A New Fluid-Flow Model for the Genesis of Banded Iron Formation-Hosted Martite-Goethite Mineralization, with Special Reference to the North and South Flank Deposits of the Hamersley Province, Western Australia

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Abstract

The North and South Flank deposits are located on the flanks of the Weeli Wolli anticline at Mining Area C in the central Hamersley Province. Supergene martite-goethite mineralization is hosted within the Marra Mamba Iron Formation and is developed over a strike length of more than 60 km. This multibillion metric ton resource has been drilled out on a 150- x 50- to 50- x 50-m grid, thus providing us with an unprecedented data set for analysis. This study synthesizes the drill hole data and presents a physical process model that can account for the observed distribution of mineralization.

A fluid and mass flux model is proposed which envisages a three-stage process: (1) leaching of Fe from banded iron formation (BIF) in the vadose zone by reduced, acidic, meteoric-derived fluids; (2) penetration of an Fe-rich supergene-fluid plume, driven by gravity and focused by bedding-parallel permeability into the body of ambient alkaline groundwater, effecting nonredox, mimetic replacement of magnetite by hematite and of the gangue minerals (carbonate, silicate, and chert) by goethite coupled with the release of silica into the fluid phase; and (3) a change from silica leaching to silica deposition on the downdip margins of the system before the ore-fluid plume is eventually diluted and becomes indistinguishable from the surrounding body of groundwater.

Despite the undoubted secondary role played by structurally enhanced permeability, the primary control on ore-fluid hydrology is gravity-driven flow along bedding planes. This central observation explains every observed feature of the three-dimensional distribution of martite-goethite mineralization, and the inherited structural architecture simply provides the context for this process to play out. This type of control is by no means obvious—the ingress of meteoric fluids during later lateritic weathering of the mineralization does not show this control and produces broadly subhorizontal, bedding-discordant zones of overprinting.

The fundamental control exerted on the distribution of martite-goethite mineralization by bedding-plane permeability within BIF horizons suggests that the supergene ore-fluid plume created its own porosity via the relevant ore-forming reactions, and that these were in turn controlled by bedding. A corollary of the pseudomorph replacement process, both the generation of hematite after magnetite and goethite after gangue phases, is that it typically introduces porosity. The mineralizing process thus creates porosity (and potentially permeability) and is likely to be self-propagating as long as there is continuous supply of ore fluid. This putative active porosity-generation process may be an important clue as to the unique conditions of martite-goethite ore formation. Indeed, it may be that the distribution of magnetite is the critical controlling feature of these ore systems, as the nonredox transformation to hematite not only releases Fe\textsuperscript{3+} to the fluid phase but concurrently introduces porosity. Further research is required to formulate a comprehensive chemical (as opposed to physical) process model for supergene martite-goethite ore formation.

Based on the physical process model presented here, the development of a large-scale martite-goethite mineralizing system requires continued delivery of unleached BIF (and, perhaps ultimately, previously mineralized martite-goethite ore) into the vadose zone. The Hamersley Province has been undergoing significant uplift since at least 60 Ma. Preliminary dating of martite-goethite ores from Mining Area C indicates that they formed at about 45 Ma, at a time when the local climate was temperate and wetter than today. The combination of ongoing uplift and a wet, temperate climate is likely to be the key to the widespread formation of martite-goethite deposits in the Hamersley Province.

Introduction

Australia is one of the world’s biggest exporters of high-grade, direct-shipping iron ore with resources in excess of 70 billion metric tons (Gt; Knight et al., 2018). This resource base comprises three distinct ore types. Two of these are hosted by banded iron formation (BIF) units: Proterozoic, hypogene martite-microplaty hematite (M-mplH) ore and Cenozoic, martite-goethite (M-G) ore (Clout and Simonson, 2005). The third ore type includes a variety of Proterozoic, Mesozoic, and dominantly Cenozoic detrital ores, of which only the Mio- cene, fluvialitic, pisolitic, channel iron deposits are mined as stand-alone operations (Kneeshaw and Morris, 2014).

Although hypogene martite-microplaty hematite ores make up a very small proportion of the current resource base, the first mines to open in the Hamersley Province were of this type (Mount Tom Price opened in 1966, followed closely by...
Mount Whaleback in 1968, and then by Paraburdoo in 1972), and these operations have been the mainstay of the Western Australian iron ore industry for the past 50 years. As a result, much of the research conducted over the past two decades has concentrated on this deposit type (Barley et al., 1999; Powell et al., 1999; Taylor et al., 2001; Brown et al., 2004; Thorne et al., 2004, 2005, 2009, 2014; Webb et al., 2004, 2005; Dalstra, 2014; and Perring and Crowe, 2017).

In contrast, rather little has been published on the martite-goethite style of mineralization despite there now being close to 50 billion metric tons (Gt) in resource of this ore type in Western Australia (Knight et al., 2018). Deposit descriptions are provided by Paquay and Ness (1998; Hope Downs), Hannan et al. (2005; Chichester Ranges), Hodkiewicz et al. (2005; C deposit, North Flank), Lascelles (2006; Hope Downs), Bodycoat (2007; E deposit, North Flank), Clout and Fitzgerald (2011; Roy Hill), and Knight et al. (2018; South Flank).

The martite-goethite style of mineralization was first described petrographically by Morris (1980). One of the key diagnostic features of this ore type is the way in which the primary texture of the BIF is preserved during the iron enrichment process: original magnetite euhedra are replaced by hematite (this pseudomorphic hematite is termed "martite") and the gangue phases are replaced by goethite (carbonate rhombs are replaced first, followed by Fe silicate phases, and finally chert). Ramanaidou and Morris (2010) pointed out that although this ore-forming process is supergene, there is a clear distinction between the "mimetic" textural preservation typical of martite-goethite ores and the texturally destructive iron enrichment associated with lateritic weathering.

Morris et al. (1980) presented an "electrochemical cell" model of ore genesis, which predicts that the mineralization forms from the bottom upward. Lascelles (2006), in contrast, proposed that the precursor lithologies to martite-goethite ores are a unique "chert-free BIF" facies. Both of these models for martite-goethite ore genesis agree that the ore fluids were essentially of meteoric origin.

Routine use of three-dimensional geologic modeling packages to create resource and mining models allows us to interrogate the distribution of martite-goethite mineralization on a camp scale in a way that would have been impossible only a few years ago. We thus have a unique opportunity to analyze orebodies in terms of mass flux and fluid flow and to compare these results with current ore genetic models for martite-goethite mineralization. At Mining Area C (Fig. 1), the North and South Flank deposits cover a combined 60 km in strike extent, have been drilled out on a 150- × 50- to 50- × 50-m grid and to depths in excess of 300 m in places, and provide an ideal case study. They have a combined premining resource of about 3 Gt of iron ore (Hodkiewicz et al., 2005; Knight et al., 2018).

Fig. 1. Regional geology of the Central Hamersley Province, showing the location of Mining Area C with inset showing the entire Province and the major towns.
Regional Geology

The Hamersley Group comprises a 2.5-km-thick sequence of sedimentary and igneous lithologies, deposited in a passive margin setting on the southern margin of the Pilbara craton between about 2630 and 2445Ma (Trendall and Blockley, 1970; Ewers and Morris, 1981; Trendall et al., 2004; Fig. 1 inset). The group is divided into eight formations (Fig. 2). A brief summary of each formation is given below, ordered in the direction of stratigraphic younging.

1. Marra Mamba Iron Formation: intercalated meter-scale beds of BIF and shale grouped into three members, the central more shale-rich MacLeod Member separating the lower Nammuldi Member from the upper Mount Newman Member (Blockley et al., 1993).

2. Wittenoom Formation: a dolomite-rich sequence divided into the lower BIF-bearing, shale-rich West Angela Member, the central dolomitic Paraburdoo Member and the upper Bee Gorge Member, which comprises intercalated shales, cherts, volcanioclastics, and turbiditic dolomites (Simonson et al., 1993).


4. Mount McRae Shale: pyritic and carbonaceous shales with thin interbedded BIF units toward the top and irregularly distributed carbonate turbidites.

5. Brockman Iron Formation: the lowermost Dales Gorge Member BIF is sequentially overlain by the Whaleback Shale Member, the Joffre Member BIF, and the Yandicoogina Shale Member.

6. Weeli Wolli Formation: dominated by doleritic sills which intrude an alternating sequence of BIF and shale.

7. Woongarra Rhyolite: felsic to intermediate volcanic rocks dominate the upper and lower portions of this unit and the central portion comprises BIF, dolerite, and shale.

8. Boolgeeda Iron Formation: dominated by BIF, intercalated with more shale-rich units, this formation has a distinctive central package comprising chert-rich BIF, green mudstone and siltstone, and diamicite (Warchola et al., 2018).

The Hamersley Group was deposited conformably on the Jeerinah Formation, the uppermost unit of the sequence of flood basalts and intercalated sedimentary rocks, which make up the Fortescue Group, and is conformably overlain by the shallowing-upward sedimentary sequence of the Turee Creek Group (Powell et al., 1999). The Fortescue Group records rifting of the southern margin of the Pilbara craton, followed by deep-water sedimentation on a passive margin (lower Hamersley Group; Morris and Horwitz, 1983; Blake and Barley, 1992; Trendall et al., 2004; Johnson et al., 2013). The southern cratonic margin was converted to an active margin, accompanied by back-arc rifting and bimodal volcanism, from ~2450 Ma (upper Hamersley Group; Blake and Barley, 1992; Barley et al., 1997; Krapez et al., 2017). This was followed by subsidence and the development of a retro-arc foreland basin.
The earliest orogenic event to have affected the Hamersley Province took place at ~2430 to 2400 Ma (Martin et al., 2000; Rasmussen et al., 2005) and was probably associated with inversion of the back-arc magmatic belt. A recent study of the Paulsens orogenic gold deposit describes gold lodes, dated at 2403 ± 5 Ma, crosscutting a folded and faulted gabbroic dike with an age of 2701 ± 11 Ma (Fielding et al., 2017). The larger scale geodynamic context of this event currently remains unknown. Collision of the Glenburgh terrane with the southern margin of the Pilbara craton resulted in closure of the southern ocean and the second phase of orogeny, the Ophthalmian orogeny, between ~2215 and ~2145 Ma (Rasmussen et al., 2005; Johnson et al., 2013). Subsequent collision between the combined “Pilboyne” craton and the Yilgarn craton at ~2050 to 1950 Ma (Johnson, 2013; Johnson et al., 2013) produced the West Australian craton and resulted in the Glenburgh orogeny.

The rocks of the Fortescue, Hamersley, and Turee Creek Groups have undergone burial metamorphism on a regional scale (Smith et al., 1982). The metamorphic isograds trend broadly parallel to the southern and western margins of the Pilbara craton, with grades decreasing from green schist facies (epidote-actinolite) along the cratonic margin to prehnite-pumpellyite facies in the interior. Metamorphic ages decrease from south to north and the metamorphic event was probably driven by the northward advance of the Ophthalmian fold-and-thrust belt (Rasmussen et al., 2005). In the west of the Hamersley Province there is evidence of local metamorphism associated with the intrusion of 2208 Ma dolerite sills, which predate and are folded by the Ophthalmian orogeny (Shibuya et al., 2010).

The Pilbara craton was amalgamated into the larger entity of the West Australian craton in the Palaeoproterozoic and this has behaved as a coherent body since the Mesoproterozoic. However, minor internal geodynamic activity, during the Mesozoic and Cenozoic, has played an important role in the evolution of the landscape as it exists today, the related regolith development, and the formation of iron deposits.

Australia finally separated from Gondwana in the Cretaceous and drifted north. A major regional-scale plate reorganization occurred at about 45 Ma (Hall, 2002), probably driven by the initial collision of India with Asia. Significant long-wave-length uplift of the western part of Australia began at about this time (Czarnota et al., 2014). The greatest amount of this uplift was localized within the Hamersley region, resulting in the formation of a broad, approximately elliptical area of anomalous elevation referred to as the Hamersley Dome. This has approximate dimensions of 400 × 150 km, with its long axis oriented west-northwest; its highest point is 1,248 m above sea level (asl). It is broadly coincident with the currently preserved extent of the Palaeoproterozoic Hamersley Basin. Localized uplift of this magnitude is unusual within stable cratons and the explanatory mechanism is currently not understood. This coincidence of anomalous Cenozoic uplift with the largest outcropping area of BIF in the world is, however, likely to be the critical regional-scale factor responsible for the iron endowment of this province. Significantly, the commencement of the Hamersley Dome uplift is coeval with the oldest dates of martite-goethite ore formation (Ramanaidou et al., 2019).

