagioclase. Chlorite (diabasite) and epidote form fibrous aggregates that alter orthopyroxene and plagioclase; leucoxene occurs as an alteration product of titanium-bearing minerals including titanite, perovskite, and leucoxene. Tremolite and chlorite alter amphibole. Epidote forms euhedral to subhedral crystals that replace clinopyroxene.

Proximal alteration zone

The proximal alteration zone is generally 4 to 5 m in width and has a typical assemblage of chlorite-dolomite-talc-tremolite with traces of leucoxene, epidote, and hematite. Relict primary chalcopyrite, plagioclase, and quartz collectively make up <15 vol % of the dolerite. Talc and chlorite form pseudomorphs after feldspars. Weak carbonate alteration associated with V4 and V5 veins shows that dolomite as small (20–60 μm; Table 2), euhedral to subhedral crystals disseminated within the wall rock.

Supergene alteration zone

The supergene alteration zone in the 4E deposit rarely extends to depths of approximately 100 m below the modern land surface. Above this depth, the dolerite is completely weathered to pink kaolinite-rich clays. Within the 4EE orebody, supergene alteration of the dolerite dikes is only present along fault zones in the upper portions of the orebody (e.g., DDH46; Fig. 8B). Here, alteration is restricted to <1 m at the contact between high-grade ore and the dolerite.

Fluid Chemistry of Hydrothermal Alteration Zones

Analytical methods

Microthermometric measurements were conducted at the Centre for Exploration Targeting at the University of Western Australia using the fully automated Linkam THMSG 600 heating and freezing stage with a TMS 93 temperature programmer. The stage was calibrated with synthetic inclusions supplied by SynFlinc. The liquid nitrogen pump (LNP) cooling system operates directly from a 2-1 dewar of unpressurized liquid nitrogen. Measurements were collected...
by presetting temperature profiles that comprise a sequence of stepwise consecutive heating or cooling rates. In addition to the temperature profile, the cycling technique (Goldstein and Reynolds, 1994) was used to obtain accurate temperatures for high- and low-temperature phase transitions. The precision for microthermometric measurements was ±0.2°C for temperatures below 30°C and ±2°C for temperatures above 30°C. The accuracy for the ice melting temperature of 0.0°C was 100%.

For crush-leached ion-chromatography, 109 samples were taken from the Mt. Tom Price, Paraburdoo, and Channar deposits as part of a regional sampling program (Fig. 1; Thorne et al., 2007b). Samples were obtained from hydrothermal alteration assemblages and high-grade iron ore as described by Thorne et al. (2004, 2007b, 2008). Samples from the 4E deposit for leachate analysis were chosen from those samples studied previously by microthermometry. Chemical analysis of fluid inclusions in quartz, carbonate, and hematite was carried out on samples studied by microthermometry at the University of Leeds using the bulk crush-leached method of Banks et al. (2000). Quartz, carbonate, and hematite samples were crushed to between 1 and 2 mm, heated in 18.2 MΩ water to almost boiling, and washed several times with similar-quality water, and the procedure was repeated. Dried samples between 0.5 and 1 g were crushed to a fine powder in an agate pestle and mortar and transferred to a sample container, and 5 to 6 ml of 18.2 MΩ water was added to redissolve the dried salts. Prior to analysis, the samples were filtered through a 0.2-μm nylon filter. The anions Cl and Br were analyzed by ion chromatography; Na, K, and Li were analyzed by an emission spectroscopy (LaCl3) was added to give a concentration of 200 ppm La in each sample to prevent redissolve the dried salts. Prior to analysis, the samples were sent the fluid inclusion contents and not host mineral concentrations, only spectra having coincident Na and other cation peaks were processed. Inclusion-free quartz was analyzed as a check for possible interference and produced no signal.

Fluid inclusion microscopy and microthermometry

Sample locations: Eleven 100- to 200-μm-thick, doubly polished sections of quartz and carbonate were examined petrographically and eight were selected for microthermometric analysis. The samples chosen are from V2 to V6 veins (Table 2) hosted in distal, intermediate, and proximal hydrothermal alteration zones (Figs. 6–9).

Nature of fluid inclusions: Fluid inclusions were measured in quartz from V2 and V6 veins and dolomite from V3, V4, and V5 veins (Table 2). The sequence of fluid entrapment was determined from the distribution of inclusion assemblages in veins of different paragenetic stages (Table 2; Figs. 7–9, 10, 11).

Fluid inclusions within dolomite from V3, V4, and V5 veins and in quartz from V2 veins (Table 2) are isolated or form three-dimensional clusters. Dolomite crystals are typically saturated with fluid inclusions that form random groupings; importantly, dolomite is largely devoid of fluid inclusion trails that cut internal or crystal boundaries. Quartz crystals typically contain fewer inclusions than found in dolomite crystals. These fluid inclusions in both quartz and dolomite crystals are interpreted as being primary as they form three-dimensional clusters and isolated inclusions (Roedder, 1984; Wilkinson, 2000).

Petrographic observation of fluid inclusions within quartz from V2 veins shows inclusions that form three-dimensional clusters and/or internal trails (Fig. 12A). These are interpreted as pseudosecondary inclusions based on the criteria of Roedder (1984). Fluid inclusions present as trails that cut crystal boundaries (Fig. 12B) are interpreted as secondary inclusions.

Two types of fluid inclusions were distinguished according to the number, nature, and volume proportions of phases present at room temperature (Fig. 12A): aqueous two-phase (L-V) inclusions and aqueous three-phase (L-V-S) inclusions with halite daughter crystals. Aqueous two-phase inclusions occur as primary, pseudosecondary, and secondary inclusions. Aqueous three-phase inclusions occur as primary inclusions. Both aqueous inclusion types have a high liquid/vapor ratio (avg 85% liquid, 15% vapor) and occur as irregularly shaped to rounded inclusions in quartz and dolomite crystals.

Petrography and all measurements of single inclusions were carried out on fluid inclusion assemblages, defined as the most finely discriminated fluid inclusion trapping event that can be identified based on petrography (Goldstein and Reynolds, 1994). Assemblages contain both aqueous two-phase (L-V) inclusions and aqueous three-phase (L-V-S) inclusions. Two main types of assemblages are found in both the...
Table 3. Microthermometric Results of Primary, Pseudosecondary, and Secondary Fluid Inclusion Assemblages from the 4E Deposit

<table>
<thead>
<tr>
<th>Hypogene alteration zone</th>
<th>Vein type</th>
<th>Mineral type studied</th>
<th>Vein Mineral inclusion assemblage Phase(s)</th>
<th>Tmice (°C) Range</th>
<th>TmTOT(L) (°C) Range</th>
<th>Tm-halite (°C) Range</th>
<th>Salinity (wt % NaCl/CaCl₂ equiv) Range</th>
<th>Density (g/cc) Range</th>
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</thead>
<tbody>
<tr>
<td>Distal: Synchronous with ore formation</td>
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<tr>
<td>Magnetite-quartz</td>
<td>PS 2 L-V 2</td>
<td>Magnetite-quartz</td>
<td>V2 Quartz PS 1 L-V 1</td>
<td>-39.1 ± 1.2 -40.3 to -38.0</td>
<td>113 ± 9 103 to 120</td>
<td>27.8 ± 0.2 27.6 to 28.1</td>
<td>1.20 ± 0.01 1.20 to 1.21</td>
<td></td>
</tr>
<tr>
<td>Stilpnomelane-hematite-pyrite</td>
<td>PS 3 L-V 2</td>
<td>Stilpnomelane-hematite-pyrite</td>
<td>V2 Quartz PS 2 L-V 2</td>
<td>-36.7 ± 0.1 -36.8 to -36.6</td>
<td>119 ± 3 117 to 122</td>
<td>27.3 ± 0.1 27.3 to 27.4</td>
<td>1.18 ± 0.02 1.17 to 1.19</td>
<td></td>
</tr>
<tr>
<td>Magnetite-quartz-dolomite-stilpnomelane</td>
<td>PS 4 L-V 1</td>
<td>Magnetite-quartz-dolomite-stilpnomelane</td>
<td>V2 Quartz PS 3 L-V 3</td>
<td>-36.5</td>
<td>130</td>
<td>27.3</td>
<td>1.19</td>
<td></td>
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<tr>
<td>Synchronous with ore formation</td>
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<tr>
<td>Dolomite</td>
<td>P 10 L-V 10</td>
<td>Dolomite</td>
<td>V3 Dolomite P 10 L-V 10</td>
<td>-7.1 ± 4.3 -15.1 to -3.5</td>
<td>166 ± 30 151 to 187</td>
<td>10.0 ± 4.9 5.6 to 18.7</td>
<td>0.97 ± 0.04 0.93 to 1.03</td>
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<td>Magnetite-dolomite-chlorite-quartz-stilpnomelane</td>
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<td>Magnetite-dolomite-chlorite-quartz-stilpnomelane</td>
<td>V4 Dolomite P 11 L-V 1</td>
<td>-4.1</td>
<td>163</td>
<td>6.52</td>
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<tr>
<td>Magnetite</td>
<td>P 12 L-V 2</td>
<td>Magnetite</td>
<td>V5 Dolomite P 21 L-V 4</td>
<td>-15.9 ± 1.7 -17.4 to -13.5</td>
<td>136 ± 3 135 to 141</td>
<td>19.3 ± 1.4 17.3 to 20.5</td>
<td>1.06 ± 0.01 1.04 to 1.07</td>
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<tr>
<td>Chlorite-quartz-stilpnomelane</td>
<td>P 13 L-V 3</td>
<td>Chlorite-quartz-stilpnomelane</td>
<td>V5 Dolomite P 22 L-V 7</td>
<td>-15.2 ± 2.0 -17.2 to -12.4</td>
<td>133 ± 3 130 to 137</td>
<td>18.7 ± 1.7 16.3 to 20.3</td>
<td>1.06 ± 0.02 1.04 to 1.07</td>
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</tr>
<tr>
<td>Magnetite</td>
<td>P 14 L-V 4</td>
<td>Magnetite</td>
<td>V5 Dolomite P 23 L-V 6</td>
<td>-16.7 ± 0.8 -17.9 to -15.9</td>
<td>130 ± 9 127 to 132</td>
<td>20.1 ± 0.6 19.3 to 20.9</td>
<td>1.07 ± 0.01 1.06 to 1.09</td>
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<tr>
<td>Chlorite</td>
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<td>Chlorite</td>
<td>V5 Dolomite P 24 L-V 9</td>
<td>-14.8 ± 2.1 -17.6 to -9.9</td>
<td>131 ± 6 119 to 141</td>
<td>26.6 ± 4.6 17.5 to 30.1</td>
<td>1.19 ± 0.05 1.09 to 1.24</td>
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</tr>
<tr>
<td>Chlorite-quartz-hematite</td>
<td>P 16 L-V 1</td>
<td>Chlorite-quartz-hematite</td>
<td>V5 Dolomite P 25 L-V 11</td>
<td>-10.2 ± 1.6 -14.8 to -9.3</td>
<td>128 ± 2 125 to 133</td>
<td>14.1 ± 1.5 13.2 to 18.4</td>
<td>1.02 ± 0.01 1.02 to 1.06</td>
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<tr>
<td>Magnetite-quartz-hematite-chlorite-quartz-stilpnomelane</td>
<td>P 17 L-V 2</td>
<td>Magnetite-quartz-hematite-chlorite-quartz-stilpnomelane</td>
<td>V5 Dolomite P 26 L-V 14</td>
<td>-15.2 ± 2.8 -19.8 to -9.2</td>
<td>131 ± 4 125 to 140</td>
<td>18.6 ± 2.5 13.1 to 22.2</td>
<td>1.06 ± 0.02 1.01 to 1.10</td>
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<tr>
<td>Chlorite-quartz-hematite-stilpnomelane</td>
<td>P 18 L-V 3</td>
<td>Chlorite-quartz-hematite-stilpnomelane</td>
<td>V5 Dolomite P 27 L-V 17</td>
<td>-17.9 ± 0.8 -18.8 to -16.9</td>
<td>130 ± 9 127 to 132</td>
<td>20.5 ± 0.6 19.7 to 21.6</td>
<td>1.07 ± 0.01 1.05 to 1.09</td>
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</tbody>
</table>

**Notes:**
- P = primary,
- PS = pseudosecondary
Fig. 10. Core samples, fluid inclusion samples, and fluid inclusion maps showing analyzed fluid inclusion assemblages. 
A. Diamond drill core displaying V2 vein and sampling location. B. Quartz sample showing mapping of fluid inclusions. 
C. Inset displaying fluid inclusion sample map with assemblage number, salinity, and homogenization temperature and final halite melting. D. Diamond drill core displaying V3 vein and sampling location. E. Sketch of diamond drill core showing mineralogy. F. Dolomite sample showing mapping of fluid inclusions. G. Inset displaying fluid inclusion sample map with assemblage number, salinity, and homogenization temperature and final halite melting. Full microthermometry data are presented in Table 3. Abbreviations: Chl = chlorite, Cht = chert, Dol = dolomite, Mag = magnetite, Qtz = quartz.
Fig. 11. Core sample, fluid inclusion samples, and fluid inclusion maps showing analyzed fluid inclusion assemblages. A. Diamond drill core displaying V; vein and sampling location. B. Fluid inclusion sample displaying areas used for analysis. C. Photomicrograph of area A, assemblage 21 in dolomite crystal showing primary two-phase (L-V) and three-phase (L-V-S) inclusions. D. Sketch of photomicrograph salinity, homogenization temperature and final halite melting. E. Photomicrograph of area B, assemblage 22 in dolomite crystal showing primary two-phase (L-V) and three-phase (L-V-S) inclusions. F. Sketch of photomicrograph salinity, homogenization temperature, and final halite melting. G. Photomicrograph of area C, assemblage 22 in dolomite crystal showing primary two-phase (L-V) and three-phase (L-V-S) inclusions. H. Sketch of photomicrograph salinity, homogenization temperature, and final halite melting. Full microthermometry data are presented in Table 3. Abbreviations: Dol = dolomite, Hem = hematite.
dolomite and quartz crystals. The first type contains less than four inclusions and forms random tightly grouped clusters of both two- and three-phase inclusions (Fig. 10A; assemblage A4). The second type of assemblage contains larger groups of two- and three-phase inclusions, typically between 5 and 15 (Fig. 10A; assemblage A6).

Results of microthermometry: Final ice melting (Tm ice), total homogenization (Th TOT(L)), and halite melting temperature (Tmhalite), salinity (wt % NaCl and CaCl2 equiv), and density (g/cc) are provided for each assemblage (Table 3) and displayed in Figures 10 and 11. The eutectic temperature of H2O-NaCl (–21.2°C) was used to discriminate between NaCl and CaCl2 systems.

Salinity, bulk composition, and bulk density of all fluid inclusion types were calculated using the MacFluinCor program (Brown and Hagemann, 1995). Equations of state by Bodnar and Vityk (1994) were applied for the H2O-NaCl-KCl system, and by Zhang and Frantz (1987) for the H2O-CaCl2-NaCl-KCl system.

The highest temperature measured in an assemblage of aqueous two-phase (L-V) and three-phase (L-V-S) inclusions is 360°C (Fig. 10; Table 3: assemblage 26). Fluid inclusion assemblages hosted in V2 veins show differences of approximately 20° to 60°C in average homogenization temperatures between two-phase (L-V) and three-phase (L-V-S) inclusions, whereas fluid inclusion assemblages trapped in V2 to V5 veins show a difference of 110° to 190°C (Figs. 10, 11).

Similarly, the salinity range of primary fluid inclusions occurring within one assemblage displays a clear difference between those trapped in V2 veins and those present in later V3 to V5 veins. Assemblages (A1–A6; Figs. 10A, 13; Table 3) of aqueous two-phase and three-phase inclusions in V2 veins have small salinity variations of ±2.5 wt % CaCl2 equiv, whereas aqueous two-phase and three-phase inclusions in V3 to V5 veins show large salinity variations of up to ±35 wt % NaCl equiv (Figs. 10, 11; Table 3: assemblages 13, 15) within one assemblage of aqueous inclusions having similar liquid/vapor ratios (0.10–0.20). The secondary aqueous (L-V) inclusion assemblages trapped in V2 veins display similar temperature and salinity ranges as the aqueous (L-V) inclusions trapped in V5 and V6 veins, but lack aqueous three-phase (L-V-S) inclusions.

Interpretation of fluid inclusion data: Three major processes can explain heterogeneous fluid chemistry and trapping conditions in fluid inclusion assemblages (Goldstein and Reynolds, 1994). First is the cooling of a homogenous fluid (Roedder, 1984; Diamond, 1990) that produces inclusions that have trapped different proportions of liquid and vapor at differing homogenization temperatures. Second is the boiling of a homogeneous fluid that results in the trapping of two end-member fluids (Shepherd et al., 1985) and in a series of inclusions that have trapped different proportions of liquid and vapor at similar homogenization temperatures. Third is the heterogeneous trapping of two or more originally partly miscible, homogeneous fluids as a result of partial mixing (Anderson et al., 1992) that results in the trapping of inclusions with an assemblage with different homogenization temperatures and salinity.

No petrographic or microthermometric evidence exists for fluid immiscibility (unmixing) of an aqueous fluid within fluid inclusion assemblages trapped in veins at the 4EE orebody. Single fluid inclusion assemblages trapped within V3, V4, V5, and V6 veins display large variations of salinity and temperatures of homogenization/final halite melting (Figs. 10, 11, 13). The fluid inclusion assemblages are typified by (1) a small number of low-temperature (120°–180°C) and medium-salinity (10–15 wt % NaCl equiv) inclusions, (2) a small number of high-temperature (250°–350°C) and high-salinity (30–40 wt % CaCl2 equiv) inclusions, and (3) a large number of moderate-temperature (150°–250°C) and high-salinity (20–30 wt % NaCl equiv) inclusions. Petrography and microthermometric data suggest the heterogeneous mixing of two distinct fluids. One fluid has low temperatures and salinities (250°–350°C, 30–40 wt % CaCl2 equiv, respectively). The second fluid has higher temperatures and salinities (250°–350°C, 30–40 wt % CaCl2 equiv, respectively). Similar examples of heterogeneous mixing of two fluids are uncommon but have been documented in the Mt. Tom Price
Fig. 13. Equivalent wt % NaCl and CaCl₂ versus homogenization temperature (°C) data for fluid inclusions trapped in dolomite and quartz. A. V₂ pseudosecondary and secondary fluid inclusion assemblages. B. V₃/V₄ primary fluid inclusion assemblages. A small population of Ca-rich inclusions with salinities >21% is present. C. V₅ fluid inclusion assemblages. D. V₆ primary fluid inclusion assemblages. E. All fluid inclusion assemblages, V₂ to V₆.
iron deposit (Hagemann et al., 1999) and the Serra Norte iron deposit (Figueiredo e Silva et al., 2013), thus suggesting that fluid mixing is important in the formation of BIF-hosted, high-grade iron (>63 wt % Fe) deposits, both in the Hamersley province and the Carajás province.

The lack of evidence for fluid boiling suggests that the homogenization temperatures of the fluid inclusion assemblages do not represent trapping temperatures (cf. Pichavant et al., 1982; Roedder, 1984). Therefore, a pressure (temperature) correction is applied to the homogenization temperatures (Table 3) in order to define the trapping temperatures of the fluid inclusions (Roedder, 1984). At the time of vein formation, the host rocks to the veins were most likely at lithostatic pressures of up to 1.5 kb, based on stratigraphic reconstruction of the Hamersley province by Smith et al. (1982). Isochores for the fluid inclusion assemblages from V2 to V6 veins were calculated using the MacFlinCor program (Brown and Hagemann, 1995) to estimate pressure conditions during each stage of vein formation (Fig. 14). A temperature correction of 55°C is required to determine the true trapping conditions of primary fluid inclusion assemblages present in V2 to V6 veins (Fig. 14).

**Crush-leach analyses**

The halogen contents of the fluid inclusions can be used to discriminate different hydrothermal fluids and provide information on their origin (Kesler et al., 1995). In the absence of halite as a daughter mineral, Cl and Br are not significantly involved in fluid-rock interactions, to the extent that the Cl/Br ratio of the fluid is drastically changed, thus allowing the use of Cl/Br ratios in fluid inclusions as a guide to the origin of the salinity (Kesler et al., 1995; Banks et al., 2002; Lüders 2002).

**Fig. 14. Isochore plots for salinity data from V2 to V6 fluid inclusion assemblages.**
et al., 2005). The evaporation of seawater to halite saturation does not change the molar Cl/Br ratio of 655; however, as halite precipitation proceeds, the molar Cl/Br ratio decreases to 430 at the end of halite precipitation (Banks et al., 2002).

Further evaporation results in fluids that are even more Br rich (Lüders et al., 2005). In contrast, the dissolution of halite (e.g., from sedimentary evaporite deposits) produces Br-deficient fluids with a large range in molar Cl/Br ratios from 5,000 to 20,000 (Fontes and Matray, 1993).

Results and interpretation: The fluids in this study form two distinct but overlapping populations (Fig. 15; Table 4). The first population (fluid 1) contains fluids from hydrothermal alteration assemblages and high-grade hematite ore. With few exceptions, fluid 1 has Na/Br ratios of 90 to 430, which are lower than that of seawater. Their position indicates that these fluids were derived from seawater that had precipitated halite and that there has been exchange of Na for other cations.

The second population of fluids, present only within high-grade hematite ore (fluid 2), displays a wide range of Cl/Br ratios from 800 to 8,000, the highest of which are likely to result from halite dissolution (cf. Fontes and Matray, 1993). The magnitude of Na loss is much greater in the halite-derived (Cl-rich) fluid (fluid 2) than the high-Br fluid (20–30 wt % CaCl₂ equiv), due to the lower salinity making it easier for the fluid to lose most Na through water-rock interaction. Sodium loss is likely to have been caused by interaction with Ca in plagioclase (Davison and Criss, 1996), possibly within overlying mafic rocks of the Weeli Wooli Formation.

The distribution of Cl/Br ratios can be explained by the mixing of a low-salinity, halite-derived, Cl-rich fluid (i.e., fluid 1), which lost significant Na through extensive wall-rock interaction, with a high-salinity, Br-rich brine (fluid 2), as indicated by the mixing line in Figure 16. This model is supported by microthermometric data from primary fluid inclusion assemblages trapped in quartz and dolomite (Figs. 10, 11).

The mixing line was calculated using salinities of high-salinity fluid 2 (40 wt % CaCl₂ equiv) that we interpret as basinal brine and of low-salinity fluid 1 (5 wt % NaCl equiv), interpreted as modified meteoric water. The data are compatible with formation of the distal and proximal hypogene alteration zones by mixing of a predominantly high salinity brine (fluid 2) with a minor component of the halite-derived fluid (fluid 1). Fluid inclusions in the high-grade ore samples have Cl/Br ratios that overlap those from the hypogene alteration zones, thus demonstrating a large input of halite-derived fluid and suggesting that fluid mixing was important in producing the high-grade iron ore.

Fluid 1 and fluid 2 have a wide range in Na/Li ratios of 20 to 720 (Fig. 17), which is a marked increase in the Li content...
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</tr>
<tr>
<td>HG26</td>
<td>Hematite</td>
<td>359</td>
<td>0.12</td>
<td>n.d.</td>
<td>152</td>
<td>11.9</td>
<td>0.1</td>
<td>355</td>
<td>130</td>
<td>0.42</td>
<td>12.8</td>
<td>-</td>
<td>158</td>
<td>-</td>
<td>Fluid 1</td>
</tr>
<tr>
<td>HG27</td>
<td>Hematite</td>
<td>101</td>
<td>0.39</td>
<td>6.90</td>
<td>42</td>
<td>5.7</td>
<td>0.1</td>
<td>259</td>
<td>108</td>
<td>0.42</td>
<td>7.3</td>
<td>6</td>
<td>579</td>
<td>15</td>
<td>Fluid 1</td>
</tr>
</tbody>
</table>

n.d. = not determined
of the fluids compared to Br-rich brines and halite, the maximum Na/Li ratios of which are 6,500 and 55,000 (Fontes and Matray, 1993). Sources of Li may include Li that resides in micas (Fontes and Matray, 1993; Banks et al., 2002), in evaporites (Thorne et al., 2010; Evans et al., 2013), and in magnesian clays (Moine et al., 1981). It is likely that water-rock reactions interacted with these Li-bearing rocks and produced a marked increase in the Li concentration within both the low and high Cl-Br fluids. There is no evidence of Li-bearing rocks proximal to the 4EE deposit, thus suggesting that both fluids 1 and 2 interacted with Li-bearing rocks distal from the 4EE deposit, possibly within the Wittenoom Formation (Kargel et al., 1996). The ratio of Na, K and Li in fluids 1 and 2 (Table 4) may be more sensitive to fluid-rock interaction. This is evident in samples from Paraburdoo, where Na/Li ratios range from 100 to 500 in both hydrothermal alteration assemblages and high-grade ore.

The compositions of the fluid inclusions point to the presence of two fluids at the time of ore formation. Fluid 1 is interpreted to originate from evaporated seawater that had reached halite saturation. This fluid lost Mg and Na and gained Li and Ca through interactions with Li-bearing rocks and feldspars within rocks of the Wittenoom Formation, and through dolomitization of carbonate successions of this formation (Banks et al., 2002). The δ¹⁸Ofluid values (−3.5 to +3.2‰) of fluid in equilibrium with hematite from hydrothermal alteration assemblages at the Paraburdoo and Mt. Tom Price deposits (Hagemann et al., 1999; Thorne et al., 2009) support this interpretation of a basinal brine origin for fluid 1.

Fluid 2 is interpreted to be modified meteoric water derived from the interaction with halite at the surface or during descent through the stratigraphy, where it lost Na through extensive interaction with wall-rock plagioclase, possibly in overlying mafic rocks of the Weeli Wooli Formation. Low oxygen isotope values for hematite and quartz associated with high-grade iron ore at the Paraburdoo and Mt. Tom Price deposits (Hagemann et al., 1999; Thorne et al., 2009) support this interpretation of a modified meteoric water origin for fluid 2.

**Laser ablation-ICP-MS**

LA-ICP-MS intensity ratios for fluid inclusions were normalized to Na and converted to concentration ratios by...
external calibration against NIST 610 standard reference material. Data reduction (calibration, background and drift corrections, signal integration) was conducted with the Matlab-based SILLS program (Guillong et al., 2008). Because several fluid inclusions decrepitated during microthermometric analysis, individual salinities were not used for the determination of Na concentration; instead, the average salinity of fluid inclusions in a sample was used.

Detection limits vary according to inclusion volume but for most elements are 1 to 100 μg/g. For a detailed description of the LA-ICP-MS analytical protocol, refer to Allan et al. (2005). Due to matrix interference in dolomite samples, Mg2+, Ca2+, Mn2+, and Fe2+/3+ were not analyzed from V3, V4, and V5 veins. The Na/Ca ratios of inclusions were calculated from results of the bulk crush-leach study on samples taken from the same rock samples as those used for the study of fluid inclusions and LA-ICP-MS work (Table 4).

Results of analyses: Analytical results for inclusions from V2, V3, V4, and V5 veins show that Na+, Ca2+, Mg2+, and K+ are the dominant components, constituting an average of 91.8 ± 8.2% (62.0–99.7; 1σ, n = 75) of the entire solute content (Table 5). Due to the analytical error associated with the LA-ICP-MS process and the use of average salinities for each vein type, the hydrothermal fluids are discriminated best by normative plots of element concentrations (Fig. 18).

Quantitative LA-ICP-MS microanalysis of pseudosecondary fluid inclusion assemblages (Table 5) in quartz from dolomite-quartz-pyrite veins (V2; Table 5) yields two distinct populations of data. All assemblages (Figs. 10C, 18A-C) contain inclusions that are largely Ca dominant (population A: Ca > Na > Mg > K) or K dominant (population B: K > Ca > Na > Mg; Fig. 18A-C, Table 5). Fluid inclusions trapped in V2 veins are enriched in Cu relative to Zn (Fig. 18A-C). The maximum Fe concentration in an assemblage (no. 3; Table 5) is 2.8 wt %.

Primary fluid inclusion assemblages (no. 11–17; Table 5; Fig. 10G) in dolomite from dolomite-chlorite veins (V3 and V4; Tables 2, 3) are characterized by low K concentrations and near-equal concentrations of Ca-Mg and Na. The solute chemistry is noteworthy for enrichment in Cu (Fig. 18G, I), Sr (Fig. 18F, G), and minor enrichment in Ba (Fig. 19H) relative to fluid inclusions from V2 and V5 veins (Table 5). Analysis of primary fluid inclusion assemblages in dolomite from dolomite-quartz-pyrite veins (V5; no. 21–23; Table 5) are Ca-Mg dominant (Fig. 18E) and characterized by enrichment in Zn (Fig. 18I) relative to the other inclusion types.
The variance of fluid compositions with different fluid inclusion assemblages within dolomite in V5 veins is displayed in Figure 11C to F. Data for three assemblages (no. 21–23; Table 5) measured within different crystals from a single sample display a solute chemistry that is within experimental error.

Interpretation of LA-ICP-MS data: The presence of monovalent and divalent cations (Ca\(^{2+}\), Mg\(^{2+}\), Na\(^{+}\), K\(^{+}\)) within primary, high-salinity (>20 wt % NaCl equiv) inclusions suggests that the aqueous fluids are complex and have certain similarities to hydrothermal fluids or basinal brines that formed both Mississippi Valley-type (MVT) and Irish-type Zn-Pb-Ag-Ba deposits (Wilkinson, 2000). Fluid inclusions in MVT deposits typically contain Na > Ca >> K, whereas those from iron ore from this study have Ca > Na > K > Mg. The salinities and solute chemistry of the fluids from this study (Tables 3, 5) also are similar to those in Irish-type Zn-Pb-Ag-Ba deposits, where fluid inclusions typically fall into three broad groups: (1) a moderate-salinity (10–15 wt % NaCl-dominant) fluid; (2) a high-salinity (20–25 wt % NaCl-CaCl\(_2\)) fluid; and (3) a low-salinity (5 wt % NaCl-dominant) fluid.

Concentrations of Pb, Mn, and Cu in all fluid inclusions studied at the 4E deposit are comparable to those of sedimentary...
formation brines (cf. Yardley, 2005, Fig. 3) with Zn and Fe in inclusions trapped in V2 veins having elevated contents relative to such brines (Carpenter et al., 1974). Marie et al. (2001) showed that brines that formed MVT deposits generally have intermediate Pb/Zn ratios and Fe/(Pb + Zn) ratios of 0.2 to 1. The Pb/Zn ratios and Fe/(Pb + Zn) ratios for V2 fluid inclusions range from 0.08 to 1.95 and from 0.22 to 23.95, respectively (Table 3), reflecting the high Fe and low Zn concentrations (Fig. 19C) of the fluids. When compared to calculated Fe, Zn, and Pb contents (Fig. 19) of Pb-rich brines from central Mississippi (Kharaka et al., 1987) and southwest Arkansas (Moldovan and Walter, 1992), most samples from the 4E deposit plot near the Fe apex. The Fe-rich fluids of the 4E deposit are comparable to those that formed Fe-rich MVT deposits, whereby Fe enrichment increases with depositional temperature (Marie et al., 2001). Such Fe-rich deposits typically occur on the structural margins of basins that generate hot, acidic brines with low sulfur contents (Marie et al., 2001). Such a structural setting may be analogous to the structural position of the 4E deposit (Fig. 2).

Primary fluid inclusions in dolomite from the V3 and V4 veins are characterized by high Sr concentrations (Fig. 18) up to 1.6 wt % (Table 5). The Sr-rich inclusions are interpreted to record increased solubility of Sr at high temperatures (Honor, 1994) and thermodynamic buffering of the basinal brines with silicate-carbonate (+halide) mineral assemblages (Honor, 1994). The Wittenoom Formation that underlies the 4E deposit contains primary fluid inclusions in some V2 veins and all V5 veins (Tables 2, 5) are similar to those of Sr. The higher Zn contents of primary fluid inclusions in some 4E deposit are comparable to those that formed Fe-rich MVT deposits, whereby Fe enrichment increases with depositional temperature (Marie et al., 2001). Such Fe-rich deposits typically occur on the structural margins of basins that generate hot, acidic brines with low sulfur contents (Marie et al., 2001). Such a structural setting may be analogous to the structural position of the 4E deposit (Fig. 2).