Paleoclimatic variation during the Cenozoic also provides important context for understanding martite-goethite mineralization. The generally established view of the paleoclimatic history of the West Australian craton is summarized by Anand and Paine (2002) and considers that this period comprised two distinct phases: (1) seasonally humid, subtropical to tropical conditions during the Cretaceous to mid-Miocene; and (2) drier, arid to semiarid climates since the mid-Miocene. The implication of this for supergene iron mineralization is that the initial period of Hamersley Dome uplift probably coincided with the presence of abundant vegetation, capable of generating reduced, acidic surface waters with the capacity to mobilize iron.

Geology of Mining Area C

The North and South Flank deposits are located at Mining Area C in the center of the Hamersley Province of Western Australia (Fig. 1). At the camp scale, the outcrop pattern is dominated by a series of large-scale, open, upright folds with wavelengths on the order of 20 km. These are typically E-W-trending and doubly plunging, forming a series of domes of which the Weeli Wolli anticline at Mining Area C is a typical example (Fig. 3). The cores of domes form low ridges composed of the Marra Mamba Iron Formation and shales of the uppermost Jeerinah Formation. The intervening synclines crop out as ranges of the more resistant Brockman Iron Formation. The Wittenoom Formation appears to have undergone significant karstic erosion and is rarely exposed in outcrop. It forms the subcrop to a series of E-W-trending valleys (the “strike valleys” of Kneeshaw and Morris, 2014) filled with a variety of Mesozoic to Cenozoic sedimentary rocks. A detailed overview of the stratigraphy and structure of Mining Area C is presented by Knight et al. (2018) and only a summary is given here.

Stratigraphy

The outcropping portions of the Weeli Wolli anticline are dominated by the Mount Newman, MacLeod, and Nammuldi Members of the Marra Mamba Iron Formation, and outcrops of the underlying Jeerinah Formation are restricted to the antcinal hinge zones (Fig. 3). The MacLeod and Nammuldi Members contain few iron oxide micro- and mesobands (Kepert, 2001). The fresh, unmineralized Nammuldi Member consists of a lower carbonate-dominated sequence overlain by chert-rich carbonate and BIF, with minor interbedded (tuffaceous) shales. The MacLeod Member is characterized by prominent, podded units and a thick central tuffaceous bed composed largely of stilpnomelane. The podded units are largely composed of chert, partly after silicified carbonate, and chert-rich BIF. The Mount Newman Member is characterized by close-spaced Fe oxide-chert ± minnesotaite and chert-silicate ± carbonate mesobands. There are eight shale macrobands composed dominantly of either stilpnomelane or carbonate. The majority of the mineralization at the North and South Flank deposits is restricted to the Mount Newman Member. This member is further subdivided into three units, formally termed (from bottom to top) N1, N2, and N3: the N2 unit is more shale rich (24% shale macrobands) than the N1 and N3 units (<5% shale macrobands; Kepert, 2001; see Fig. 2).
Some mineralization also occurs within the overlying West Angela Member of the Wittenoom Formation. The West Angela Member varies significantly in composition across the Hamersley Province, from interbedded dolomite and shale-rich dolomite in the west (e.g., Rocklea) to residual, interbedded manganiferous shales and cherts with rare dolomite in the central and eastern parts of the province (Kepert, 2001). At Mining Area C, the manganiferous shale component increases and the dolomitic component decreases in proximity to martite-goethite mineralization. The West Angela Member is informally subdivided into a lower unit, WA1, which consists of intercalated chert, BIF, and shale, and an upper unit, WA2, which consists of ferruginous and manganiferous shales where intersected in proximity to martite-goethite mineralization, but which is composed of dolomite and dolomitic shale away from the ore (Blockley et al., 1993; Kepert, 2001).

The Phanerozoic sedimentary sequence comprises valley-fill deposits of three distinct ages (Kepert, 2001; Kneeshaw and Morris, 2014). The oldest package, referred to in the literature as Cenozoic detrital unit 1 (CzD1) may, in fact, be of Late Jurassic to Early Cretaceous age (Hannaford, 2016; Charles, 2019). It comprises hematite-rich conglomerates and siltstones with intercalated kaolinitic clay-rich lenses. The middle detrital package, and the most voluminous (CzD2), is Oligocene to Miocene in age and includes the channel iron deposits found around the margins of the Hamersley Province (Morris and Ramanaidou, 2007; Kneeshaw and Morris, 2014). In the vicinity of the Weeli Wolli anticline the CzD2 package consists largely of sideritic and lignite-rich clays, with minor pisolitic material present in the Hope Downs area (Lascelles, 2006). The sequence is strongly overprinted by the effects of intense late-Miocene weathering. A well-developed lateritic weathering
Profile is preserved with mottled saprolitic clays overlain by a vuggy, ferruginous material variably cemented by vitreous goethite (Kepert, 2001). Pedogenic calcrite cements the uppermost part of the sequence. The youngest sedimentary package (CzD3) is of Plioecene age and consists of ferruginous siltstone with more gravel-rich material accumulated close to the valley margins. Quaternary colluvial fans of BIF-rich scree are continuing to form today at the mouths of steep gullies incised into the BIF ranges, particularly the Packsaddle Range located on the northern side of the valley from North Flank.

Structure

Mining Area C is dominated by the regional-scale Weeli Wolli anticline. The anticline is doubly plunging and has an “M”-shaped profile in cross section. Second-order, mesoscale folds have sinuous hinge lines (both in the horizontal and in the vertical plane) and are uniformly N-verging. The combined effect of the fold generations results in a complex outcrop pattern that reveals a number of smaller domes superimposed on the broader Weeli Wolli anticline (Fig. 3). The mesoscale folds clearly predate the Weeli Wolli anticline, but both fold generations can plausibly be linked to progressive deformation during the Ophthalmian orogeny (Kepert, 2001).

A series of subhorizontal thrusts have developed locally in response to overtightening of the mesoscale asymmetric folds. The thrusts typically nucleate in the West Angela Member with displacement rapidly decreasing down stratigraphy into the lower Mount Newman and MacLeod Members. Faulting has largely been accommodated by bedding-parallel shearing in shale units and along stratigraphic boundaries, commonly making discrete fault planes difficult to identify within a broader zone of shearing about 20 m in thickness. A shallowing of fault dips from the south to north limb of the Weeli Wolli anticline suggests that thrust faults were formed prior to or synchronous with regional-scale folding. In places, particularly in the North Flank, the thrusts exhibit convex-up morphologies with portions of some of the thrust planes now dipping to the north. Many of these thrusts appear to have considerable strike continuity, at least at the resolution of the current drilling, but the degree of offset can vary considerably along strike, suggesting a series of closely spaced en echelon thrusts and/or a degree of strain partitioning across crosscutting structures.

Two mafic dikes crosscut the Weeli Wolli anticline (Fig. 3). The Waterfall Gully dike strikes northwest-southeast through the Packsaddle Range and its southeastern extension can be traced into the Weeli Wolli anticline between the C and D deposits in the North Flank. This dike potentially belongs to a series of closely spaced en echelon thrusts and/or a degree of strain partitioning across crosscutting structures.

Martite-goethite mineralization

Martite-goethite mineralization is hosted principally by the Mount Newman Member of the Marra Mamba Iron Formations and is developed extensively along the northern (North Flank) and southern (South Flank) limbs of the Weeli Wolli anticline. The North Flank deposits, from east to west, are A, B, C, D, E, and F, and Dead End; the South Flank deposits, from west to east, are Highway, Grand Central and Vista Oriental; the R deposit is located on the northern flank of a subsidiary dome at the southeastern end of the Weeli Wolli anticline (Fig. 3). Note that the deposit boundaries shown in Figure 3 are somewhat arbitrary; they are used for operational purposes and are not an accurate reflection of the distribution of mineralization. Mining at North Flank started at the C deposit in 2003 (Hodkiewicz et al., 2005) and at South Flank, at the Grand Central deposit, in 2018 (Knight et al., 2018).

Both Hodkiewicz et al. (2005) and Bodycoat (2007) noted the significant role played by faults and stratigraphic contacts in controlling the flow of mineralizing fluids at the C and E deposits. One of the aims of this study is to analyze these controls in more detail and across the full strike extent of mineralization at both the North and South Flank deposits.

The age of the martite-goethite mineralization is poorly constrained and is the subject of ongoing research. A recent pilot study conducted for BHP by the Commonwealth Scientific and Industrial Research Organization (CSIRO) indicates that the original martite-goethite ore-forming event took place in the mid-Eocene at ~45 Ma, based on (U-Th)/He dating on coexisting martite and goethite (Ramanaidou et al., 2019). This places it intermediate in age between the CzD1 and CzD2 detrital units.

There has been at least one and possibly multiple episodes of intense lateritic weathering in the Hamersley Province since ~40 Ma (Morris and Ramanaidou, 2007). The martite-goethite mineralization has been overprinted by this weathering, as seen in the development of a ferruginous duricrust or “hardcap” on the mineralization (Clout, 2003, 2006). The hardcap extends up to 60 m below surface and consists of highly porous or coarse cellular-textured brown or vitreous goethite. Vitreous goethite is characteristic of this zone. A hydrated zone occurs immediately below the hardcap and here most micropores and cavities are lined with colloform secondary goethite and hydrohematite. The ore is denser with lower porosity and higher goethite content compared with primary ore.

Method

This study used the GOCAD™ three-dimensional modeling software package to integrate, query, and analyze a large body of two- and three-dimensional data. Two-dimensional datasets include various geophysical images (principally airborne magnetic and Falcon™ gravity data), 1:20,000 BHP mapping data and orthophotographs. Three-dimensional datasets comprise exploration drilling data (stratigraphy, major element chemistry, and semi-quantitative mineralogy derived from hyperspectral logs, collected on 3-m sample intervals using reverse-circulation drilling), resource block models, topographic surface (digital elevation model, DEM), and premining water table. Additional low-resolution regional-scale surfaces were generated specifically for this study, as detailed below.

Drill spacing is 50 m × 50 m over North Flank and the central part of South Flank, and 150 m × 50 m elsewhere. Drilling took place over a period of about 50 years (see Knight et al., 2018, for a summary of the exploration history) but less than...
6% of the 17,527 holes incorporated in the model were drilled prior to 2000. It is company policy to pulverize and analyze 100% of drill sample material making it impossible to collect additional data in areas of interest/uncertainty.

**Base of CzD1, CzD2, and CzD3:** The starting surface was the low-resolution DEM surface for the Weeli Wolli anticline area on which a region had been created that corresponded to the extent of mapped detrital material. Point sets were generated from the drill logs to reflect the drilled base of CzD1, CzD2, and CzD3. To create a CzD1 base surface, the low-resolution DEM surface was warped to the CzD1 base control points in the areas of detrital valley-fill only. The distribution of drilled control points is extremely irregular, being concentrated close to the North and South Flanks of the bedrock dome formed by the Weeli Wolli anticline. The shape of the CzD1 base surface is therefore poorly constrained, particularly in the western and eastern portions of the model volume. This process was repeated to create preliminary surfaces for the CzD2 and CzD3 base surfaces. Where there were no control points and the preliminary CzD1 base surface trended below the CzD2 or CzD3 base surface, it was warped upward using the corresponding CzD2 or CzD3 base point. A similar process was applied to the CzD2 base surface where it trended below the CzD3 base surface. No allowance has been made for the ongoing erosion of the bedrock anticline, although this has clearly been contributing to the sedimentary sequence filling the valleys that surround the Weeli Wolli anticline. The surfaces thus generated are very approximate models of the shape of the paleotopography, at the start of CzD1, CzD2, and CzD3 sedimentation, respectively, and exclude the effect of any karst formation.

**Top of N2:** Since ~70 to 85% of high-grade martite-goethite ore developed on the flanks of the Weeli Wolli anticline is hosted by the N2 and N3 units of the Mount Newman Member of the Marra Mamba Iron Formation, the N2 top surface is particularly well-constrained by exploration drilling. Mapped stratigraphic contacts across the Weeli Wolli anticline area have been used to generate a series of pseudocontours, assuming an average thickness for each of the stratigraphic units, and these are used to control the shape of the N2 top surface in the absence of hard data, such as mapping or drill control points.

**Relationship Between Paleodrainages, Bedrock Topology, and Mineralization**

**Paleodrainage patterns**
The approximate location of the main CzD1, CzD2, and CzD3 drainages have been digitized onto each surface using the Z contours on each surface as a guide, in effect marking out the valley positions in each paleotopographic surface. The CzD1 drainage pattern may be biased due to the preservation potential of CzD1 material being highest in areas with the steepest dip on the Marra Mamba Iron Formation/Wittenoom Formation contact. Nonetheless, this is thought to be the most likely location of the main drainage channels at the time. The three modeled paleodrainage systems are illustrated in Figure 4A-C.

A striking feature of the paleodrainage patterns is the consonance between them and the strong control exerted by bedrock stratigraphy and structure. The major drainages trend...
east-west and follow subcrop of the dolomitic Wittenoom Formation on the flanks of the Weeli Wolli anticline (Fig. 3). Other significant drainages are localized by mesoscale bedrock synclines, such as the Syncline Valley at Grand Central and similar tributary valleys at the B and Vista Oriental deposits (Fig. 4D). At the regional scale the paleodrainages overlap to a large extent and reveal a subtle drainage divide at the western end of the Weeli Wolli anticline. The position of this divide lies within a corridor of NNE-trending lineaments, visible in the Bouguer gravity data, the FalconTM gravity data (gD) and, to a lesser extent, the aeromagnetic data. This NNE-trending corridor marks the southern continuation of the Lalla Rookh-West Shaw tectonic zone, which is exposed in the Pilbara craton to the north (Van Kranendonk et al., 2001).

An unusual and characteristic feature of all the drainages is the tendency for them to link a series of deeper, doubly plunging depressions. These have been termed depocenters as they correspond with the thickest drilled intercepts of the respective detrital units. Note that stratigraphic intervals have been carefully filtered to exclude any that terminate either in the drill collar or the end of hole (EOH) and thus the intervals closely reflect true stratigraphic thicknesses (however, no allowances have been made for angled holes, which are relatively uncommon, or dipping stratigraphy). There is a strong apparent bedrock control on the depth of the CzD1 depocenters in that the thickest CzD1 intercepts occur adjacent to the flanks of the Weeli Wolli anticline in areas where the bedrock stratigraphy dips most steeply. The dip of CzD1 beds measured in outcrop within the Syncline Valley is ~45°, much greater than the angle of repose for detrital material. Furthermore, the top of the calcrete horizon that is widely developed at the top of CzD2, and which is regarded as either pedogenic or lacustrine in origin (Kepert, 2001), is also depressed above these depocenters.

**Bedrock topology**

Topology is used here to mean the way in which different parts of a surface are connected to one another in three-dimension through the shape of that surface. The topology of bedding contacts exerts a major control on the flow of supergene fluids, which are driven by gravity along bedding planes. The N2 top surface typically occurs within or at the base of martite-goethite mineralization and is therefore well controlled by drilling, at least within the deposits. It is taken as illustrative of the topology of the Mount Newman Member and the following analysis relates to this surface.

The topology of this surface reflects the net result of Proterozoic deformation (Fig. 5). The main feature of the surface at the 10-km scale is the E-W-trending, doubly plunging Weeli Wolli anticline with its distinctive M-shaped cross-sectional profile. Older, mesoscale folds with wavelengths on the order of 100 to 500 m trend from west-northwest, through east-west, to east-northeast but are curvilinear in a vertical plane, plunging both east and west and being characterized by a number of doubly plunging depressions. From the point

![Topology diagram](http://pubs.geoscienceworld.org/segweb/economicgeology/article-pdf/115/3/627/5015616/4734_perring_et_al.pdf)

**Fig. 5.** Topology of the N2 top surface (shown as contours), illustrating the control exerted by bedrock stratigraphy on groundwater flow. The surface can be divided up into five catchments separated by watersheds. The resulting ore domains are labeled: North Flank, Highway Valley, R Deposit Valley, Syncline Valley, and South Flank. Mesoscale, synclinal keels (colored by RL) highlight significant changes in fold plunges and reveal the common occurrence of doubly plunging fold axes.
of view of understanding bedding-parallel fluid flow, this surface can be divided into catchments separated by watersheds, which feed the main drainages as shown in Figure 5.

Distribution of martite-goethite mineralization

The shape of the N2 top surface defines five catchments or ore domains: North Flank, Highway Valley, R Deposit Valley, Syncline Valley, and South Flank (Fig. 5). The stratigraphic distribution, quality and depth of high-grade (>57 wt % Fe) mineralization in each ore domain is compared in Table 1. The North Flank and Syncline Valley ore domains contain significantly larger volumes of ore that is, on average, higher grade and that extends to greater depths than that in the other three domains.

The tight synclinal keels in Figure 5 belong to the N-verging, asymmetric, mesoscale, Ophthalmian-age folds. The steep limbs of these folds are typically thickened by thrusting on shallowly S-dipping, en echelon faults. Much of the high-grade mineralization developed around the Weeli Wolli anticline is associated with these structural positions. There is also an apparent association between areas of steeper bedrock dip and economic mineralization (Fig. 6). At the scale of entire deposits (kilometer-scale), average bedding dip (i.e., dip of the regional N2 top surface) appears to determine the presence/absence of high-grade mineralization. Where bedding dips are <10° and a mesoscale synclinal keel is absent, prominent barren zones occur within the Mount Newman Member stratigraphy on the flanks of both the main Weeli Wolli anticline and the subsidiary A and B deposit domes.

Controls on Deposit Morphology

Bedding topology

The most obvious way in which bedding topology controls ore deposit morphology occurs as a result of a primary lithologic control on martite-goethite mineralization: high-grade mineralization is largely confined to the Mount Newman Member (~80–90%), and within the Mount Newman Member, the N2 and N3 units are preferentially mineralized over the N1 unit (~30–45% for N2 and N3 vs. ~10–15% for N1; percentages calculated on the basis of the number of samples reporting >57 wt % Fe in each different stratigraphic unit, see Table 1).

In addition to the regional-scale role that bedrock dip appears to have on orebody development (Fig. 6), there are numerous areas within individual deposits where bedding dip appears to control orebody morphology at the scale of 100s of meters. Toward the eastern end of the E deposit there are two pockets of extremely deep mineralization (~250 m deep) just to the east of the NE-trending dolerite dike (Fig. 7; see Fig. 6 for location). Both of these are associated with the steepest portions of the steeply dipping southern limb of a mesoscale syncline (these folds typically have wavelengths of 20–200 m and wavelengths of 200 m–5 km: Kepert 2001). Figure 8 illustrates the central portion of the Grand Central deposit and again the importance of bedding dip is evident. Two broad orebody morphologies are present: synclinal in the north, associated with the Syncline Valley, and tabular in the southern flank of the Weeli Wolli anticline. The tabular body of mineralization thins very rapidly downdip to the south as the dip of the Mount Newman Member shallows from ~45° near surface to <10° at depth. In contrast, the northern limb of the Syncline Valley dips consistently at ~30° over ~275 m from surface and the entire N2/N3/WA1 package is consistently mineralized. At this depth there is a small, subsidiary anticline (amplitude ~25 m) within the overall mesoscale syncline. The mineralization becomes discontinuous and terminates to the south over a distance of ~75 m in response to the dip reversal that accompanies this small anticline.

A further control exerted by bedding is evident in Figures 8 and 9. Where mineralization peters out downdip, it typically does so in a predictable manner, favoring the N3/WA1 contact in particular, but also associated with the N2/N3 and WA1/WA2 contacts. This leads to bifurcating ore terminations as shown in Figures 8 and 9. Note that these contacts are all marked by the transition from BIF to more shale-rich material.

Synclines

Some of the thickest ore intercepts are associated with mesoscale, synclinal keels, as can be seen by comparing the po-

| Table 1. Distribution of High-Grade Mineralization (samples reporting >57% Fe) between the Five Camp-Scale Domains (see Fig. 5 for geographic location of domains) in Terms of Host Stratigraphic Unit, Fe Grade, and Depth of Mineralized Material Below Topographic Surface |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mineralized domain | Style | Count | MM (%) | N1 (%) | N2 (%) | N3 (%) | MN (%) | WA1 (%) | WA2 (%) | Fe (%) | Depth (m) |
| | | | Median | IQR | Median | IQR | Median | IQR | Median | IQR | Median | IQR |
| North Flank | Synclinal | 582,396 | 2.6 | 13.0 | 29.3 | 38.9 | 81.2 | 11.4 | 3.1 | 63.3 | 57.0–64.8 | 61 | 32–99 |
| Syncline Valley | Synclinal | 24,554 | 8.3 | 17.2 | 29.7 | 34.8 | 81.7 | 9.1 | 0.8 | 62.3 | 59.9–64.3 | 51 | 28–83 |
| South Flank | Planar and synclinal | 20,506 | 2.2 | 15.4 | 35.1 | 35.7 | 86.2 | 10.8 | 0.9 | 62.0 | 59.8–63.8 | 35 | 17–64 |
| Highway Valley | Planar and synclinal | 5,767 | 1.1 | 9.2 | 44.6 | 39.9 | 93.8 | 4.8 | 0.3 | 62.2 | 60.0–63.8 | 29 | 17–47 |
| R Deposit Valley | Planar and synclinal | 6,362 | 1.6 | 7.6 | 33.8 | 45.2 | 86.6 | 10.9 | 0.8 | 62.2 | 59.9–64.1 | 27 | 12–47 |

Abbreviations: MM = MacLeod Member, Marra Mamba Iron Formation; N1 = lowest unit of the Mount Newman Member, Marra Mamba Iron Formation; N2 = central unit of the Mount Newman Member, Marra Mamba Iron Formation; N3 = uppermost unit of the Mount Newman Member, Marra Mamba Iron Formation; MN = Mount Newman Member, Marra Mamba Iron Formation; WA1 = lower unit of the West Angela Member, Wittenoom Formation; WA2 = upper unit of the West Angela Member, Wittenoom Formation; IQR = interquartile range
sition of these keels with the Fe meters map (Fe meters = grade*intercept for sample intervals where Fe >50 wt% over a downhole width of >6 m with less than 3-m dilution by internal zones of <50 wt% Fe: a metals meters parameter is commonly used in the mining industry as a measure of the quality of an ore-grade drill intercept, see Fig. 10). In part this reflects the thrust-thickening of the steep, N-facing limbs of these synclines (see Figs. 7–9), which results in anomalously thick packages of Mount Newman Member rocks. As a direct result of the folding and thrusting, these synclinal positions and their steep southern limbs are likely to be more fractured and thus more permeable. Indeed, small-scale shears are a very common feature in folded areas, and dissolution-precipitation creep, occurring at the time of the Ophthalmian orogeny (Eggsleder et al., 2016), has almost certainly increased the permeability. Where synclines are oriented subparallel to the regional hydrologic gradient and to the regional bedrock strike (e.g., at the C and E deposits in North Flank and Grand Central in South Flank) the keels appear to act as gutters and can focus the ore fluids over plunge extents in excess of a kilometer. An example is illustrated in Figure 8: the entire western end of the syncline at Syncline Valley is mineralized from surface over a plunge extent of 3 km (Fig. 8A), but farther to the east the mineralization extends only part way down the northern limb, and the lower part of the southern limb and keel are shielded from the ore fluid by the presence of shallowly dipping thrusts (Fig. 8B).

Although the mineralization is broadly stratabound, the base of the mineralization is commonly strongly discordant in the center of these synclines. This discordance is likely to be due to enhanced permeability in zones where more intense fold cleavage has developed (Figs. 7–9). A further factor that will influence orebody geometry, particularly the base to mineralization in synclinal keels, is that at some point the dominant fluid flux needs to become lateral rather than downward (the regional hydrologic gradient drives groundwater flow from west to east)—we would not predict mineralization below that interface, which is likely to be discordant.