The data presented above suggest that the hydrothermal chemistry of the veins (V2–V5) record different stages in the complex evolution of the enclosing basin. An initial Ca- and K-rich basinal fluid, also enriched in Cu and Fe, was involved...
in hydrothermal alteration of low-grade BIF, producing a magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite assemblage. Continued evolution of the basin saw a change in brine chemistry with brines characterized by low K, near-equal concentrations of Ca, Mg, and Na, and high contents of Cu, Ba, and Sr. These later basal fluids interacted with the earlier hydrothermal alteration assemblage, resulting in the formation of the hematite-dolomite-magnetite-stilpnomelane alteration assemblage.

**Structural, Hydrothermal, and Fluid Chemical Model for the 4EE Deposit**

**Timing and structural setting**

Hydrothermal alteration at the Paraburdoo 4E deposit occurred ca. 2.05 to 2.00 Ga during continental extension (Müller et al., 2005). Enhanced heat flow within the basin to the south of the Paraburdoo Range (Johnson et al., 2011) resulted in the migration of basal brines through the sedimentary successions (Fig. 20A). SHRIMP U-Pb dating of xenotime and monazite intergrown with hematite in high-grade ores at the Mt. Tom Price deposit (Kasmussen et al., 2007) indicates that hydrothermal flow within the southern Hamersley province was likely multistage and may have begun as early as ~2.15 Ga, with subsequent fluid flow, remobilization, and/or mineralization at ~2.05, ~1.68, ~1.59, ~1.54, ~1.47, and ~0.86 Ga.

Near the Paraburdoo deposits, basal brines were focused by low-angle faults (Figs. 2, 20A) where reactivation of the 18E fault and development of the NW-striking, steeply SW-dipping 4E and 4EE normal faults provided the fluid pathways for heated basal brines. We infer that these brines were derived from or interacted with the underlying Witwenoo Formation to react with BIFs of the Dales Gorge and Joffre members. Hydrothermal alteration is temporally constrained by NW-trending dolerite dikes (2008 ± 16 Ma; Müller et al., 2005) that intruded the 4E and 4EE faults prior to this alteration. These dikes are hydrothermally altered and control the distribution and areal extent of iron ore along the Paraburdoo Range.

Supergene alteration at Paraburdoo likely began in the Cretaceous and lasted for at least 50 m.y., and is ongoing today (Vasconcelos, 1998; Taylor et al., 2001; Cochrane, 2003; Thorne et al., 2008). Weathering of shale layers resulted in considerable volume reduction and the formation of kaolinite-rich clay minerals that constitute the main impurity in the presently mined orebody.

**Model**

The timing relationships outlined above, in combination with distinction of hypogene and supergene hydrothermal alteration zones and related geochemical data on the ore fluids, provide constraints on a preliminary structural-hydrothermal-fluid chemical model for the Paraburdoo 4E
deposit, including the concealed 4EE orebody. In our model, a clear distinction is made between progressive stages of hydrothermal alteration and mineralization (stages 1a-c; Fig. 20) that can be distinguished in the transition from low-grade BIF (30–35% Fe) to high-grade iron ore (>63% Fe), and supergene ore (stage 2; Fig. 20) that presently is exploited in the 4E and 4N North open pits.

Stage 1a early hypogene alteration: Initial hypogene alteration (Fig. 20a) at the 4E deposit occurred within the dominant magnetite-chert layers of the Dales Gorge and Joffre members. Low-grade magnetite-chert BIF wall rocks (Fig. 3A) are transformed laterally and vertically into magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite BIF (Fig. 9B, C). This latter assemblage reflects the partial replacement of chert layers by anhedral magnetite, dolomite, stilpnomelane, and euhedral pyrite. Magnetite-rich layers contain both euhedral magnetite and fine-grained anhedral hematite. The distal alteration zone ranges in width from several centimeters to approximately 30 m and forms an alteration halo of up to 600 m from the high-grade hematite orebody. Dolomite-quartz ± pyrite veins (V3) cut and parallel BIF bands. The adjacent wall rock locally is brecciated with clasts of magnetite-rich mesobands in a matrix of anhedral magnetite, dolomite, stilpnomelane, and hematite.

Fluid inclusion evidence suggests that the magnetite-quartz-dolomite-stilpnomelane-hematite ± pyrite alteration involved a warm to hot (160°–255°C), Ca-rich (26.0–31.9 wt % CaCl2 equiv) brine. The hydrothermal brine likely originated from evaporated seawater that had lost Mg and Na, and gained Li and Ca through fluid-rock reactions with Li-bearing rocks and feldspars in volcanlastic rocks within the Wittenoom Formation and through dolomitization of carbonate successions of the Wittenoom Formation, respectively. The presence of two different solute phases, one Ca dominant (Ca > Na > Mg > K) and the other K dominant (K > Ca > Na > Mg), together with similar proportion of metals (Mn, Cu, Fe, Pb, and Ba) within fluid inclusions trapped in V3 quartz veins, suggests that early hypogene alteration resulted from the mixing of multiple pulses of variably saline brines. The difference in solute chemistry of the two fluids is interpreted to reflect interactions of the hypogene fluids with different rock types (carbonates, shale, volcanlastic rocks) within the sedimentary successions of the Wittenoom Formation along their fluid pathway.

Early basal brines are rich in Fe (up to 2.8%), which had equilibrated with evaporites and likely formed on structural margins of developing sedimentary basins to the south of the Hamersley province (Marte et al., 2001; Fig. 20A).

The first incursion of deeply circulating heated (106°–201°C), modified meteoric water is recorded by the presence of low-to-medium-salinity (5.8–9.5 wt % CaCl2 equiv; V4), secondary inclusions that cut primary high-salinity inclusions within quartz from V3 veins (Fig. 42A). The lower temperature and salinity of this hydrothermal fluid are interpreted to represent the dilution of basal brines by heated, modified meteoric water (Fig. 20B-2). The modified meteoric water had lost Na through extensive wall-rock interaction with plagioclase in dolomite of the Weeli Woolf Formation that directly overlies the Joffre and Dales Gorge members.

The incursion of modified meteoric water may have been in response to uplift of the southern Hamersley province (McLennan et al., 2003).

Stage 1b advanced hypogene alteration: Continuing reactions between the ascending hypogene basinal brine and the pre-existing magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite alteration zone produced both the intermediate magnetite-dolomite-hematite-chlorite-pyrite and the proximal hematite-dolomite-stilpnomelane alteration assemblages. Hematite, chlorite, and pyrite were progressively replaced by quartz and magnetite. Remnants of the former alteration are preserved at the periphery of the hydrothermal system (Fig. 20C).

Iron oxides including microplaty (10–80 μm), platy (100–250 μm), and anhedral hematite increasingly replaced magnetite in the intermediate alteration zone during formation of the proximal alteration zone (Fig. 9D-F). The dominant veins in the intermediate alteration zone consist of dolomite ± pyrite (V4) and dolomite-chlorite ± pyrite (V5), and dolomite-hematite ± pyrite (V5) in the proximal alteration zone. All vein types are abundant, cut BIF and shale layers, and locally cut the 4E basal fault dolerite. Breccias are observed within the fault zones, are matrix supported, and consist of angular and rotated clasts of altered wall rock within a dolomite-hematite ± pyrite ± chlorite ± stilpnomelane matrix (Fig. 21C, D).

Fluid inclusions in various stages of evolution in V4, V5, and V5 veins within the intermediate and proximal alteration zones represent the variable mixing of a hot (250°–400°C), high-salinity, Ca-rich (30–40 wt % CaCl2 equiv) basinal brine, and a low-temperature and low-salinity (~5 wt % NaCl equiv) fluid that is interpreted to be modified meteoric water that was heated (~100°–200°C) during its descent into the upper crust (Fig. 20C). Mixing of the two fluids (Figs. 10, 11) resulted in the trapping of fluid inclusion assemblages containing primary two-phase (L-V) and three-phase (L-V-S) fluid inclusions with a wide range of trapping temperatures (up to 200°C). Occurrence of salinity variations of up to 25 wt % NaCl equiv within primary fluid inclusion assemblages further supports this model of heterogeneous mixing of the two end-member fluids. The Δ18O values calculated to be in equilibrium with hematite (Thorne et al., 2009) from the proximal alteration assemblage are ~0.9‰ to ~0.6‰, compared to those
HIGH-GRADE 4EE IRON OREBODY AT THE PARABURDOO 4E DEPOSIT, HAMERSLEY PROVINCE, WESTERN AUSTRALIA

Mt Whaleback

D Stage 1c Late hypogene alteration
2020-2005Ma

- Hypogene 160 to 255°C, (26.8 to 31.9 eq wt. % CaCl₂) brine
  - (i) K-dominated (K>Ca>Na>Mg)
  - (ii) Ca-dominated (Ca>Na>Mg>K) pulses
- Meteoric 105 to 205°C, Na-rich (5.8 to 9.5 eq wt. % NaCl) water

C Stage 1b Advanced hypogene alteration

- Hypogene hot (250 to 400°C), high-salinity (30-40 eq wt. % CaCl₂) brine
  - Ca-dominated [Ca>K>Na]; Sr-rich
- Meteoric (100 to 200°C), low-salinity (<5 eq wt. % NaCl) water

E ~1770 Ma

- Active erosion of horsts, formation of hematite conglomerate
- Capricorn Orogeny (D4) rotation of 4E and 4EE deposits 40-50° SSW

F Stage 2 Supergene alteration
80 Ma-Present

- Descending shallow meteoric fluids (<50°C) in cold crust

Basinal brines
2020-2005Ma

E ~1770 Ma

- Active erosion of horsts, formation of hematite conglomerate
- Capricorn Orogeny (D4) rotation of 4E and 4EE deposits 40-50° SSW

Mt McGrath Formation (Upper Wyloo)
- Proximal Alteration (hematite-dolomite-chlorite-pyrite+magnetite)
- Intermediate Alteration (magnetite-dolomite-hematite-chlorite-quartz-stilpnomelane)
- Distal Alteration (magnetite-quartz-dolomite-stilpnomelane)

Beasley Formation (Lower Wyloo)
- Weeli Wooli Formation
- Joffre Member
- Whaleback Shale Member
- Dales Gorge Member
- Mt Sylvia and McRae Shale Formation
- Wittenoom Formation
- Marra Mamba Formation
- Fortescue Group

N 4EE deposit

A
for magnetite from low-grade BIF of the Dales Gorge and Joffre members that range from 12.7 to 21.7%. The values indicate that hematite within the proximal alteration zone equilibrated with basinal brines that contain a component of meteoric water.

The mixing of basinal brine and modified meteoric water resulted in hydrothermal fluids that are Ca rich (Ca > Na > K) and concentrated in Sr relative to early stage 1a hypogene fluids (Fig. 19F). This increase in the proportion of Sr is interpreted to reflect buffering of the hypogene fluids with Sr-bearing, marine carbonates of the Wittenoom Formation (Figs. 5, 6). NW-trending dikes (ca. 2008 Ma; Müller et al., 2005) limit the areal extent of hydrothermal alteration and are altered to chlorite-tremolite and chlorite-dolomite-talc-tremolite assemblages.

Stage 1c late hypogene alteration: The final stage of hypogene fluid flow is characterized by the downward migration of modified meteoric water (Fig. 20-S). Interaction of low-temperature (<110°C), low-salinity (~5 wt % NaCl equiv), modified meteoric water with the proximal hematite-dolomite-magnetite-stilpnomelane alteration assemblage resulted in the formation of a porous, hematite-apatite assemblage. Hematite from high-grade ore has δ18Owater values of ~9.6 to ~6.0‰ that suggest equilibrium of hematite with meteoric water (Thorne et al., 2009).

Hydrothermal alteration removed dolomite, chlorite, and most of the apatite from precursor BIF. Martite and anhedral hematite replace magnetite and iron silicates, respectively. Intergranular porosity increases significantly to about 10%. Apatite is located predominantly within the intergranular space of maritie mesobands and in matrix minerals such as montonite and seronite. The formation of post-ore quartz-hematite veins (Vb) reflects the continued mixing of warm to hot (200°–415°C) brines with heated meteoric water (100°–200°C). These mixed fluids may correspond with the waning stages of a large hydrothermal system or be part of a separate system unrelated to ore formation.

Uplift and active erosion of horsts along the Paraburadoo Range (Fig. 4; 11 4West and Channar) occurred during the deposition of the Upper Wyloo Group and resulted in the formation of the hematite-clast conglomerate (Fig. 20-7). The 4E orebody at the 4E deposit formed in a downthrown graben, at least several hundred meters below the surface, and therefore survived erosion prior to deposition of the Mt. McGrath Formation. The Capricorn orogeny (D4) rotated the 4E deposit and hematite-clast conglomerates of the Upper Wyloo Group 40° to 50° to the south-southwest (Fig. 20-S).

Stage 2 supergene alteration: Supergene alteration is likely to have begun in the Cretaceous and persisted for at least 80 my, and is ongoing today (Vasconcelos, 1998). Supergene alteration resulted in the destruction of hypogene alteration zones that are only preserved below the depth of modern weathering (Fig. 20-B). Shale bands are reduced in volume by up to 60% by removal of carbonates, oxidation of pyrite to limonite, and replacement by kaolinitic clays. Goethite, which decreases in abundance with depth, commonly forms a fine-grained groundmass surrounding hematite crystals and locally forms microplaty hematite crystals. Stilpnomelane is replaced by anhedral hematite and remnant magnetite is replaced by goethite. Intergranular porosity is estimated to be about 20%. Phosphorus preferentially precipitated in Al-rich goethite and at lithologic contacts such as ore-shale and ore-dolomite contacts. The distribution of phosphorus and presence of goethite within the alteration zone suggest that the formation of this stage involved cool (<100°C), presumably shallow surface fluids.

Discussion

Fluid processes

The Paraburadoo, Mt. Tom Price, and Mt. Whaleback iron ore deposits of the Hamersley province are all characterized by formation involving the interaction of basinal brines with the host BIF (Taylor et al., 2001; Webb et al., 2004; Dalstra, 2005; Thorne et al., 2008). Hypogene basinal brines also have been invoked for the formation of iron ore deposits at Thabazimbi Republic of South Africa (Netshioski, 2002); Magnetically derived fluids are interpreted to be responsible for the formation of the NSE deposit, Carajás, Brazil (Figueiredo e Silva et al., 2008, 2013), and the Koolyanobbing deposit, Australia (Augener and Hagenauf, 2010).

The early-stage hypogene hydrothermal fluids (stage la, Fig. 20) in the North deposit at Mt. Tom Price and the Paraburadoo deposit show distinct differences in salinities and trapping temperatures. At the North deposit, pseudosecondary and secondary fluid inclusions were trapped during cooling of hot (300°–350°C) and cool (170°C) hypogene fluids. Deep circulation of meteoric water is interpreted to postdate the formation of carbonate alteration assemblages. In contrast, stage lb (Fig. 20) primary fluid inclusions at the 4E orebody show a wide range of salinities but a limited range of trapping temperatures; this pattern is interpreted to record the mixing of a hot saline brine with warm, deeply circulating, modified meteoric water during the formation of the carbonate alteration assemblages. The early ingress of heated meteoric water at the deposits may be due to their uplift and erosion during hypogene alteration, ultimately resulting in deposition of the clasts of hematite within the McGrath Formation.

Research at the Thabazimbi deposit supports a structural and hydrothermal model similar to that proposed here for the 4E and North deposits. At Thabazimbi, warm (160°C) highly saline, Ca- and Mg-rich (27 wt % NaCl equiv) brine and cooler (130°C), Na-dominant (10 wt % NaCl equiv) meteoric water, focused by steep normal faults, are proposed to have interacted with the Perke Iron Formation to form a hypogene carbonate-hematite alteration assemblage (Netshioski, 2002). Subsequent leaching and karstification of this assemblage by shallow meteoric water produced the hematite orebodies. The Paraburadoo 4E deposit can be considered a higher-temperature analogue of the Thabazimbi deposit.

Similarities to the 4EE deposit are also shown from research at the Koolyanobbing deposit hosted within the Mesoarchean lower succession BIF of the Koolyanobbing greenstone belt, in the Yilgarn craton in Western Australia. Early Fe-Mg < Ca metamatism caused local ferroan carbonate and ferroan talc alteration of the metamorphosed quartz-magnetite BIF protolith. Fluids invoked for this hydrothermal alteration involve fluids exchanged with proximal mafic rocks as well as fluids
derived from granites. Talc alteration is also recognized at the Mt. Tom Price deposit (Thorne et al., 2009) where the presence of talc is interpreted to represent a higher-temperature equivalent of the Paraburadoo deposit (Dalstra and Guedes, 2004).

Recent work at the Carajás deposit (Figueiredo e Silva et al., 2008, 2013) has demonstrated that the interaction of hypogene hydrothermal fluids with jaspilite along steep faults resulted in the formation of high-grade iron orebodies. For example, in the N5E deposit at Carajás, the hypogene fluids exhibit a wide range of salinities (0–30 wt % CaCl₂ equiv) but a restricted range of trapping temperatures (200°–330°C), interpreted to reflect the mixing of hot magmatically derived fluid and heated meteoric water. The fluid characteristics of the N5E deposit suggest that it is a slowly warming temperature analogue of the Paraburadoo 4E deposit. Results of this study and of the N5E deposit support a temperature continuum model for high-grade iron ore mineralization, as first described by Dalstra and Guedes (2004). Within this model, the Paraburadoo 4E deposits can be considered a lower-temperature equivalent of the Mt. Tom Price deposit.

**Formation of ore types**

The global classifications of iron ores, from province to province and from mine to mine, are usually based on hardness and degree of induration (friability-hardness) with iron ores categorized using terms such as hard, friable, soft powdery, and blue dust (Beukes et al., 2008). The high-grade iron formed at the 4E deposit as well as Mt. Tom Price and Mt. Whaleback, and below the depth of modern weathering can be considered to be homogeneous at a deposit scale and classified as hard ore. Within the supergene weathering profile, the relative physical hardness of high-grade hematite ore at the 4E deposit varies considerably within the deposit and is a function of ore texture and the extent of supergene weathering processes such as the secondary leaching of martite, goethite, and microplaty hematite or precipitation of late secondary brown goethite within the supergene alteration profile (Clout, 2005). These processes largely result in an increase in the hardness of the ore closer to the modern topographic surface.

In comparison, Indian iron ore deposits and those of the Serra dos Carajas district in Brazil (Beukes et al., 2008) that are proposed to have formed by hydrothermal ore-forming processes have undergone significant supergene modification, probably resulting in further enrichment of the hydrothermal iron ore. Two ore types dominate these deposits: hard ore and soft friable ore. The hard ores within these deposits are believed to have formed by the extensive replacement of the BIF by massive magnetite during early hydrothermal events (Mukhopadhyay et al., 2005; Figueiredo e Silva et al., 2013). This was followed (in most deposits) by the widespread replacement of magnete by martite. The hard ores display variable destruction of the primary banding of the BIF during magnetite-martite ore formation, which differs from hard ore of the Pilarba deposits, which preserves the primary banding of the BIF.

While there is common acknowledgement of the role of hydrothermal fluids in the formation of hard ores (Dalstra and Guedes, 2004), there is conjecture regarding the development of the soft saprolitic ores. Within India deposits, the soft hematite ores that surround the hard orebodies of hydrothermal origin essentially represent saprolite of hard ore that developed in the bleached zone of ferricrete soil profile due to intense acidic groundwater circulation (Mukhopadhyay et al., 2008). Dalstra and Guedes (2004) provide an alternate hypothesis for the formation of soft hematite ore of the Carajás orebody whereby the soft hematite ore was developed directly by supergene leaching of porous hematite-carbonate protore formed during early hydrothermal alteration. As a result, the hematite-martite layers are condensed, with some porosity remaining where there was previously carbonate. At present, there is a lack of research to determine the role of early hydrothermal alteration that ultimately results in the formation of soft iron ores. If soft hematite ores are formed from the weathering of a hematite-carbonate protore, they may form without the presence of hard hematite orebodies. Conversely, if soft hematite ore was formed by weathering of hard ore, then the presence of soft ore directly implies the presence of hard orebodies, possibly at depth. Each of these models for the formation of soft ore has direct consequences for exploration of iron ore deposits and requires further research.

**Fluid sources**

The source of hypogene fluid responsible for hypogene alteration in iron deposits of the Hamersley province has long been debated. Morris (2002) argued that the Wittenoom Formation could not contain the volume of basinal brines required for hypogene alteration and ore formation of the Mt. Tom Price, Paraburadoo, and Mt. Whaleback deposits. Crush-leached ion chromatography of carbonate and hematite from hypogene alteration assemblages and high-grade ore at the Paraburadoo 4E and the North deposits supports the critical role of hypogene fluids in their formation. The low Cl/Br ratios of the samples indicate that basinal brines derived from evaporated sea water were responsible for the hypogene alteration and, ultimately, for formation of the high-grade iron ore of the 4E deposit. Recent work on the tectonic setting of the southern Pillbara craton (Cawood and Tyler, 2004; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo at the time of the hypogene alteration (>2005 Ma; Barley et al., 1999; Müller et al., 2005; Johnson et al., 2011) advocates the existence of a back-arc basin to the south of Paraburadoo (Johnson et al., 2011). Enhanced heat flow driven by back-arc extension and volcanism is interpreted to have resulted in thermally driven flow of basinal fluids along low-angle faults, northward into the banded iron formations of the southern Hamersley province. Here, the hot basinal fluids mixed with deeply circulating meteoric water to ultimately form the high-grade iron deposits. Given this tectonic setting, there likely would be a sufficient volume of hypogene fluids to form the high-grade deposits at Paraburadoo, Mt. Tom Price, and Mt. Whaleback (Fig. 1).
Petrographic evidence (Fig. 9D) from the distal magnetite-quartz-dolomite-stilpnomelane-hematite-pyrite alteration assemblage shows that the corrosion and pitting of euhedral magnetite results in the formation of anhedral magnetite adjacent to euhedral magnetite. This localized dissolution and recrystallization of Fe is interpreted to reflect a local source of the iron measured in fluid inclusions from within the host BIF. The dissolution and precipitation of magnetite indicates that the hydrothermal fluids are likely to have low pH (Ohmoto, 2003).

**Similarities with Mississippi Valley-type (MVT) deposits**

Netshioski (2002) and Thorne et al. (2004) suggested that the chemistry of hydrothermal fluids responsible for forming the North deposit at Mt. Tom Price and the Thabazimbi deposit is analogous to that of MVT mineralizing fluids. The concentrations of Fe, Mn, Pb, and Zn in hydrothermal fluids from the 4E deposit (Fig. 19, Table 5) correspond to higher-temperature equivalents of sedimentary formation brines involved in the formation of low-temperature Pb-Zn deposits (Fig. 20; Carpenter et al., 1974; Sverjensky, 1986). However, the Paraburdoo deposit differs significantly when compared to MVT deposits, based on its lack of any significant base metal mineralization. The high salinities, elevated metal contents (Cu, Pb, Zn), and high temperatures of hydrothermal fluids inferred for the Paraburdoo deposit have clear MVT ore-forming potential (Fig. 20), yet sulfide mineralization there is restricted to minor amounts of pyrite (<1 vol %) within hypogene alteration assemblages.

Given the lack of sulfur within the primary BIF (Trendall and Blackley, 1970), significant base metal mineralization in the southern Hamersley province would require the sulfur to be transported either with metals in an oxidized form (SO₄) with reduction occurring at the site of ore deposition or separately from the metals and sulfur in two different fluids, thus requiring fluid mixing at the site of ore deposition (e.g., Beales and Jackson, 1986). Exploration for base metals in the southern Hamersley province is likely to be conducted systematically and stratigraphic traps where such fluid mixing or oxidation may have occurred.

**Exploration model**

The structural setting of the Paraburdoo deposits proposed here implies that the formation of high-grade hematite deposits in the Brockman Iron Formation is directly linked to fluid pathways (i.e., fault zones) that link the enclosing sedimentary basins to the depositional site. Results of this study, together with data from Mt. Tom Price (Barley et al., 1999; Hagemann et al., 1999; Taylor et al., 2001; Thorne et al., 2004, 2005, 2008) and Mt. Whaleback (Webb et al., 2004; Oliver, 2007), indicate that the formation of hypogene alteration assemblages and, ultimately, high-grade iron ore deposits is restricted to the southern Hamersley province. These deposits all have genetic models that invoke heated (150–300°C) hypogene brines in their formation. We suggest that the hypogene fluids required to upgrade BIF to high-grade iron ore must have sufficient heat and salinity for the replacement of quartz with carbonate (Fournier, 1985) and the formation of hypogene alteration assemblages. Heated basinal brines interpreted to have been sourced from developing sedimentary basins to the south of the Hamersley province are likely to have cooled and diminished during their inferred migration northward and upward into the BIFs. Consequently, a spatial limit for the formation of high-grade iron ore deposits can be applied to the Hamersley province. A distance of 150 km, based on known occurrences of hypogene alteration in the Pilbara, measured from the southern limit of the Hamersley province (Fig. 1), is estimated to be the extent over which hypogene alteration can result in the formation of high-grade iron ore deposits.

Steep normal faults have been recognized as important exploration criteria (Taylor et al., 2001) as conduits for ascending hypogene fluids to interact with BIF of the Dales Gorge and Joffre members. The results of this study at the Paraburdoo 4E deposit further confirm the critical role that steep normal faults play in the formation of high-grade iron ore deposits. Our structural reconstruction of the deposit has shown that rotation of steep normal faults contributed to the preservation of “shaded” orebodies below unmineralized cover sequences. However, although the presence of steep normal faults is a valuable exploration tool, they should be used together with knowledge of the structural history of the area, in order to determine the best strategy for exploration.

**Acknowledgments**

We thank Rio Tinto Exploration and Rio Tinto Iron Ore, which allowed us access to current data on their deposits. Without their logistical and geological support, many of the details and geological plans of their deposits would not have been available. Thorne particularly acknowledges Raul Bitten-court, who organized this research project. Rio Tinto Iron Ore geologists Wayne Zarb and Joffre Buswell are thanked for invaluable assistance during fieldwork and also for willingly sharing detailed knowledge of the 4E deposit. We also thank David Sepe for his contribution to the research as part of his honors thesis on the Paraburdoo deposit. We also thank Jydpil Mulkapaddiy, Francisco Javier Rios, and John Slack for constructive criticism of an earlier version of this manuscript.

**REFERENCES**

CHAPTER 5: CONCLUSIONS, FUTURE WORK, AND IMPLICATION FOR EXPLORATION.

In the past decade significant advances have been made in the understanding of the formation of high-grade iron ore deposits of the Hamersley province. Advances include the structural reconstruction of deposits, mapping and detailed geochemistry of hypogene alteration zones, and a greater understanding of the processes and controls of supergene alteration. Despite this greater understanding, many fundamentals questions related to the formation of these deposits remained: (1) Which structures control hydrothermal fluid flow? (2) Where is the source of hydrothermal fluids that transformed unmineralised BIF to high-grade iron ore? (3) Can new applied geochemical techniques be used to decipher the processes that formed these deposits, and (1) Do the high-grade martite-microplaty hematite deposits of the Hamersley Province share a structural-hydrothermal history, and if so can this be used to predict the location of further deposits? This thesis was aimed at answering these questions.

In the Hamersley Province the martite-microplaty hematite deposits are located within the Dales Gorge and Joffre Member of the Brockman Iron Formation, and the Marra Mamba Iron Formation. High-grade iron ore is spatially associated with normal faults and thrust faults that formed during D2 deformation (~2200 Ma), and have later been affected by D3 deformation (~1650 Ma). At Mt Tom Price and Paraburdoo, it is the second-order faults (southern batter fault, 4E fault) that are the critical pathways for transporting basinal brines from below, and meteoric waters from above, into the host BIF. The re-activation of these second-order faults is constrained by dating of dolerite dykes at Paraburdoo to a period extension at ~2000 Ma. This period of extension allowed the release of hot basinal brines from the underlying sedimentary packages of the Wittenoom Formation and was the first step in a a hydrothermal process that ultimately formed the high-grade iron ore deposits of the Mt Tom Price and Paraburdoo. When large volumes of ascending basinal brines are focused into the magnetite-chert–rich BIF, the rocks are transformed into a hypogene carbonate-altered BIF. Interaction of descending meteoric waters with the carbonate-altered BIF produces a porous martite-microplaty hematite-apatite ore. The final stage of late Mesozoic supergene alteration removed much of the apatite from the martite-microplaty hematite-apatite assemblage, forming martite-microplaty high-grade ore and converted shale layers to kaolinite-rich clay.

Distinct alteration haloes are present at both the Mt Tom Price and Paraburdoo deposits and they provide not only a vector towards mineralization but information of the processes that lead to their formation. Hydrothermal alteration haloes at Mt Tom Price and Paraburdoo led to the application of several geochemical techniques (LA-ICPMS of fluid inclusions, crushed-leach ion chromatography,
microthermometry, \(^{18}\)O isotope geochemistry), some of which have not previously been used in the study of iron ore deposits in the Pilbara. The results of these studies were vital to the understanding of the thermodynamics, fluid flow within deposits, fluid sources and fluid mixing. Crushed-leach ion chromatography of quartz, dolomite from hydrothermal alteration assemblages and high-grade iron ore showed that basinal brines likely originated from evaporated seawater that had lost Mg and Na and gained Li and Ca through fluid-rock reactions with volcaniclastic rocks and carbonate successions within the Wittenoom Formation. The source of the evaporated seawater is interpreted to be from developing sedimentary basins to the south of the Hamersley Province. The study of the oxygen isotope geochemistry of magnetite and hematite during the transition from BIF to high-grade ore in the central Hamersley province provides support for the hydrothermal formation of the deposits and the role of both basinal brines and meteoric waters in their formation.

The high-grade iron ore deposits of the Hamersley Province share a similar structural-hydrothermal history with localised geology and fluid geochemistry being important in the development of the mineralogy and geochemistry of high-grade iron ore at the deposits.

**Exploration significance**

The awareness of hydrothermal alteration zones, surrounding high-grade BIF-related iron deposits, allows companies to place emphasis on the identification and the spatial distribution of hypogene carbonate alteration and replacement textures. These can then be used as vectors toward high-grade iron ore in exploration drilling. The understanding that faults and thrusts are important controls in the localization of supergene martite-goethite and supergene-modified hydrothermal martite-microplaty hematite deposits provides an important tool for exploration beneath cover. Conventional structural and stratigraphic mapping and reconstructions of the structural history of entire iron districts might allow the identification of areas where fluids can be focused by faults into structural traps (i.e., synclines). The concealed 4EE orebody of the 4E deposit demonstrates the critical role that structural geology plays in the exploration for high-grade iron ore bodies. Structural reconstruction should be considered a critical exploration activity in structural complex terrains where more concealed ore bodies may exist.

**Further Work**

Several key questions remain within the supergene-modified hypogene models proposed for the Mount Whaleback, Mount Tom Price, and Paraburdoo deposits. One of these is the relative roles and composition of ascending and/or descending hypogene fluids and meteoric waters, respectively. Further work to integrate and expand the fluid data sets, in particular, within fault zones where evidence of fluid mixing is likely to be preserved, is needed. The mechanism and P-T-X condition for the crystallization of microplaty
hematite also requires further work. Several mechanisms are invoked, both redox and non-redox, based on the requirement for oxygen to transform magnetite to hematite. If oxygen is not required to convert magnetite to hematite then descending oxygenated hydrothermal fluids are not required to transform magnetite to martite or form microplaty hematite at the Mount Tom Price, Mt Whaleback and Paraburdoo deposits. This significantly changes the relationship(s) of the deposit to the Proterozoic weathering surface and increases the depth at which high-grade orebodies may be found.