**Thrusts**

One or more thrusts are typically developed on the steep, N-facing limbs of mesoscale synclines at both North and South Flank (Figs. 7–9). These thrusts are Proterozoic in age and formed in association with the regional folds, but they do show some evidence of later modification. They typically have a rather unusual convex-up morphology and the apparent offset across them increases from the upper MacLeod/lower Mount Newman Members, where it is typically <75 m, to the upper Mount Newman/lower West Angela Members, where apparent offsets of a couple of hundred meters can be measured. The thrusts were probably initiated at the Mount Newman Member/West Angela Member contact: prior to dissolution of carbonate in the West Angela Member this contact would have been rheologically weaker than the MacLeod Member/Mount Newman Member contact. The initial fault plane was likely oriented at a high angle to bedding, as is typically observed where these thrusts cut the upper MacLeod/lower Mount Newman Members, but become progressively more concordant within higher stratigraphic units, typically becoming essentially parallel to bedding within the WA2 unit, a feature that
will have been accentuated in the West Angela Member by the carbonate dissolution that accompanied mineralization.

On the one hand, these structures can be associated with anomalously thick portions of the orebody, primarily due to the structural thickening they impose on the BIF protolith. On the other hand, there are numerous instances where these thrusts appear to act as aquitards, particularly where they are oriented subparallel to bedding and where Mount Newman Member BIF is thrust over WA2 shale (Figs. 7–9). Faults are most likely to become aquitards when they form fine-grained clay-rich gouge; naturally, this is more likely when they cut through phyllosilicate-rich host rocks, such as shale, than more siliceous rocks, such as BIF. The thrusts may act to compartmentalize the flow of supergene fluids by creating an extreme permeability gradient. In the Syncline Valley at Grand Central the thrusting has effectively shielded the lower portion of the steep limb of the syncline from the effects of vertically driven supergene fluid flow over the central ~4-km strike length of the valley (the eastern half of that portion of the Syncline Valley shown in Fig. 8A). It should be noted that in the far western and eastern portions of the Syncline Valley, where the synclinal keel is located at shallower depths (within ~150 m of surface), the entire syncline is mineralized, including beneath thrusts in the western portion (see Fig. 8A).

Figure 8A also illustrates a pocket of very deep mineralization (>250 m) associated with the syncline keel where mineralization extends down the entire length of the northern limb into the keel. The northern limb does not dip more steeply here relative to the portions immediately west or east but the keel itself does change position markedly (RL increases by ~100 m and northing, by ~175 m) immediately to the east. It is possible that this jog marks the position of a subvertical fault that has not been intercepted in drilling. Subtle features in the magnetic image, the Fe meters map and the shape of the N2 top surface indicate a possible NE-trending fault. If present, this could explain the development of the deep pocket of mineralization immediately to the west, through the fault (1) acting as a recharge zone, delivering surface-derived supergene fluids directly to the deep keel position, and (2) focusing lateral groundwater flow into the keel position (i.e., fluids moving along the groundwater gradient interact with and are redirected along the fault plane and into the keel).

**Dikes**

Two major dikes crosscut the Weeli Wolli anticline: a NE-trending dike in the west and a NW-trending dike in the east. Mineralization is particularly well-developed on the eastern side of both dikes, at the E and C deposits, and at the western end of the B deposit (Figs. 6, 7), reflecting offset of the synclinal keel in an east-side-down manner across the dike-hosting fractures. These dikes have a striking impact on the premining water table, with drops of ~30 m occurring from west to east across both dikes in areas of good drill control (Fig. 11). While the dikes are clearly acting as aquitards, impeding...
Fig. 8. Grand Central showing geologic controls on ore deposit morphology and inferred supergene fluid-flow pathways. A. Plan view of the high-grade ore shell (>57 wt % Fe) at Syncline Valley (location of the plan view is shown in Fig. 6). The high-grade ore shell is colored by depth from surface and the overthrust southern portion of the ore shell has been removed to reveal what is happening at depth. B. Cross section through the entire deposit at 695855mE. The cross-sectional view shows all mineralization (i.e., Fe >48 wt %); coordinates are MGA zone 50.

Fig. 9. Cross section through the southeastern end of the B deposit at 706910mE, showing geologic controls on ore deposit morphology. Location shown in Figure 6; coordinates are MGA zone 50.
the regional flow of groundwater from west to east, they are steeply dipping and their contacts could thus act as effective recharge zones. This appears to have been the case at the E deposit (Fig. 7) where a particularly deep (>250 m) and localized pocket of mineralization is developed within a synclinal keel position immediately to the east of the NE-trending dike.

**Topography**

A supergene ore system, by definition, is tied intimately to the topographic surface. Table 2 and Figure 12 illustrate the way in which the North and South Flank martite-goethite mineralization (taken as all samples containing >48 wt % Fe) is distributed across different stratigraphic units and how this distribution changes with depth slice. In all cases the N2 and N3 units are the most favorable host rocks. The distribution is most symmetrical for near-surface samples (0–100 m vertical depth from the current topographic surface) and becomes increasingly skewed toward the N3 unit with depth.

Most of the North and South Flank orebodies are now drilled to 150 m E x 50 m N and many are drilled to 50 m E x 50 m N. It is therefore possible to make with confidence some empirical observations about the depth extents of the martite-goethite ore systems. For high-grade mineralization (samples with >57 wt % Fe): maximum vertical depth is ~290 m; maximum down-dip extent is ~650 m, and for mesoscale synclines the entire current topographic surface and ~3 km down the plunge of the keel (measured from surface); beyond this the limbs may be partially mineralized, but the keel is commonly barren. Low-grade mineralization (samples with >48 wt % Fe) may extend to ~300 m vertical depth and ~820 m down-dip. The most extensive down-dip mineralization occurs in the tabular body of ore that occupies the southern flank of the Weeli Wolli anticline in the southeast of Vista Oriental. The orebody thins with depth and thin, low-grade mineralization continues beyond the economic extents of the deposit, marking the N3/WA1 contact.

Despite the considerable depth extent of martite-goethite mineralization, the vast majority of ore is located within 200 m of the current land surface. For North and South Flank bedrock samples returning >48 wt % Fe, as a proportion of all bedrock samples and normalized by the number of samples drilled in each depth bin, the following depth distribution is observed: 0 to 100 m = 43%, 100 to 200 m = 38%, and 200 to 300 m = 18%. Furthermore, for the high-grade (>57 wt % Fe) ore shell the 90th percentile value for the depth below current topographic surface is 116 m; less than 5% of the surface vertices (i.e., the nodes on the triangulated surface) lie at depths of greater than 150 m and less than 1% of the vertices lie at depths below 200 m.

There is also a tendency for mineralization to become discordant to bedding where it is shallow (<50 m deep) and predominantly restricted to the hard-capped zone (i.e., the zone of ferruginous duricrust associated with younger lateritic weathering). In Figure 9, for example, the relatively constant thickness (~25 m) of mineralization developed within the Mount Newman Member at the northern, outcropping end of the section also has a base that is distinctly discordant to
As elaborated below, this probably reflects later, overprinting lateritization.

Controls on Ore Composition

In the following the drill dataset is divided into barren (<48 wt % Fe) versus mineralized (>48 wt % Fe) categories. The mineralized category is further subdivided based on the depth from surface, into shallow (<100-m vertical depth) versus deep (>100-m vertical depth). The Fe histogram for all samples from the Marra Mamba Iron Formation and the West Angela Member of the Wittenoom Formation is strongly bimodal with a pronounced minimum at ~48 wt % Fe. The boundary between mineralized and barren BIF is thus taken to be 48 wt % Fe. Samples containing vitreous goethite and S >0.02 wt % are common within 100 m of surface. Both of these constituents are characteristic of the lateritic weathering zone (Ramanaidou and Morris, 2010; Banerjee et al., 2017) and thus a depth of 100 m from surface is used to separate potentially weathered from predominantly fresh martite-goethite mineralization.

Primary bedrock stratigraphy

As noted by Knight et al. (2018), the composition of the bedrock protolith exerts a strong control on the martite-goethite ore composition (Figs. 13–14). Despite the median absolute Fe content more than doubling as a result of the mineralizing process, the ranking of the stratigraphic units by median Fe content remains unchanged from barren BIF through shallow to deep martite-goethite ore (N3 > N2 > N1 > WA1 > MM > WA2: compare Fig. 13A to B and C). The other signifi-

Table 2. Stratigraphic Distribution of Mineralization at North and South Flank Across Three Different Depth Slices, Estimated Using the Percentage of Each Stratigraphic Unit Represented in the Exploration Drill Hole Sample

<table>
<thead>
<tr>
<th>Depth bin (m)</th>
<th>MM (%)</th>
<th>N1 (%)</th>
<th>N2 (%)</th>
<th>N3 (%)</th>
<th>WA1 (%)</th>
<th>WA2 (%)</th>
<th>Total sample count</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–100</td>
<td>7</td>
<td>14</td>
<td>31</td>
<td>31</td>
<td>13</td>
<td>4</td>
<td>151,215</td>
</tr>
<tr>
<td>100–200</td>
<td>1</td>
<td>8</td>
<td>28</td>
<td>37</td>
<td>19</td>
<td>7</td>
<td>28,392</td>
</tr>
<tr>
<td>200–300</td>
<td>0</td>
<td>9</td>
<td>32</td>
<td>40</td>
<td>14</td>
<td>2</td>
<td>653</td>
</tr>
</tbody>
</table>

Abbreviations: MM = MacLeod Member, Marra Mamba Iron Formation; N1 = lowermost unit of the Mount Newman Member, Marra Mamba Iron Formation; N2 = central unit of the Mount Newman Member, Marra Mamba Iron Formation; N3 = uppermost unit of the Mount Newman Member, Marra Mamba Iron Formation; WA1 = lower unit of the West Angela Member, Wittenoom Formation; WA2 = upper unit of the West Angela Member, Wittenoom Formation

Fig. 11. Modeled topology of the premining water table (2-m contours, colored by RL in m): note that the regional hydrologic gradient drops from west to east. The steep gradients that are broadly coincident with dolerite dikes imply that the dikes are acting as aquitards and damming the groundwater flow. Note: the modeled shape of the water-table surface relies increasingly on the interpolant function with increasing distance from a drill pierce-point and thus becomes increasingly uncertain. The areas discussed in the text are highlighted in pink.
cant compositional change that accompanies martite-goethite ore formation is the near-quantitative stripping of SiO$_2$ in all mineralized units except WA2 the median silica content is reduced by an order of magnitude (compare Fig. 14A to B and C). Mineralization is accompanied by hydration of all units except WA2 (compare Fig. 13D to E and F), and by enrichment in Al$_2$O$_3$ (compare Fig. 14D to E and F). The WA2 unit is carbonate-bearing where unmineralized and loss of this primary carbonate appears to accompany mineralization (Blockley et al., 1993; Haukohl, 2014, unpub. report), explaining the apparently anomalous loss-on-ignition (LOI) pattern. Phosphorus is also enriched in the ores over the background values present in unmineralized BIF, approximately doubling in content from ~0.02 to 0.05 wt % in barren BIF to ~0.05 to 0.08 wt % in mineralized BIF.

Variation with depth below current topographic surface

In addition to bedrock stratigraphy, depth below surface also plays a significant role in controlling ore composition. As can be seen in Figures 13 and 14, shallow martite-goethite mineralization (defined as being within 100 m of surface) has consistently lower Fe contents and higher LOI, SiO$_2$ and Al$_2$O$_3$ contents, irrespective of stratigraphic unit.

The depth-related compositional trends are investigated in more detail in Figure 15, which plots the changes in the composition of Mount Newman Member ore (i.e., combined N1, N2, and N3) with depth in 50-m bin increments. Loss-on-ignition, S percentage, and goethite/(goethite + hematite) ratio decrease steadily with increasing depth bin. Similar trends are seen for Al$_2$O$_3$ contents and estimated lump percentage (“lump” being defined as that portion of the mined material exceeding 6.3 mm in diameter) but there is essentially no further decrease in either factor below 150-m depth. Iron content increases markedly and the interquartile range (IQR, i.e., the difference between the 25th and 75th percentile values) decreases between the 0- to 50- and 50- to 100-m-depth bins. Iron content then remains relatively constant with depth although the IQR increases with depth for samples taken from >150-m depth. Silica content is highest in the 0- to 50-m bin, drops sharply in the 50- to 100-m bin, and then increases gradually with bin depth. The IQR for silica is moderate in the 0- to 50-m bin, drops sharply for the 50- to 100-m bin, and then increases with depth. Estimated lump percentage is significantly higher for the shallowest samples (those in the hard-capped zone of 0–50 m) but remains fairly constant with depth below 100 m. There is a gradual increase in density with depth for the deepest two bins (150–200 and 200–250 m). Median S contents decrease sharply with depth over the top 100 m and then remain constant with depth below 100-m vertical depth.

The WA2 unit appears to have a significantly different composition regionally, where it is composed of up to about 100 m of dolomite and shaly dolomite (WRL1 type section: Blockley et al., 1993), compared to the thinner, often manganiferous, shaly material found adjacent to martite-goethite mineralization (median thickness of 49 m on the flanks of the Weeli Wolli anticline). This manganiferous, shaly material appears to be residual after the leaching of carbonate during martite-goethite mineralization. Mineralized (>57 wt % Fe), manganiferous and carbonate-bearing WA2 samples form discrete compositional/mineralogical zones at increasing depth and distance from outcrop at the structurally simple, tabular-style A deposit. The most carbonate rich WA2 samples are found at greatest depth (i.e., greatest downdip distance from the current topographic surface), typically at vertical depths in excess of 100 m, and indicate the termination of the supergene, Fe-mineralizing system. The downdip transition from mineralized WA2 to manganiferous WA2 is illustrated in the cross-sectional view presented in Figure 16.

Control exerted by the premining water table

The premining water table is likely to be lower than that at the time of mineralization as a result of ongoing uplift and erosion of the Hamersley Dome during the Cenozoic, compounded by the development of an increasingly arid climate from the Miocene onward. The position of the water table also exerts
an important control on development of the weathering overprint on primary martite-goethite mineralization.

The modeled premining water table is located at between 60 and 100 m in the vicinity of the North and South Flank deposits. Several compositional variables of the North and South Flank ores (all of which are vitally important to iron ore quality control) appear to be controlled by the water table (either the modern premining one or possibly a slightly older and more elevated water table stand). These include:

1. **P content**: Samples containing >0.1 wt % P cluster just above the modeled position of the premining water table (Fig. 17A), indeed for all mineralized (>48 wt % Fe) bedrock samples containing >0.1 wt % P, the interquartile range for distance to the water table is 2 to 66 m above the water table.

2. **S content**: Typically >0.02 wt % above the premining water table (Fig. 17B) and correlating with K_2O for S contents >0.2 wt %, suggesting the presence of alunite.

3. **Vitreous goethite**: Samples containing >5% are located almost exclusively above the premining water table (Fig. 17C).

4. **Estimated lump percentage**: Samples located above the premining water table typically have a predicted lump content of >55% (and up to ~60%), whereas those below the water table have a predicted lump content typically <40%.

5. **Reverse-splitting behavior**: The geometallurgical algorithm predicts the distribution of various major elements between the lump (>6.3 mm) and fines (<6.3 mm) fractions: samples which show reverse splitting behavior (fine Fe content greater than lump Fe content) tend to be siliceous and to occur below the premining water table.

**Ore types**

Hyperspectral data indicate that the mineralogical make-up of martite-goethite ore is simple and dominated by hematite and goethite with lesser kaolinite and quartz, which is in agreement with the historic petrographic work of Morris (1980, 1985). In geochemical space, most samples plot either close to the hematite-goethite control line or within the quartz-goethite-hematite or kaolinite-goethite-hematite fields. Using the S content as a proxy for the base of weathering (cf. Banerjee et al., 2017) and simple geochemical clips to separate the Fe(hydr)oxide-rich samples from those with appreciable quartz or kaolinite, five categories of mineralization have been developed: martite-goethite high-grade, martite-goethite aluminous, martite-goethite siliceous, lateritic high-grade, and lateritic low-grade. The stratigraphic, mineralogical, geochemical, and estimated geometallurgical composition of each of the categories is summarized in Table 3.

High-grade martite-goethite mineralization is developed almost exclusively after Mount Newman Member rocks (i.e., after BIF) and the near-surface portions of this mineralization are weathered to lateritic high-grade mineralization. The West Angela Member mineralization falls into the aluminous category. Siliceous mineralization is typically preserved along the lower margins and at the downdip terminations of the body of mineralization. Low-grade lateritic mineralization can
BIF-HOSTED MARTITE-GOETHITE MINERALIZATION, HAMERSLEY PROVINCE, WA

Analysis of the difference in stratigraphic thickness between mineralized and unmineralized units around the Weeli Wolli Dome suggests that martite-goethite mineralization was accompanied by volume loss (mineralized intercepts are typically 20–25% thinner than their unmineralized counterparts). Gresens (1967) described a method that can be used to estimate the relative gains and losses of different elements during mineralization and that method has been applied here in the form of the graphic isocon solution proposed by Grant (1986). The average compositions of different stratigraphic units, both mineralized (Fe >48 wt %) and unmineralized (Fe <48 wt %), are presented in Table 4 (all samples taken from >100-m vertical depth to avoid possible weathering overprint). Isocon plots for the principal mineralized stratigraphic units, Mount Newman Member N1 to N3 and West Angela Member WA1, are presented in Figure 18. The plots are similar: P and TiO₂ are taken to be immobile and are used to construct the isocon (note that, in contrast to the immobile behavior of P in primary martite-goethite mineralization, there is good evidence for P being somewhat mobile in the weathering zone: see Fig. 17A). This leads to estimates of the volume loss on mineralization of between 35% (WA1) and 43% (N3), which fits well with the observed degree of stratigraphic thinning. Iron, Al, water (LOI) and Mn (except for WA1) are all added during mineralization while Si, K, Na, Ca, and Mg are lost. This is in keeping with the petrographic observation that the gangue phases (carbonates, silicates, and chert) are replaced by goethite.

**Fluid Flow Model**

The mimetic replacement of gangue phases (carbonate, silicate, and chert) by goethite is a defining feature of martite-goethite ores and separates their genesis from the normal lateritic weathering process, which is texturally destructive and in which Fe is concentrated by residual enrichment (Ramanaidou and Morris, 2010). It is therefore important to understand the composition of the starting material. The mineralogy of fresh Marra Mamba iron formation has been described by Trendall and Blockley (1970), Ewers and Morris (1981), Klein and Gole (1981), and Morris (1991). The BIF units consist almost exclusively of two mesoband types: chert and chert-iron oxide. Minnesotaite occurs as sprays and sheaves and is closely associated with chert. Ferroan talc is also common and has a more bladed habit compared with the acicular habit of minnesotaite. Minor constituents include riebeckite, chlorite, greenalite, apatite, and sulfide. The statistical analysis of Morris (1991) reveals a very strong association between magnetite and quartz. Riebeckite and talc are moderately associated with quartzite-magnetite layers, and siderite and minnesotaite are weakly associated.

The S bands typically contain carbonate (calcite and members of the dolomite-ankerite series) and stilpnomelane with minor feldspar and mica. Carbonates occur as thin continuous...
laminae between bands of magnetite or Fe silicate or are present as disseminated euhedra in silicate-magnetite-carbonate assemblages. Stilpnomelane is the most common Fe silicate in shale bands of the Mount Newman Member and it is typically very fine grained. It occurs as massive bands and as thin laminae intercalated with magnetite- or carbonate-rich laminae. Petrographic analysis indicates that typically more than 75% of the original gangue component is replaced by goethite.

Fig. 15. Changes in various chemical and physical properties of Mount Newman Member mineralization (>48 wt % Fe) with depth. A. Fe wt %, B. LOI wt %, C. SiO₂ wt %, D. Al₂O₃ wt %, E. S wt %. F. Goethite/(goethite + hematite) derived from hyperspectral data. G. Density g/cc. H. Lump % (generic geometallurgical estimation). Gt = goethite. Ht = hematite.
during martite-goethite mineralization. The common gangue phases contain between zero (quartz, calcite, dolomite) and approximately 30 wt % Fe (stilpnomelane, minnesotaite) compared to 57 to 63 wt % Fe for goethite from martite-goethite ores (Manuel and Clout, 2017). It therefore seems probable that, unlike the hypogene martite-microplaty hematite ores (Taylor et al., 2001; Egglseder et al., 2018), significant amounts of Fe must have been added during martite-goethite ore formation, in addition to efficient SiO₂ removal. These assumptions of Fe addition and SiO₂ stripping are borne out by the Gresens analysis described above.

A fluid and mass flux model is proposed for the martite-goethite ore system (summarized in Fig. 19) which envisages a three-stage process: (1) leaching of Fe from BIF in the vadose zone by reduced, acidic, meteoric-derived fluids; (2) penetration of a Fe-rich supergene fluid plume, driven by gravity and focused by bedding-parallel permeability (particularly BIF/shale contacts) into the body of ambient alkaline groundwater, effecting nonredox, mimetic replacement of magnetite by hematite and of the gangue minerals (carbonate, silicate, and chert) by goethite coupled with the release of silica into the fluid phase; and (3) a change from silica leaching to possible silica deposition on the updip margins of the system before the ore-fluid plume is eventually diluted and becomes indistinguishable from the surrounding body of groundwater (much of the silica is probably lost to the hydrosphere).

The proposition that the ore fluids were acidic is based on the following observations: (1) carbonate appears to have been removed from the West Angela Member where the underlying Mount Newman Member is mineralized, (2) the first gangue phase to be replaced by goethite during martite-goethite mineralization is carbonate, and (3) martite-goethite mineralization is accompanied by alteration of the remaining aluminosilicate fraction to kaolinite. A reduced ore-fluid composition is deduced based on: (1) the fact that Fe³⁺ is highly insoluble and mass transport of Fe is likely to have occurred as the Fe²⁺ ion, and (2) sulfides are rarely observed in the ore (e.g., chalcopyrite in equilibrium with kenomagnetite in a martite-goethite ore sample from Marandoo; Morris, 1985). The climate of the Hamersley region is likely to have been warmer and wetter in the Eocene to early Miocene, as evidenced by the accumulation of lignite and siderite, which is preserved at the base of the CzD2 sequence in valleys that flank the Weeli Wolli Dome. Reduced, acidic run-off would be likely under these conditions.

Rare pockets of leached BIF are still preserved in areas protected from erosion. The very fine, powdery nature of the leached BIF highlights its very low preservation potential, providing an explanation for why it no longer occurs in large volumes and does not occur as clasts of leached BIF in the Cenozoic sedimentary record. Furthermore, the sandy component of the CzD2 package (Kepert, 2001; Knight et al., 2018) could potentially be derived from leached BIF. It is also likely that in many places only partial Fe leaching of the BIF occurred (see Fig. 20).

Fluid flow

The ore fluids involved in supergene martite-goethite mineralization are ultimately sourced from meteoric precipitation. Two processes are important here: (1) groundwater recharge through the focussing of meteoric run-off and its percolation through the vadose zone, and (2) subsurface groundwater flow...
beneath the water table. Surficial flow will be driven by paleotopography and channelled by paleodrainages. In the Weeli Wolli anticline area both paleotopography and paleodrainages are largely controlled by bedrock structure and stratigraphy (see Fig. 4). Fault zones, particularly subvertical ones, can present zones of enhanced permeability that contribute significantly to groundwater recharge. With the exception of the NE- and NW-trending dike-filled fracture zones, there are few confirmed camp-scale subvertical structures intersecting the Weeli Wolli anticline, but there are a number of deposit-scale ones. Where intruded by dolerite dikes, faults can also act as significant barriers to lateral flow of groundwater due to the production of clay-rich saprolite. In the Weeli Wolli anticline area the regional hydrologic gradient drives groundwater flow from west to east, from a water table high under the Milli Milli Dome in the west to a low at Weeli Wolli spring in the east (despite the drainage divide shown in Fig. 4, which is still in place). Both the NE- and the NW-trending dikes act as barriers to groundwater flow and cause stepwise changes in the level of the premining water table, suggesting ponding and reduced groundwater flow on the western side of the dikes (Fig. 11). At both the B and E deposits (see Fig. 7), better ore accumulations occur on the eastern side of the dolerite dikes where the water table is deeper: this makes sense in terms of our mass flux model since it implies a greater vadose zone thickness available for leaching of Fe. Groundwater flow is driven by the hydrologic gradient (i.e., the shape of the water table, which is in turn, partly controlled by topography) and flow is focused along zones of enhanced bedrock permeability. In a highly banded rock, like BIF, permeability will be much greater in the plane of bedding than orthogonal to it (i.e., strongly anisotropic permeability) and the zones of highest permeability appear to be the contacts between BIF units and more shaly layers which act as aquitards.