The supergene origin of martite-goethite deposit has previously been widely accepted (Paquay and Ness, 1998). However, recent work by Lascelles (2006a) provided an alternative supergene-modified syngentic model for the Hope Downs deposit. Stable isotope analyses of hematite, goethite, manganese, and carbonates can assist in elucidating the origin of fluids responsible for the transformation of BIF to high-grade martite-goethite ore at the deposit. Dating of alteration phases (e.g., manganese oxihydroxides) may also help resolve the timing and ultimate formation of the Hope Downs deposit.
APPENDICES

Appendix 1:
Refereed co-authored journal articles, refereed extended abstracts and non-refereed abstracts

Appendix 2 (electronic):
1. Sample Register
2. Fluid Inclusion Microthermometry Analysis
3. Fluid Inclusion Laser Ablation ICP-MS Data
4. Iron Oxide Laser Ablation Analysis
5. Carbonate Carbon-Oxygen Isotope Analysis
6. D/H Talc Analysis
7. Crushed Leach Analysis
8. Structural and Geological Maps
9. Conference Presentations
Chapter 4

Hypogene Alteration Associated with High-Grade Banded Iron Formation-Related Iron Ore

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Abstract

Hydrothermal alteration in structurally controlled, high-grade banded iron formation (BIF)-related iron deposits at Carajás (Brazil), Hamersley (Australia), and Thabazimbi and the Zeekoebaart prospect (South Africa) exhibit significant similarities and differences in geologic setting and hypogene alteration. In Carajás, Paleoproterozoic hematite deposits are hosted in low-metamorphic grade Archean jaspilites that are encased in metabasalts. The Paleoproterozoic BIF-hosted deposits of the Hamersley district, the Thabazimbi deposit, and the Zeekoebaart prospect are surrounded by shales.

At Carajás, the hydrothermal alteration of jaspilites is characterized by a distal alteration zone with magnetite-calcite-quartz-pyrite where the primary microcrystalline hematite \( \rightarrow \) magnetite (±kenomagnetite). The intermediate alteration zone consists of martite-microplaty hematite-quartz with magnetite \( \rightarrow \) martite, whereas the proximal alteration zone contains hematite ± carbonate ± quartz with martite \( \rightarrow \) microlamellar hematite \( \rightarrow \) anhedral hematite \( \rightarrow \) euhedral-tabular hematite. The proximal alteration zone represents the high-grade ore (i.e., porous hard to soft and hard ores). Hydrothermal alteration also affected mafic wall rocks with chlorite-quartz-carbonate ± hematite in distal alteration zones, and chlorite-hematite-quartz-albite-mica-carbonate ± titanite ± magnetite ± sulfides and hematite-chlorite-quartz-albite-mica-carbonate ± titanite ± magnetite ± sulfides in intermediate and proximal alteration zones, respectively.

At the Mount Tom Price deposit in the Hamersley district, three spatially and compositionally distinct hydrothermal alteration zones are distinguishable: (1) distal magnetite-siderite-iron silicate, where the shape of the magnetite is suggestive of it being pseudomorphous after preexisting minerals, likely siderite; (2) intermediate hematite-ankerite-magnetite, with euhedral and bladed magnetite showing minor replacement by martite along crystal boundaries and replacement of iron-silicates by anhedral and microplaty hematite; and (3) proximal martite-microplaty hematite zones, where carbonate is removed. Martite and anhedral hematite replace magnetite and iron silicates of the intermediate alteration assemblage, respectively.

The Thabazimbi deposit and the Zeekoebaart prospect lack unequivocal evidence for the formation of paragenetically early hydrothermal magnetite. Chert in ore zones has been replaced by microplaty hematite or has been leached, giving rise to porosity. Veins contain coarse tabular hematite and coarse crystalline quartz. High-grade hematite-martite orebodies are the result of SiO\(_2\) leaching and associated volume loss that created widespread brecciation of the high-grade hematite ore. In addition to high-grade hematite-martite ores, four mineralogically distinct types of iron ore have been recognized: (1) goethite-rich, (2) low-grade dolomite-hematite, (3) low-grade calcite-hematite, and (4) talc-hematite.

The comparison of hydrothermal alteration characteristics in the three case study areas revealed: (1) a similar paragenetic sequence of iron oxides, marked by an abundance of open-space filling and replacement textures; (2) distinct lack of a penetrative fabric in alteration lithologic units and high-grade ores; and (3) the importance of porosity and brecciation to accommodate volume loss. Differences include: (1) the formation of carbonate in different hydrothermal alteration zones of each deposit; (2) the presence of stilpnomelane in BIF that is surrounded by shales and hosted in sedimentary basins but absence in BIF that is bounded by mafic rocks; (3) the presence of significant amount of siderite in distal alteration zone in the Hamersley deposits but absence in the Carajás and Thabazimbi deposits; (4) the presence of significant amount of sulfides in the Carajás deposits but absence in the Hamersley and Thabazimbi deposits; and (5) significant amounts of chlorite, talc, white mica, and albite in basalt-hosted iron ore deposits (e.g., Carajás) or mafic dikes that are spatially and temporally associated with iron mineralization (e.g., in the Hamersley province).

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The systematic documentation of hydrothermal alteration mineralizations and assemblages has significant implications for the exploration of concealed high-grade iron orebodies, because key hydrothermal alteration minerals such as chlorite, talc, carbonates or iron silicates are an expression of the hydrothermal footprint of the BIF iron-ore mineral system and, therefore, can be used as mineral vectors.

**Introduction**

This contribution aims to critically review the effects of hydrothermal alteration on BIF-hosted iron deposits and associated wall-rock lithologic units in the Carajás (Brazil), Hamersley (Australia), Thabazimbi, and Zeekoebaart (South Africa) areas. The study of the hydrothermal alteration has provided important insights into the nature of hydrothermal fluids and the formation of hydrothermal ore deposits (i.e., Reed, 1997). Recent studies have revealed the presence of hydrothermal alteration zones associated with and surrounding high-grade BIF-hosted iron ore deposits of the Carajás district (Figueiredo e Silva et al., 2008), the Hamersley district (Thorne et al., 2008), the Thabazimbi deposit (Netshiozwi, 2002), and the Zeekoebaart prospect (Harding, 2004). Although the effects of hydrothermal alteration may be recorded by BIF, they are often more prominent in associated lithologic units such as volcanic rocks (Lobato et al., 2005b) or sills and dikes (Dalstra and Guedes, 2004). The scarcity of studies of the hydrothermal alteration associated with most high-grade BIF-hosted iron deposits can be attributed to the difficulty of recognizing it through the effects of intense supergene weathering (i.e., Beukes et al., 2003).

This contribution describes and compares the effects of hydrothermal alteration on BIF and associated wall-rock lithologic units in the Carajás and Hamersley districts, Thabazimbi deposit, and Zeekoebaart prospect. The parageneses and textures of hydrothermal alteration zones, veins and breccias related to hydrothermal alteration, and lateral and vertical transitions of the alteration zones are described for these districts and/or deposits. Finally, implications for exploration are provided and questions for further research are discussed.

**Hydrothermal Alteration in BIF-Hosted Iron Deposits**

High-grade iron orebodies are almost monomineralic and composed essentially of hematite (and martite-textured hematite), with minor magnetite, goethite, quartz, and trace amounts of apatite (Taylor et al., 2001). In most mining districts, high-grade orebodies are in sharp contact with oxidized BIF that consists of subequal amounts of hematite and quartz. This oxidized BIF has been categorized as altered in only a few districts where it has been traced laterally into unaltered BIF host rock (Barley et al., 1999; Thorne et al., 2004). We have attempted to delineate zones of hydrothermal alteration with distinct textures and/or mineral associations and apply the terms distal, intermediate, and proximal to indicate their position relative to high-grade iron orebodies. In general, distal alteration is weaker and proximal is stronger.

Iron oxides change in form through the mineralogical sequence and hydrothermal alteration zones from diagenetic microcrystalline or dusty hematite, locally followed by hydrothermal magnetite (Mag) in distal alteration zones. Magnetite may be replaced by martite (Mt) exhibiting kenomagnetite (KMag) nuclei and finally hematite, both characterizing the intermediate zones. Hematite types are classified according to their morphology into microplaty (MpHem), platy, euhedral to anhedral, patchy (similar to anhedral), bladed, and tabular (see Fig. 1), which are typical of the proximal alteration zones.

**Carajás Iron District, Brazil**

**Regional geologic setting**

The Carajás mineral province is in an Archean through Paleoproterozoic portion of the Amazon craton. The Carajás mineral province is dominated by metavolcano-sedimentary rocks and granitoids, formed from 2.76 to 2.68 Ga, and also includes the Meso-Archean igneous and metamorphic Pium and Xingu Complexes (Santos, 2003).

The Carajás mineral province is host to a wealth of mineral deposits (Figueiredo e Silva et al., 2008), which are associated with a range of metavolcano-sedimentary rocks that overlie...
what is considered as the tonalite-trondhjemite-granodiorite basement terranes of the Xingu Complex (ca. 2.86 Ga; Silva et al., 1974; Tassinari et al., 2000). The Carajás metavolcano-sedimentary and metasedimentary units belong to the Itacuaitnas Supergroup (DOCEGEO, 1988), formally subdivided into the Grão Pará, Igarapé Salobo, and Igarapé Piquena Groups, although a better stratigraphic definition is still required (Figueiredo e Silva et al., 2008). Geochemical data indicate two main volcano-sedimentary periods, at 2.76 Ga and 2.75 to 2.73 Ga (Machado et al., 1991; Galarza et al., 2001). Jaspilites (plus the iron ores) of the Carajás Formation and their surrounding magmatic rocks (Beisiegel, 1982) are part of the Grão Pará Group dated at 2740 ± 5 Ma by Trendall et al. (1999). The volcano-sedimentary rocks are covered by sedimentary psammo-pelite rocks (e.g., arenites, calcarenites, siltites, and conglomerates) of the Águas Claras Formation (fig. 1 of Figueiredo et al., 2008), which yielded zircon U-Pb ages of 2708 ± 37 Ma (Mougeot, 1996) and 2645 ± 12 Ma (Dias et al., 1996).

A recent summary of the structural framework of the Carajás area is provided by Resoêre et al. (2006). The authors proposed that the dominant structure is a flattened flexural fold system intersected by several strike-slip faults subparallel to the axial plane of the fold system (fig. 1 of Figueiredo e Silva et al., 2008). Further details of the geology of the Carajás mineral province are presented by Figueiredo e Silva et al. (2008).

**Hydrothermal alteration zones**

Hydrothermal alteration associated with the transformation of jaspilite to high-grade iron ore affected jaspilites to varying degrees at different times during the evolution of the iron deposits (Table 1A; Guedes et al., 2002; Figueiredo e Silva, 2004; Figueiredo e Silva et al., 2004, 2007, 2008; Lobato et al., 2004, 2005a, b, 2007; fig. 7 of Figueiredo e Silva et al., 2008). The mineralogical sequence for iron oxides in the distal hydrothermal alteration zone is microcrystalline hematite → magnetite (kenomagnetite; Fig. 1A, D, E). The intermediate alteration zone is characterized by widespread martitization, and the proximal alteration zone has martite → microlamellar hematite → anhedral hematite → euhedral-tabular hematite (Fig. 1H, M-O; fig. 7 of Figueiredo e Silva et al., 2008). Microcrystalline hematite (~4 µm; Fig. 1A) is typical of the jasper layers in nonmineralized jaspilites. It represents the original iron oxide and is stable even in the more advanced alteration stages in desilicified, banded ore-breccia fragments.

**Table 1.** Characteristic Hydrothermal Alteration Assemblages in Distal, Intermediate, and Proximal Alteration Zones for Selected Iron Deposits in the Carajás and Hamersley Iron Districts in Brazil and Australia, Respectively, Thabazimbi Deposit and Zerkoebaart Prospect in South Africa1

<table>
<thead>
<tr>
<th>Alteration zones</th>
<th>Serra Norte deposits (in jaspilite)</th>
<th>North deposit</th>
<th>Southern Ridge deposit</th>
<th>Thabazimbi deposit</th>
<th>Zerkoebaart prospect</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>BIF</strong></td>
<td>Jasper-MiHem</td>
<td>Mag-Chert-Sd/Fe-Dol-StP-Py</td>
<td>Qtz-Mag-Stp-Dol</td>
<td>Chert/Micro Qtz-Mag-Stp-Carb-Gru</td>
<td>Hemi-Mag-Sd/Ank-Chert-StP</td>
</tr>
<tr>
<td>Distal</td>
<td>Mag-Carb-Qtz-Py(Sep)</td>
<td>Mag-Sd-FeSila (Py-Ap-Qtz)</td>
<td>Mag-Sd-StP</td>
<td>KMag-Mt-Flaty Hem + Red Chert-TlHem</td>
<td>Hemi(KMag)-Chert-Mega Qtz</td>
</tr>
<tr>
<td>Intermediate</td>
<td>Mt-MpHem(Qtz(Sulf-Mag)</td>
<td>Hemi-Mag(Tl-Car)</td>
<td>Hemi-Mag-MyHem-Ap</td>
<td>Hemi-Carb</td>
<td>Patchy Hemi-MpHem-Mt</td>
</tr>
<tr>
<td>Proximal and/or</td>
<td>high-grade ore</td>
<td>Mt-MpHem-Ap</td>
<td>Mt-MpHem-Ap</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1 Adapted from data by Thorne (2001); Netshiozwi (2002); Harding (2004); Lobato et al. (2005b); Thorne et al. (2007); Zucchetti, 2007; Figueiredo e Silva et al. (2008); minerals in parentheses represent minor phases
2 These include martite, anhedral, tabular, microplaty, and microcrystalline hematite
3 These include martite, patchy, microplaty, anhedral, and tabular hematite

Mineral abbreviations: Ab = albite, Ank = ankerite, Ap = apatite, Ang = anhydrite, Carb = carbonate, Ccp = chalcopyrite, Chl = chlorite, Dol = dolomite, FeSila = iron silicate, Gru = grunerite, Ilm = ilmenite, KMag = kenomagnetite, Mag = magnetite, MiHem = microcrystalline hematite, MpHem = microplaty hematite, Mt = martite, Ol = olivine, Pl = plagioclase, Py = pyrite, Qtz = quartz, Sd = siderite, StP = stilpnomelane, Sulf = sulfide, THem = tabular hematite, Tlc = talc, Tm = titanite
deposit). The proximal alteration zone includes the porous hard to soft ore and the hard, massive ore itself (the latter is about 20 to 50 m in width). Their mineralogical composition and general characteristics are depicted in Tables 1 and 2.

Distal alteration zone—magnetite-calcite-quartz-pyrite:

The distal alteration zone (Tables 1A, 2A; Fig. 3) contains microcrystalline hematite, which is overgrown by magnetite (± martite) “blasts” (Fig. 1A). Magnetite-rich jaspilite ores (Fig. 1D) are present in the N5S deposit. Jasper bands may display discoloration resulting in clear halos of recrystallized chert or fine-grained, clear granoblastic quartz (Fig. 2A-C) that surrounds martite (Fig. 2C). The halos may represent leaching of iron from jasper due to chert recrystallized into quartz.

Sulfide minerals, mainly pyrite and chalcopyrite, may be located in vein breccias in association with carbonate minerals, viz. calcite (Figueiredo e Silva et al., 2008), and magnetite and quartz that are discordant to layering in the jaspilite. Pyrite-magnetite assemblage locally overgrows microcrystalline hematite-rich jaspilite layers; monomineralic pyrite veins are also observed.

Jasper recrystallization to quartz is associated with the formation of abundant quartz veins and veinlets (Fig. 9, Table 3 of HYPOGENE ALTERATION ASSOCIATED WITH HIGH-GRADE BIF-RELATED IRON ORE 111.
Figueiredo e Silva et al., 2008) and porosity development (Fig. 2A-E). The dominant V1a and V1b veins (Table 3 of Figueiredo e Silva et al., 2008) in the distal alteration zone contain quartz ± sulfide or calcite ± chalcopyrite (pyrite)-magnetite, and either crosscut or are parallel to jaspilite bands. The jaspilite adjacent to V1 veins is locally brecciated.

Intermediate alteration zone—martite-microplaty hematite-quartz:
Martite characterizes the martite-microplaty hematite-quartz assemblage of this alteration zone (Tables 1, 2; Fig. 3), with or without kenomagnetite. Martite (grain size 100–300 μm, exceptionally 600 μm) may preserve kenomagnetite relics (e.g., at the N4W deposit; Fig. 1E). Quartz-jasper dissolution led to porous, locally brecciated rocks, and to the precipitation of lamellar (microplaty or platy) hematite (Fig. 1H) that fills the open spaces or vugs (Fig. 2D, E). The size range of this hematite is 4 to 8 μm (microplaty) and 100 to 200 μm, locally 400 μm (platy). The microplaty or platy hematite also formed at the expense of martite edges. Rare gold particles are included in quartz or associated with martite in altered jaspilite (Figueiredo e Silva, 2004).

There are two vein types (Fig. 9, Table 3 of Figueiredo e Silva et al., 2008) in the intermediate alteration zone: V2—quartz ± microcrystalline hematite bedding-discordant and vuggy veins, and V3—microcrystalline hematite-quartz veins that crosscut and/or are parallel the jaspilite bedding. Quartz is granoblastic, zoned, and comb textured in the vugs. Carbonates are absent.

Proximal alteration zone—hematite (carbonate-quartz):
The proximal alteration zone (Tables 1A, 2; Fig. 3) is characterized by various types of hematite and/or by the presence of martite. The destruction of martitic magnetite “blasts” results from their recrystallization forming abundant eye-shaped agglomerates of anhedral hematite (~20 μm), especially in the N5E deposit hard ores (Fig. 1M). Some anhedral hematite

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**Fig. 3.** Paragenetic sequence for the Carajás N4 and N5 iron ore deposits. Table at bottom displays variation from hard hematite (Hem) ore zone through proximal, intermediate, and distal alteration zones. Note that chert recrystallization to quartz and their dissolution are associated with the intermediate and proximal zones, whereas quartz vein deposition is observed mainly in the zone of hydrothermally altered jaspilites (intermediate and distal alteration zones). Top table displays oxide mineral variation from early- to late-stage hydrothermal alteration. Mineral abbreviations: AHem-SHem = anhedral and subhedral hematite, MiHem = microcrystalline hematite, MpHem = microplaty hematite, Mt = martite, THem = tabular hematite (adapted after Lobato et al., 2005b).
**Table 2. Summary Hydrothermal Alteration Chart, Displaying Commonalities and Differences in Hypogene Alteration for Selected Iron Ore Deposits of the Carajás and Hamersley Iron Districts in Brazil and Australia, Respectively, Thabazimbi Deposit and Zeekoebaart Prospect in South Africa**

<table>
<thead>
<tr>
<th>Alteration zones</th>
<th>Serra Norte deposits (in jaspilite)</th>
<th>North deposit (Tom Price)</th>
<th>Thabazimbi deposit</th>
<th>Zeekoebaart prospect</th>
</tr>
</thead>
<tbody>
<tr>
<td>Distal</td>
<td>(1) Chert-Qtz leached; abundant vein Qtz ± Carb ± Sulf (2) Mag after MiHem &amp; in veins (3) Qtz-Carb veins may have Py and Cx but lack Hem</td>
<td>(1) Distal magnetite-siderite-iron silicate zone (2) Chert bands partially replaced by bladed Mag, Sd and Fe silicates (3) Mag after Dol or Sd (4) Brecia and vein formation</td>
<td>(1) Fe silicates, Carb and Mag replaced by KMag and Mt (2) Chert is partly leached and also replaced by MpHem (3) Collapse breccias cemented by late THem, with angular fragments of oxidized BIF (4) Microcrystalline Qtz and Hem replace Gru and Ank</td>
<td>(1) Mag replaced by fine-grained Hem to form KMag that develops into Mt (2) Leaching of silicate layers results in collapse structures (3) MpHem in iron-poor cherty layers (4) PH replaces Mt (5) Qtz veins</td>
</tr>
<tr>
<td>Intermediate</td>
<td>(1) Intense Mag martitization (2) Advanced chert recrystallization to Qtz and chert-Qtz leaching (3) Intense Qtz-Hem and H-Qtz formation with MpHem and less Sulf (4) Significant porosity increase; vugs partially filled by MpHem</td>
<td>(1) Intermediate hematite-anhydrite-magnetite zone (2) Sd replaced by Ank; liberated Fe precipitates (3) Brecias, veins; complete absence of chert mesobands (4) MpHem</td>
<td>Not described</td>
<td>Not described</td>
</tr>
<tr>
<td>Proximal and/or high-grade ore</td>
<td>(1) Progressive martitization to form AnHem (2) Rare Qtz veins with MpHem (3) Stockwork-style, ‘Christmas-tree’-type veins contain THem, and crosscut rocks of all alteration stages (4) Hematite-martite and hematite ore types, with MpHem, AnHem, Them, and preserving MiHem (5) Intense Carb alteration results in ore breccias cemented by Dol (N5 deposit) and Kt (6) Late-stage, AnHem- and Them-rich ore lacks Qtz and Carb veins</td>
<td>(1) Proximal martite-microplaty hematite-apatite zone (2) Mt and AnHem replace Mag and Fe silicates, respectively (3) Proportion of Mt increases with proximity to high-grade ore (4) Increased oxidation of Mt is accompanied by increase in voids and Gt formation (5) MpHem important constituent of both hard and soft high-grade iron ores (6) All Carb are dissolved (7) Ap occurs mainly within the inter-granular space of Mt mesoband</td>
<td>(1) Efficient leaching and replacement of chert (2) Two hypogene, high-grade iron ore types that grade into one another: (a) hard hematite that may be massive, laminated, and brecciated; (b) carbonate-hematite (3) (a) Contain fine-grained Hem and minor Mt-laminated ores dominated by MpHem, which is abundant in brecciated and matrix ores; (b) Encompasses various breccia ore types, cemented by sparry Carb, and developed in response to fracturing of high-grade hematite ores (4) Hematite-martite and high-grade ore (5) Intense Carb alteration (6) Late-stage, AnHem- and Them-rich ore lacks Qtz and Carb veins</td>
<td>(1) Complete replacement of KMag to form Mt (2) Efficient leaching of SiO₂ led to laminated and massive hematite ores (3) Mt recrystallization resulting in PH in former Fe-rich bands in BIF (4) MpHem develops within Fe-poor or microcrystalline Qtz microbands at the expense of chert that is effectively leached (5) Small patches of chert almost entirely replaced by MpHem</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Alteration zones</th>
<th>Various deposits in the Serra Norte (in mafic wall rock)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Distal</td>
<td>(1) Igneous, variolitic texture partially preserved (2) Abundant Chl replaces prealteration silicates forming chloritites (3) Qtz-Carb veins (4) Some Hem, white mica and comb-textured Ab (5) Also Mag, minor MpHem, fibrous vein filling Tlc, Tn, and Bkh in veins; rare Zrn</td>
</tr>
<tr>
<td>Intermediate</td>
<td>(1) Persuasive Chl, also Hem development along veins and replacing anygylases, locally forming Hem chloritites (2) Important Carb and Qtz development (3) Also white mica, Ab, Tn, Mag, subordinante Sulf</td>
</tr>
<tr>
<td>Proximal</td>
<td>(1) Persuasive and abundant Hem; further Chl development (2) Presence of Hem chloritites (3) Also Carb, Qtz, white mica, Ab, Tn, Mag, subordinate Sulf</td>
</tr>
</tbody>
</table>

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Mineral abbreviations: Ab = albite, AHem = anhedral hematite, Ank = ankerite, Ap = apatite, Carb = carbonate, Cep = chalcopyrite, Chl = chlorite, Dol = dolomite, FeSīl = iron silicate, Grn = granulite, Gt = goethite, Hem = hematite, KMag = K-Magnetite, Kut = kutnahorite, MiHem = microcrystalline hematite, Mag = magnetite, MpHem = microplaty hematite, Mt = martite, PH = patchy hematite, Py = pyrite, Qtz = quartz, Sd = siderite, Sulf = sulfide, THem = tabular hematite, Tlc = talc, Tn = titanite, Zrn = zircon
also developed by recrystallization of microcrystalline hematite, where it was in contact with euhedral and/or tabular hematite veins. Fine- to medium-grained (200–300 μm, locally up to 500 μm), euhedral and bladed-tabular hematite (Fig. 1N, O) forms internal selvages to discordant and layering-parallel milli- to centimeter veins. The euhedral and bladed-tabular hematite appears to correspond to the latest oxide stage and is located in discordant veins and veinlets without quartz. They are commonly comb-textured, exhibit growth zones, and further fill the vugs created by chert-quartz dissolution. Growth lamellae are common in euhedral hematite (Fig. 1M) but also are observed in anhedral, tabular, and, rarely, in lamellar (microplaty or platy) hematite.

Pyrite is located as inclusions in hematite (martite and/or anhedral types) and, in the case of mafic rocks, also as rings around amygdales fillings. Locally pyrite is surrounded by chalcopyrite, which in places is overgrown by covellite. Rare gold is encountered in this zone as inclusions in tabular hematite. Rare chalcopyrite is observed in anhedral hematite bands of high-grade ore.

The proximal alteration zone is characterized by vein filling (vein types in table 3 of Figueiredo e Silva et al., 2008) in breccias classified as V4—carbonate-quartz breccia, and V5—quartz ± microplaty hematite breccia; both are located in high-grade ore (Table 3). In this alteration zone, carbonate is predominantly dolomite impregnated by hematite dust (fig. 9 of Figueiredo e Silva et al., 2008) that may be associated with (crystalline) hematite. Quartz and dolomite may contain hematite inclusions.

**Mafic rocks:** Hydrothermal alteration also affected mafic wall rocks surrounding jaspilites and ores, including mostly basalts, and to a lesser degree gabbro and diabase. Basalts display features indicative of seawater hydrothermal alteration (e.g., Meirelles, 1986; Meirelles and Dardenne, 1991; Teixeira et al., 1997). Amphiboles (± epidote), relict plagioclase laths and augite may occur as prealteration minerals (Tables 1B, 2B, Fig. 4) in samples that still preserve their original textures. These rocks were subsequently affected by hydrothermal alteration associated with the formation of high-grade iron ore (Zucchetti and Lobato, 2004; Lobato et al., 2005a; Zucchetti, 2007; Zucchetti et al., 2007).

In the distal alteration zone, igneous textures are partially preserved, such as variolites and amygdales (Table 2B, Fig. 5). The hydrothermal alteration is defined by chlorite-quartz-carbonate ± hematite assemblage (Table 1B). The color of the rocks is light pinkish-green due to the impregnation of very fine-grained hematite (Fig. 5A–F). This alteration was initiated via fluid influx in amygdalas or variolites, which acted as nuclei to fluid dispersion (Fig. 5A–C).

The intermediate alteration zone is characterized by chlorite-hematite-quartz-albite-mica-carbonate ± titanite ± magnetite ± sulfides assemblage, whereas the proximal alteration zone displays hematite-chlorite-quartz-albite-mica-carbonate ± titanite ± sulfides assemblage (Figs. 4, 5; Tables 1B, 2B). In both intermediate and proximal alteration zones hematite is dominantly acicular, microlamellar, and lamellar. Anhedral hematite (developed after martite) agglomerates in veins and random crystals located in the breccia matrices; carbonate, quartz, and rare sulfides dominate in veins. As shown in Figure 5D and E, the rupturing of amygdales margins gave place to hematite-rich chloritites, with concomitant to late development of hematite ± chalcopyrite-filled fractures.

**Hamerlys iron District, Australia**

**Regional geologic setting**

Banded iron formation-hosted high-grade iron ores of the Hamersley province in northwestern Australia (fig. 1 of Thorne et al., 2008) are located in the Neo-Archean to Paleoproterozoic (2.6–2.4 Ga) Mount Bruce Supergroup, a succession of volcanic and sedimentary rocks that overlie the granitoids and greenstone belts of the Archean Pilbara craton (e.g., summary in Martin et al., 1998). The Hamersley province contains a resource of about 25 billion metric tons (Bt) of iron ore and includes several deposits dominated by martite-microplaty hematite ore. The two most important examples are the giant Mount Tom Price and Mount Whaleback deposits (Thorne et al., 2008).

<table>
<thead>
<tr>
<th>Serra Norte deposits: Carajás</th>
<th>North deposit: Hamersley</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vein type</td>
<td>Vein mineralogy</td>
</tr>
<tr>
<td>V₁</td>
<td>Qtz, Py; Cal, Ccp, Py, Mag</td>
</tr>
<tr>
<td>V₂</td>
<td>Qtz, MpHem (Ccp)</td>
</tr>
<tr>
<td>V₃</td>
<td>MphHem, Qtz</td>
</tr>
<tr>
<td>V₄</td>
<td>Kut-Dol (Cal), MphHem</td>
</tr>
<tr>
<td>V₅</td>
<td>Qtz, MphHem (Ccp)</td>
</tr>
</tbody>
</table>

Table 3. Characteristics of Vein Types in the Serra Norte N4 and N5 Deposits in the Carajás Province and in the North Deposit in the Hamersley Province

Mineral abbreviations: Ank = ankerite, Cal = calcite, Ccp = chalcopyrite, Dol = dolomite, Kut = kutnahorite, MphHem = microplaty hematite, Py = pyrite, Qtz = quartz, Sd = siderite

Adapted with data from Thorne (2001), Lobato et al. (2005b), Thorne et al. (2007), Figueiredo e Silva et al. (2008)
The Neo-Archean Marra Mamba (ca. 2.6 Ga, Beukes and Gutzmer, 2008) and Early Paleoproterozoic Brockman (ca. 2.46–2.5 Ga, Beukes and Gutzmer, 2008) iron formations of the Hamersley Group, Mount Bruce Supergroup are host to the main iron ore deposits (Martin et al., 1998). A thick succession of shales and dolostones separate these two iron formations (e.g., Martin et al., 1998). Minor amounts of shale and tuffaceous beds are intercalated with the iron formations (Trendall and Blockley, 1970; Blake and Barley, 1992; Barley et al., 1997). The iron formations are typically microbanded, composed of microcrystalline quartz (chert), iron oxides, iron silicate, and carbonate minerals (Trendall and Blockley, 1970; Barley et al., 1999) that reflect green-schist-facies metamorphism. Dolerite sills and dikes locally intrude into the iron formations. Some of these mafic intrusions predate ore formation and are affected by intense hydrothermal alteration (Dalstra and Guedes, 2004), whereas others postdate ore formation (Müeller et al., 2005).

Iron ore deposits and types

The iron ores exploited in the Hamersley province can be subdivided into three distinct types: (1) high-grade martite-microplaty hematite deposits developed predominantly within the Brockman Iron Formation, with minor deposits within the Marra Mamba Iron Formation; (2) martite-goethite deposits hosted within the Marra Mamba and Brockman Iron Formations; and (3) detrital deposits and goethite-rich pisolitic accumulations in Tertiary paleochannels. The latter two ore types are rather convincingly related to geologically recent weathering processes, whereas a hydrothermal origin, followed by supergene modification, is now widely accepted for most high-grade martite-microplaty hematite ores.

The identification and investigation of hydrothermal alteration is summarized based on detailed work by a number of authors on the Mount Whaleback (Powell et al., 1999; Brown et al., 2004; Webb et al., 2004; Oliver et al., 2007), Mount Tom Price (Barley et al., 1999; Hagemann et al., 1999; Ridley, 1999; Taylor et al., 2001; Dalstra et al., 2002; Cochrane, 2003; Thorne et al., 2004, 2005, 2007, 2008) and Paraburdoo deposits (Taylor et al., 2001; Dalstra, 2005).

With a premining resource in excess of 1800 Mt at 65 wt percent Fe the Mount Whaleback deposit is the largest known iron ore accumulation in Australia (Clout, 2005). The deposit is structurally complex (fig. 4 of Thorne et al., 2008), with the orebody constrained by the westerly plunging overturned East and South synclines (Powell et al., 1999; Taylor et al., 2001; Webb et al., 2003, 2004; Thorne et al., 2004, 2005, 2007, 2008) and Paraburdoo deposits (Taylor et al., 2001; Dalstra, 2005).