The depth of penetration of supergene ore fluids can be understood in terms of a combination of: (1) a gravity-related driving force propelling the Fe-bearing fluid plume down into the ambient groundwater reservoir, and (2) the locus of zones of enhanced permeability within the bedrock. The driving force will increase with bedrock dip as the vertical component of the

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**Fig. 17.** Cross plots illustrating the relationship between ore constituents and the premining water table. A. P wt % (samples are from the combined Mount Newman Member and belong to the high-grade martite-goethite mineralization category, they are colored by Al₂O₃ content). B. Vitreous goethite % (samples are colored by stratigraphic unit and sized by S content).
### Table 3. Comparison of the Composition of the Five Categories of Mineralization at South Flank with Respect to Stratigraphic Distribution, Mineralogy, Geochemistry, and Estimated Geometallurgical Composition

<table>
<thead>
<tr>
<th>Stratigraphic Unit</th>
<th>Martite-goethite high grade</th>
<th>Martite-goethite aluminous</th>
<th>Martite-goethite siliceous</th>
<th>Lateritic high grade</th>
<th>Lateritic low grade</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Number of samples</td>
<td>% of ore category</td>
<td>% of strat unit</td>
<td>Number of samples</td>
<td>% of ore category</td>
</tr>
<tr>
<td>WA1</td>
<td>29,989</td>
<td>16,686</td>
<td>20,744</td>
<td>56,888</td>
<td></td>
</tr>
<tr>
<td>WA2</td>
<td>18,094</td>
<td>32.5</td>
<td>32.8</td>
<td>18,051</td>
<td>13</td>
</tr>
<tr>
<td>N1</td>
<td>31,499</td>
<td>33.4</td>
<td>33.4</td>
<td>31,499</td>
<td>33.4</td>
</tr>
<tr>
<td>N2</td>
<td>18,827</td>
<td>30.1</td>
<td>30.1</td>
<td>18,827</td>
<td>30.1</td>
</tr>
<tr>
<td>N3</td>
<td>3,802</td>
<td>6.1</td>
<td>6.1</td>
<td>3,802</td>
<td>6.1</td>
</tr>
<tr>
<td>MM</td>
<td>540</td>
<td>9.5</td>
<td>9.5</td>
<td>540</td>
<td>9.5</td>
</tr>
<tr>
<td>Total</td>
<td>62,645</td>
<td>29,989</td>
<td>16,686</td>
<td>20,744</td>
<td>56,888</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Property</th>
<th>Martite-goethite high grade</th>
<th>Martite-goethite aluminous</th>
<th>Martite-goethite siliceous</th>
<th>Lateritic high grade</th>
<th>Lateritic low grade</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gibbsite (%)</td>
<td>23</td>
<td>26</td>
<td>18.3</td>
<td>1.5</td>
<td>1.3</td>
</tr>
<tr>
<td>Goethite (%)</td>
<td>10,538</td>
<td>54</td>
<td>63.5</td>
<td>51.7</td>
<td>49.7</td>
</tr>
<tr>
<td>Hematite (%)</td>
<td>10,539</td>
<td>42</td>
<td>30.3</td>
<td>31.7</td>
<td>29.8</td>
</tr>
<tr>
<td>Kaolinite (%)</td>
<td>10,535</td>
<td>2.1</td>
<td>8.0</td>
<td>9.8</td>
<td>8.7</td>
</tr>
<tr>
<td>Quartz (%)</td>
<td>674</td>
<td>13</td>
<td>9.0</td>
<td>8.3</td>
<td>1.3</td>
</tr>
<tr>
<td>Fe (wt %)</td>
<td>62,645</td>
<td>64.5</td>
<td>63.5</td>
<td>62.9</td>
<td>61.6</td>
</tr>
<tr>
<td>LOI (wt %)</td>
<td>62,645</td>
<td>5.0</td>
<td>5.0</td>
<td>4.7</td>
<td>4.5</td>
</tr>
<tr>
<td>SiO2 (wt %)</td>
<td>62,645</td>
<td>1.3</td>
<td>1.3</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>Al2O3 (wt %)</td>
<td>62,645</td>
<td>0.8</td>
<td>0.8</td>
<td>0.6</td>
<td>0.5</td>
</tr>
<tr>
<td>P (wt %)</td>
<td>62,645</td>
<td>0.05</td>
<td>0.05</td>
<td>0.05</td>
<td>0.05</td>
</tr>
<tr>
<td>density (g/cc)</td>
<td>45,388</td>
<td>2.9</td>
<td>2.9</td>
<td>2.9</td>
<td>2.9</td>
</tr>
</tbody>
</table>

Notes: Detection limits for mineralogical estimates: 0.2% for gibbsite and kaolinite, 1% for goethite, hematite and quartz; the geometallurgical properties are estimated based on the generic geometallurgical algorithm.

Abbreviations: n = number of samples; MM = MacLeod Member, Marra Mamba Iron Formation; N1 = lowermost unit of the Mount Newman Member, Marra Mamba Iron Formation; N2 = central unit of the Mount Newman Member, Marra Mamba Iron Formation; N3 = uppermost unit of the Mount Newman Member, Marra Mamba Iron Formation; WA1 = lower unit of the West Angela Member, Wittenoom Formation; WA2 = upper unit of the West Angela Member, Wittenoom Formation; IQR = interquartile range
Table 4. Comparison of Unmineralized and Mineralized Stratigraphic Units (all samples >100-m depth, unmineralized samples contain <48 wt % Fe, mineralized samples contain >48 wt % Fe)

<table>
<thead>
<tr>
<th></th>
<th>Unmineralized</th>
<th>Mineralized</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Median</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(n)</td>
<td>3,179</td>
<td>2,469</td>
</tr>
<tr>
<td>(Fe(%))</td>
<td>26.20</td>
<td>57.46</td>
</tr>
<tr>
<td>(SiO_2(%))</td>
<td>52.95</td>
<td>5.14</td>
</tr>
<tr>
<td>(Al_2O_3(%))</td>
<td>0.88</td>
<td>2.11</td>
</tr>
<tr>
<td>(TiO_2(%))</td>
<td>0.040</td>
<td>0.064</td>
</tr>
<tr>
<td>(Mn(%))</td>
<td>0.030</td>
<td>0.064</td>
</tr>
<tr>
<td>(MgO(%))</td>
<td>0.092</td>
<td>0.080</td>
</tr>
<tr>
<td>(CaO(%))</td>
<td>0.037</td>
<td>0.019</td>
</tr>
<tr>
<td>(Na_2O(%))</td>
<td>0.016</td>
<td>0.009</td>
</tr>
<tr>
<td>(K_2O(%))</td>
<td>0.027</td>
<td>0.003</td>
</tr>
<tr>
<td>(P(%))</td>
<td>0.035</td>
<td>0.007</td>
</tr>
<tr>
<td>Density (g/cc)</td>
<td>2.58</td>
<td>2.27</td>
</tr>
</tbody>
</table>

Abbreviations:
- MM = MacLeod Member, Marra Mamba Iron Formation;
- N1 = lowermost unit of the Mount Newman Member, Marra Mamba Iron Formation;
- N2 = central unit of the Mount Newman Member, Marra Mamba Iron Formation;
- N3 = uppermost unit of the Mount Newman Member, Marra Mamba Iron Formation;
- WA1 = lower unit of the West Angela Member, Wittenoom Formation;
- WA2 = upper unit of the West Angela Member, Wittenoom Formation.

Although the shape of the water table will play a role in determining supergene fluid-flow vectors, bedrock topology is likely to be the preeminent control, not only from the point of view of determining permeability variations but also in terms of the magnitude of the vertical (gravitational) driving force. The regional water table drops by ~200 m west-to-east across the 30-km strike of the Weeli Wolli anticline, and the maximum local drops, which are associated with NE- and NW-trending dikes, are on the order of 30 m over a horizontal distance of a few tens of meters. In contrast, the bedding may drop by over 200 m across a horizontal distance of ~200 m (e.g., Grand Central syncline, Valley and E deposits; see Figs. 7–8). The asymmetric, N-verging, mesoscale synclines appear to act as gutters, focusing down-flowing supergene fluids from the limbs into the synclinal keel and then channelling the fluids downplunge along the keel (Fig. 8). The steep (southern) limbs of these folds are typically thrust-thickened, producing structurally thickened packages of Mount Newman Member BIF with enhanced structural permeability (Figs. 7–9). Together, these two structural elements produce particularly favorable sites for supergene enrichment.

Structural elements can be either constructive or destructive of deep ore potential because of the different ways in which they affect groundwater flow. Subvertical structural elements such as faults, dike contacts, joint sets, and zones of axial-planar cleavage can act as zones of focused groundwater infiltration and can introduce bedding-orthogonal per -
of axial-planar cleavage can act as zones of focused ground-
water flow. Subvertical structural elements such as faults, dike contacts, joint sets, and zones of axial-planar cleavage can act as zones of focused groundwater infiltration and can introduce bedding-orthogonal permeability, which greatly enhances the downward flow of the ore-fluid plume (Figs. 7–9). In contrast, the shallow thrusts that are commonly developed on steeply-dipping fold limbs tend to act as aquitards (particularly where Mount Newman Member BIF is thrust over WA2 shale), preventing further downward percolation of the ore fluids (Figs. 7–8). Note that the keel position beneath a thrust may nonetheless be mineralized in part, providing the three-dimensional geometry of the syncline allows ore-fluid access either from the shallowly dipping northern limb or along the plunging keel itself (Figs. 7–9).

Despite the undoubted role played by structurally enhanced permeability, the fundamental concept that is central to this process model is that the primary control on ore-fluid hydrology is gravity-driven flow down bedding planes. This central observation explains every other observed feature of the three-dimensional distribution of martite-goethite mineralization and essentially the inherited structural architecture just provides the context for this process to play out. This type of control is
by no means obvious. We would normally expect the hydrology of consolidated rock sequences (such as Proterozoic sedimentary rocks) to be dominated by fractures, with limited bedding-parallel porosity/permeability. For example, as discussed below, the ingress of meteoric fluids during later lateritic weathering does not show this control and produces broadly subhorizontal, bedding-discordant zones of overprinting.

Iron flux
Fe mobility is the crux of the proposed mineral system. The vast areas of BIF that occur in the Hamersley Province contain significant volumes of Fe that have remained relatively immobile during the region's geologic history. The proposed fluid-flow model predicts that Fe is mobilized during a period of anomalous Fe mobility by leaching of BIF within the vadose zone and redistributed downdip within the BIF unit. In order to maintain the mineralizing system through a steady influx of Fe, new BIF (and, ultimately, previously mineralized BIF) must constantly be entering the vadose zone. The Hamersley Province is unusual in that despite its great age and cratonic basement, the highest mountain in Western Australia (Mount Meharry) is located in the center of the Province, just to the west of the Weeli Wolli anticline. The entire Province forms an elliptical topographic high and river profiling indicates that significant uplift has occurred over the last 60 Ma, accelerating at about 45 Ma (Czarnota et al., 2014), resulting in inverted topographic features (e.g., mesas of cemented, pisolithic, fluviatile material which make up the Miocene-age channel iron deposits; Morris and Ramananidou, 2007) and the incision of deep gorges into the peneplanned land surface (e.g., Karijini National Park).

The remainder of this section presents an order-of-magnitude mass-flux model for Fe in the martite-goethite ore systems of the Weeli Wolli anticline. The model takes into consideration Fe source regions, their volume and the efficiency with which groundwater flow and meteoric run-off may have transported and concentrated iron. The ore fluids are largely constrained to flow parallel to bedding within the BIF host unit and thus the topology of this unit can be analyzed in terms of watersheds, resulting in the definition of five catchments or ore domains (Fig. 5), which must be treated individually. Table 5 records the principal elements of the mass-flux model. The main assumptions are outlined in the following paragraphs.

Iron available in the source region: It is assumed that the main source region for the Fe contained in the orebodies was the now-eroded Mount Newman Member (based on the

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Fig. 18. Isocon diagrams showing the gains and losses of different chemical constituents as a result of martite-goethite mineralization for the principal ore-bearing units. A. Mount Newman Member N1. B. Mount Newman Member N2. C. Mount Newman Member N3. D. West Angela Member WA1.
premise that groundwater flow is likely to be predominantly bedding-parallel). The model assumes a conservative rate of erosion of 1 m/Ma and a maximum age for the mineralization of 50 Ma (Ramanaidou et al., 2019), resulting in a modeled mid-Eocene topographic surface that is 50 m higher than the modern land surface; erosion is assumed to be by peneplanation rather than pediplanation. Note that the conservative assumption that half the volume of the CzD2 and CzD3 sediments preserved in the surrounding valleys was sourced from the outcropping area of bedrock in the core of the Weeli Wolli anticline (the remainder being sourced from the Brockman Iron Formation ridges to north and south) and that no sediment has been washed out of these local valleys, gives a comparable figure of a minimum of 30 m of erosion since the mid-Eocene. The area of Mount Newman Member BIF removed since the mid-Eocene is the area of the N2 top surface between the modelled top of CzD1 (the CzD1 package appears to be Late Jurassic to Early Cretaceous in age, based on palynological dating, and would have been deposited around the flanks of the Weeli Wolli anticline prior to the martite-goethite mineralizing event) and the modelled mid-Eocene land surface (i.e., the present-day topographic surface raised through 50 m). This source area has then been divided into five ore domains on the basis of the N2 top watersheds: North Flank, Syncline Valley, South Flank, Highway Valley, and R Deposit Valley. The volume of material eroded in each domain is then calculated by multiplying the respective areas of N2 top by 65 m (the median unmineralized stratigraphic thickness of the Mount Newman Member in the Weeli Wolli anticline area). The contained Fe is calculated using the median Fe content and density of fresh, unmineralized Mount Newman Member in the Weeli Wolli anticline area (29 wt % Fe and 2.98 g/cc, respectively).

Iron added to martite-goethite deposits: The volume of the mineralizing systems is based on the modelled high-grade shells (>57 wt % Fe). These have been used because the effects of lateritic weathering give the low-grade shells (>48 wt % Fe) a blanket-like morphology near surface, which obscures the bedrock controls on orebody shape. The modelled high-grade ore shells have been grouped according to the ore domains and volumes are calculated accordingly. The amount of Fe added to each mineralized domain has been estimated using a typical background Mount Newman Member Fe content of 29 wt % and a density of 2.98 g/cm³ and the median Fe content and density of >57 wt % Fe samples within each domain (see Table 5). The estimated amount of Fe added is thus a minimum since it does not take into account the low-grade mineralization nor any mineralization that has been eroded.

A measure of the relative intensity of the ore system developed in each domain is given by the percentage of Fe added to create the high-grade mineralization as a proportion of the amount of Fe available in the source region (Table 5). The ore systems can be ranked in declining order: North Flank, Syncline Valley, South Flank, Highway Valley, and R Deposit Valley, with the Fe-fixing efficiency of the ore systems declining from about 30% for North Flank and Syncline Valley to about 10% for Highway Valley and R Deposit Valley. This
ranking corresponds with other measures of orebody quality such as maximum wt % Fe content, maximum depth of high-grade mineralization, and maximum Fe meters, which all decrease from North Flank/Syncline Valley to Highway Valley/R Deposit Valley (see Fig. 10, Table 5). It should be noted that the two least intense ore systems, Highway Valley and R Deposit Valley, occupy the gently-dipping western and eastern flanks of the domed Weeli Wolli anticline and that bedrock dips are typically much steeper on the northern and southern flanks of the anticline.

In summary, the results of this simple mass-balance analysis support the proposed fluid-flow model and predict the observed ranking of orebody quality at geologically reasonable levels of mineralizing efficiency.

Silica flux

Analyzing the complementary silica flux predicted by the mass flux model described above is more difficult. Silica distribution is analyzed in the context of the main ore horizon, N3, in an attempt to reduce the background variability that is due to stratigraphic position. Fresh, unoxidized, unmineralized N3 in the Weeli Wolli anticline area has a median SiO$_2$ content of ~50 wt % and an interquartile range of 45 to 55 wt % SiO$_2$. A value of SiO$_2$ >60 wt % is thus taken as anomalous for N3, although some such intervals could potentially be a primary cherty lithofacies. Furthermore, elevated silica values could be due either to Fe leaching in the source region or to silicification on the outflowing ore-fluid pathway. Hand specimen examples of both leached and silicified BIF are recognized, with the former being friable and the latter having clear evidence of silica (chalcedony) deposition in cavities (see Fig. 19). By analyzing the position of the siliceous N3 samples relative to mineralized N3, it is possible to identify subsets of the siliceous population that clearly lie up- or downdip of the mineralization, potentially corresponding to the leached and silicified cases described above. The two subsets of siliceous N3 have subtly different density, Fe content, and Al$_2$O$_3$/TiO$_2$ ratio, consistent with their putative origins (Fig. 20). The deeper samples have compositions close to that of unaltered N3 that has been diluted by ~30

![Silica flux analysis](image-url)

Fig. 20. Contrasts in physicochemical parameters between silica-rich N3 (unmineralized, SiO$_2$ >60 wt %): updip of martite-goethite mineralization (potentially due to Fe leaching) and downdip of martite-goethite mineralization (potentially due to silicification).
wt % silica addition, whereas the shallow samples are less dense and have a broad range in Fe and Al2O3/TiO2 values, consistent with variable degrees of leaching.

**Lateritic overprint**

The martite-goethite ores flanking the Weeli Wolli anticline are thought to be mid-Eocene in age (Ramanaidou et al., 2019) and are thus likely to have been exposed to at least one major episode of tropical lateritic weathering: the mid-Miocene event associated with channel iron deposit ore genesis (Morris and Ramanaidou, 2007). The weathering-related variation of both chemical and physical parameters with depth may have significant and predictable effects on the geometalurgical and materials handling properties of the ore and deserves to be better understood. Ramanaidou (2009) presented a detailed study of the lateritic ore developed on top of fresh itabirite (~51.6 wt % Fe, ~44.7 wt % SiO2) at the Capanema mine in Brazil. Five distinctive weathering fronts are recognized, each of which will be dynamic, gradually moving downward as weathering proceeds.

From surface downward these fronts are: (1) the erosion surface which is immediately underlain by hardcap, (2) the duricrust front that forms the base of the hardcap zone, (3) the cemented zone marked by the precipitation of secondary hematite and goethite, (4) the Fe oxide dissolution front characterized by the intense dissolution of primary Fe oxides (hematite and martite at Capanema) and the development of increased porosity, and (5) the desilification or weathering front.

Note that for lateritic weathering superimposed on martite-goethite mineralization, which is already desilicified, this latter front will be difficult to pick.

Vitreous goethite is typically developed during lateritic weathering and can be present in significant volumes in hard-capped ore (Clout, 2003, 2006; Crowe, 2012, unpub. report).

Vitreous goethite development is most prominent within 50 m of surface in the North and South Flank orebodies but a significant number of samples with >5% vitreous goethite occurs down to 100-m vertical depth. Several chemical and physical properties of the mineralized material also show a marked difference in composition with depth, particularly over the top 100-m vertical interval (Figs. 13–15): for a given stratigraphic unit, Fe content tends to be lower, and LOI, SiO2, Al2O3, and S contents tend to be higher in this depth interval than at greater depth.

Samples from the most shallow 50 m tend to have a more variable composition than those from deeper intervals (with the exception of Fe, SiO2, and density in the deepest interval, where the increased variability is reflecting incomplete mineralization on the distal, outflow portion of the supergene mineralizing system; Fig. 15). These observations are in keeping with the well-documented characteristics of hardcap but true hardcap is typically 30 to 40 m thick and reaches a maximum thickness of 80 m in the Hamersley Province (Crowe, 2012, unpub. report).

It is not only the chemical effects of lateritic weathering that extend to at least 100 m, physical properties also show a relationship with depth from surface, particularly over the top 100 m, with the goethite to goethite + hematite ratio and estimated lump percent decreasing and density increasing with depth bin (Fig. 15). These chemical and physical changes can be attributed to the mineralogical changes observed in the Capanema study, such as increasing proportion of goethite, increasing Al content of goethite and hematite, and increasing porosity within the weathering zone compared to the fresh rock.

These depth-related variations are interpreted to reflect variable degrees of lateritic overprinting of the primary martite-goethite mineralization. In the Weeli Wolli anticline area, S is a particularly good indicator of weathering with values >0.02 wt % occurring almost exclusively within 100 m of surface. This has been used to divide the mineralized sample population into primary martite-goethite mineralization (S <0.02 wt %) and lateritic mineralization types.

In contrast to the other BIF-rich portions of the mineralizing system (N1, N2, N3, and WA1) the MacLeod Member mineralization is almost completely restricted to the top 100 m (Fig. 12), has uniformly elevated S contents (>0.02 wt %) and falls consistently into the lateritic low-grade category of mineralization. In the lateritic Capanema deposit, increased availability of Al2O3 in the form of proximity to dolerite dikes, greatly increases the thickness of the lower part of the laterite profile and effectively increases the thickness of the mineable orebody (Ramanaidou, 2009).

It would appear as though the MacLeod Member mineralization may be purely lateritic in origin and that the more shale-rich nature of this BIF member has facilitated development of lateritic ore in this stratigraphic unit.

The shallow mineralization which extends as a zone of relatively constant thickness, discordant to stratigraphy, under the outcropping portion of the northern limb of the B deposit syncline (Fig. 9), may also represent mineralization which is largely lateritic in origin. If the N3 mineralized samples within 50-m depth are categorized very simply as having >60 or <60 wt % Fe there is a tendency for the lower grade samples to occur within drill holes that are collared in outcropping N3,
whereas those samples containing >60 wt % Fe typically occur in drill holes that are collared in WA2 or WA1, and hence are partially shielded from the effects of lateritic oxidation and hardcap development. There is, however, a fair amount of intermingling of the two samples types in many drill holes, suggesting that the effects of lateritization are complex and deserve further study.

**Orebody zonation**

As described in the section detailing controls on ore composition, five types of mineralization are recognized in martite-goethite deposits of the North and South Flank areas. Original stratigraphic differences between Fe-rich BIF of the Mount Newman Member and the more aluminous, shale-rich West Angela Member results in contrasting primary styles of martite-goethite mineralization described as martite-goethite high-grade and martite-goethite aluminous types, respectively. Variations in the intensity of the mineralizing martite-goethite system results in a third type, martite-goethite siliceous, which is restricted to the lower margins and downdip extremities of the ore system and reflects incomplete removal of silica during mineralization (Fig. 16). These three categories of mineralization reflect the character of the primary, mid-Eocene, martite-goethite ore-forming event.