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High-grade iron ore is equally distributed between the Dales Gorge and Joffre Members of the Brockman Iron Formation. Lesser amounts of iron ore are associated with flat-lying faults within the Yandicoogina Shale and the lowermost section or the Weeli Wolli Formation (fig. 6 of Thorne et al., 2008). Structural reconstruction of the Paraburdoo deposit suggests that the flat fault zones that now underlie the deposit were steeply dipping normal faults, prior to late tectonism (Dalstra, 2005). Northwest-trending dikes (ca. 2008 Ma; SHRIMP 207Pb/238U baddeleyite; Müller et al., 2005), which limit the axial extent of iron orebodies at Paraburdoo (fig. 6 of Thorne et al., 2008), intrude extensional faults active during the mineralization processes and were hydrothermally altered to a talc-chlorite assemblage during ore formation (Barley et al., 1999; Dalstra, 2005) but do not intrude the Mount McRae Formation. This indicates that the hypogene alteration and iron ore formation took place during or shortly after emplacement of the Paleoproterozoic dikes.

Hydrothermal alteration zones

Evidence for hydrothermal alteration is surprisingly scant at the Mount Whaleback deposit. Webb et al. (2004) reported that the chemical and mineralogical composition of the Mount McRae Shale beneath the iron orebody (fig. 4 of Thorne et al., 2008) at Mount Whaleback at depth is chemically and mineralogically distinct from the regional black and red shales produced during regional metamorphism. The shale is locally intensely carbonate altered and adjacent to high-grade ore consists of fine- to medium-grained dolomite cut by numerous chlorite and ferroan-dolomite and/or ankerite veinlets (cf. Webb et al., 2004; fig. 4 of Thorne et al., 2008). Fine-grained ferroan-dolomite and ankerite crystals are spatially associated with many of these veins (fig. 3A of Thorne et al., 2008). Powell et al. (1999) and Brown et al. (2004) studied at least five generations of quartz ± hematite veins surrounding the Mount Whaleback deposit. At least one generation of hematite ± quartz veins that is locally enclosed by microplaty hematite ore (Fig. 1K) was suggested to be synchronous to the main ore-forming event.

Three compositionally distinct zones of hydrothermal alteration have been described at the Mount Tom Price deposit (Taylor et al., 2001; Thorne et al., 2004, 2005, 2008). distal magnetite-siderite-iron silicate (Fig. 6A, B), intermediate hematite-ankerite-magnetite, and proximal martite-microplaty hematite zones (Figs. 6C-F, 7A, B; Tables 1A, 2). The distal alteration zone (Fig. 6D, Fig. 7A) is about 30 m wide and characterized by replacement of chert layers by bladed magnetite (50–200 μm), siderite (800–1,500 μm), and iron silicates (20–110 μm; Fig. 1B). The shape of the magnetite is atypical, suggesting that it is pseudomorphous after preexisting minerals, likely siderite (Fig. 1C). Where preserved, microcrystalline quartz is intergrown with euhedral magnetite and forms inclusions within siderite blades. Pyrite crystals (50–300 μm) are finely disseminated within shale layers. Apatite is observed as euhedral crystals intergrown with anhedral chlorite (Fig. 7A). Veins in the distal alteration zone (V2 veins, Table 3) are observed in orientations both crosscutting and parallel to the lamination of the BIF host rock and are filled by siderite and stilpnomelane (Table 3). The wall rock adjacent to these veins is locally brecciated with clasts (5–60 mm) of magnetite-rich mesobands in a matrix of bladed magnetite, siderite, and iron silicates.
HYPOGENE ALTERATION ASSOCIATED WITH HIGH-GRADE BIF-RELATED IRON ORE

FIG. 6. Core samples from the North deposit, Mount Tom Price. A. Least altered BIF showing chert and magnetite banding. B. Least altered BIF showing hematitized portion. Note that hematite invades bands of BIF. C. Hard hematite ore, with preservation of bands characterized by alternating martite- and microplaty hematite-rich bands. D. Magnetite-siderite-iron silicate distal alteration zone showing preservation of bands. E. Hematite-ankerite-magnetite intermediate alteration zone with localized brecciated magnetite bands. F. Martite-microplaty hematite ore (supergene). Note preservation of banding and goethite infill. Abbreviations: FeSil = iron silicate, Gt = goethite, Mag = magnetite, MpHem = microplaty hematite, Mt = martite, Sd = siderite (adapted after Thorne et al., 2004).

FIG. 7. A. Paragenetic alteration sequence in BIF at the North deposit, Mount Tom Price. Proximity to iron mineralization increases to the right. Zone widths shown here do not relate to actual widths observed in the field (adapted after Thorne et al., 2004). B. Paragenetic sequence of oxide minerals between nonmineralized BIF and high-grade ore for various deposits in the Hamersley province (Thorne et al., 2007).
Intermediate alteration zones are about 15 m in width (Figs. 6E, 7A) and characterized by the mineral assemblage with hematite-ankerite-martite that replaces quartz, siderite, magnetite, and iron silicates (Fig. 6). Microplaty hematite (10–60 μm) forms both individual blades and dense clusters as overgrowth on magnetite (Fig. 1C) and as individual plates within ankerite crystals. Euhedral and bladed magnetite show minor replacement by martite along crystal boundaries, whereas iron silicates are replaced by anhedral and microplaty hematite (Fig. 7B). Ankerite crystals (0.20–4 mm) form variably recrystallized mosaics with irregular grain boundaries. Apatite is present as inclusions within magnetite and microplaty hematite and as anhedral crystals within ankerite crystals.

There are two vein sets associated with the intermediate alteration zone. These are ankerite-hematite (V3, Table 3) and pyrite veins (V4, Table 3). The V3 veins are especially abundant and crosscut BIF and shale bands, whereas pyrite veins are rare and are in the fracture zones that postdate hematite-ankerite-magnetite alteration. Brecias are matrix supported and consist of angular and rotated clasts (5–40 mm) of altered iron formation within an ankerite-microplaty hematite matrix.

The proximal alteration zone is about 15 to 30 m wide and characterized by the mineral assemblage martite-microplaty hematite-apatite. Martite and anhedral hematite replace magnetite and iron silicates of the intermediate alteration assemblage, respectively. Minor amounts of ankerite and chlorite are located in thin layers, together with some fine-grained martite. Intergranular porosity increases significantly. Locally, late-stage talc alteration crosses the proximal alteration zone.

The high-grade ore consists of martite and microplaty hematite, with the ore preserving the meso- and microlayering of the primary BIF bands. The ore is characterized by porosity and of randomly oriented, fine-grained platy hematite and martite. Individual microplates (0.001–0.25 mm) are bladed or cubic shaped with sharp terminations (Taylor et al., 2001) that exhibit intensive overgrowth on subhedral martite grain margins. High-grade martite-microplaty hematite ore was formed by the interaction of supergene fluids with the proximal martite-microplaty hematite-apatite assemblage. Magnesium silicates were converted to a kaolinitic residue, thereby significantly thinning the shale layers, apatite was dissolved, and both calcium and phosphorus were leached from the ore. Dolerite dikes that bound the high-grade ore are intensely altered and largely composed of chlorite and talc, with accessory leucoxene and pyrite. The chlorite (dibantite) forms pseudomorphs after primary phases such as pyroxene and amphibole, whereas talc or chlorite form pseudomorphs after feldspar.

As for Paraburadoo, preliminary work at the deposit has identified at least one hydrothermal alteration zone transitional between the BIF protolith and the high-grade iron ore (Thorne et al., 2007; fig. 6 of Thorne et al., 2008). The proximal alteration zone is characterized by a hematite-dolomite-chlorite-pyrite assemblage, as is at least 20 m wide, and is emplaced below the depth of modern weathering. It forms by the replacement of magnetite by hematite and the replacement of quartz by dolomite, chlorite, and pyrite. Matrix-supported BIF breccias are observed locally and consist of angular and rotated clasts of altered BIF in a dolomite-microplaty hematite-chlorite-pyrite cement. The mineral assemblage is similar to that of the intermediate alteration zone at Mount Tom Price. The entire orebody at Paraburadoo now lies within the Tertiary weathering profile and some high-grade martite-goethite ore is mixed with the martite-microplaty hematite ore. The latter ore is also very similar to that described at Mount Tom Price.

South African Iron Deposits

Regional geologic setting

The Late Archean to Paleoproterozoic (2.06–2.65 Ga) Transvaal Supergroup on the Kaapvaal craton is comprised of lowermost protobasinal volcano-sedimentary deposits, followed by platform carbonate rocks (dolostone, minor limestone), and BIF of the Ghaap-Taupane-Chuniespoort Groups. These are unconformably overlain by clastic sedimentary and volcanic rocks of the Postmasburg-Segwagwa-Pretoria Groups (Eriksson et al., 1995). Voluminous iron formations of the Asbesheuwels Subgroup in the Griqualand West subbasin and the laterally correlative Penge Formation of the Transvaal Supergroup (2.43–2.45 Ga, Pickard, 2003) host all economically important high-grade iron ore deposits of South Africa. The Asbesheuwels Subgroup thickness ranges from 250 (platform facies) to 750 m (basal facies), with a complex mineralogical and textural facies architecture that is documented in detail by Beukes (1983) and Beukes and Gutzmer (2008). Mineralogical and textural iron formation facies can be correlated over large distances, suggesting deposition on a continental platform (i.e., a submerged Kaapvaal craton) in a very stable depositional and/or tectonic environment (Beukes, 1983).

The most important South African high-grade hematite ore deposits are hosted by the Manganore Iron Formation, an extensively oxidized remnant of the Asbesheuwels Subgroup, in the Griqualand West subbasin. The Manganore Iron Formation and associated high-grade iron ores are preserved in depressions that are related to Paleoproterozoic (2.1–2.2 Ga) peneplanation, lateritic weathering, and karstification (Gutzmer and Beukes, 1998). High-grade iron ores within the Manganore Iron Formation are developed immediately below the erosional unconformity that marks this peneplanation event and are thought to be of ancient supergene origin (Beukes et al., 2003). These are, therefore, not the subject of this contribution.

In contrast, high-grade iron orebodies exploited at the Thabazimbi deposit in Limpopo province, which are hosted by the basal portion of the Penge Iron Formation, as well as the Zeekoeibaantop prospect of the basal portion of the Kuruman Iron Formation near Nauga in Griqualand West, are regarded to be of hydrothermal origin. The orebodies of the Thabazimbi deposit are particularly well documented (Strauss, 1964).

The Thabazimbi deposit

The Thabazimbi iron deposit comprises a series of tabular, strata-bound orebodies aligned along the faulted basal contact between the Neo-Archean to Paleoproterozoic Penge Iron Formation and the dolostones of the Malmani Subgroup, immediately above a chest-rich, black carbonaceous shale unit that is intensely sheared and highly oxidized (Van...
Deventer et al., 1986; Gutzmer et al., 2002). The concentration of high-grade hematite orebodies near the structural, basal contact of the Penge Iron Formation suggests an ore-forming process that was both structurally and lithologically controlled. Formation of the high-grade iron ores took place after contact metamorphism related to the intrusion of the Bushveld complex (2.05 Ga) but before the deposition of red beds of the Waterberg Group at ca. 1.9 Ga (age of ore formation by de Kock et al., 2008).

The hydrothermal alteration attending ore formation has been described by Netshiozwi (2002) and Netshiozwi et al. (2002) and much of the present review is based on their work (Tables 1, 2). The origin of the Thabazimbi ores is also discussed by Beukes et al. (2002) and (Gutzmer et al., 2002, 2006).

The metamorphosed Penge Iron Formation is very fine grained with alternating dark iron oxide bands and light gray laminae containing chert, grunerite, and ankerite, with the following main assemblage (Fig. 8A, B): chert and/or microquartz-magnetite (±dusty hematite inclusions)-stilpnomelane-calcite-ankerite-grunerite. Variations in the amounts of these phases characterize different mineralogical facies of the metamorphosed orebodies. Reddish chert (QTZ-I) and microquartz (QTZ-II) are the most abundant constituents (Fig. 9), and these enclose dusty hematite (He-I). Diagenetic magnetite probably formed by the reduction of dusty hematite is the dominant iron oxide. Stilpnomelane and closely associated calcite (CC-I, Fig. 9) and ankerite are also interpreted to be of diagenetic origin. Carbonates and stilpnomelane predate the formation of grunerite that is thought to be a product of contact metamorphism related to the intrusion of the giant Bushveld layered igneous complex. Minor to trace amounts of pyrite,apatite, and chlorite are also present (Fig. 9).

All high-grade iron orebodies at the Thabazimbi deposit are surrounded by a halo of oxidized and partly enriched BIF, i.e., alteration of the Penge Iron Formation (Figs. 1F, SC; Tables 1A, 2A). Platy hematite (He-II, Fig. 9) is disseminated in chert, imposing a distinctive pinkish color, and probably represents recrystallization of primary dusty hematite and replacement of grunerite and ankerite by hematite and the formation of martite at the expense of magnetite. Locally, chert has been either replaced by microplaty hematite or it has been leached, thus giving rise to porosity. Veins and veinlets are filled by coarse tabular hematite (referred to as specularite by...

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**Fig. 8.** Photographs illustrating the typical appearance of the least altered and oxidized Penge Iron Formation, South Africa (from Netshiozwi, 2002). A. Mesobanded iron formation characterized by thick grunerite-chert nesobands closely intercalated by thick magnetite-rich laminae. This type of iron formation is likely to produce a massive hematite ore during the ore-forming process. B. Magnetite-rich grunerite-chert iron formation. Note that grunerite-chert nesobands are characterized by thin laminae of magnetite. C. Oxidized cherty facies of the Penge Iron Formation. Note porosity developed where chert was leached. D. Laminated hematite ore comprised of massive hematite beds closely alternating with thin porous hematite laminae. Note minor brecciation toward bottom right of the sample. The breccia fragments are cemented by porous hematite. Photographs (E) to (J): display different types of high-grade hematite ore at the Thabazimbi mine. E. Typical appearance of massive hard hematite ore, mainly comprised of fine-grained martite (Mt) and patchy hematite. F. Finely laminated hematite ore characterized by very thin (millimeter-thick) massive hematite laminae alternating with dusty hematite laminae of distinctly greater thickness (3–5 mm). G. Folded, laminated high-grade hematite ore. Note that the folding of the ore-precipitated mineralization. Photos (H), (I), and (J) show samples of carbonate-hematite ore. H. Hematite-calcite breccia with rectangular to angular hematite fragments cemented by sparry calcite. I. Dolomite breccia defined by irregular outlines of hematite ore fragments, cemented and replaced by sparry dolomite. J. Hydraulic hematite breccia with well-defined angular fragments of hematite cemented by sparry calcite.
Gutzmer et al., 2002, and Netshiozwi, 2002) and coarse crystalline quartz fills fractures and veinlets. The effects of hydrothermal alteration are similar to the intermediate alteration zone in the Carajás iron district.

The formation of high-grade hematite orebodies at the Thabazimbi deposit are the result of the gradually increasing degree of SiO₂ leaching (Fig. SD-1) and closely associated volume loss that created widespread brecciation of the high-grade hematite ore (Fig. SD; Gutzmer et al., 2005). The presence of finely laminated and massively textured iron ores devoid of brecciation, on the other hand, suggests either the introduction of iron and replacement of silica-rich laminae by porous aggregates of microplaty hematite or the remobilization of iron during the ore-forming process. The formation of such high-grade hematite-martite ores at the Thabazimbi deposit corresponds to the proximal alteration zones described in the Hamersley and Carajás iron districts.

Hard hematite ores dominate at the Thabazimbi deposit. They may be massive, laminated, and brecciated and are mainly composed of martite, patchy (He-IV), and microplaty hematite. Massive ores may have originated by replacement of the banded Penge Iron Formation, including fine-grained microplaty hematite. These ores are either devoid of sedimentary laminarization or may display traces of mesobanding. Patchy hematite and martite constitute the massive bands of these laminated hematite ores. Anhedral, patchy hematite is defined by lobate grain boundaries, and it is interpreted to have formed during the recrystallization of martite. Tabular
hematite is observed in laminated hematite ore, infilling porosity, and also in crosscutting quartz veins. There is a gradual transition from laminated through weakly fragmented to brecciated ore types. Breccias evolved from hard hematite ores that were reduced by collapse and secondary tabular hematite infill. Breccia fractures are variously cemented by very fine grained anhedral hematite, microplaty hematite, and tabular hematite.

In addition to high-grade hematite-martite ores, four mineralogically distinct types of iron ore have been recognized: (1) goethite-rich, (2) low-grade dolomite-hematite, (3) low-grade calcite-hematite, and (4) talc-hematite. Goethite-rich ores are restricted to the geologically recent weathering environment, whereas carbonate-rich ores interfinge with high-grade hematite-martite ore at depth. Tale-rich ores are found locally along the fringes of the high-grade iron orebodies and along the contact of high-grade iron ore with the basal shale.

The iron oxides in the low-grade carbonate- and talc-rich hematite ores are of very similar texture and mineralogy (i.e., hematite, martite, remnant magnetite) to high-grade ores and are intimately intergrown with sparry carbonates and talc (Fig. 5H-J; Tables 1A, 2A). This led Netshiozwi (2002) and Gutzmer et al. (2005) to the conclusion that ore formation was genetically associated with the formation of calcite, dolomite, and talc. It is interesting to note that despite the fact that both sparry calcite and dolomite appear to be co-genetic with the ore-forming microplaty hematite, these carbonates are not observed in close spatial association or even intergrown. Indeed, they appear always to be separated by either an oxidized shale bed or a bed of high-grade hard hematite ore.

Supergene alteration of both high-grade hematite and carbonate-hematite ores is common and leads to friable, often goethite-bearing high-grade ores.

The Zeekoebaart prospect

The Zeekoebaart prospect is in an area of intense deformation along the southwestern margin of the Kaapvaal craton in Griqualand West (Harding, 2004). It consists of a strata-bound lens of hard hematite ore located near the basal contact of the iron formation to the underlying Klein Naute Shale (Beukes and Gutzmer, 2008). The present account of the high-grade iron ores of the Zeekoebaart prospect and its relationship to the surrounding Kuruman Iron Formation is based on the work by Harding (2004).

The high-grade iron ore at the Zeekoebaart prospect is surrounded by a wide halo of oxidized iron formation. This oxidized BIF is comprised essentially of alternating layers of iron oxide (hematite and martite) and fine-grained quartz (Fig. 10; Tables 1A, 2A). Microplaty hematite laths form in the iron-poor or cherty layers, whereas patchy hematite is observed to replace martite in the iron oxide-rich layers. Overall, the oxidized BIF comprises well to oxidized BIF at the Thalasimbi deposit and the intermediate alteration zone in the Carajás iron district. Short veins of coarse-grained quartz crosscut the oxidized BIF; hematite and carbonates are conspicuously absent. Unaltered BIF is known to crop out at a distance of ca. 1 km from the Zeekoebaart prospect. This unaltered basin facies of the Kuruman Iron Formation encompasses alternating iron-rich magnetite-hematite-siderite (ankerite) and iron-poor chert-siderite-stilpnomelane mesolayers (Beukes, 1983). Comparison of this unaltered protolith to oxidized BIF reveals the effects of marked oxidation as well as SiO2 leaching (Fig. 10).

Proximal alteration at the Zeekoebaart prospect around the high-grade iron orebody involved leaching of SiO2 from oxidized iron formation and formation of laminated and massive hard hematite ores (Fig. 11; Tables 1A, 2A). The laminated ore is composed of dense layers of fine-grained patchy hematite (former iron oxide bands) that alternate with layers of coarser grained, microplaty hematite (former chert layers; Fig. 11). Patchy hematite is interpreted to be the product of recrystallization of martite. Massive hematite ores, in contrast, develop at the expense of thicker mesobanded BIF and lacks the clear separation of microplaty and patchy hematite. Thin quartz veins, up to 5 cm wide and containing minor amounts of coarse blades of hematite, are randomly distributed in the high-grade ore.

Goethite appears only in minor amounts and is attributed to the influx of oxidizing meteoric water along pore spaces and microfractures. The goethite forms as open-space infill but also as a replacement product of hematite and even quartz.

Commonalities and Differences in Hyopgene Alteration of BIF

Mineral associations

The most obvious common attribute of all deposits included in this review (Tables 1, 2; Figs. 3, 7, 9, 11) is a very similar paragenetic sequence of iron oxides. Microcrystalline hematite and fine-grained magnetite, associated with Fe-rich
silicates, siderite, and ankerite, dominate the least altered protolith BIF. In distal alteration zones of the Carajás and Hamersley districts hematite and locally iron-rich silicates and carbonates are replaced by euhedral hydrothermal magnetite (Tables 1, 2). This magnetite is typically more abundant and coarse grained than magnetite formed during metamorphism. The Thabazimbi deposit and Zeekoebaart prospect lack unequivocal evidence for the formation of paragenetically early hydrothermal magnetite (Figs. 9, 11). At the Hamersley and South African districts, extensive oxidation is reflected by martitization and the formation of hematite at the expense of Fe²⁺-rich carbonates and silicates, marking the intermediate alteration zones, which in Carajás is characterized by oxidation of magnetite. Several textural types of hematite, including martite, are located in the oxidized BIF and are similar in appearance to hematite in associated high-grade iron ores. This may be used to support the cognetic relationship between oxidized BIF and high-grade hematite ore.

In order to transform an altered and/or oxidized BIF to high-grade ore very effective removal of SiO₂ needs to take place. Our case studies revealed that the removal of SiO₂ can take place in two ways. The simplest process involves direct leaching of SiO₂, which is accompanied by widespread oxidation. The second process involves leaching of hydrothermal carbonate that replaced SiO₂ during hydrothermal alteration. The dissolution of these carbonates can be caused by meteoric water influx into fault zones or geologically recent deep chemical weathering.

Hydrothermal carbonates are observed in all our case studies, located in veins and breccias (Table 3), with the notable exception of the Zeekoebaart prospect. They are present in the distal (siderite) and intermediate (ankerite) alteration zones at Mount Tom Price, whereas only the proximal

![Mineral paragenesis chart](https://example.com/mineral_paragenesis.png)

**FIG. 11.** Mineral paragenesis chart for the hematitized iron formation and hematite iron ore types from the Zeekoebaart deposit, as determined from the study by Harding (2004). The distribution of oxide minerals along the alteration path is shown in the upper figure; the 1st and the 2nd alteration stages represent the distal and proximal alteration zones, respectively. Mineral abbreviations: Hem = hematite, MH = microplaty hematite, PH = patchy hematite.
alteration zone at Thabazimbi displays calcite and dolomite. In both deposits, carbonates effectively replaced chert in altered BIF surrounding high-grade orebodies; in the proximal alteration zones carbonate is then replaced by hematite. In Carajás, calcite dominates in the distal alteration zone, with dolomite and kyanoharite in the proximal alteration zone (the complete hydrothermal alteration assemblages are listed in Table 1). Carbonate-altered BIF has, as yet, not been identified at the Mount Whaleback deposit. This is tentatively attributed to a lack of geologic information on the deeper parts of the ore-forming system, as the closely associated Mount Silvia shale has at least locally provided evidence for extensive carbonate alteration (Webb et al., 2004). Textural relationships between the iron oxides in carbonate-altered BIF and high-grade iron ores in the Hamersley district strongly suggest that carbonate alteration preceded the formation of high-grade iron ore, and that the transformation of carbonate-altered BIF to high-grade hematite-martite ore is due to oxidation and complete decomposition of the hydrothermal carbonates.

In the Carajás iron district, only the intermediate alteration stage (martite-microplaty hematite-quartz assemblage) lacks carbonate. Calcite dominates in the distal alteration zone (magnetite-calcite-quartz-pyrite assemblage; Fig. 3) in veins and breccias. The proximal zone (hematite ± carbonate-quartz assemblage) has dolomite and kyanoharite, covered by hematite dust, in equilibrium with hematite and rare quartz. In addition, carbonate-bearing late-stage veins and breccias may also be associated with the proximal alteration zone, replacing the martite-rich layers. Sulfides are only significant in the Carajás district where these, together with rare gold, form in quartz-carbonate veins associated mainly with distal hydrothermal alteration zones.

Finally, it is important to note common patterns in the hydrothermal alteration of mafic wall rocks, dikes, and sills, and even shales associated with our case studies. Dalstra and Guedes (2004) were the first to point out that in several important deposits and/or districts mafic dikes and sills that intruded BIF protolith prior to hydrothermal ore formation display marked hydration as expressed by the formation of abundant Mg chlorite and talc. Mafic wall rocks, sills, and dikes in the Carajás district show very similar alteration effects, as they were transformed into hematite- and talc-bearing chloritites. Even the formation of talc-rich iron ores along the contact between high-grade iron orebodies and basalt shale at the Thabazimbi deposit may be attributed to intensive hydration. At Carajás, the widespread and dominant chlorite, talc, white mica, and albite in the hydrothermally altered mafic wall rocks that surround the high-grade ores is a function of the mafic wall rocks which are typical of the granite-greenstone belt setting. This is in contrast to the Hamersley district and Thabazimbi deposit, where iron silicates are common and reflect the sedimentary host rocks typical for the host sedimentary basins. Iron silicates are absent at Carajás. 

**Vein formation, brecciation, and volume loss**

Evidence for multiple events of hydrothermal fluid flow and associated remobilization is provided at all the studied sites by different sets of veins. It is important to point out, however, that veins and breccias can also form pre- and post-the transformation of BIF to high-grade iron ore. Detailed geologic mapping, diamond core logging, and a combination of petrographic and geochemical analyses of veins in the three case studies provide specific evidence for the contemporaneous emplacement of particular vein and breccia types during the hydrothermal alteration and mineralization processes in the BIF, but also in the surrounding wall rocks. Veins emplaced synchronous with iron mineralization processes display typically crack-seal, massive, or brecciated textures, and their formation may be genetically linked to deformation that can coincide with the ore-forming fluid-flow event. However, veins rarely are observed within high-grade iron ore. The formation of high-grade iron ore is related to the advanced stage of the mineralization process where desilification, with or without carbonate alteration, makes the preservation of veins unlikely.

Moreover, it is pervasive rather than focused fluid flow that is required to transform a large volume of BIF into high-grade iron ore. Because unaltered BIF is typically impervious, the generation of porosity (i.e., produced by the dissolution of quartz and/or carbonate) is regarded as instrumental for the progress of alteration and iron ore formation. It comes thus as no surprise that a marked increase in porosity (and, by inference permeability) is associated with all hydrothermal alteration zones in all the districts under consideration.

Provided that an average iron formation contains between 25 and 45 wt percent SiO2 (Klein, 2005), and that quartz and silicates have a distinctly lower density than magnetite and hematite, a volume loss of 60 to 70 vol percent is expected to be associated with the complete removal of SiO2 from the protolith BIF. Similarly, this would apply for the dissolution of hydrothermal carbonate that may have replaced chert prior to high-grade ore formation. This massive volume loss could be accommodated by a combination of (1) porosity increase, (2) collapse brecciation, (3) bedding-parallel compaction, and (4) the introduction of additional iron by the hydrothermal fluid. A marked porosity increase and collapse brecciation have been described in all of the deposits included in this review. Bedding-parallel compaction has to be inferred to explain the origin of high-grade iron ores sporting laterally continuous fine lamination reminiscent of the protolith. In deposits for which simple SiO2 leaching is invoked (Carajás, Thabazimbi, Zeekoevaart), the increase of porosity was gradual. However, although it is readily noted during incipient alteration, a rapid increase of porosity accompanies the transition from intermediate to proximal alteration zones, i.e., the contact of oxidized BIF to high-grade iron ore. The marked porosity of the high-grade ore is due to a very open, scaffold-like intergrowth of microplaty or platy hematite that is thought to have formed as open-space infill. This very porous intergrowth often is observed along distinct layers that are tentatively identified as former chert layers, which are separated by much less porous hematite-martite laminae that are thought to represent iron-rich layers of the protolith BIF. In contrast, no such gradual increase of porosity has been noted for deposits that have experienced carbonate metasomatism as an integral part of high-grade ore formation (Hamersley district, possibly Thabazimbi). Instead, the increase of porosity is almost entirely associated with the contact of carbonate-altered BIF to high-grade iron ore. This increase is due to the dissolution of the hydrothermal carbonate. Carbonate dissolution may be
attributed either to most intense hydrothermal alteration (Hamersley district, Thabazimbi deposit) or, alternatively, be due to geologically recent chemical weathering processes.

Dissolution collapse brecciation is another important process that accommodates the volume loss during the hydrothermal transformation of BIF protolith to high-grade hematite ore. It has been described in all the deposits and/or districts, with the notable exception of the Zeekoebaart prospect (Taylor et al., 2001). Typically, brecciation is noted already during incipient alteration, although it may be restricted to the immediate vicinity of veins (Hamersley and Carajás districts). At Thabazimbi brecciation is usually found to increase gradually toward the high-grade iron ore, likely reflecting the increasing amount of volume loss (Gutzmer et al., 2005). In the high-grade iron orebodies, dissolution collapse brecciation usually abounds. It is observed at different scales and intensity, ranging from millimetric brecias to hand specimens of large volumes of iron ore marked by chaotic collapse brecciation (Nethshiozi, 2002). An open scaffold of microplaty or somewhat coarser grained platy hematite usually forms the cement of such brecias. At the Thabazimbi deposit, abundant brecciation is not restricted to the high-grade iron orebody but is also locally observed in low-grade ores with hydrothermal carbonate cement (Nethshiozi, 2002; Gutzmer 2005). The presence of talc as breccia cement appears to be unique to the Thabazimbi deposit. In Thabazimbi, this is a synore process. In Carajás, brecciation is associated with the early to advanced, including late, stages, with these brecias being cemented by quartz, carbonate, and hematite in the ore-proximal zone; locally, this carbonate cement constitutes close to 50 vol percent, thereby decreasing the ore grade (Figueiredo e Silva et al., 2008). Interpretations regarding the relationship of such hydrothermal carbonate and talc cement differ considerably and indicate the need for more detailed investigations.

Discussion

Three aspects of the hypogene alteration processes described for the selected case studies above warrant further discussion. These are the (1) significance of the early hydrothermal magnetite, (2) origin of soft ore in the Carajás district, and (3) question of whether iron is introduced or simply remobilized with the Carajás iron deposits as an example.

Hydrothermal magnetite

Regardless of the iron oxide sequence, magnetite is an early oxide phase in the alteration paragenesis (Tables 1, 2), followed by various types of hematite. The South African deposits are an exception, since in these magnetite is considered to have formed as a prealteration phase (Figs. 9, 11). The formation of hydrothermal magnetite suggests that the early-stage fluid had a fO2 at the magnetite equilibrium conditions (Lobato et al., 2005a) and interacted with rocks that variably contained dominant ferric (Carajás) or ferrous (Mount Tom Price) phases.

Silica leaching

The leaching of SiO2 can be attained by solutions that are oxidizing enough to render ferric oxide immobile; such is the case for hydrothermal fluids or low-temperature supergene fluid. In the case of hydrothermal fluids mildly alkaline and silica-undersaturated conditions enhance effectively the SiO2 transport (Barnes, 1997). Remobilization of SiO2 to form quartz-rich veins, paired by a distinct increase in porosity (documented for the Carajás district, Thabazimbi and Zeekoebaart), lend important support to the concept of SiO2 leaching. Carbonates, silicates, and sulfide minerals are conspicuously absent from the oxidized BIF. This model would require very high fluid/rock ratios and is thus likely to apply to orebodies emplaced along major dilational structures acting as important fluid pathways. The Zeekoebaart prospect may be regarded as the most plausible example for simple SiO2 leaching. For Carajás, Lobato et al. (2005a) and Figueiredo e Silva et al. (2008) suggest relatively hot magmatic fluids interacting with a colder, exhumed rock sequence, to account for SiO2 leaching, with some parallels to a model proposed by Gruner (1937, 1938) for the Lake Superior region.

Hypogene soft ore in Carajás

We propose that a hypogene, porous hard to soft, but relatively cohesive hematite ore evolved as a result of porosity enhancement and quartz-chert leaching at Carajás. The ore preserves martite; microplaty and anhedral hematite partially fills the existing pores (Lobato et al., 2005a, b). Following early-stage jaspilite alteration, with significant SiO2 leaching, jasper discoloring and porosity development (Fig. 2), vugs were initially filled predominantly by microplaty hematite as seen in figure 9E of Figueiredo e Silva et al. (2008). The hard hematite ore that is enveloped by this soft ore, which itself is in contact with jaspilite (Fig. 6 of Figueiredo e Silva et al., 2008), would thus represent an evolutionary stage of the hypogene soft ore. This is because it developed due to advanced cavity filling via precipitation of additional microplaty (after martite), platy (Fig. 1H), and also euhedral and comb-textured, bladed and/or tabular hematites (Fig. 1M-O), with the latter ones especially in veins. The possibility for a hypogene origin for soft hematite ores was also indicated by Gruner (1937) and Mann (1953) for the Lake Superior region and the Marquette and Gogebic ranges, respectively.

Conclusions

Comparison of hydrothermal alteration at Carajás, Hamersley, Thabazimbi, and Zeekoebaart revealed important similarities (Tables 1, 2), including: (1) a similar paragenetic sequence of iron oxides, marked by an abundance of open-space filling and replacement textures; (2) the lack of a penetrative fabric of alteration lithologic units and high-grade ores; and (3) the importance of porosity and brecciation to accommodate volume loss.

Marked differences include the following:

1. Carbonate metasomatism in hydrothermal alteration zones: Hydrothermal carbonate is observed in all three case studies and located in veins and breccias. At Tom Price in distal and intermediate alteration zones, in the Thabazimbi deposits in the proximal alteration zone, and in Carajás in the distal and proximal alteration zones in jaspilites. The Mount Whaleback deposit lacks significant carbonate alteration; however, this may be due to lack of information on the deeper
parts of the orebody. At Tom Price the transformation of carbonate-altered BIF in the intermediate alteration zones to high-grade martite-microplaty hematite ore in the proximal alteration zone is due to oxidation of magnetite and leaching of carbonate. In contrast, at Carajás dolomite and lathalumite veins and breccias are stable in the proximal alteration zone and locally replace martite-rich layers. Significant amount of siderite as the dominant carbonate species is only recorded in the Hamersley iron deposits.

2. Silicate-alteration phases such as stilpnomelane are only present where hydrothermally altered BIF is part of a sedimentary basin, such as the Hamersley basin or the Asbestos Group in the Griqualand West subbasin, and the laterally extrusive Ponge Formation of the Transvaal subbasin of the Transvaal Supergroup. The basalt-hosted Carajás iron deposits are devoid of these alteration minerals.

3. Chlorite, talc, white mica, and albite are part of the alteration assemblage in basalt-hosted iron deposits, for example, in Carajás (Tables 1B, 2B) or mafic dikes that are spatially and temporally associated with iron mineralization, for example, in the Hamersley province.

4. Hydrothermal sulfides present in significant quantities are recorded only in Carajás. Their appearance may relate to the more sulfur-rich basaltic wall rocks (Zucchetti, 2007) and/or to the magmatic hydrothermal fluids postulated to be one major fluid source for the ore-forming event in the Carajás district (Lobato et al., 2005a; Figueiredo e Silva et al., 2008). Possible similarities of these iron ore-forming fluids with those associated with the formation of hydrothermal magmatic deposits, dominantly rich in the ore elements Cu and Au, and that are present in the same district (including IOCG-type deposits), need to be carefully evaluated (Lobato et al., 2005a, b).

**Exploration Significance**

The acknowledgment that complex and possibly polyphase hydrothermal systems, including both ascending and descending hydrothermal fluids, cause the transformation of BIF (including jaspilite) to high-grade hematite-martite ore has significant implications for the exploration of blind orebodies (Hagemann et al., 2006). Structures such as fault zones, shear zones, and/or isoclinal folds serve as fluid pathways for the hydrothermal fluids and, therefore, can be used to constrain areas of maximum fluid flow and associated hydrothermal alteration and iron mineralization. Hydrothermal alteration minerals, such as carbonates, chlorite, and magnetite, may form distinct alteration zones, both lateral and above and below the high-grade iron orebodies, thereby providing a hydrothermal alteration footprint that can be used in the search for new high-grade iron deposits or extensions of existing orebodies; in particular, those that are located concealed below cover sequences or nonmineralized BIF. These hydrothermal alteration indicator minerals in both nonmineralized BIF and surrounding wall rocks can thus be used as vectors toward iron mineralization.

In the Hamersley-type deposits, the identification of hydrothermal alteration minerals, veins, and breccias at the contact between nonmineralized BIF and low- and high-grade iron ore are commonly subtle and require detailed mapping, core logging, and petrographic and textural work. Of particular importance is the petrography and geochemistry of carbonate and quartz-carbonate veins. With increasing proximity to high-grade iron ore, the carbonate species change from siderite to ankerite and calcite to dolomite at the Mount Tom Price (Fig. 7) and Paraburdo deposits, respectively. The manganese content of the carbonates (up to 10 wt %) also increases with increased proximity to high-grade ore (Hagemann et al., 2006).

The hydrothermal alteration of dolerite dikes that are spatially and temporally related to iron mineralization (cf. Taylor et al., 2001) provide another vector toward iron ore mineralization. The dikes commonly display intense talc-chlorite alteration, particularly where in close proximity to the iron orebodies. Although rarely containing significant iron enrichment, they bound the mineralization at several orebodies at the Paraburdo, Chamor, and Mount Tom Price deposits (Dalstra et al., 2005; Taylor et al., 2005, Thorne et al., 2006). The identification of quartz ± carbonate veins and/or breccias that are emplaced contemporaneous with the iron mineralizing event, and often in distal alteration zones, serves as another vector toward iron mineralization.

In the Carajás deposits the contact between nonmineralized and hydrothermally altered and mineralized jaspilites can be gradational. In this case a distal alteration zone can be defined by magnetite, quartz, and carbonate veins that may locally be layer-parallel or crosscutting with respect to the jaspilite (Fig. 3). The intermediate alteration zone is made up of variable proportions of quartz ± hematite veins, magnetite, microplaty hematite, anhedral-subhedral hematite, euhedral-tabular hematite, and locally sulfides and gold (Fig. 3). Hydrothermal alteration minerals and zonation are particularly well distributed in the basalts that surround the nonmineralized jaspilites and iron orebodies (Zucchetti et al., 2007; Figueiredo e Silva et al., 2008). The distal alteration zone, which contains carbonate and locally hematite, can extend up to 80 m away from the iron orebodies and, therefore, can be used as vectors toward iron mineralization. The intermediate alteration zones are characterized by chlorite, quartz, albite, white mica, carbonate, hematite, and traces of magnetite and sulfides (Fig. 4). Therefore, any drilling program that aims to identify either deep extensions of existing orebodies or new, concealed iron mineralization can utilize these distal alteration minerals and veins as vectors toward iron mineralization.

Geologic mapping, core logging, and understanding key hydrothermal alteration (i.e., indicator) minerals in three dimensions may provide important constraints on the geometry of the hydrothermal footprint of any given iron mineral system (Hagemann et al., 2006; Figueiredo e Silva et al., 2008; Thorne et al., 2008). Therefore, the rapid identification of these indicator minerals in the field and in diamond core is crucial. Besides sound petrographic analyses, methods such as PIMA analyses, airborne HyMap™, and satellite ASTER imagery, both handheld on diamond core and airborne have been developed in the past years. These techniques significantly assist not only in identifying these often fine-grained hydrothermal alteration minerals but also constraining their three-dimensional distributions. The analyses of diamond core using the new Fe HyLogger™ system can assist in the rapid identification and calculation of modal mineralogy, rapidly differentiating between minerals and products of
chemical weathering (e.g., clay and goethite). The presence and distribution of the latter is particularly important in exploring in arid areas or areas with significant lateritic cover, such as in Brazil, India, and Australia (Morris, 1985).

**Future Work**

This contribution highlights the limited amount of detailed mineralogical and petrographic studies of high-grade BIF-hosted iron deposits that take cognizance of the presence of alteration halos. Detailed delineation of the distribution of distinct alteration mineral assemblages are required to delineate the true size and extent of distinct, mappable alteration zones in three dimensions both in BIF and wall rocks. The processes that characterize the transformation of unaltered to hydrothermally altered BIF need to be better constrained, in particular with respect to the role of iron-rich carbonates ( siderite and ankerite) that often abound in unaltered BIF protolith or ore formation (Hamersley). Detailed alteration studies need to be extended to deposits and districts that have been affected by high-grade, sometimes polyphase regional metamorphism, i.e., the Quadrilátero Ferrífero (Rosière and Rios, 2004; Rosière et al., 2008) and Krivoy Rog districts (Plotnikov, 1994; Dalstra and Guedes, 2004). Finally, the possibility that not all porous, friable and soft high-grade iron ores are supergene in origin, as suggested for the Carajás district by Lobato et al. (2005a, b), needs to be further investigated.

Geochronological data are urgently required to advance our understanding of the different alteration zones, as defined in this study, and their relationship to each other. Case in point is the strongly divergent opinions regarding the genetic significance of hydrothermal carbonate metasomatism. Geochronological constraints can be obtained by radiogenic isotope dating of hydrothermal alteration minerals (i.e., SHRIMP U-Pb dating of monazite in high-grade ores from Mount Tom Price, Rasmussen et al., 2007). Alternatively, ores and alteration assemblages may be indirectly dated by careful paleomagnetic analyses (cf. de Kock et al., 2008).

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Prolonged history of episodic fluid flow in giant hematite ore bodies: Evidence from in situ U–Pb geochronology of hydrothermal xenotime

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Abstract

Absolute ages for hydrothermal mineralization and fluid flow are critical for understanding the geological processes that concentrate metals in the Earth’s crust, yet many ore deposits remain undated because suitable mineral chronometers have not been found. The origin of giant hematite ore deposits, which are hosted in Precambrian banded-iron formations (BIFs), remains contentious. Several models have been formulated based on different sources and timing for the mineralizing fluids; supergene-metamorphic, syn-orogenic, late-orogenic extensional collapse and syn-extensional. Precise geochronology of the ore offers a means of discriminating between these models. In this study, two U–Pb chronometers, xenotime and monazite, have been identified in high-grade hematite ore bodies from the Mount Tom Price mine in the Hamersley Province, northwestern Australia. Both phosphate minerals occur as inclusions within the hematite ore and as coarser crystals intergrown with martite (hematite pseudomorph after magnetite) and microplaty hematite, indicating that the xenotime and monazite precipitated during mineralization. In situ U–Pb dating by ion microprobe indicates that both phosphate minerals grew during multiple discrete events. Our results suggest that ore genesis may have commenced as early as ∼2.15 Ga, with subsequent hydrothermal remobilization and/or mineralization at ∼2.05 Ga, ∼1.84 Ga, ∼1.67 Ga, ∼1.59 Ga, ∼1.54 Ga, ∼1.48 Ga and ∼0.85 Ga. The location of the ore bodies along ancient fault systems, and the coincidence of at least some of the U–Pb phosphate dates with episodes of tectono-thermal activity in the adjacent Proterozoic Capricorn Orogen, implies that fluids were channelled through major structures in the southern Pilbara Craton during discrete phases of tectonic compression and extension. Our results show that the hematite ore bodies formed at sites of repeated focussed hydrothermal fluid flow. In contrast to the aforementioned models, our results imply that iron-ore formation was probably a long-lived, multi-stage process spanning more than one billion years.

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1. Introduction

High-grade (>65% Fe) hematite deposits hosted in Precambrian banded-iron formations represent the world’s most important source of iron ore, with notable examples in Brazil, South Africa, India and Australia [1]. The best-studied examples of these deposits are located in the Hamersley Province of Western Australia (Fig. 1), but despite their economic significance and decades of research, their origin remains unclear [2–7]. Uncertainty about formation of the ore partly stems from its simple composition (comprising almost exclusively hematite, a mineral that gives few clues about its origin), and from strong overprinting by deformation and weathering [1]. Until recently, the ores were widely regarded to have formed in the Proterozoic through supergene enrichment of BIFs followed by recrystallization during burial metamorphism [2,8]. In this model, the age of supergene mineralization is considered to correspond with the development of a major unconformity surface between the Turee Creek Group and Lower Wyloo Group (between ~2.2 and 2.03 Ga) [2]. However, recent petrographic, isotopic and fluid-inclusion studies have recognized the presence of hydrothermal fluids (~250–400 °C) in the ore environment [9–14]. Hydrothermal models for ore formation have linked the timing of mineralization to fluid movement related to tectonic compression and orogeny (i.e., the Ophthalman Orogeny; 2.2–2.1 Ga) [6,10], late-orogenic extensional collapse (~2.1–2.05 Ga), or flood-basalt volcanism and continental extension (2.05–2.0 Ga) [5,9,15].

Determining the precise age of the mineralization can be used to test competing genetic models, and provide insights into the geological controls on ore formation. In this study, we present ion microprobe U–Pb data for xenotime and monazite intergrown with microplaty hematite and martite in high-grade iron-ore from the Mount Tom Price mine (Fig. 1). This deposit was selected because it preserves hypogene alteration zones below the limit of modern weathering [5,9,13]. In addition, it is less intensely deformed than other major deposits in the Hamersley Province (e.g., Paraburdoo and Mount Whaleback near Newman). Therefore, Mount Tom Price represents the most suitable deposit to determine the timing of mineralization and to test different models.

2. Geological setting and tectonic evolution

High-grade hematite ore deposits are located in the Hamersley Province along the southern margin of the Pilbara Craton (Fig. 1), and are spatially associated with Proterozoic normal fault systems [5]. The province

Fig. 1. (A) Map of the Hamersley Province along the southern margin of the Pilbara Craton, northwestern Australia. (B) Stratigraphic column and tectonic history of the Hamersley Province and adjacent Capricorn Orogen. Vertical scale is in Ga.
contains volcano–sedimentary successions spanning nearly one billion years in depositional age, from the Fortescue Group (∼2.77 Ga) to the Upper Wyloo Group (∼1.80 Ga). The sequences, which comprise BIFs, shale, sandstone, dolomite, and felsic and mafic volcanic and intrusive rocks, increase in deformation intensity and metamorphic grade from north to south. The tectonostratigraphic history of the province has been summarized by Taylor et al. [5].

The Hamersley Province is bounded to the south by the Proterozoic Capricorn Orogen. The orogen has a long and complex history spanning ∼1.5 Ga and records the amalgamation of the Archean Yilgarn and Pilbara Cratons to form the West Australian Craton. The geological history of the orogen is yet to be fully documented, but it involved continental collision, accretion, extension and repeated intracratonic reactivation [16]. Major phases of deformation and metamorphism are recorded at 2.21–2.15 Ga (Ophthalmian Orogeny), 2.0–1.96 Ga (Glenburgh Orogeny), 1.83–1.78 Ga (Capricorn Orogeny), 1.68–1.62 Ga (Mangaroo Orogeny) and 1.03–0.95 Ga (Edmundian Orogeny) [16–19]. The orogen has also undergone extension and mafic magmatism at 2.05–2.0 Ga, 1.47 Ga, 1.07 Ga and 0.75 Ga [15,20]. The southern Pilbara margin was deformed during the Ophthalmian and Capricorn orogenies, producing a series of foreland fold-and-thrust belts. However, effects of the other tectonic events on the Pilbara craton have not been documented.

The Mount Tom Price mine is one of several giant iron-ore deposits in the Hamersley Province that formed through the conversion of BIFs of the Hamersley Group.

Fig. 2. (A) Map showing the geology around the Mount Tom Price iron-ore deposit. (B) Cross-section through the centre of Mount Tom Price deposit along line A–B (see A) (after Taylor et al. [5]).
to martite and microplaty hematite [5]. Most of the ore is hosted in the Dales Gorge Member of the Brockman Iron Formation, and at Mount Tom Price occurs as a folded east–west trending, southward-dipping sheet within the main closure of the Turner Syncline [5,13]. Two normal faults are prominent within the mine (Fig. 2), but rapidly disappear vertically and laterally away from the deposit [5]. Mineralization is broadly parallel with the fault trends, with most of the ore restricted to the north of the main northwest-trending, southwest-dipping fault (Southern Batter Fault) (Fig. 2). The timing of mineralization at Mount Tom Price (and other high-grade hematite deposits in the region) is uncertain. Although there are no direct dates, mineralization must postdate 2.5 Ga, the depositional age of the Mount McRae Shale [21], which is footwall to ore

![Fig. 3. Reflected light images of hematite ore from Mount Tom Price deposit. (A) Banded hematite ore comprising layers of densely packed microplaty-hematite (mph) crystals. (B) Coarse, randomly oriented microplaty-hematite crystals. Note the high porosity (black). (C) Blocky and massive martite (mrt) ore with a minute xenotime crystal (xt) in the matrix. (D) Coarse xenotime with minute inclusions of hematite surrounded by martite, microplaty hematite and pores (black). (E) Irregular, elongate xenotime enclosed by martite.](image-url)
bodies hosted in the Brockman Iron Formation, and it must predate 0.75 Ga, the age of a dolerite dyke that metamorphosed the ore in a nearby mine [5].

3. Sample details and analytical methods

3.1. Petrographic and analytical methods

Thirty polished thin sections were prepared from samples of high-grade hematite ore from the Southeast Prong, West Pit and North Deposit of the Mount Tom Price mine (Fig. 2A). Each polished thin section was examined by optical microscope and a JEOL JSM-6400 scanning electron microscope (SEM) fitted with a Link ISIS energy dispersive spectrometer (EDS) located in the Centre for Microscopy, Characterisation and Analysis (CMCA) at the University of Western Australia (UWA).

Mineral analyses and element mapping were carried out by electron microprobe, using an automated JEOL JSM-6400 SEM fitted with three WDS crystal spectrometers located in the CMCA at UWA. Operating conditions of 20 kV accelerating voltage and 100 nA beam current give a spatial resolution of 2–3 μm. Synthetic phosphates and glasses, and natural minerals were used as standards. Methodology for the analysis of xenotime followed that of Williams [22]. Counting times of 50 ms per pixel were employed for the element maps. Data reduction and image manipulation used software from Moran Scientific.

3.2. Sample description

Most of the samples studied here are typically banded, with microplaty hematite-rich layers alternating with bands rich in martite. The microplaty-hematite ore comprises a mesh of randomly oriented, platy hematite crystals (Fig. 3A, B). The martite-rich ore comprises subhedral, equant pseudomorphs after magnetite (Fig. 3C) that commonly display pronounced hematite overgrowth along their margins. The samples are relatively porous (up to 30%), which is typical of the ore body, and is interpreted to relate to multi-stage dissolution of silica and carbonate in precursor BIFs [5].
Most samples contain trace amounts of xenotime (from $<10 \mu m$ up to $100 \mu m$) intergrown with microplaty hematite and martite crystals (Figs. 3D, E, 4A–D, 5A). Xenotime is typically irregular in shape and commonly contains minute inclusions of hematite, and in some cases, irregular specks of zircon. Xenotime crystals display complex compositional zoning in back-scattered electron (BSE) images (Fig. 5C). X-ray element mapping and WDS point analysis indicate that the zoning is mainly related to substitution of heavy rare earth elements (e.g., Yb, Lu) for Y (Table 2 in the Appendix). Contents of Yb$_2$O$_3$ vary from 3.08 wt.% in the darkest areas to 5.10 wt.% in the lightest zones (Fig. 5D). All WDS point analyses contain $\sim 1–3$ wt.% Fe (Table 1 in the Appendix), reflecting the ubiquitous presence of very fine hematite inclusions within the xenotime. The large grain shown in Fig. 5 has formed through coalescence of five or more smaller grains and shows evidence for several episodes of resorption and re-precipitation.

Monazite was also observed in many samples, but is generally smaller ($<20 \mu m$) and less abundant than co-existing xenotime. It occurs as minute inclusions in martite and microplaty-hematite ore (Fig. 6), as irregular cement intergrown with hematite and as a late-stage precipitate in cross-cutting fractures.

### 3.3. SHRIMP U–Pb geochronology

Xenotime and monazite sufficiently large for in situ analysis ($>10 \mu m$) were cut or drilled from polished thin sections and cast in five 25 mm epoxy discs. In situ U–Pb analysis was carried out using a sensitive high-resolution ion microprobe (SHRIMP), with $\leq 0.5$ nA $O_2^-$ primary beams focused onto 5–10 $\mu m$ spots. The fine spatial resolution allowed compositionally distinct zones to be targeted. Established SHRIMP operating procedures were followed for xenotime [23,24] and monazite [25].

Xenotime was analysed during five separate analytical sessions, sometimes in combination with mounts from other studies. Xenotime standards MG-1, BS-1 and z6413 [24] were analysed in separate mounts, one of which was re-polished for each analytical session and Au-coated with the sample mount(s). MG-1 was used as the primary Pb/U standard and z6413 was used to
monitor and re-normalise $^{207}\text{Pb}/^{206}\text{Pb}$. The data acquisition sequence followed that of Fletcher et al. [23], that is, Pb/Th was not independently calibrated (cf., [24]) and the retardation lens was activated, requiring a small re-normalisation of $^{207}\text{Pb}/^{206}\text{Pb}$ data.

Data reduction was carried out using Squid 1.00 [26]. Pb/U reproducibility for the standards was assessed from data from MG-1 and z6413 and propagated to the sample data. Matrix corrections to Pb/U and Th/U followed those of Fletcher et al. [24]. Matrix corrections for REE are more difficult. Measuring REE for the volume of material analysed (therefore ablated) by SHRIMP is impossible, and it is impractical to acquire electron microprobe data from adjacent material for the large number of analyses reported here. The matrix corrections for REE, for all analysed grains, were made using the average REE content for a large multi-aged grain (05–07C; Tables 1 and 2 in the Appendix). Data from this grain demonstrate no correlation between age and REE contents.

Almost 200 analyses were made on ~40 xenotime crystals (Table 3 in the Appendix). Prior to making any other assessments, analyses with >1% common Pb were discarded, regardless of the possible sources of the common Pb and its location in relation to the sampled volume of xenotime. The critical consideration for the other 183 analyses is their U–Pb concordance. In the present context, discordance is likely to result from age-mixing, that is, analysis of a volume that includes xenotime of differing ages. Applying strict concordance criteria to individual analysis is difficult, for reasons detailed in the Appendix, but we have used a fairly stringent ±5% discordance limit for older grains. For dates of ~1000 Ma this would exclude some data that are within precision of concordance, so data that give dates <1200 Ma and are within 2$\sigma$ of Concordia have also been retained. A small number of analyses from discordant samples may have been retained and, conversely, data from some concordant samples may have been disregarded. Overall though, the 114 analyses that lie within the accepted concordance band are considered sufficient to give a valid, if not highly precise, record of major periods of xenotime growth. The 60 normally discordant analyses (concordance <95%) with low common Pb are likely to be mixtures of xenotime from those (and perhaps other younger) growth events. There is no simple explanation for the few reversely discordant points.

4. Xenotime and monazite age data

The 114 analyses that are considered to give reliable geochronological data, from ~40 xenotime crystals from six samples, give $^{207}\text{Pb}/^{206}\text{Pb}$ dates from ~2220 Ma to ~800 Ma (see Table 3 in the Appendix). The distribution of dates within individual xenotime crystals generally displays a pattern of younger rims around old cores. However, in one single large grain (05–07C), there is complexity in the spatial distribution of the measured ages reflecting the intergrowth of numerous discrete crystals that have been infilled and enlarged by later generations of xenotime, juxtaposing regions of different age.

Amongst the concordant data, several $^{207}\text{Pb}/^{206}\text{Pb}$ age populations can be distinguished (Fig. 7). Three of these may reflect discrete events at 2145 ± 28 Ma (mean square of weighted deviates [MSWD] = 0.86; n = 7), 2052 ± 13 Ma (MSWD = 0.74), and 845 ± 20 Ma.
(MSWD = 1.14; n = 13), with a possible fourth event at \( \sim 1840 \) Ma. Most of the dates fall in a complex cluster between 1723 Ma and 1460 Ma (Fig. 7). The unmixing algorithm of Sambridge and Compston [27] suggests three well-defined populations within this cluster, with ages of 1673 ± 15 Ma, 1589 ± 11 Ma and 1538 ± 12 Ma. There is a smaller peak at 1480 ± 20 Ma. The uncertainties in all these dates are minima, since the dates are all dependent on data selection. The real uncertainties are possibly \( \sim 5 \) Ma greater than the calculated values.

Five analyses of four monazite crystals from two samples yield imprecise \( {^{207}}\text{Pb} / {^{206}}\text{Pb} \) dates because of the very low U contents (typically < 10 ppm). The corresponding Th–Pb dates are more precise because of higher Th contents (0.1 – 1.4%), and suggest growth at 2.07 Ga, 1.75 Ga and 0.94 Ga (Table 4 in the Appendix). However, because of the very low U contents of monazite from the iron-ore samples, xenotime was analysed in preference to monazite in this study.

5. Discussion

5.1. Interpretation of xenotime dates

The age spectrum obtained from analyses of xenotime crystals from the Mount Tom Price ore bodies is complex, with up to eight populations identified in a span of more than one billion years. Xenotime, like monazite, is considered to be extremely resistant to diffusive Pb loss in most geological environments [28]. The isotopic robustness of xenotime is supported by the preservation of concordant and precise \( {^{207}}\text{Pb} / {^{206}}\text{Pb} \) ages in detrital and diagenetic xenotime in mid-Archean (~ 3.2 Ga) metasedimentary rocks, despite their antiquity and complex post-depositional history [29]. In the Mount Tom Price deposit, fluid-inclusion data show that xenotime experienced only low-temperature conditions (< 400 °C) [13], so Pb diffusion is unlikely to be responsible for the complex age spectrum. This is further supported by the preservation of Yb zoning (Fig. 5C and D), because rare earth elements are significantly less resistant to diffusional alteration than Pb in xenotime [28]. Thus, the xenotime age data does not reflect diffusive Pb loss. All of the xenotime dates are younger than the depositional age of the precursor BIF, the Brockman Iron Formation, and they therefore represent multiple episodes of growth, and probably dissolution and regrowth, during the circulation of hydrothermal fluids.

The timing of fluid movement and associated xenotime growth is not random. Some of the episodes of hematite and xenotime growth and regrowth in the ore bodies are synchronous with tectonic events in the Capricorn Orogen and adjacent southern Pilbara Craton. The oldest population (2145 Ma) overlaps with the Ophthalmian Orogeny (2.21 – 2.15 Ga), and fluid flow associated with this event has been mapped across most of the Pilbara Craton [30]. The presence of 2.15 Ga xenotime in the ore suggests that the first stages of mineralization commenced during the Ophthalmian Orogeny. This interpretation is consistent with the occurrence of rare clasts of microplaty hematite in conglomerates from the basal Beazley River Quartzite (Lower Wyloo Group) [31]. A second xenotime age population at ~ 2.05 Ga is synchronous with crustal extension and extrusion of the Cheela Springs Basalt (Lower Wyloo Group), which is probably only slightly older than 2031 ± 6 Ma, the age of a tuff bed in the conformably overlying Wooly Dolomite [15]. A small population of ~ 1.84 Ga xenotime broadly coincides with the timing of the Capricorn Orogeny, which affected the southern margin of the Pilbara Craton between ~ 1.83 Ga and ~ 1.78 Ga [16]. Therefore, the three major Paleoproterozoic tectonic events recorded in the southern Pilbara Craton and Capricorn Orogen were accompanied by the growth of xenotime and microplaty hematite in the Mount Tom Price ore bodies.

The main cluster of xenotime dates spans 1.7 Ga to 1.45 Ga, with four possible peaks identified at ~ 1.67 Ga, ~ 1.59 Ga, ~ 1.54 Ga and ~ 1.48 Ga. The oldest of these
dates corresponds with the Mangaroon Orogeny (1.68–1.62 Ga), which is characterized by deformation, metamorphism and granite magmatism in the Capricorn Orogen [18]. There is some evidence that the effects of this orogenic event are more widespread as growth of monazite at 1.65 Ga in the Soanesville Group of the Pilbara granite–greenstone terrain has been linked to fluid movement related to the Mangaroon Orogeny [29]. The youngest date coincides with the emplacement of 1.47 Ga dolerite sills into the Edmund Group of the Bangemall Supergroup [20]. No information about the geodynamics of this event is currently available. Structures related to the Mangaroon Orogeny and mafic magmatism in the Edmund Group are not recorded in the southern Pilbara Craton or Hamersley Group. The growth of xenotime in the Mount Tom Price ore bodies shows that fluids were moving through pre-existing fault structures in the southern Pilbara margin at the same time that deformation and magmatism were occurring further to the south. This suggests that the movement of crustal fluids is related to tectonic events, and that while it is most probably channelled through pre-existing structures, its effects can be discerned over a larger area than that in which deformational fabrics can be detected.

Almost one third of the concordant ages recorded from xenotime in the Mount Tom Price deposit fall between 1.53 Ga and 1.62 Ga. No tectonic or magmatic events of this age have yet been recorded from the Hamersley Province or Capricorn Orogen, although given the state of knowledge it is entirely possible that events of this age are present but have not yet been dated. As many of the episodes of hematite and xenotime growth are synchronous with recognized tectonic events, and as cryptic fluid flow related to tectonism has been identified well within the Pilbara Craton, it is possible that the 1.53–1.62 Ga dates are recording mineral growth related to fluid flow produced by tectonic events that have not yet been properly defined. For example, a well-developed extensional fault system that forms the northern margin of the Bresnahan Basin was active at some time between ∼1.8 Ga (the age of the Capricorn Orogen) and ∼1.5 Ga (the interpreted age of the Edmund Group) [32]. Approximately 4 km of conglomerate and sandstone were deposited in alluvial fan deposits in the Bresnahan Basin, and it is possible that the 1.6–1.5 Ga ages recorded in the iron ore at Mount Tom Price are related to this episode of extensional faulting in the Capricorn Orogen. This time span (1.6–1.5 Ga) was also a period of convergence along the eastern margin of proto-Australia, and is recorded in the northern and southern Australian cratons by the Isan Orogeny in the Mount Isa Block and Olarian Orogeny in the Curnamona Province [33]. Fluid flow at Mount Tom Price may have been associated with cryptic movements in the Capricorn Orogen related to tectonic events in eastern Australia but this question cannot be resolved with currently available data. The youngest xenotime from Mount Tom Price (∼0.85 Ga) is from the aureole of an undated WNW-trending dolerite dyke and may reflect the age of contact metamorphism.

5.2. Implications for iron-ore mineralization

The intergrowth of xenotime and monazite with microplaty hematite and martite in high-grade ore implies that phosphate and hematite growth was synchronous. Authigenic monazite in the footwall at the Mount Tom Price mine (Mount McRae Shale) records tectonothermal events at ∼2.4 Ga and ∼2.2 Ga [30]. The absence of younger generations (<2.15 Ga) of xenotime and monazite in the footwall shale suggests that the flow of subsequent hydrothermal fluids was restricted to the ore horizons. Thus, the ore bodies, which are linked by a series of Proterozoic normal faults [5], were the focus of repeated pulses of hydrothermal fluid flow. It is likely that the heated fluids were channelled through major structures in the southern margin of the Pilbara Craton during repeated reactivation related to extension and compression in the adjacent orogen. While most models for hydrothermal ore genesis invoke two or three stages of hypogene mineralization related to a single tectonothermal event, isotopic dating of xenotime intergrown with the ore shows that the Mount Tom Price iron-ore deposit experienced up to seven hydrothermal events, spanning more than one billion years. What is not apparent is whether most of the ore formed during the Ophthalmian Orogeny and experienced only minor remobilization during subsequent hydrothermal events, or whether the ore formed through progressive episodic upgrading over one billion years.

A long-lived history of hematite growth is supported by a number of structural, paleomagnetic and petrographic studies. For instance, petrographic examination of ores from Mount Tom Price reveals the presence of up to five compositional zones of magnetite–hematite on magnetite cores [5]. Also, paleomagnetic studies of hematite ore from the Southeast Prong ore body in the Mount Tom Price deposit suggest a complex history involving at least two major intervals of martite and microplaty-hematite formation and late-stage dyke intrusion [34]. From the Mount Whaleback district, at least five groups of veins have been linked to events spanning the onset of the Ophthalmian Orogeny to post-
Capricorn Orogeny faulting (after ∼1.78 Ga). Analysis of the veins suggests that large volumes of heated fluids (200–300 °C) were focussed into the Mount Whaleback district and that one or more of the vein groups probably represented pathways for genesis of the Mount Whaleback deposit [12]. A prolonged history of microplaty-hematite precipitation is indicated by the presence of coarse hematite crystals in the matrix of conglomerates of the basal Wyloo Group that contain hematite clasts, and by hematite veins in sandstones of the Upper Wyloo Group [34]. These studies point to hydrothermal iron precipitation during, and hundreds of millions of years after, the Ophthalmian Orogeny [5,12,34].

Our work on xenotime and monazite intergrown with hematite ore at the structurally controlled Mount Tom Price deposit implies that ore genesis was a prolonged, multi-stage process that commenced as early as ∼2.15 Ga, with subsequent episodes of iron mineralization and remobilization. The location of other giant hematite ore bodies along ancient fault systems (e.g., Mount Whaleback, Hamersley Province; Carajas area, Brazil; Thabazimbi, South Africa) [5,7,35], suggests that these deposits may also have undergone a protracted history involving repeated episodes of fluid flow. Indeed, it is possible that these deposits only develop in structurally controlled sites where iron-rich host rocks (e.g., BIFs) are intersected by major faults with an extended history of hydrothermal fluid circulation.

6. Conclusions

Xenotime and, to a lesser extent, monazite are widespread trace minerals in high-grade hematite ore bodies from the Mount Tom Price deposit in the Hamersley Province. The presence of xenotime as inclusions within hematite ore and its intergrowth with microplaty hematite and martite, indicates that xenotime growth was synchronous with iron mineralization. In situ U–Pb dating of xenotime reveals a complex history involving multiple episodes of growth. Our results imply that the Mount Tom Price deposit was the site of repeated episodes of hydrothermal fluid flow that commenced at ∼2.15 Ga and continued intermittently for more than one billion years. It is likely that fluids were focussed through structures in the southern margin of the Pilbara Craton during continental reactivation and reworking in the bordering Capricorn Orogen.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2007.03.033.

References


Hydrothermal Alteration Zonation and Fluid Chemistry of the Southern Ridge and North Deposits at Mt Tom Price

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ABSTRACT

The Mt Tom Price deposit is a world-class high-grade hematite deposit in the Hamersley Province of Western Australia with an original resource of 900 Mt of almost pure hematite, averaging 63.9 wt per cent Fe. Petrological and geochemical studies at both the Southern Ridge and the North deposit at Mt Tom Price have identified three hypogene alteration zones between unmineralised BIF and high-grade iron ore:
1. distal magnetite-siderite-stilpnomelane;
2. intermediate hematite-magnetite-ankerite-talc-chlorite; and
3. proximal martite-microplaty hematite-magnetite-apatite alteration zones.

Fluid inclusions trapped in siderite within the distal magnetite-siderite-stilpnomelane alteration zone at the North deposit revealed primary high salinity (25.5 eq wt per cent NaCl-CaCl₂) inclusions that homogenised between 181° and 257°C. Fluid inclusions trapped in quartz from quartz-hematite veins from the Southern Batter Fault, revealed that primary high salinity (7.0 - 9.0 eq wt per cent NaCl) apatite within talc-microplaty hematite veins in the intermediate hematite-magnetite-talc-chlorite alteration zone at the North deposit revealed mostly H₂O-NaCl pseudosecondary and secondary inclusions with salinities of 22.4 - 25.4 eq wt per cent and 22.9 - 25.9 eq wt per cent CaCl₂, respectively. Pseudosecondary inclusions homogenised between 153° and 449°C and secondary inclusions homogenised between 103° and 157°C. Fluid inclusions trapped in apatite within talc-microplaty hematite veins in the intermediate hematite-magnetite-talc-chlorite-apatite alteration zone at Southern Ridge revealed that primary, medium salinity (7.0 - 9.0 eq wt per cent NaCl) fluid inclusions homogenised between 118° and 257°C into liquid. Microthermometric analysis of quartz from quartz-hematite veins from the Southern Batter Fault, Southern Ridge show a complex fluid inclusion history. Primary fluid inclusions consist of:
1. low- and high-salinity H₂O-NaCl inclusions trapped at temperatures of approximately 140° to 230°C;
2. vapour-rich inclusions of unknown compositions; and
3. complex salt-rich (Ca, Mg, K and Na inclusions).

Secondary inclusions consist of medium-salinity fluid inclusions trapped at temperatures of about 140° to 280°C.

A two-stage hydrothermal model is proposed for the formation of both the Southern Ridge and North deposits. Early 1a hypogene alteration involved the release of hydrothermal NaCl-CaCl₂-rich (25.5 eq wt per cent) basinal brines (110° - 150°C) from the underlying Wittenoom Formation and directed upward along normal faults and focused within the silica-rich rocks of the Dales Gorge Member, by the shafts of the underlying Mt McRae Shale Member and overlying Whalebake Formation. Within the Dales Gorge Member hydrothermal basinal brines migrated laterally within large-scale folds with permeability controlled by shale bands and the NW trending dolerite dyke sets. Fluid rock reactions transformed unmineralised BIF to magnetite-siderite-iron silicate BIF, with subsequent desilification of the chert bands.

Stage 1b hypogene involved an increase in temperature of the hydrothermal, CaCl₂-rich saline (24 eq wt per cent) basinal brines (250° - 300°C) resulting in formation of hematite-ankerite-magnetite-talc-chlorite alteration and the crystallisation of microplaty hematite.

INTRODUCTION

At the North deposit at Mt Tom Price the preservation of hypogene alteration zones below the limit of modern weathering (Barley and Pickard, 1997; Barley et al., 1999; Thorne, 2001; Thorne, Hagemann and Barley, 2004) provides a unique opportunity to expand the knowledge of the processes that formed the high-grade (≥65 wt per cent Fe) iron deposit. This paper provides a detailed description of the hypogene alteration zones that surround the high-grade iron orebody at the North deposit (Barley et al., 1999; Cochrane, 2003; Thorne, 2003; Thorne, Hagemann and Barley, 2004) and Southern Ridge deposit (Ridley, 1999; Hagemann et al., 1999; Taylor et al., 2001; Cochrane, 2003) and their paragenetic sequence within the mineralisation system. Fluid inclusions and carbon and oxygen isotope compositions of carbonates in the major alteration zones at the North deposit were analysed in order to establish the pressure, temperature and composition of the hydrothermal fluids. Geochemical analyses including fluid inclusion microthermometry, oxygen and hydrogen isotope analyses and ion-chromatography studies on quartz-hematite veins, related to late-brittle extensional movement on the Southern Batter Fault, Southern Ridge deposit (Ridley, 1999; Hagemann et al., 1999) were also analysed in order to establish fluid regimes, fluid sources and the composition of cation and anions transported by the hydrothermal fluids. This paper compares and contrasts the hydrothermal alteration and geochemical analysis at the North and Southern Ridge deposits and discusses the implications for high-grade iron mineralisation in BIF-related deposits.

MT TOM PRICE IRON DEPOSIT

The Mt Tom Price orebody (Figure 1) extends for seven kilometres from the North deposit in the NW to the South East Prong deposit in the SE is up to 1.6 km wide (average 600 m), with a maximum depth of 250 m below the surface (Taylor et al., 2001). Surface outcrop of the orebody occurs north of the WNW trending Southern Batter Fault. The detailed stratigraphy and geology of the mine area (Harmsworth et al., 1990; Ridley, 1999; Taylor et al., 2001) are presented in Figure 1. High-grade hematite ore preserves the meso and microbanding of the host BIF. It consists essentially of randomly oriented fine-grained platy hematite (10 - 100 μm) and martite (20 - 250 μm). Martite is subhedral to euhedral, and exhibits intensive overgrowth of microplaty hematite from their grain margins. The intervening S-Bands are pale pink clay seams, and constitute the main remaining impurities in the ore (Taylor et al., 2001).

GEOLOGY OF THE NORTH DEPOSIT

The North deposit is located NW of the Southern Ridge, Synclines and Centre deposits at Mt Tom Price (Figure 1). The strata consist of the Dales Gorge Member, the Whalebake Shale...
Member and the Joffre Member. Previous work (Barley et al., 1999; Taylor et al., 2001; Thorne, 2001; Thorne, Hagemann and Barley, 2004) describes the North deposit. The North deposit is concealed below colluvium, canga and low-permeability unmineralised shale that limits the depth of weathering and preserve hypogene alteration zones both below and lateral to the deposit (Figure 2b).

Two zones of high-grade mineralisation are present within the North deposit: martite-microplaty hematite (Mr-mpH) ore (>65 wt per cent Fe) with low P levels (<0.05 wt per cent) and Mr-mpH ore with high phosphorus levels (>0.05 wt per cent). Supergene Mr-mpH (low P) occurs above the depth of weathering and extends to near surface on the northern limb of the syncline where intense weathering makes identification of hypogene alteration and structures impossible. The distribution of supergene low P, Mr-mpH above and hypoge high P, Mr-mpH mineralisation below the depth of weathering is similar to Southern Ridge (Thorne, Hagemann and Barley, 2004).

PETROGRAPHY OF HYPOGENE ALTERATION ZONES

Host rocks

Unmineralised, unweathered BIF is characterised by alternating magnetite-chert mesobands and microbands with subordinate carbonate, iron silicates, and pyrite (Figure 3a). Magnetite occurs as euhedral crystals (30 - 250 μm) with rare, anhedral hematite inclusions (5 - 15 μm). Finely granular to bladed of crystals stilpnomelane (20 - 70 μm), rare fibrous riebeckite (20 - 100 μm), and carbonates (50 - 400 μm, siderite to dolomite) occur as thin bands and intergrown within chert mesobands (Figure 4a). Unweathered shale bands, which are metamorphosed tuffaceous mudstones (Pickard, 2002) within the Dales Gorge Member are green/black, laminated (mm to cm) contain iron-rich chlorite, stilpnomelane, massive dolomite, euhedral pyrite (50 - 300 μm) and local magnetite-rich bands.

NORTH DEPOSIT

Extensive core logging and petrological studies (Barley et al., 1999; Cochrane, 2003; Thorne, 2001; Thorne, Hagemann and Barley, 2004) identified a laterally extensive pervasive hypogene footwall alteration zone below low P, Mr-mpH ore (<65 wt per cent Fe, P < 0.05 wt per cent; Figure 2). The alteration comprises three zones; distal magnetite-siderite-iron silicate; intermediate hematite-ankerite-magnetite-talc-chlorite, and proximal martite-microplaty hematite-apatite. Hypogene alteration is restricted to BIF bands with shale bands preserving their original mineralogy and textures. Hypogene alteration is restricted to the Dales Gorge Member, and strongly developed within DG3 and upper DG2. The outer distal alteration zone becomes restricted to the proximity of BIF/shale band contacts (Thorne, Hagemann and Barley, 2004).

Distal

Unmineralised BIF grades into distal alteration zones that are approximately 30 m in width (Figure 2b) and characterised by the mineral assemblage magnetite-siderite-stilpnomelane (Figure 3b). The assemblage reflects the partial replacement of chert bands by
bladed magnetite (50 - 200 μm), siderite (800 - 1500 μm) and iron silicates (20 - 110 μm; Figure 4b, c, d). Pyrite crystals (50 - 300 μm) are finely disseminated within shale bands. Apatite occurs as euhedral crystals intergrown with anhedral chlorite. Talc forms radiating and fibrous aggregate that replace siderite. Chlorite is intergrown with talc and contains microplaty hematite and anhedral to euhedral magnetite grains. Talc-chlorite-microplaty hematite veins (V4; Thorne et al, in prep) are common and cross-cuts magnetite-siderite-stilpnomelane wallrock.

Intermediate
Intermediate alteration zones, are about 15 m in width and are characterised by the mineral assemblage hematite-ankerite-magnetite-talc chlorite, as the result of ankerite and microplaty hematite replacing quartz, siderite, and iron-silicates (Figure 3c). Microplaty hematite (10 - 60 μm) has crystallised as both individual blades and dense clusters that form overgrowths on magnetite, and as individual plates within ankerite crystals. Anhedral and microplaty hematite replace iron-silicates (Figure 4e, f, g). Pyrite veins are rare and occur within fracture zones that postdate hematite-ankerite-magnetite alteration. Talc and chlorite form anhedral intergrown masses within magnetite mesobands.

Proximal
The proximal alteration zones are about 15 m in width and characterised by the mineral assemblage marlrite-microplaty hematite-apatite (Figure 3d). Martite and anhedral hematite replace magnetite and iron silicates, respectively. With increased proximity to the topographic surface, the abundance of goethite and skeletal and cellular martite increases. Fibrous quartz and colloform goethite is locally developed within fracture zones (Figure 4b). Intergranular porosity increases significantly to about 15 per cent.
SOUTHERN RIDGE

Detailed core logging and petrological studies (Ridley, 1999; Taylor et al., 2001; Cochrane, 2003; Thorne et al., in prep) identified a laterally extensive pervasive hypogene alteration zone on the downthrown side of the Southern Batter Fault at Southern Ridge. Hypogene alteration comprises three zones, the distal magnetite-siderite-iron silicate, the intermediate hematite-magnetite-talc-chlorite, and the proximal martite-microplaty hematite zones. Hypogene alteration is strongly developed within the hanging wall Dales Gorge Member and locally in the Joffre Member where the Southern Batter Fault juxtaposes it against the Dales Gorge Member. Within the Joffre Member, located above high phosphorus hematite mineralisation, Taylor et al. (2001) report irregular hematite bodies that are surrounded by hematite rich jaspilite BIF and quartz veins.

Distal

Unmineralised BIF (quartz-magnetite-stilpnomelane-dolomite) grades into distal alteration zones that are approximately 35 m in width and characterised by the mineral assemblage magnetite-siderite-stilpnomelane. Textures and mineralogy are identical to distal alteration at the North deposit.

Intermediate

The intermediate alteration zone is characterised by the assemblage hematite-magnetite-talc-chlorite and is up to 80 m in width and preserved both in the Dales Gorge Member and Joffre Member. The alteration results from the crystallisation of microplaty hematite and the replacement of siderite/stilpnomelane alteration by talc and chlorite. Talc occurs as fine grained (5 - 20 μm) feathery masses replacing siderite and intergrown with martite and microplaty hematite. Magnetite shows variable replacement by martite. Pyrite (10 - 2000 μm) occurs as larger subhedral grains in the matrix within martite layers or along fractures which cut the microplaty hematite-martite (magnetite) matrix. Some pyrite contains inclusions of remnant magnetite. Chlorite (5 - 20 μm) occurs as intergranular masses in magnetite bands and on the margins of talc veins. Apatite (2 - 20 μm) is present as randomly oriented subhedral grains within fibrous, fine-grained talc-microplaty hematite veins and intergrown with martite in martite mesobands (Cochrane, 2003). Intergranular porosity in the intermediate alteration zone is estimated at ten per cent.

Proximal

The proximal alteration zone is about 30 m in width and characterised by the mineral assemblage martite-microplaty hematite-apatite. Martite and anhedral hematite replace magnetite and iron silicates, respectively. Minor amounts of apatite and chlorite are preserved and occur intergrown in distinct layers, together with some fine-grained martite (Thorne, in prep). Intergranular porosity increases significantly to about 15 per cent. The intervening S-bands are black or green shales and preserve the mineralogy of the host rocks (Taylor et al., 2001).

PETROGRAPHY OF SUPERGENE ALTERATION ZONES

Areas of supergene alteration at the Southern Ridge and the North deposit are characterised by the mineral assemblage martite-microplaty hematite-goethite. This assemblage reflects the replacement of remnant magnetite to goethite and the removal of most of the apatite. With increased proximity to the topographic surface, the abundance of goethite and skeletal and cellular martite increases. Fibrous quartz and colloform goethite is locally developed within fracture zones (Thorne, Hagemann and Barley, 2004) Intergranular porosity is estimated to be about 30 per cent (Taylor et al., 2001). Talc and chlorite, primarily at Southern Ridge, is replaced by montomorrillonite (Thorne et al., in prep). Shale bands are reduced in volume by up to 60 per cent by the removal of carbonates, the oxidation of pyrite to limonite, and the replacement of shales by pink, kaolinitic clays.

Fig 3 - Core samples from the North deposit. (A) Unmineralised BIF showing chert (Qtz) and magnetite (Mt) banding. (B) Magnetite-siderite-iron silicate (Mt-Sid-FeSil) alteration showing preservation of banding. (C) Hematite-ankerite-magnetite alteration with localised brecciated martite bands. (D) Martite-microplaty hematite (Mr-mpHm) ore. Note preservation of banding and goethite (Goe) infill (modified after Thorne, Hagemann and Barley, 2004).
FIG 4 - Microphotographs showing hypogene alteration mineralogy. (A) Unmineralised magnetite (Mt) and chert (Qtz) microbands. Note microcrystalline dolomite (Dol) within chert microband. (B) Intergrown platy siderite (Sid) and iron silicates (FeSi) pseudomorphing chert mesoband. (C) Brecciated magnetite microband with matrix of siderite (Sid) and iron silicates (FeSi). (D) Radial and individual bladed magnetite (Mt), partially oxidised to microplaty hematite (mpHm) within siderite (Sid) matrix. (E) Hematite-ankerite-magnetite alteration with ankerite (Ank) replacing siderite and microplaty hematite (mpHm) replacing iron silicates. Euhedral magnetite (Mt) remains unoxidised. (F) Ankerite (Ank) veins, V3, cross-cutting magnetite (Mt) mesobands with wallrock crystallisation of microplaty hematite (mpHm) on magnetite. (G) Martite (Mr) and microplaty hematite (mpHm) within ankerite (Ank). (H) Skeletal martite (Mr) crystals within goethite (Goe) matrix. Note minor interstitial quartz (Qtz) (from Thorne, Hagemann and Barley, 2004).
FLUID CHEMISTRY OF HYPOGENE ALTERATION ZONES

North deposit

Microthermometric data
Detailed microthermometric analysis of ankerite from magnetite-siderite-stilpnomelane alteration (Cochrane, 2003) and ankerite from ankerite-microplaty alteration zones (Thorne, Hagemann and Barley, 2004) were conducted. Fluid inclusion analysis on siderite from magnetite-siderite-stilpnomelane distal alteration indicates that primary high salinity \(H_2O-NaCl\) and \(H_2O-CaCl_2\) inclusions (maximum salinity of 25.5 eq wt per cent CaCl\(_2\)) were trapped between 107° and 142°C. Fluid inclusions trapped in ankerite in ankerite-hematite veins in the hematite-ankerite-magnetite-talc-chlorite intermediate alteration zone revealed mostly \(H_2O-CaCl_2\) pseudosecondary and secondary inclusions with salinities of 22.4 - 25.4 and 22.9 - 25.9 eq wt per cent CaCl\(_2\), respectively. Pseudosecondary inclusions homogenised between 153° and 449°C (253 ± 60°C; 1σ; n = 34) and secondary inclusions between 103° and 157°C (117 ± 10°C; 1σ; n = 66). The decrepitation of pseudosecondary inclusions above 350°C suggests that their trapping temperatures are likely to be higher (ie 400°C).

Primary, pseudosecondary and secondary aqueous fluid inclusions all have similar salinities but vary significantly in their homogenisation temperatures. The 100° to 250°C difference in trapping temperatures between the fluid inclusion types is compatible with an evolving, fluid source. An initial warm 120°C fluid that crystallised siderite which was overprinted by a saline (24 eq wt per cent CaCl\(_2\)) a hot (>300° to 350°C) brine that circulated during the formation of the hematite-ankerite-magnetite-talc-chlorite alteration, and a later stage (170°C) saline (24 eq wt per cent CaCl\(_2\)) brine. The latter fluid could represent either the final cooling stage during hematite-ankerite-magnetite alteration or a separate fluid phase.

Isotopic data
Depleted δ\(13C\) values of ankerite (δ\(13C\); -4.9 ± 2.2 ‰; 1σ, n = 15) from the hematite-ankerite-magnetite-talc-chlorite alteration zone indicate that the bulk of the carbon within the alteration zone is not derived from the BIF sequence. Similar oxygen isotope compositions, but increasingly heavy carbon isotopes from magnetite-siderite-iron silicate alteration (δ\(13C\); -8.8 ± 0.7 ‰, 1σ, n = 34) to hematite-ankerite-magnetite-talc-chlorite alteration, and a later stage (170°C) saline (δ\(13C\); -4.9 ± 2.2 ‰, 1σ, n = 17) zones suggest the progressive exchange (mixing) with an external fluid with a heavy carbon isotope signature (Thorne, Hagemann and Barley, 2004). It is likely that an ascending, saline fluid mixed with the Wittenoom Formation (δ\(13C\); 0.9 ± 0.7 ‰, 1σ, n = 15) provided such a fluid source. Evidence from deep drilling at the Mt Tom Price deposit (Taylor et al., 2001) suggests that the Wittenoom Formation is stratigraphically thinned below the Mt Tom Price deposit and is structurally linked via the Southern Batter Fault.

Southern Ridge
Geochemical work at the Southern Ridge comprises fluid inclusion microthermometry on apatite from tafel-apatite veins from the hypogene intermediate alteration zone (Cochrane, 2003) and detailed geochemical analyses of quartz-hematite veins (SR-V4; Ridley, 1999; Hagemann et al, 1999) including fluid inclusion microthermometry, oxygen isotope analyses on quartz and hematite, hydrogen isotope analyses, laser-Raman analyses and ion chromatography on fluid inclusions (Hagemann et al, 1999). The veins strike and dip parallel to the Southern Batter fault zone and display crack-seal texture with slivers of hematite and magnetite oriented parallel to the wallrock and characterised by up to a 3 cm wide halo of hematite alteration.

Microthermometric results
Detailed microthermometric analysis was conducted on apatite from the intermediate hematite-magnetite-talc-chlorite-apatite alteration zone. Apatite occurs as individual crystals within talc-microplaty hematite-apatite veins (Cochrane, 2003) indicates that primary medians of fluid inclusions (δ\(18O\) = 7.0 ± 9.0 eq wt per cent NaCl) trapped at temperatures between 181° and 257°C and primary high salinity 22.8 - 25.9 eq wt per cent CaCl\(_2\) fluid inclusions homogenised between 118° and 257°C. Microthermometric analysis of quartz from quartz-hematite veins (SR-V4; Hagemann et al, 1999) show a complex fluid inclusion history: Primary fluid inclusions consist of:

1. low- and high-salinity \(H_2O-NaCl\) inclusions trapped at temperatures at approximately 140° to 230°C;
2. vapour-rich inclusions of unknown compositions; and
3. ‘complex’ salt-rich inclusions.

‘Complex’ salt-rich inclusions contain a mixture of mono and divalent cations such as Ca, Mg, K, and Na. Secondary inclusions consist of medium-salinity fluid inclusions trapped at temperatures of about 140° to 230°C. Aqueous-carbonic inclusions are rare and their exact composition and timing relationship with the other inclusion types remains uncertain at present.

The occurrence of primary low-salinity inclusions indicates that possibly an additional fluid source, such as seawater or meteoric water, may have been present during the formation of the quartz veins. The possible trapping of hematite in fluid inclusions, the close spatial relationship between hematite and primary fluid inclusions, and recent experimental data (Barton and Johnson, 1996) that suggests iron can be transported in solution by sulfur-poor and Cl-rich brines, indicating that hematite and quartz vein formation could have been related to the circulation of complex hydrothermal fluids.

Stable isotope analyses
Stable isotope analyses revealed that magmatic and metamorphic fluids can largely be discounted as fluid source for the quartz-hematite veins. The contribution of seawater (δ\(18O\) = 0) cannot be completely disregarded as some of the δ\(18O\)vein values for quartz are close to zero. However, hydrogen values are lighter than seawater (on average 38 ‰), therefore, are not compatible with seawater as a major source for hydrogen (Hagemann et al, 1999).

Ion chromatography
Ion-chromatography investigations on fluid inclusion liquids revealed that quartz-hematite (SR-V4) veins contain Na>Mg>Ca>K as major cations, with elevated Li, Mn, and B content as trace elements. Anion ratios such as Br/Cl, I/Cl and Cl/SO\(_4\), are not compatible with ratios for bulk earth, igneous and metamorphic fluids and rocks, or seawater. The fluids show a close affinity to Canadian Shield brines and, in the case of Cl/SO\(_4\) versus Na/K ratios, brines that are related to Columbian emerald formation (Hagemann et al, 1999). This study of fluid inclusion petrography and microthermometry, stable isotope and ion chromatography analyses suggests that the majority of aqueous fluids trapped in fluid inclusions in quartz-hematite veins are related to complex basinal brines.

STRUCTURAL AND HYDROTHERMAL MODEL FOR THE NORTH DEPOSIT AND SOUTHERN RIDGE DEPOSIT, MT TOM PRICE
Thorne et al., in prep) from the North deposit together with detailed petrology provides evidence to support a two-stage hydrothermal model for the formation of the hypogene alteration zones at the North deposit and Southern Ridge.

**Stage 1a – early magnetite-siderite-iron silicate hypogene alteration**

Initial hypogene alteration (Figure 5) occurred within the dominantly magnetite-chert layers of the Dales Gorge Member. Unmineralised, magnetite-chert BIF wallrock is transformed laterally and vertically into magnetite-siderite-iron silicate BIF with subsequent desilicification of the chert bands. Fluid inclusion analysis on siderite from magnetite-siderite-stilpnomelane alteration at the North deposit indicates that the magnetite-siderite-stilpnomelane alteration involved warm (107° to 142°) saline (25 eq wt per cent CaCl2) brine (Cochrane, 2003). The trend from heavy carbon isotope compositions of dolomites from the Wittenoom Formation that underlies the Mt Tom Price deposit, to progressively heavier carbon isotope compositions of siderite in the stage 1a hypogene alteration zone, suggests that hydrothermal fluids (brines) were released from the underlying Wittenoom Formation and directed upward along normal faults. Stable isotope and ion chromatography analyses from quartz-hematite veins from the Southern Batter Fault, Southern Ridge deposit show that the majority of aqueous fluids trapped in fluid inclusions in quartz-hematite veins are related to complex basinal brines. These basinal brines were focussed within the silica-rich rocks of the Dales Gorge Member by the shales of the underlying Mt McRae Shale Member and overlying Whaleback Formation, which acted as an aquitard. Within the Dales Gorge Member hydrothermal fluids migrated laterally within large-scale folds with permeability controlled by shale bands and the NW trending dolerite dyke sets (Thorne, Hagemann and Barley, 2004).

**Stage 1b – hematite-ankerite-magnetite-talc-chlorite hypogene alteration**

Continuing reactions between the ascending hydrothermal fluids and magnetite-siderite-iron silicate alteration produced the hypogene hematite-ankerite-magnetite-talc-chlorite alteration at the North deposit leaving only the remnants of the former. The replacement of iron silicates by hematite and the replacement of siderite by ankerite accompanied continued desilicification of the chert bands. Microplaty hematite has crystallised as both individual blades and dense clusters that form overgrowths on magnetite, and as individual plates within ankerite crystals and talc-chlorite veins. Fluid inclusion evidence suggests that the hematite-ankerite-magnetite alteration at the North deposit involved a hot (>300° to 400°C) saline (24 eq wt per cent CaCl2) brine (Figure 5). Talc-microplaty hematite-apatite veins from hematite-magnetite-talc-chlorite-apatite alteration at Southern Ridge indicate alteration involved a hot, up to 257°C saline (24 eq wt per cent CaCl2) brine. The similar compositions of hypogene brines at both the North and the Southern Ridge deposits suggest a common fluid source in both deposits. Rare pyrite veins (V4; Thorne, Hagemann and Barley, 2004) that cross-cut all other vein types and the crystallisation of microplaty hematite suggest locally a late-stage influx of sulfide-bearing fluid (Figure 5). The increase in heavy carbon isotope values from stage 1a magnetite-siderite-iron silicate alteration to stage 1b hematite-ankerite-magnetite alteration (Thorne, Hagemann and Barley, 2004) suggests the ongoing progressive isotopic exchange via the influx of hydrothermal brines sourced from the underlying Wittenoom Formation.

**Stage 1c – late martite-microplaty hematite-apatite hypogene alteration**

The final stage of hypogene alteration involved the transformation of magnetite and iron silicates to hematite, and the dissolution of the ankerite from the precursor stage 1b hematite-ankerite-magnetite rock. At the Southern Ridge this dissolution is widespread and structurally controlled by dolerite dykes and the Southern Batter fault. This phase of hydrothermal fluid is responsible for the high intergranular porosity in the intermediate and proximal alteration zones at Southern Ridge. The preservation of low-temperature, saline secondary fluid inclusions (~120°C and 24 eq wt per cent CaCl2; Thorne, Hagemann and Barley, 2004) preserved in relict ankerite at the North deposit, suggests that ankerite dissolution occurred late in the hydrothermal evolution of the North deposit (Figure 5).
The high salinity of the fluids precludes meteoric water as a fluid source and suggests that these brines likely relate also to the release of fluids from the underlying Wittenoom Formation. This study of fluid inclusion petrography and microthermometry, stable isotope and ion chromatography analyses suggests that the majority of aqueous fluids trapped in fluid inclusions in quartz-hematite veins are related to complex basinl brines.

**Stage 2 – martite-microplaty hematite-goethite supergene alteration**

The second stage resulted in the removal of most of the phosphorus from the BIF bands with goethite and anhedral hematite replacing martite, and the beds were weathered to clay with a considerable reduction in volume. The distribution of phosphorus and presence of goethite within the alteration zone suggest that this stage involved cool (<100°C), presumably shallow meteoric fluids.

**DISCUSSION**

The transformation of unmineralised BIF to high-grade iron ore provides evidence for the interaction of the progressive alteration of BIF with heated saline brine. The mineralogy of the distal to proximal alteration zones reflects this interaction. The two differences between the North deposit and the Southern Ridge deposit are firstly, the preservation of the intermediate alteration zone, and secondly the structural control on the preservation of hypogene alteration zones.

At the Southern Ridge deposit the intermediate hematite-ankerite-magnetite-talc-chlorite alteration zone present at the North deposit is not observed. The presence of talc-chlorite-microplaty hematite veins at both deposits replacing siderite suggest that similar processes where occurring at both deposits. The crystallisation of microplaty hematite at both deposits by hot (250°C - 300°C), saline (24 wt per cent CaCl₂) suggests a coeval timing of hydrothermal alteration. It is likely that the pervasive hematite-ankerite-magnetite-tal-chlorite alteration did form at Southern Ridge, however, but later hydrothermal fluids removed this alteration zone leaving only remnant talc and chlorite. This resulted in the characteristic high intergranular porosity (ten to 15 per cent) of the intermediate and proximal alteration zones at the Southern Ridge deposit. The preservation of low-temperature, saline secondary fluid inclusions (~120°C and 24 eq wt per cent CaCl₂) Thorne, Hagemann and Barley, 2004) in relict ankerite at the North deposit, suggests that ankerite dissolusion occurred late in the hydrothermal evolution of the North deposit. The lack of structural control at the North deposit may have allowed preservation of the hematite-ankerite-magnetite alteration zones. Similar salinities of this fluid suggest that the dissolution late of ankerite from the intermediate alteration zone at Southern Ridge occurred from hypogene alteration with no need for a meteoric fluid input as suggested by Taylor et al (2001) in the formation of any of the alteration zones.

Microthermometric data from the North and Southern Ridge deposit show strong similarities in the temperatures and salinities of hydrothermal brines responsible for the transformation of unmineralised chert-magnetite BIF to martite-microplaty hematite-apatite hypogene alteration. Both deposits show that the transformation occurs from low-temperature (107°C to 142°C), Ca-rich (25 eq wt per cent CaCl₂) brines during magnetite-siderite-iron silicate hypogene alteration to higher temperature (250°C - 300°C), Ca-rich (25 eq wt per cent CaCl₂) brines. Fluid inclusions trapped in quartz-hematite veins from the Southern Batter Fault, Southern Ridge deposit possibly record this fluid evolution. The salinities of primary and secondary inclusions vary in such a way that the primary inclusions display high- and low-salinity inclusions, whereas the majority of the secondary inclusions are of medium-salinity suggesting that the pattern of primary inclusions may be a result of boiling of medium-salinity brine.

The evidence from the Southern Ridge and North deposits indicates that hydrothermal brines are likely to be solely responsible for the transformation from unmineralised BIF to martite-microplaty hematite-apatite ore. At present, more detailed work on the Mt Tom Price deposit needs to be completed to confirm this. If true, it opens up a new exploration approach with the knowledge that ‘blind’, unweathered martite-microplaty hematite-apatite deposits potentially exist within the Hamersley Province.

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Oxygen Isotope Compositions of Iron Oxides from High Grade (>63 wt% Fe) BIF-related Iron Deposits of the Pilbara Province, Western Australia

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ABSTRACT: The δ18O values of iron oxides (magnetite and haematite) were analysed from unmineralised BIF, hypogene carbonate-altered BIF (proto-ore), talc altered BIF and high-grade iron ore from the Paraburdoo, Channar and Mt Tom Price deposits and Ratty Springs prospect. Iron oxides from hypogene proto-ore and high-grade ore samples are strongly depleted in 18O relative to unmineralised BIF, with supergene and talc-altered ore samples having intermediate δ18O values between unmineralised BIF and high-grade ore. Samples with the lowest δ18O values occur proximal to fault zones and are less depleted in 18O towards the periphery of the ore body as a result of increased hypogene fluid flow and fluid:rock interaction within fault zones. The δ18O values of iron oxides provide a useful tool for discriminating between supergene and hypogene ores and act as a vector for the structures that controls mineralization.

KEYWORDS: Oxygen, Isotope, Iron, Oxide, Hypogene, Pilbara

1 INTRODUCTION

This paper reports results of an oxygen isotope investigation of the iron oxides from the Mt Tom Price, Paraburdoo and Channar high-grade iron ore deposits in the Pilbara Province, Western Australia. The δ18O values of iron oxides have been studied to determine: (1) the spatial distribution of the oxygen isotope composition of iron oxides within and between each deposit; (2) the relationship of oxygen isotope values to faults zones that control hypogene fluid flow and (3) the variation of oxygen isotope composition of oxide minerals during the paragenetic steps from unmineralized BIF to high-grade iron ore.

2 GEOLOGICAL SETTING OF THE MT TOM PRICE, PARABURDOO AND CHANNAR IRON ORE DEPOSITS

The Mt Tom Price deposit occurs as a southerly dipping sheet within the eastern fold closure of the Turner syncline with the majority of the Fe ore hosted within the Dales Gorge Member of the Brockman Iron Formation. Studies of the Mt Tom Price deposit (Barley et al., 1999; Taylor et al., 2001; Thorne et al., 2004; Thorne et al., 2005) identified three hypogene alteration zones between unmineralized BIF and high-grade iron ore (Fig. 1). Firstly, a distal magnetite-siderite-stilpnomelane zone; secondly, an intermediate haematite-magnetite-ankerite-talc-chlorite and thirdly a proximal martite-microplaty haematite-magnetite-apatite alteration zone. These hypogene carbonate alteration zones are referred to as 'proto-ore' because they result in the crystallization of microplaty haematite, essential in the formation of high-grade martite-microplaty haematite ore (Thorne et al., 2004). Supergene weathering of the proto-ore, removes relict carbonate andapatite from the proto-ore. Pervasive talc alteration post-dates the proto-ore and locally forms pods of high-grade iron ore in the hangingwall of the Southern Batter Fault.

Supergene ores (i.e. Marra Mamba ore; Fig. 1) are formed solely by deep supergene enrichment of the precursor banded iron formations with no hypogene alteration (Taylor et al., 2001).

Mineralization at the Paraburdoo deposit is controlled by flat faults that now underlie both the 4W and 4E orebodies, and NW trending
eral dolerite dykes (Taylor et al., 2001). Mineralisation is equally distributed between the Dales Gorge and Joffre Members and consists of martite-microplaty ore, similar in style to the Mt Tom Price high-grade ore but with distinctly higher goethite content. A proximal haematite-ferroan dolomite-chlorite-pyrite alteration zone (proto-ore) occurs between unmineralized BIF and high-grade ore occurs at the 4W deposit (Fig. 1; Thorne et al., 2007 in press).

The Channar orebody is located on the southern limb of the Bellary Anticline with mineralisation controlled by northwest trending normal faults and dolerite dykes. Hosted within both the Dales Gorge and Joffre Members the high-grade ore is similar in texture to Paraburdoore ore but with noticeably more goethite (Fig. 1; Taylor et al., 2001). No proto-ore has been identified at the deposit.

3 SAMPLE SELECTION

A total of 49 samples were collected from diamond drill core and hand samples from the Mt Tom Price (n=36), Paraburdoore (n=9) and Channar (n=3) deposits and Ratty Springs prospect (n=1). Samples from unmineralised Dales Gorge and Joffre Member BIF, proto-ore, and high-grade iron ore from varying pit locations and depths within the deposit were sampled.

4 ANALYTICAL METHODS

The oxygen isotope composition (\(^{18}O, ^{17}O, ^{16}O\)) of the samples were measured at the Institute of Mineralogy and Geochemistry, University of Lausanne, Switzerland using a method similar to that described by Sharp (1990) and Rumble & Hoering (1994) and is described in more detail in Kasemann et al. (2001). Oxygen isotope compositions are given in the standard \(\delta\)-notation, expressed relative to VSMOW in per mil (%). Replicate oxygen isotope analyses of the standards used (NBS-28 quartz (n=13) and UWG-2 garnet (n=7) have an accuracy of \(\delta^{18}O\) values better than 0.2‰ compared to accepted \(\delta^{18}O\) values for NBS-28 of 9.64‰ and UWG-2 of 5.8‰.

\[
\begin{array}{|c|c|c|c|c|}
\hline
 & n & \text{Min} & \text{Max} & \text{Ave} & \text{In} \\
\hline
\text{Unmineralized BIF} & & & & & \\
Dales Gorge Member & 2 & 4.1 & 6.1 & 5.0 & 1.4 \\
Joffre Member & 2 & 8.8 & 13 & 10.9 & 3.0 \\
\hline
\text{Proto-ore} & & & & & \\
Mt Tom Price & & & & & \\
Distal & 3 & -9.0 & -7.7 & -8.4 & 0.7 \\
Intermediate & 8 & -6.2 & -3.5 & -4.9 & 1.0 \\
Proximal & 5 & -9.5 & -4.1 & -6.0 & 2.2 \\
Talc altered & 6 & -2.2 & 1.5 & -0.8 & 1.3 \\
Paraburdoore & & & & & \\
Proximal & 1 & -4.1 & & & \\
\hline
\text{High-grade Ore} & & & & & \\
Channar & 3 & -7.2 & -6.9 & -7.2 & 0.4 \\
Ratty Springs & 1 & -5.6 & & & \\
Paraburdoore & 6 & -7.3 & -4.9 & -5.9 & 0.8 \\
Mt Tom Price & 1 & -8.0 & -2.9 & -4.4 & 1.5 \\
Supergene Ore & & & & & \\
Marra Mamba South & 1 & -1.5 & & & \\
\hline
\end{array}
\]

Table 1. The \(\delta^{18}O\) values of iron oxides during upgrade from unmineralized BIF to high grade ore, for various deposit (Thorne et al. 2007 in press)

5 RESULTS AND INTERPRETATION

All the proto-ore, high-grade iron ore, talc-altered BIF and supergene ore samples are depleted in \(^{18}O\) relative to unmineralised BIF (Table 1). The \(\delta^{18}O\) values of iron oxides from hypothen proto-ore at the Paraburdoore deposit have a similar isotopic composition as the high-grade ore whereas Mt Tom Price proto-ore samples are similar or more depleted in \(^{18}O\) relative to unmineralized BIF. Haematite from talc-altered samples at the Mt Tom Price deposit has a similar isotopic composition as supergene ore. The \(\delta^{18}O\) value of high-grade iron ore is lowest at Channar (ave. -7.2‰), Paraburdoore (ave. -5.9‰) and Ratty Springs (ave. -5.6‰) display similar \(\delta^{18}O\) values, and the Mt Tom Price deposit (ave. -4.4‰) has the highest \(\delta^{18}O\) values.
6 CONCLUSION

The δ18O values of proto-ore and high-grade ore from the studied deposits are much lower (~9 %) relative to unmineralized BIF. This depletion in 18O occurs predominantly during the initial carbonate alteration with subsequent carbonate removal and supergene overprint having little effect on the isotopic composition of the iron oxides.

The depletion of 18O in iron oxides is strongly controlled by fault zones at the Mt Tom Price Deposit (Fig. 2), with the most depleted samples close to fault zones and less depleted samples, with respect to unmineralized BIF further from fault zones. This is interpreted as the result of higher fluid flow and isotopic exchange close to faults, with the faults serving as fluid conduits. δ18O values of iron oxides from high grade ore increase away from the faults and approach values similar to those in the host rock. The δ18O values of iron oxides thus provide a useful vector towards the structures controlling the mineralization.

As supergene weathering of hypogene samples does not appear to considerably change the δ18O values of iron oxides, this potentially may be a useful tool to discriminate between hypogene and supergene ore samples, a task that is not always easily done.

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Halogen and Cation Constraints on the Origin of Hydrothermal Fluids Forming High-Grade Iron Ore Deposits in the Hamersley Province, Western Australia

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Abstract. The numerous marteite-microplaty hematite ore deposits of the Hamersley Province represent one of the world’s premier iron ore provinces. Despite significant advances in the understanding of these deposits little is known about the source region of the basinal brines that are responsible for the transformation of BIF to high-grade ore. Detailed crush-leach analyses on fluid inclusions defined two fluids: Fluid 1 originated from evaporated seawater that had reached halite saturation. This fluid lost Mg and Na and gained Li and Ca through interactions with carbonates, feldspars, and micas, respectively. Fluid 2 was a meteoric water that interacted with halite; subsequently this fluid lost significant Na through extensive wallrock interaction with plagioclases and micas. The Cl/Br ratios of Fluids 1 and 2, responsible for hypogene alteration minerals including iron oxides, form overlapping populations, reflecting fluid mixing at the time of the formation of high-grade iron.

Keywords: Ion Chromatography, Iron Ore, Hydrothermal Alteration, Hamersley Province

1 Introduction

Significant advancement to the understanding of the high-grade (>60 wt. % Fe) martite-microplaty hematite orebodies of the Hamersley Province have been made in the last decade. Several genetic models (Taylor et al., 2001; Thorne et al., 2004, 2008) have described successive stages of Paleoproterozoic hydrothermal alteration in the transformation of banded iron formation (BIF) to high-grade martite-microplaty ore, with each stage being the necessary precursor to the next (Thorne et al., 2008).

Hydrothermal alteration at the Mt. Tom Price, Paraburdoo and Channar deposits form a well-defined vertical and lateral transition between BIF and high-grade ore. Detailed geochemistry, including microthermometry, laser-ablation of fluid inclusions, confirms (Hagemann et al., 1999; Thorne et al., 2004, 2007) that both basinal brines and meteoric waters were responsible for the formation of carbonate-altered BIF and, ultimately, the formation of high-grade martite-microplaty ore.

Despite these significant advances, the relative roles of both basinal brines and heated meteoric fluids.

2 Samples and Analytical Methods

A total of 119 samples from hydrothermal alteration assemblages and high-grade iron ore were taken from the Mt Tom Price, Paraburdoo and Channar deposits. Chemical analysis of fluid inclusions in quartz, carbonate and hematite were carried out at the University of Leeds using the bulk crush-leach method as detailed by Banks et al. (2002). Quartz, carbonate and hematite samples were crushed to a fine powder in an agate pestle and mortar, transferred to a sample container and 5-ml of 18.2 MΩ water added to re-dissolve the dried salts. Prior to analysis, the samples were filtered through a 0.2-μm nylon filter. The anions, Cl, and Br, were analysed by ion chromatography (Dionex DX-500 with an AG14 anion column) and Na, K, and Li, were analysed by flame emission spectroscopy (LaCl3 was added to give a concentration of 200 ppm La in each sample to prevent ionization of the alkalis during analysis). For both methods of analysis the typical precision was 5% RSD.

3 Crushed Leach Analysis

The halogen content of the fluid inclusions can be used to discriminate between different fluids and provides information on their origin. In the absence of evaporitic minerals, Cl and Br are not significantly involved in fluid-rock interactions, to the extent that the Cl/Br of the fluid is drastically changed, allowing the use of Cl/Br ratios in fluid inclusions to provide information on the origin of salinity in brines (Kesler, 1996; Banks et al., 2002; Lüders et al., 2005). The evaporation of seawater to halite saturation does not change the molar Cl/Br ratio of 655, however, as halite precipitation proceeds, the molar Cl/Br ratio decreases to 240 at the end of halite precipitation (Banks et al., 2002). Further evaporation results in fluids that are even more Br-rich (Lüders et al., 2005). In contrast the dissolution of halite produces Br-deficient fluids with a large range in molar Cl/Br ratios from 20,000 to 5000 (Fontes and Matray, 1993).
The fluids in this study form two distinct but overlapping populations (Fig. 1). The first population contains fluids from hydrothermal alteration assemblages and high-grade hematite ore. With a few exceptions, they have Na/Br ratios lower than seawater and plot to the left of a line in Figure 1, representing the progressive evaporation of seawater, indicating that there has been exchange of Na in the fluid for other cations. The second population from high-grade hematite ore has a wide range in Cl/Br ratios from 800 to 8000, the highest of which are likely to result from halite dissolution (c.f., Fontes and Matray, 1993). The magnitude of Na loss is much greater in the halite derived (Cl-rich) fluid (6-12 eq. wt% NaCl) than the high-Br fluid (20-26 eq. wt % NaCl) due to the lower salinity making it easier for the fluid to lose most of the Na through water-rock interaction (WRI). Sodium loss is likely to have been caused by interaction with Ca from plagioclase (Davisson and Criss, 1996) possibly contained in the overlying mafic rocks of the Weeli Wooli Formation. The distribution of Cl/Br ratios (Fig. 2) can be explained by the mixing of a low salinity, Cl-rich Fluid 1 with a high-salinity, Cl-poor brine as indicated by the mixing line in Figure 2. The mixing line was calculated using salinities of fluid inclusions and Na-Cl-Br concentrations from Tom Price (Thorne et al., 2008). The mixing line shows that the hypogene alteration zones formed by the mixing of a predominantly high-salinity brine with a minor component of meteoric water. High-grade ore samples have Cl/Br ratios that both overlap with those from hypogene alteration zones and those that indicate large input of meteoric water (Fig. 2). This suggests that fluid mixing is critical in the formation of high-grade iron. This is further supported by a salinity decease from hydrothermal alteration zones to high grade iron ore.

Figure 1: Cl/Br versus Na/Br values of hydrothermal alteration assemblages and high-grade ore.

Figure 2: Cl/Br versus Na/K ratios of fluid inclusions trapped in hydrothermal alteration minerals and iron oxide species of high-grade iron ore.
(Thorne et al., 2008).

The fluids have a wide range in Na/Li ratios of 20-720 (Fig. 3), a marked increase in the Na/Li ratios of Br- rich brines and halite. According to Fontes and Matray (1993) the maximum Na/Li ratios of Br-rich brines and halite are 6500 and 55,000, respectively. As Li is associated with micas, it is likely that the high WRI destroyed micas and produced a marked increase in the Li concentration in the fluids in both the low and high Cl-Br fluids.

4 Implications and Conclusions

The evidence from the composition of the fluid inclusions reported here indicates the presence of two fluids at the time of mineralisation. Fluid 1 originated from evaporated seawater that had reached halite saturation. This fluid lost Mg and Na and gained Li and Ca through interactions with micas and feldspars, respectively, and through dolomitisation. Fluid 2 was meteoric water that interacted with halite and subsequently lost significant Na through extensive wallrock interaction with plagioclase and micas.

The involvement of evaporated seawater as one source of the hydrothermal fluid involved in the upgrade of BIF to high grade iron ore raises the question about the source region of this fluid. Thorne et al., (2008) suggested the existence of a rift-basin to the south of the Hamersley province at the time of ore formation; which could provide a likely heat and fluid source for hydrothermal fluids. Although not preserved within the stratigraphic record in the Ashburton basin, widespread quartz pseudomorphs after evaporitic minerals are interbedded with stromatolites in 2.20 Ga sedimentary rocks in the Yerrida rift basin of Western Australia (El Tabakh et al., 1999). The Yerrida basin, together with other Early Proterozoic basins of Western Australia are inferred to have developed in a rifted passive margin between the Pilbara and Yilgarn cratons where extensive hydrothermal and volcanic activity took place (Pirajno et al., 1998). The Yerrida basin is consequently a possible heat and fluid source for the basinal brines that are critical in the formation of the high-grade martite-microplaty hematite deposits.

The low salinity and low temperature of Fluid 2 suggests that these fluids were deeply circulating, likely heated meteoric water.

The data presented here suggests that the mixing of Fluids 1 and 2 is critical to the formation of high-grade iron deposits and supports several models (Taylor et al., 2001; Thorne et al., 2004) that suggest that the mixing of two fluids is responsible for the crystallisation of microplaty hematite and ultimately the formation of low-phosphorus martite-microplaty hematite ore. Given the similarities in the temperatures (~100-300°C) and compositions of the hydrothermal fluids across the Mt Tom Price, Paraburdoo and Channar deposits the significant differences in ore quality, ore mineralogy and textures may be a result of the amount of relative mixing of the two fluid types within each deposit.

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Recent Advances in BIF-related Iron Ore Models and Exploration Strategies


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ABSTRACT

Recent research on BIF-related high-grade iron ore mineralization has resulted in new genetic models that emphasize the structurally controlled hypogene alteration and upgrade of BIF to high-grade (>65% Fe) iron ore. Conventional structural and stratigraphic mapping and reconstructions of the tectonic history of iron districts, in combination with high-tech geochemical analyses such as laser ICP-MS analyses of in situ oxides and fluid inclusions, stable (C-O-H) and radiogenic (Sr) isotopes, provide the iron explorationists with an invaluable set of tools to discover concealed iron ore bodies, deposits and districts. Two case studies from Western Australia illustrate: (1) the power of a tectonic reconstruction of the Paraburdoo Ranges and its significance for the location of high-grade hematite mineralization, and (2) the interpretation of structural controls on iron mineralization in the C deposit and its implications for resource estimation.

INTRODUCTION

The past 15 years has seen significant new research conducted on BIF-related iron ore mineralization leading to new genetic models that emphasize the role of hypogene alteration and structurally controlled hydrothermal fluid flow in the upgrade of BIF (35% Fe) to high-grade iron ore (>65% Fe). The application of hydrothermal alteration techniques such as stable isotopes, fluid inclusions, 3-D structural and ore geometry visualization (e.g., Leapfrog modeling) and geophysical methods paired with state-of-the-art numerical modeling has provided exploration geologists with significant new sets of tools to discover concealed ore bodies in established iron deposits, and hopefully new deposits in existing, and to be discovered iron districts.

This contribution highlights the recent developments in the genetic models for BIF-related iron mineralization, provides an overview of geophysical and geochemical methods that can be used to explore for, and distinguish between high-grade iron deposits and styles, and presents two case studies highlighting the role of fault zones and associated fluid flow in the formation of high-grade BIF mineralization and improved resource estimation.

RECENT ADVANCES IN GENETIC MODELS

In the mid-1990’s Hamersley Iron established a task force to investigate the potential of exploring for concealed iron ore bodies in the Hamersley Province of Western Australia. Emphasis was placed on the regional analysis of the evolution of the Hamersley Province, and review of the genesis of the high-grade iron ores. The ultimate objective was to decipher the key processes, which had created the known giant high-grade deposits, and to use this understanding to explore in locations where similar processes had occurred. This work re-ignited academic interest in the genesis of high grade iron ore deposits, and eventually led to a renewed “hypogene” ore genesis concept for deposits in the Hamersley Province in Australia and worldwide (Barley et al., 1999; Taylor et al., 2001; Dalstra et al., 2002, Hagemann et al., 2005).

Conventional wisdom in the early to mid 1990’s was that high-grade Fe ores were formed by the supergene upgrading of BIF (35% Fe) to high-grade ore (>65% Fe) during the Mesozoic time (Morris, 1980, 1985). The validity of the “young” age for the mineralization (upgrade) process was questioned by Martin et al. (1998) who applied SHRIMP U-Pb zircon age dating on zircons from volcanioclastic breccias within the lower Wyloo Group and indicated that hematite detritus in that group was derived from hematite ores older than 2209 +/- 15 Ma. This result casted doubt on the timing of the supergene model of
Morris (1980, 1985) who proposed a long period of surficial exposure between deposition of the Turiec Creek and Wyloo Groups. This date is compatible with the enrichment of iron through hypogene fluids as first proposed by Li (1993). Barley et al. (1999) used textural evidence paired with fluid inclusion data on quartz from quartz-hematite veins to propose a largely hydrothermal origin for high-grade iron mineralization at Mt Tom Price. Powell et al. (1999) used new geological mapping, basin analysis, and calculated fluid compositions (based on stable isotope data) to propose that the microplaty hematite, and possibly martite-goethite ore bodies in the Hamersley Province were formed by heated fluids (200-400°C) driven by early Paleoproterozoic Ophiolitan orogenesis (~2450 - 2200 Ma).

Hagemann et al. (1999) applied detailed microthermometry, ion-chromatography and oxygen and hydrogen isotopes on quartz, carbonates, oxides and sheet silicates to propose a 2 stage hypogene hydrothermal model of oxidation and leaching reactions for the upgrade of BIF to high-grade iron ore. Ascending fluids were derived from basinal brines whereas late, descending fluids were likely meteoric waters. Oliver and Dickens (1999) conducted stable isotope and fluid inclusion analyses of sets of regional samples from the Hamersley Province and concluded that largely meteoric waters penetrated BIF layers during active foreland fold- and-thrust belt formation and exhalation. As these fluids descended and warmed, they removed silica, locally transported iron, and oxidised magnetite-rich BIF’s to produce hematite. In 2001, Taylor et al. (2001) published the so far most comprehensive genetic model for the high-grade hematite ore bodies in the Hamersley province. They used an integrated structural, hydrothermal alteration and fluid chemistry approach to show that the hematite ores resulted from a multi-stage, sequential removal of gangue minerals and oxidation processes through hypogene fluids. The final, purely supergene stage of upgrading penetrated deep below the present surface and produced the final product of highly porous hematite ore characterized by martite and microplaty hematite interbedded with kaolinitic shale bands. In 2002, the Australian Institute of Mining and Metallurgy (AUSIMM) conducted a conference on iron ore, with a conference proceedings containing 10 papers that dealt with various aspects of hypogene alteration and mineralization in the Hamersley (Brown and Oliver, 2002; Webb et al., 2002), Iron Quadrangle (Pires, 2002; Verissimo et al., 2002), Carajas (Guedes et al., 2002) and Northern Cape Province (Carney and Mienie, 2002) Three papers evaluated the genetic models (Beukes et al., 2002; Morris 2002, and Lascuelles, 2002). Webb et al., (2003) used detailed mineralogy and geochemical analyses on samples from the Dales Gorge Member and surrounding shales from type sections near Wittenoom and the Mt Whaleback mine near Newman which revealed that: (1) no single process can produce all of the altered rocks at Mt Whaleback, (2) oxidation of magnetite to hematite can occur independently of silica removal or replacement, and (3) the main mineralization event postdates metamorphism. Spier et al. (2003) published the first comprehensive geological and geochemical description of the worldclass Agua Clara and Pico iron mines in the Quadrilatero Ferrifero. They propose that the soft high-grade ores and iron-rich tabrites are related to supergene processes whereas the hard high-grade ores are of hypogene origin. Ohtomo (2003) published an alternative mechanism for the transformation (upgrade) of magnetite-rich BIF to hematite-rich ores:

\[
\text{Fe}_3\text{O}_4(s) + 2\text{H}^+ \rightarrow \text{Fe}_2\text{O}_3(s) + \text{Fe}^{2+} + \text{H}_2\text{O}
\]

i.e., the leaching of Fe²⁺ from magnetite through hydrothermal fluids. This reaction has significant implications for any epicigenetic model as now the same “deep” hydrothermal fluids that leached silica (de-silification) would have leached Fe²⁺ from magnetite and converted it to hematite. Consequently the downwards penetrating O₂-rich meteoeric waters and resulting oxidation reactions are not anymore a necessary process for the formation of high-grade iron ore. Brown et al. (2004) linked detailed fluid chemistry data to the deformation events at the Mt Whaleback iron ore district and concluded that large amount of heated fluids (200-300°C) were expelled during D2 regional folding and thrusting. This fluid flow either continued through to the time of D3 regional extension, or two or more pulses of fluid corresponded with these deformation events. Webb et al. (2004) investigated the carbonate alteration beneath the martite-microplaty hematite ore deposit at Mount Whaleback and concluded that the lack of carbonate-rich, silica-poor rocks in the overlying Dales Gorge Member at Mount Whaleback is consistent with pervasive oxidation of most rocks in the region during or after ore genesis, a process that removed carbonates. McLellan et al., (2004) applied for the first time numerical modeling to simulate surface- and basin-derived fluids through sites of iron ore during the Proterozoic deformation in the Hamersley province. Their model supports mixing of deep basinal brines and heated meteoric fluids and formation of ore in a 2 stage process via deep “basement”-penetrating, high permeability faults.

Thorne et al., (2004) proposed a two-stage hydrothermal-supergene model for the formation of the North ore body in the Mt Tom Price deposit: Early 1a hypogene alteration involved the upward movement of hydrothermal, CaCl₂-rich brines (150-250°C) from the carbonate-rich Wittenoom Formation within large-scale folds of the Dales Gorge Member. Fluid-rock reactions transformed unmineralised BIF to magnetite-siderite-iron silicate BIF, with concomitant de-silification of the chert bands. Stage 1b hypogene alteration is characterised by an increase in temperature (possibly to 400°C) and the formation of hematite-ankerite-magnetite alteration and finally the crystallization of microplaty hematite. Late Stage 1c hypogene alteration involved the interaction of low temperature (120°C) basin brines with the hematite-ankerite-magnetite hydrothermal assemblage leaving a porous martite-microplaty hematite-apatite mineral assemblage. Stage 2 supergene enrichment in the Tertiary resulted in the removal of residual ankerite and apatite and the weathering of the shale bands to clay. Rosiere and Ross (2004) investigated fluid inclusions in hematite using infra-red microthermometry technique. They showed the Conceicao ore deposit in the Iron Quadrangle contains two types of high-grade ore bodies: hard massive, practically concordant bodies developed during the Transamazonian orogeny (2.1 - 2.2 Ga), and tabular schistose bodies controlled by shear zones that formed during the Brasiliano orogeny (0.8 to 0.6 Ga). The evolution of the hard massive ore is complex and contains two types of hematite with different fluid characteristics. No fluid data for hematite I are
available. Hematite II contains low temperature, low to medium salinity hydrothermal fluids of possible (modified) meteoric fluids. The shear zone related ore bodies contain a third hematite type formed by low temperature, high salinity fluids. A final fluid pulse characterized by high temperature and high salinity facilitated deformation and produced specularite. Dulstra and Guedes (2004) proposed that BIF-hosted high-grade hematite deposits form a coherent genetic group with proto-ores forming under relatively oxidizing conditions at temperatures near the surface to reduced conditions at temperatures above 500°C at depth. Mueller et al. (2005) constrained the maximum age of iron ore mineralization (i.e., upgrade of BIF via hypogene hydrothermal fluids) in the Hamersley Province to 2008 Ma, thereby strongly linking it to continental extension possibly related to the breakup of Paleoproterozoic Australia. A second AUSIMM conference on iron ore contained conference proceedings with 17 iron related geological papers. These included: a review paper about iron formation-hosted iron ores in the Hamersley Province (Clout 2005), and a paper that presented discrete genetic models for the Hamersley, Carajas, and Iron Quadrangle iron ore provinces (Hagemann et al., 2005). In addition, there are several regional papers about iron ore deposits or districts in Africa (4), Brazil (4), Iran (1), and Australia (6) (see AUSIMM Publication Series No 8/2005 for detailed references). Gutzmer et al. (2006) published a summary paper about oxygen isotope composition of hematite and genesis of high grade BIF-hosted iron ores. They concluded that when oxygen isotope composition of magnetite and hematite in the BIF protolith are compared with hypogene hydrothermal ore the latter are significantly depleted in 18O. Importantly, this shift is not evident in hematite ores of ancient supergene origin. The oxygen isotope composition may become a tool to categorize high-grade iron ore deposits of unknown origin or, especially in geologically complex or poorly exposed areas. Lascelles (2006) published the results of the first modern geological investigations on the iron ore in the Archean Mount Gibson BIF in the Yilgarn craton of Western Australia. He suggests that deep saprolite in situ high-grade ore may be produced by diverse processes, including hydrothermal replacement of chert mesobands by carbonates with subsequent supergene leaching of the carbonate, and by the oxidation of chert-free BIF, in which chert bands either never developed or were apparently removed during diagenesis. Importantly, neither processes requires supergene selective leaching of quartz (clay) during deep weathering. Lascelles (2006) provides a syn-genetic model for the Hope Downs iron ore deposit in the Hamersley Province. Density currents transported reworked iron silicates and hydroxides in colloidal suspension onto an unstable sea floor. The amorphous silica produced during diagenesis of Al-poor iron silicates formed the characteristic chert bands of BIF but some of the hydrous amorphous silica was lost prior to lithification to form chert-free BIF. Weathering of the chert-free BIF produced the high-grade hematite ore that is exposed today.

In summary, the past 15 years research on high-grade iron deposits and districts worldwide resulted in a significant advance in the understanding of the transformation of BIF to high-grade ore.

RECENT ADVANCES IN EXPLORATION STRATEGIES

The recognition of a hydrothermal input to BIF-related high grade iron ore will have a profound effect on exploration strategies, specifically for the exploration of concealed iron ore bodies. Recently, some exploration groups have already applied exploration techniques to characterize high-grade iron ore and defined structural and alteration vectors towards those high-grade ore bodies.

In the past exploration for BIF-related iron mineralization has been dominated by detailed lithostatigraphic and structural mapping of iron districts at different scales. In many cases, existing outcrop of BIF was mapped in detail to better constrain the geometry and possible disruptions of the BIF in three dimensions. Subsequent drilling of the area resulted in the delineation of BIF and the high grade equivalents. Under pressure from diminishing ore reserves in the Mount Tom Price deposit, Hamersley Iron started the search for concealed ore bodies which necessitated a “fresh” approach in terms of new and innovative exploration tools. Recently, CVRD has embarked on a detailed re-appraisal of the giant Carajas iron deposit and has applied new and innovative geochemical tools to constrain the hydrothermal alteration and mineralization footprint of the mineralizing system in order to better constrain potential vectors towards high-grade ore (cf. Lobato et al., 2005). The following sections discuss recent advances in the use of geophysical and geochemical techniques in the exploration for BIF-related, high-grade iron ore mineralization, in light of the advances in understanding the genesis of the BIF-iron mineral system.

GEOPHYSICAL ADVANCES

Geophysical techniques for exploration for iron have been in existence for several centuries. At the forefront of these has been the magnetic method, given the highly magnetic nature of the magnetite rich rocks that host the deposits. As the method has technically improved, from the imprecise and hard to use “compass” style of equipment, to the precise, fast, and easily operated magnetometers of today, the method’s efficacy has dramatically improved also. Gross geology mapping surveys, in which the host rocks were being sought, has given way to structure mapping, stratigraphy identification, and finally direct detection of mineralized systems. This has been achieved through increased resolution, both spatially, i.e., data density, and spectrally, i.e., dynamic range. Taken to the air in increasingly nimble and stable platforms, entire iron-rich basins can be assessed easily, quickly and cheaply and to a resolution undreamt of in the past. Regional differences could thus be observed, making for easier comparison and anomaly recognition. That anomaly recognition moved from the “bump” detection of the early to mid 20th century, to recognition of features supporting the currently understood ore genesis models. Faults controlling hypogene and supergene fluid movement, the absence, through dissolution, of carbonate-rich sections of the stratigraphy, and reduction in magnetic intensity through the oxidation of magnetite to hematite are now easily detected (e.g., Kerr et al, 1994). An example of this is the removal of
Wittenoom Formation dolomite from between the Marra Mamba and Brockman Iron Formations, which can be directly tested by recognizing the separation between the magnetic anomalies caused by the two.

The magnetite destruction associated with magnetite oxidation and deep weathering to hematite is directly measurable with the magnetic method. An example of this can be found in the Paraburdoo Eastern Ranges where the iron ore deposits form clear magnetic lows with respect to the unmineralized iron formations.

In parallel with developments in instrumentation, the magnetic properties of host rocks and ore have become better understood. Clark and Schmidt (1994) demonstrated relationships between pre- and post-tectonic magnetic remanence and used this to predict mineralization timing relationships. Further, they observe that BIF tends from highly laminated and bedded to massive as it is upgraded to ore, leading to the identification of the destruction of magnetic anisotropy, or magnetic fabric, in the ore. The emergence of full vector and tensor magnetic surveys, together with development of sophisticated numeric modeling tools, suggests an increasing ability to directly outline areas of zero anisotropy.

The gravity method is the second oldest and most widely deployed geophysical technique used in the search for iron. Both magnetite and hematite are considerably denser than the most common gangue mineral in BIF’s, i.e., quartz. The removal of the quartz component, by both hyp- and supergene processes should have a dramatic effect on the observed gravity anomaly.

Once again, through the advent of cheaper, robust, digital gravimeters, together with GPS surveying and a burgeoning of airborne gravity gradiometry systems large tracts of perspective targets are now being assessed with the method. Direct detection and correct interpretation of iron ore is somewhat complicated, however, by the ambiguity in the density contrast. The stripping of silica from BIF’s to produce high grade iron ores can result in increased porosity, and hence a lower density, or an increased density if the process includes deposition of secondary iron (or replacement by secondary iron) in the pore spaces. In any case, the gravity method’s use as a valuable mapping tool is assured.

Most other geophysical methods have been used at some stage in the exploration cycle or as problem-specific solutions. These include radiometric, DC resistivity, induced polarisation, electromagnetic, and seismic methods. As with magnetics and gravity, their deployment in regional surveys, for example as large airborne surveys (particularly radiometrics and EM) they serve to assist in stratigraphic and structural mapping.

At a local level the most useful of these has been the downhole radiometric method. The iron rich stratigraphy in the Hamersley Basin, for example, is regularly punctuated by thin shale bands that are conveniently mapped by the method (Harmsworth et al., 1990). While not a direct exploration tool, it has been used extensively by workers to accurately correlate intercepts between holes and elucidate complex structure in an otherwise extremely regular stratigraphy. Despite its demonstrated successful application in the Hamersley Province, down hole gamma logging is rarely used in other iron ore provinces. While regionally the aeromagnetic application of the radiometric method adds little to mapping in well-known areas such as the Hamersley Basin, it provides important lithological information in less well mapped iron provinces.

An increasingly important method that is currently gaining widespread favor is the airborne Transient Electromagnetic TEM method. In its frequency domain guise (FEM) the method has been long known for its ability to differentiate magnetite (Fraser, 1981), suggesting it may be used to directly discern the spatial genesis from magnetite rich proto-ore to hematite ore. In its deeper “looking” time domain guise TEM offers the hope of direct detection of that transition.

GEOCHEMICAL ADVANCES

With the change in the genetic model for high-grade BIF-related iron ore mineralization from supergene to a mixed hypogene and supergene genetic model a change in the choice of geochemical techniques has been observed. In the past, routine XRF and ICP-MS analyses were used to constrain the iron content and contaminants of the enriched BIF. Today a more sophisticated approach is used to constrain the P-T-X conditions of the upgrade from BIF (35% Fe) to low-grade (40-50% Fe) proto-ore to high-grade (>65% Fe) iron ore. Furthermore, specific vectors to high grade iron ore are applied that may assist the explorationist to identify fertile fault or fold systems early in the exploration campaign and specifically for the exploration and exploitation of concealed ore bodies.

Geochemical techniques used include: (1) electron microprobe and laser ICP-MS analyses of oxides and carbonates in order to constrain the mineral chemistry, (2) carbon and oxygen isotopes on the different carbonate species, (3) oxygen and carbonate veins is essential in defining mineral vectors towards high-grade ore. An example is the spatial distribution of the Mn content in carbonates with respect to the proximity to high-grade iron ore (Figure 1) At both the Mt Tom Price and Paraburdoo iron deposits the Mn content increases markedly in vein hosted carbonates with proximity to the ore bodies.

The carbon isotope signature of the different carbonate species from the Hamersley Province (Figure 2) displays a systematic change in δ13C between unmineralized BIF and hypogene carbonate alteration (cf. Thorne et al., 2004). The similar oxygen isotope compositions, but increasingly heavy carbon isotopes from magnetite-siderite-iron silicate to hematite-ankerite-magnetite alteration zones, suggest the progressive exchange (mixing) with an external carbon-rich fluid with a heavy carbon isotope signature.

Recent analyses of the δ18O composition of hypogene iron oxides in the Southern Bitter fault zone located in the central portion of the Mt Tom Price deposit (Figure 3) revealed a distinct δ18O gradient from δ18O values of -9‰ in the core to -
2.5% at the edges of the fault zone (Thorne et al., in press). The spatial variation in $\Delta^{18}O$ values are interpreted to be controlled by higher hydrothermal fluid flow (high water/rock ratio) in the proximity of the fault zones. Results also show that supergene overprint of hypogene formed proto-ore (i.e., carbonate alteration) has little effect on the $\Delta^{18}O$ isotopic composition of high-grade ore (cf. Gutzmer et al., 2006). The latter may be a powerful tool to discriminate between supergene and hypogene formed iron oxide ore.

Figure 1: Mn versus Fe content plot of vein carbonates at the Mt Tom Price and Paraburadoo iron deposits with respect to the proximity to high-grade iron ore.

Figure 2: $\delta^{18}O$–$\delta^{13}C$ diagram showing the isotopic composition of the various carbonate populations in the North and Mt Tom Price deposits.

Figure 3: Sample distribution, hypogene alteration zonation and $\Delta^{18}O$ compositions of iron oxide samples through section 14000E, Mt Tom Price Deposit.
CASE STUDY 1-PARABURDOO RANGES

Structure is the most important control on high grade hematite deposits, and therefore understanding the structural framework of Proterozoic iron ore basins is critical for locating new deposits (cf. Dalstra 2005). Because high-grade iron ore deposits in the Hamersley Province formed relatively early in the tectonic history (c.f. Powell et al., 1999; Taylor et al., 2001), some of the ores have undergone several phases of deformation, metamorphism and hydrothermal alteration associated with the formation and re-activation of fault zones subsequent to ore formation. To understand the setting at the time of mineralization it is therefore critical to reconstruct the structural geometries through step by step removal of the later events. One area where a reconstruction of the fault zones has helped to identify and explain concealed high-grade iron ore bodies is the Paraburdoo Range in Western Australia. The following case study based on the detailed structural analysis by Dalstra (2005) illustrates the significance of the structural unrolling of a given mineralized area in the exploration for concealed ore bodies.

The Paraburdoo high-grade iron deposits are located about 65 km south of the Mount Tom Price deposits and contained a pre mining ore reserve of more than 300 Mt at 64% Fe and 0.08% P (Taylor et al., 2001). Several separately named deposits constitute the Paraburdoo Resource, including, among others, 11West, 4West, 4East, 23East, Eastern Ranges, 64East, and Channar (Figure 4). The Hamersley Group BIF’s host the ore, and dip steeply to the south. These are overlain unconformably by the Wyloo Group (Taylor et al., 2001). The flat 4W and 4E Basal faults and the steeply northeast-dipping 18E faults and Ratty Springs Fault displace the Hamersley Group and also cross-cut the Lower Wyloo Group unconformity (Figure 4, 5A). They are, in turn, truncated by the Upper Wyloo Group unconformity at the base of the Mount McGrath Formation, then rotated by the Capricorn folding event, which has imposed the southerly dip on all beds. Prior to the Capricorn folding the flat faults that now underlies both the 4W and 4E iron ore bodies were steep, north-dipping normal faults (Taylor et al., 2001; Dalstra 2005).

Dalstra (2005) pointed out that in order to reconstruct the setting of ore formation at Paraburdoo the hematite conglomerates beds at the base of the Mt McGrath Formation (Figure 5A) need to be back-rotated in order to assume their near horizontal depositional setting. In the western section of the Paraburdoo Ranges, the conglomerate beds dip 40-50° SSW indicating that the mineralized geometry was significantly modified during the Capricorn orogeny (Dalstra 2005). Back rotation of the 4 East deposit to synUpper Wyloo Group deposition shows that the steeply dipping reverse 18 East fault likely originated as a moderately steep SW dipping normal fault (see Figure 5B). Analysis of the geometry of the fault system depicted in Figure 5B also suggests that the hematite ore formed in an extensional graben at least 500 to 800 m below the McGrath unconformity.

Figure 4: Geological map of the BIF-related high-grade Paraburdoo deposits in the Hamersley Province of Western Australia displaying location of ore bodies and open pits. Also shown are the lithostratigraphic setting, major fault systems, and unconformities. This map is the result of detailed Hamersley Iron Pty Ltd open pit and regional mapping as well as interpretation of drill hole data.
A reconstructed long section through the Paraburdoo Ranges syn-Upper Wyloo group deposition indicates that the proto-hematite ore bodies had already formed at that time and that some were actively eroding, forming hematite conglomerates (Dalstra 2005). The long-section also shows that ore bodies at 4 West, 4 East and 64 East formed in grabens or half-grabens, at least several hundred of meters below the surface, thus were shielded from erosion and thus preserved. Only the 11 West and Channar iron deposits formed on horsts and are still preserved; ore bodies on the Ratty Springs- and 18 East Horsts that may have been present prior to the erosion before Lower-Wyloo Group deposition likely were eroded with their ores now making up the hematite conglomerates that are located immediately adjacent to these horsts.

Reconstructing the geology of the Paraburdoo Ranges, syn-Lower Wyloo group deposition, i.e., prior to the iron ore mineralization suggests that the geometry of the area is a series of half grabens progressively stepping down to the ENE (Dalstra 2005). Oblique reactivation of these early extensional normal faults prior to the Upper Wyloo Group may have been instrumental to tap into the underlying dolomites and create fluid pathways. These pathways are essential for silica-undersaturated, hypogene hydrothermal fluids to ascend upwards into the BIF, causing large-scale silica removal and subsequent iron enrichment that formed the giant iron ore bodies.

Tectonic reconstruction of the mineralized environment of the Paraburdoo Ranges results in a number of testable predictions, both for exploration and future research. The most critical prediction is that proto-ores to high-grade hematite deposits can form at least several hundreds of meters below Proterozoic unconformity surfaces, and possibly much deeper. It also predicts that there could be a range in depths for proto-ore formation, the Channar and 11West Deposits representing the shallowest systems and the 4East the deepest at Paraburdoo. All deposits in the Paraburdoo Ranges however formed most likely significantly closer to the paleosurface than the giant Whaleback and Mt Tom Price Deposits, both of which are characterized by an absence of Proterozoic unconformity surfaces nearby.

**CASE STUDY 2 THE C DEPOSIT**

Area C is located approximately 100 km northwest of Newman and contains ~1100 Mt of Marra Mamba and Brockman M-G (martite-goethite) iron mineralization. Ore-grade mineralization in the C Deposit of Area C is hosted almost entirely in the Mount Newman Member of the Marra Mamba Formation. The
dominant structures in the vicinity of Area C are variably east-west trending folds (Figure 6), which formed during two major north-south compressional tectonic events. During the Ophthalmian Orogeny (~2450 - 2200 Ma), an arcuate fold and thrust belt formed, with associated south-dipping thrust faults and north-verging asymmetric to overturned folds. During the Ashburton Orogeny (~1800 - 1650 Ma), broader, more open and upright folds formed, including the Weeli Wolli anticline at Area C. All Area C deposits are located on the northern limb of the Weeli Wolli anticline (Figure 6).

Mining operations in the C Deposit bulk sample test pit in 2002 exposed deposit-scale thrust faults and overturned folds (Figure 7), which are significant controls on the distribution of higher-grade mineralisation (> 60% Fe). In detail, these controls include: (1) south-dipping thrust faults that developed along the Mount Newman/West Angela contact, (2) folds associated with thrust fault development, (3) steeply dipping stratigraphic contacts in the hangingwall and footwall of thrust faults, and (4) gently north-dipping stratigraphic contacts. The thrust fault and folds are significant for two main reasons (Hodkiewicz et al.,

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Figure 6: Plan geology of Area C, showing regional east-west folds and the location of the C Deposit on the north flank of the Weeli Wolli Anticline. Modified from Kepert (2001).

Figure 7: View looking southeast in the C Deposit bulk sample test pit, showing sub-vertical Mount Newman units (N1, N2 and N3) in the hangingwall of the Western Thrust Fault, overlying West Angela units (WA1 and WA2) in the footwall.
2003): (1) their formation resulted in the thickening of the main ore-hosting units in the Mount Newman member, and (2) associated structures and folded lithological contacts were conduits for fluids that influenced the formation of iron mineralisation. Similar thrusts faults and small- to medium-scale north-verging folds occur throughout the district. The faults and folded contacts in the C Deposit (Fig 7) were structural pathways for fluid flow and are therefore significant controls on the distribution of mineralisation.

Mineralization domains correspond to geologically and statistically homogeneous zones (Gui bal, 2001), and the construction of accurate 3D wireframe models of mineralised domains is a critical component of any constrained resource estimation. In this study, Leapfrog™ software was used to create 3D wireframes based on drill-hole composites, in order to determine the continuity and geometry of mineralisation and to compare with existing BHPBIO geological interpretations and wireframe models.

Leapfrog™ is specialist software developed by SRK Consulting and ARANZ for 3D contouring of drill hole data and the rapid construction of wireframes. Wireframes of assay data highlight the geometry of mineralised domains and assist in the interpretation of structural and stratigraphic controls on mineralisation at a variety of scales.

A simplified cross-sectional view of wireframes through the bulk sample test pit in C Deposit is shown in Figure 8. The wireframes at 64%, 60% and 54% iron highlight structural and stratigraphic controls on mineralisation, including the sub- vertical orientation of mineralisation along folded units in the hanging wall and foot wall of the Western Thrust Fault, and the gently north-dipping mineralised domains associated with lithological contacts. Significant changes in dip of the mineralisation confirmed the locations of dip-domain boundaries that had been identified by BHPBIO geologists for use in the resource estimation.

In Figure 8 only three iron wireframes are shown for clarity. However, in this study, iron was modelled at 1% increments from 55% to +62%. The resulting wireframes clearly highlight the distribution of mineralisation and specific controls associated with a complex network of variably oriented faults and folded stratigraphic contacts, which are critical for iron mineralization.

An improved understanding of controls and the geometry of iron mineralisation at a range of cut-off grades is important for the construction of valid wireframe models of mineralised domains.

Understanding the nature of the structural controls on mineralisation is fundamental to designing appropriate data acquisition methods for resource estimation. As seen in Figure 8, initially all resource definition drill holes in C Deposit were drilled vertically, and therefore oriented parallel to stratigraphy in upright folds, where stratigraphy is also vertical. Based on an improved understanding of structural features in C Deposit, BHPBIO geologists designed angled drill holes to provide better sample coverage for geological modelling and resource estimation.

Grade interpolation methods should also take into account the stratigraphic and structural controls on mineralisation such that samples are correlated appropriately. Grade contours shown in Figure 8 show that, while the overall morphology of the mineralised domain (illustrated by the 54 %Fe contour) appears sub-horizontal, the control of higher grade (>60% Fe) mineralisation is sub-vertical where strata have been folded and displaced by thrust faulting.

Deposit-scale structural controls on iron mineralisation in C Deposit reflect regional-scale tectonic features in the Hamersley Province. The major controls are distinct fold styles associated with the Ophthalmian and Ashburton orogenies, which were dominantly north-south compressional events. Early Ophthalmian structures include an arcuate, east-west trending fold and thrust belt with associated south-dipping thrusts and north-verging folds. Broader, more open, east-west folds formed during the Ashburton, including the Weeli Wolli anticline, which hosts the Area C deposits.

Recent mining operations in C Deposit have exposed thrust faults and overturned folds that control higher-grade iron mineralisation. These structures thickened the main ore-hosting units in the Mount Newman Member and provided conduits for fluids that influenced iron mineralisation. Detailed mapping provided the basis for updated geological interpretations and the construction of valid 3D wireframe models at a range of cut-off grades. These were critical for defining domains used in the resource estimation and for designing more appropriate data acquisition programs.

Figure 8: Cross section through the C Deposit looking west, showing mineralised domains at 64%, 60%, and 54% Fe. The thicker high grade zones to the south (left side of the section) in folds associated with the Western Thrust Fault.
SUMMARY AND CONCLUSIONS

The identification of distinct hydrothermal alteration zonation around the Hamersley and Carajás deposits, and the identification of high- and low-temperature and -salinity fluid inclusions in hematite species from the Iron Quadrangle strongly suggests that structural controlled fluid flow and hydrothermal processes do indeed play a significant role in the transformation of BIF (35%Fe) into high-grade (>60% Fe) iron ore. As a consequence new genetic models have been developed that propose a dual hypogene-supergene origin for BIF-related iron ores worldwide.

These new genetic models have changed the exploration strategy for the major iron ore producers. Today many explorers target concealed, high-grade iron ore by applying a combination of traditional exploration tools such as downhole gamma logging with modern academic techniques such as laser ICP-MS analyses, microthermometry and stable isotopes. The future will undoubtedly see further refinements of the genetic models and development of high-tech analytical tools. In combination, and paired with the capability of the modern explorationist to rapidly understand the geological controls of specific iron province, this will lead undoubtedly to new and exciting discoveries of high-grade iron bodies.

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Mineralizing fluid evolution and REE patterns for the hydrothermal Carajás iron ores, Brazil, and for selected Hamersley iron deposits, Australia

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ABSTRACT: In the Carajás Northern Ridge hydrothermal iron ore deposits, different types of oxides are present, microcrystalline haematite being the most primitive, while euhedral-tabular haematite the late-stage hydrothermal phase. Relative to jaspilites, some rare earth elements (REE) patterns in ore samples show general increase in ΣREE, light-REE enrichment, a near-horizontal configuration and Eu anomalies that ranges from positive to negative. Ore samples from selected Hamersley deposits show similar features, with less pronounced enrichment in LREE. The shifts in the REE patterns are directly related to the mineralogical evolution. The overall higher REE contents of the Carajás compared to the Hamersley ore samples suggest different mineralizing fluid sources.

KEYWORDS: iron ore, Carajás, iron oxides types, rare earth elements, Hamersley

1 INTRODUCTION

Structurally controlled (Rosière et al., 2006), hydrothermal iron ore deposits (Lobato et al. 2005a, b) are located in the Carajás Province of Brazil, hosted in the Archaean metavolcano-sedimentary Grão Pará Group, Itacaiúnas Supergroup. Jaspilites (JP) host discontinuous, high-grade, soft and hard iron ore bodies (>65% Fe), which are under- and overlain by basaltic rocks (Zucchetti et al. 2007). The N4 and N5 active open pits produced in 2004 about 80 Mt of iron.

Different iron oxides in JPs and iron ores give place to various rock associations, which reflect the intensity of hydrothermal alteration. Distinctive geochemical changes, experienced during alteration attending iron mineralization, are spectacularly displayed by rare earth elements-REE data (Figueiredo e Silva 2004), and help constrain fluid source and evolution. The Carajás REE results are compared with the Tom Price, Paraburdoo and Brockman BS2 iron deposits, Hamersley, Australia.

2 PETROGRAPHIC STUDIES

Samples from drill holes (down to 300 m) were selected from the N1, N4 and N5 deposits of Carajás. These encompass haematitised JPs and hard iron ores (Figueiredo e Silva 2004), which are banded and/or brecciated, and variably porous. Jaspilites contain iron oxides, jasper, chert-quartz, whereas iron oxides dominate iron ores; carbonate and sulphide veinlets are subordinate.

For the Carajás Northern Range deposits, a mineralogical sequence for the iron oxides has been established: MiH → Mgt → Mt → MpH → AnH → EH-TH (Figueiredo e Silva 2004).

Microcrystalline haematite (MiH) – the earliest oxide, which is overgrown by magnetite (Mgt) porphyroblasts. It is typical for the jaspilite layers but stable even in the more advanced hydrothermal alteration stages.

Martitised magnetite (Mt) – has knenomagnetite (Kmg) relics, and is commonly surrounded by haloes of recrystallised jasper, resulting in fine-grained, haematite-free granoblastic quartz.

Micro-platy or lamellar haematite (MpH) – commonly partially fills vugs, or is associated with quartz and carbonate veins.

Anhedral haematite (AnH) – results either from progressive martitization, forming abundant lenticular, mosaic agglomerates, or from MiH recrystallization, mainly where in contact with euhedral and/or tabular (EH-TH) haema-
tite veins. The latest-stage oxides EH-TH haematites occur in discordant veins and veinlets. Usually comb-textured, they further cement open spaces or vugs. Growth lamellae are common in EH, and locally also occur in AnH, TH, rarely in MpH. These features indicate shallow, epithermal crustal conditions (Lobato et al. 2005a, b).

Banded, brecciated and porous ore samples from the Mt. Tom Price, Paraburdoo and Brockman deposits show the following noteworthy characteristics:

(i) Mt. Tom Price: Eight samples. AnH dominates the majority of samples, in which MpH partially occupies vacancies. Kmg relics are present in one sample, both in Mt and in MpH. Where MpH (=AnH) dominates, it may be comb textured.

(ii) Paraburdoo: Seven samples. One sample is Mt rich with MpH, the other dominated by AnH massive bands + MpH veins and vugs.

(iii) Brockman BS2: Four samples. One rich in Mt bands + MpH vugs, and the other is dominated by MpH, which is coarser grained close to the vugs.

The petrographic study clearly shows that Mt aggregates may evolve to form AnH and/or MpH. However, MpH is more common in vugs and veins.

All Carajás samples are devoid of weathered products, whereas some Hamersley samples contain goethite, suggesting weathering overprint. The petrographic study reveals a close similarity in the mineralogical and textural characteristics of the Carajás and Hamersley iron ores.

Figure 1. Chondrite-normalised REE diagrams. (A) N4W and N5E jaspilites. (B) Different types of iron ores containing oxides described in the text (most Carajás data from Figueiredo e Silva 2004). Carb: carbonate (C) BIF from the North Deposit, Dales Gorge (data in Alibert & McCulloch 1993). (D) Different types of iron ores from deposits of the Hamersley basin.
3 RARE EARTH ELEMENTS - REE

The REE patterns of the Carajás JPs (Fig. 1A) depict low ΣREE contents, light REE enrichment, and positive europium anomalies (Eu/Eu* > 1), similarly to Archaean BIF worldwide (e.g., Klein & Ladeira 2002).

Various REE populations of hematitised jaspilites and ores are identified (Fig. 1B).

(i) N1 (horizontal) and N4E (hatched lines; Fig. 1B) ore samples exhibit LREE enrichment and, like JPs, positive Eu anomaly. This resemblance may be explained by the fact that these ores have oxides similar to those of hematitized JPs (Mt-rich, MpH, MiH). The similarity is also born out by the low heavy REE signature below chondrite.

(ii) LREE-enriched N4E ore samples with negative Eu anomaly (wide crosses, Fig. 1B) are MpH-rich, with Mt ± MiH. This association is interpreted as the intermediate hydrothermal mineralization stage.

(iii) One group of N5E ores (vertical lines, Fig. 1B) displays the highest ΣREE, mainly LREE. They are composed essentially of late-stage haematites, AnH-EH-TH and MpH.

(iv) Another N5E group (dense grey, Fig. 1B) has near-horizontal REE patterns, with relative LREE and HREE enrichment.

The patterns of Hamersley BIF deposits (Fig. 1C) depict low ΣREE contents (samples ND85/79(207)BIF14 & 15 from Alibert & McCulloch 1993). While the positive Eu anomaly of the Archaean Carajás JPs is pronounced, it is discrete in the Palaeoproterozoic Hamersley BIFs.

Patterns of banded and brecciated ore samples from the Mt. Tom Price, Paraburdoo and Brockman (Fig. 1D) iron ore deposits allow the following interpretation:

(i) All samples exhibit REE enrichment, especially LREE. The HREE have a near-horizontal pattern.

(ii) Tom Price samples exhibit the highest ΣREE contents. This is the only deposit where most of the ore samples contain AnH (at the expense of Mt), and also MpH and Mt.

4 DISCUSSIONS AND CONCLUSIONS

The hydrothermal iron mineralization at Carajás was associated with: (i) early formation of Mgt; (ii) jasper recrystallisation to quartz with its cleansing with expulsion of MiH (V1 quartz vein type of Figueiredo e Silva et al. 2007); (iii) progressive leaching of chert, jasper, or quartz, leaving MiH and innumerable vacant spaces; (iv) partial space filling by MpH; and (v) continued, more efficient space filling by EH and TH.

Leaching of SiO₂ and quartz in BIF is principally a function of temperature, in the 150-250°C range (Taylor et al. 2001), certainly the case in Carajás (Lobato et al. 2005a, b). The evidence suggests that martitization is somewhat contemporaneous to chert recrystallisation and to the onset of SiO₂ leaching of jaspilites.

As pointed out by Lobato et al. (2005a, b), the Carajás porous, high-grade soft ore, rich in MpH and AnH, must have derived from interaction with this relatively hot SiO₂-leaching fluid. On the other hand, the hard ore probably developed due to closer fluid-rock equilibrium conditions allowing the precipitation of the EH-TH that cements the soft-haematite ore, typically at N5E.

Early-stage SiO₂ leaching may have resulted in a relative general REE augmentation during the incipient hydrothermal phase. However, the significant REE enrichment in ores, with associated changes in the shapes of the REE patterns, must have resulted from interaction with the mineralizing, hydrothermal fluid.

The first significant shift in LREE occurs in the Mt(±Mgt)-rich Carajás samples. This can be explained by the distribution coefficients KD of magnetite, which favours the fixation of LREE (Li 2000). As a result, MpH formation and the advance of martitization to form AnH may have favoured the relative increase of HREE in the residual fluid. This lead to the near-horizontal REE patterns associated with the later stages of mineralisation, of ores dominated by AnH-EH-TH and MpH (± MiH in relic SiO₂-leached jaspilite fragments). A general REE enrichment with near-horizontal HREE patterns are also present in ores from the Mt. Tom Price Deposit, containing AnH and MpH (Fig. 1D).

As fluid-rock interaction advanced, more oxidising conditions prevailed, and new haematite types developed. The remaining Eu²⁺ was no longer incorporated into the rocks, resulting in REE patterns with a weakly negative Eu anomaly in high-grade ores samples (Fig. 1B).

At Carajás, the development of Mgt(±Mt) as the incipient-alteration oxide suggests that the early-stage hydrothermal fluid had an fO₂ in disequilibrium with the original oxide MiH (Lobato et al. 2005a, b). The ore samples that maintained an Eu positive anomaly (Fig. 1B),
similar to JPs (Fig. 1A), probably equilibrated with a fluid capable of retaining Eu$^{2+}$.

The LREE enrichment in Hamersley ($\text{La} = \text{max. 30x chondrite, especially Mt. Tom Price}$) and most of the Carajás ($\text{La} = \text{40x chondrite}$) ore samples is similar. However, the N5E samples (vertical lines, Fig. 1B) represent an exceptional group, dominated by AnH-EH-TH + MpH and lacking MiH, with $\text{La} \approx 70x$ chondrite. Indeed, to date N5E is a distinctive deposit within the entire district, hosting mostly hard haematite ore. Its units are interpreted to have been the loci of the highest fluid:rock ratios (Lobato et al. 2005, Rosière et al. 2006). The N5E REE geochemical features may thus be used as a powerful exploration tool.

Mt. Tom Price ore samples from the Centre Pit and Southern Ridge deposits are typically more folded and harder. Also, they come from different proximities to the Southern Batter Fault (e.g., Hagemann et al. 1999). The $\delta^{18}O$ study of iron oxides shows that areas of higher fluid flow near faults have the greatest isotopic interchange (Thorne et al. 2007). Such higher fluid-rock ratios are also reflected by the REE evolution, since the Mt. Tom Price samples exhibit the highest $\Sigma$REE contents (rich in AnH).

The overall higher REE contents of the Carajás compared to the Hamersley ore samples suggest that different fluid sources may have been involved in the deposit origins. For Carajás, a magmatic fluid source has been postulated (Lobato et al. 2005; Figueiredo e Silva et al. 2007), whereas Hagemann et al. (1999) indicate that basinal brines were involved in the origin of the Mt. Tom Price deposit.

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Hypogene fluid responsible for the transformation of BIF to high-grade iron ore (>65 wt %); insights from the 4 East deposit, Paraburdoo, Western Australia

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The Paraburdoo deposit lies about 65 km south of Mount Tom Price at the southern margin of the Hamersley Province. Recent deep drilling at Paraburdoo, below the 4 East deposit, was completed to delineate deep extensions of the ore body. The drilling also intersected carbonate-altered (50-55 wt% Fe; hematite-dolomite-chlorite-pyrite) Banded Iron Formation (BIF) transitional between unmineralised BIF (~35 wt% Fe) and high-grade (>65 wt% Fe) iron ore mineralisation, below the depth of modern weathering. Structural reconstruction of the 4 East deposit suggests that the flat fault zones that now underlie the deposit were steeply dipping normal faults, prior to late tilting. Hypogene carbonate alteration formed primarily in the hangingwall of these steeply dipping normal faults. Supergene overprint (removal of hypogene carbonate) of the hypogene carbonate-altered BIF results in the formation of high-grade iron ore which consists of martite, microplaty and anhedral hematite.

Based on a careful petrographic study, primary fluid inclusions are observed in dolomite from dolomite-chlorite-pyrite veins in the hematite-dolomite-chlorite-pyrite alteration zone. Fluid inclusions in dolomite are rounded, ovate and irregular in shape. Their sizes are between 5-40 μm in diameter. They contain a liquid and vapour phase with L/V ratios of 0.6 to 0.95. Fluid inclusions in dolomite reveal only H₂O–CaCl₂ primary inclusions that show high salinity (average 20.9 CaCl₂ eq. wt%) and TH₉₀°⁻_I₉₀°⁻ = 130 ± 11 °C (1σ; n=52). Quantitative LA-ICPMS microanalysis of primary fluid inclusions from dolomite in dolomite-chlorite-veins yield 6.1-10.8 wt% Ca (ave. 8.2 %), 5.4-9.5 wt% Na (ave. 7.3 wt%), 1.0-7.7 wt% K (ave. 5.2 wt%), 0.6-5.2 ppm Cu (ave. 2.2 ppm), 0.8-12.2 ppm Zn (ave. 5.4 ppm), 1.7-10.3 ppm Sr (ave. 5.4), 0.4-2.2 ppm (ave. 1.1 ppm) Ba, and 0.2-0.9 ppm Pb (ave. 0.4 ppm). Ion chromatography leachates (9 samples) from dolomite and hematite within hematite-dolomite-chlorite-pyrite alteration and hematite from high-grade ore show similar mass ratios with Na/K = 3.9 ± 2.1 (1σ, n = 9), Cl/SO₄ = 297.1 ± 252.9, and, Na/Li = 1.1 ± 0.7. The samples have molar Cl/Br ratios of 175.1 ± 45.3, Na/Br = 45.1 ± 39.5 and K/Na of 0.2 ± 0.1.

The fluids responsible for the formation of hypogene alteration and ultimately high-grade ore at the 4 East deposit at Paraburdoo were low temperature, 130 °C, saline fluids (Ca-Na-K) with elevated Cu and Zn (1.3 and 3.2 ppm, respectively). The high Cl/Br, low Na/Br ratios, high-Ca concentrations, and base metal content of the fluids suggest that these basinal brines formed from evaporated sea water and have undergone extensive dolomitization (Luders et al., 2003). Normal faults that underlie the 4 East deposit provide the fluid pathway for hypogene fluids from the underlying carbonate sequences of the Wittenoom Dolomite into the unmineralised BIF. Such fluid flow and fluid geochemistry are typical of Mississippi Valley-Type (MVT) deposits, where faults are the principal conduits by which ascending metalliferous hydrothermal fluids access limestone units resulting in in-situ dissolution of the limestone and replacement of the host-rocks and mineralisation. The results from this work provide new insights in the fluids responsible for the transformation of unmineralised BIF to high-grade ore in the 4 East deposit at Paraburdoo. It also highlights the importance of basinal faults that link the underlying Wittenoom Dolomite with the BIF as a primary control on mineralisation.

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