Unmineralized alteration zones are developed surrounding the martite-goethite orebody and reflect the passage of the ore fluids. In addition to the Fe-leaching updp of the orebody and the silica metasomatism, downdip in the Mount Newman Member, there is a zonation developed within the West Angela Member. A systematic downdip compositional variation is evident, in the form of a transition from mineralized material (>48 wt % Fe) in the laterite zone (within ~50 m of surface), through a zone of manganiferous shales (~50- to 100-m vertical depth from surface) to pristine, carbonate-bearing WA2 at depth. The close spatial association between manganiferous, shaly WA2 and Fe mineralization is noted by a number of authors (e.g., Blockley et al., 1993; Hankohl, 2014, unpub. report). Dissolution of the carbonate fraction in WA2 adjacent to the martite-goethite mineralization developed on the flanks of the Weeli Wolli anticline has been invoked to explain: (1) the unusual convex-up morphologies of some of the thrusts, which are typically developed on the steep limbs of mesoscale, asymmetric folds; and (2) the steep bedding displayed by some outcrops of CzD1 (commonly >45° i.e., much greater than the angle of repose for fine-grained sediments; Crowe and Perring, 2017). The bulk of the carbonate dissolution must have occurred prior to deposition of the CzD2 as it has preserved CzD1 beneath the CzD2 unconformity. Also, the observed karstic effects on the CzD2 sequence (e.g., shallow depressions in the top surface of the pedogenic calcrite layer that formed at the top of the CzD2 sequence) are significantly less than those affecting the CzD1 sequence. These observations suggest a possible genetic relationship between carbonate dissolution in the WA2 unit and martite-goethite mineralization.

Later lateritic weathering, possibly mid-Miocene in age (Ramanaidou et al., 2016) and related to the genesis of the CzD2 channel iron deposits and the lateritic profile preserved by CzD2 sediments flanking the Weeli Wolli anticline (Kepert, 2001), creates a further two mineralization categories. Lateritic high-grade mineralization is developed primarily after martite-goethite high-grade mineralization, which in turn is restricted to the Fe-rich Mount Newman Member of the Marra Mamba Iron Formation. Lateritic low-grade mineralization may result from the weathering of martite-goethite aluminous and martite-goethite siliceous categories, but some of this low-grade, shallow, surface-blanketing mineralization may be of a wholly lateritic origin (e.g., the low-grade mineralization developed in the MacLeod Member; see also Fig. 9).

The current pattern of mineralization styles can thus be attributed to a two-stage genetic process: (1) primary martite-goethite ore formation in the mid-Eocene, followed by (2) overprinting lateritic Fe enrichment, probably originating in the mid-Miocene. This two-stage model for ore formation is illustrated schematically in Figure 21. The reduced, acidic, Fe-rich supergene plume, which affects martitization and goethite replacement of the gangue component in the Mount Newman Member BIF, causes carbonate dissolution and Mn + Fe enrichment in the overlying West Angela Member. This results in the steepening of the basal contact of CzD1 inset valleys and corresponding steepening of the sedimentary bedding. Both Fe and Mn appear to be mobile in these fluids (which would support the assertion that the fluids were reduced) since portions of WA1 may be manganiferous and there is apparent Mn enrichment of the basal CzD1 in several deposits (e.g., Syncline Valley; shown schematically in Fig. 21A). Later lateritic weathering has the effect of reducing the Fe grade and increasing the geochemical compositional variation and the lump percentage of the near surface (<100 m) portions of the martite-goethite system. Lateritic weathering can also cause residual Fe enrichment in BIF units that were not originally part of the martite-goethite system, creating shallow blankets of low-grade ore that mimic the topography and extend the footprint of the mineralization near surface (Fig. 21B).

**Discussion**

The ore genetic model proposed here for martite-goethite mineralization is starkly different from previously published models. A discussion of these differences forms the first part of this final section. Another point of difference concerns the nature of the magnetite to hematite conversion, with earlier models assuming an oxidative transformation. The discussion continues with a review of recent work on the nonredo transformation. The final parts of the discussion deal with the potential geodynamic and paleoclimatic drivers of this unusual style of mineralization and with the implications of the model for mineral exploration.

**Comparison with existing ore genetic models**

Morris et al. (1980) first proposed the electrochemical cell model for supergene martite-goethite ore formation (see also Morris, 1985; Ramanaidou and Morris, 2010). These authors proposed that a large corrosion cell is set up with magnetite layers in BIF acting as electron conductors. The cathode is magnetite-rich BIF in contact with oxygenated groundwaters near surface. Here the electrons, which have been released at depth and conducted up through the magnetite layers are consumed by the reduction of oxygen and the generation of hydroxyl ions. The electron-generating anodic reaction which occurs at depth is the conversion of $\text{Fe}^{2+}$ to $\text{Fe}^{3+}$ plus a free...
electron. The Fe$^{3+}$ is hydrolyzed to ferrihydrite, which eventually recrystallizes as goethite. The hydrolysis reaction generates H$^+$ ions that facilitate the dissolution of iron carbonates and silicates in the gangue layers, which in turn releases more Fe$^{2+}$. For ore formation to take place at depth the model requires a zone at depth with better fluid access, such as a fault zone cutting across the BIF. This fault plane transports additional Fe$^{2+}$, released near surface by biogenically aided dissolution, to the orebody forming at depth. From the point of view of the current analysis, the main problem with this model is that it predicts that ore formation will take place bottom up and yet the morphology of all the North and South Flank deposits clearly indicates a top-down process with a well-defined connection to the surface.

A completely different model for martite-goethite ore formation is proposed by Lascelles (2002, 2006). According to this model, certain portions of the BIF stratigraphy lose their silica component during compaction and diagenesis, becoming a distinct facies described as chert-free BIF. Supergene enrichment of this chert-free BIF (and only this chert-free BIF) results in martite-goethite mineralization. The problems with this model are threefold:

1. Despite an extensive drill database, which includes holes drilled into the Mount Newman Member downdip of martite-goethite mineralization, there is no evidence of this chert-free BIF at North or South Flank and, in fact, some Mount Newman Member material at the downdip extremities of the ore system is unusually siliceous.

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**Fig. 21.** Schematic illustrations. (A). The proposed chemical zonation developed around supergene martite-goethite mineralization in the mid-Eocene. (B). The later superimposed overprint related to lateritic weathering in the mid-Miocene.
2. The detailed petrographic work of Morris (1980, 1985) and Ramanadou and Morris (2010) clearly shows goethite replacing carbonate, silicate, and chert minerals.

3. It is not clear why a distinct BIF facies, developed on the seafloor in the Archean to Paleoproterozoic, should now show such an intimate spatial connection to the current land surface and be coincident with more steeply dipping portions of the stratigraphy.

A nonredox origin for martite-goethite mineralization

A full discussion of the ore-forming process from the perspective of grain-scale mineralogical conversions is beyond the scope of this paper, but some recent experimental work sheds new light on the magnetite to hematite conversion, which might have implications for understanding the process of supergene martite-goethite ore formation and for revealing more about the unusual conditions which led to the formation of this unique style of supergene mineralization. Zhao et al. (2018) studied the replacement of magnetite by hematite under low-temperature hydrothermal conditions (140º–220ºC) and found that (1) oxygen is not an essential factor in the replacement, although excess oxidant does trigger the oxidation reaction (as opposed to the nonredox transformation; see Ohmoto, 2003) and increases the rate of reaction; (2) even under high O2(aq) environments, some of the replacement still occurs via Fe2+ leaching from magnetite (i.e., nonredox); and (3) in nature, the overall rate-limiting step is likely to be the efficiency of Fe transport, although at the grain scale it is the dissolution of magnetite rather than the precipitation of hematite. According to Zhao et al. (2018) the redox-independent reaction can be viewed as a dissolution and reprecipitation reaction and under acidic conditions it can be written as follows:

$$\text{Fe}_3\text{O}_4(\text{mt}) + 2\text{H}^+ + 2\text{H}_2\text{O} \rightarrow \text{Fe}^{2+} + 2\text{Fe}(	ext{OH})_3(\text{aq}), \quad (1)$$

with a possible reprecipitation reaction being:

$$2\text{Fe}(	ext{OH})_3(\text{aq}) \rightarrow \text{Fe}_2\text{O}_3(hm) + 3\text{H}_2\text{O} \quad \text{(redox-independent).} \quad (2)$$

The overall nonredox transformation can be written:

$$\text{Fe}_3\text{O}_4(\text{mt}) + 2\text{H}^+ \rightarrow \text{Fe}_2\text{O}_3(hm) + \text{Fe}^{2+} + \text{H}_2\text{O} \quad \text{(redox-independent).} \quad (3)$$

Note that the dissolution of magnetite results in the formation of both Fe2+ and Fe3+ complexes within solution (reaction 1). The Fe3+ precipitates almost instantaneously as hematite (reaction 2) as a result of the low solubility of Fe3+ and the fact that nucleation occurs epitaxially, facilitated by the topotactic relationship between the lattice structure of magnetite and hematite. The Fe2+ will remain in solution and can diffuse away from the reaction front, and in martite-goethite ore systems, presumably causes goethitization of the gangue component of the BIF (i.e., the chert, Fe silicate, and carbonate bands). Zhao et al. (2018) argued that magnetite replacement takes place via an interface-coupled dissolution reprecipitation (ICDR) reaction and proceeds in a pseudomorphic manner, producing rims of porous secondary hematite. It also takes place preferentially along the (111) planes in the magnetite crystal, producing trelfis textures similar to those described by Morris (1985) from supergene Fe ores.

A corollary of the pseudomorphic replacement process, both the generation of hematite after magnetite and goethite after gangue phases, is that it typically introduces porosity. The mineralizing process thus creates porosity (and potentially permeability) and is likely to be self-propagating as long as there is a continuous supply of ore fluid. Additionally, the initial replacement of the carbonate gangue minerals would allow acidic fluids to travel farther along the flow path before being buffered.

The fundamental control exerted on the distribution of martite-goethite mineralization by bedded-plane permeability within BIF horizons suggests that the supergene ore-fluid plume created its own porosity via the relevant ore-forming reactions, and that these were in turn controlled by bedding. This putative active porosity generation process may be an important clue as to the unique conditions of martite-goethite ore formation. Indeed, it may be that the distribution of magnetite is the critical controlling feature of these ore systems, as the nonredox transformation to hematite not only releases Fe2+ to the fluid phase but concurrently introduces porosity. Further research is required to formulate a comprehensive chemical (as opposed to physical) process model for supergene martite-goethite ore formation.

A unique conjunction of stratigraphy, uplift history and climate?

The martite-goethite ore system is clearly distinct from lateritic Fe enrichment though both could be described as supergene processes. Supergene Fe ores in other parts of the world appear to be largely of lateritic weathering origin (e.g., India, Mukhopadhyay et al., 2008; Quadrilatero Ferrifero, Rosiere et al., 2008; Yilgarn craton, Angerer and Hagemann, 2010). The key to martite-goethite ore genesis may lie in the unique combination of climate and uplift history in the Hamersley Province. The local climate in the mid-Eocene was likely to have been temperate and wet enough to support temperate rainforest species (e.g., Nothofagus and Casuarinaceae spores preserved in mid-Eocene rocks at Glen Florrie Station near the southwestern margin of the Hamersley Province; Martin, 2006). General circulation models suggest that the early-Eocene climate in northwest Australia was cool (average summer temperature of 18ºC and average winter temperature of 12ºC) with an annual rainfall of approximately 1,200 mm (Frakes and Barron, 2001). Planktonic oxygen isotope analyses also suggest a sharp drop in ocean surface temperatures at ~50 Ma (Frakes et al., 1994; summarized by McGowan and Li, 1998). In a wet ecosystem with abundant rotting organic matter covering the ground it is possible that some of the Fe2+ in solution owes its presence to organic processes, as is likely to be the case in iron bogs (Stan- ton et al., 2007). Certain microbes exist (in both aerobic and anaerobic environments), which derive energy from the oxidation of organic compounds coupled with reduction of Fe3+ held in the lattice of iron oxide minerals. This type of bacterial respiration will release additional Fe2+ into the groundwater (over and above that released through the inorganic reaction 3) and has been mooted as a possible factor in channel iron deposit genesis (Fritz and Yapp, 2018). There was a major geodynamic change at ~45 Ma when the northward drift of Australia away from Antarctica accelerated, probably in response to broader scale plate re-organization processes (Cockburn, 2014). This would have resulted in rapid climate change and may have
been responsible for the termination of the martite-goethite ore-forming event.

The second critical component, and one that is particularly important to the generation of supergene ores generally, is having sufficient concurrent uplift to continuously bring new rock mass into the vadose zone where it can be leached and provide an ongoing metal input into the supergene system (Note that uplift will also enhance the regional hydrologic gradient). River-profile modeling by Czarnota et al. (2014, fig. 20) predicts the onset of vigorous uplift of the Hamersley Dome at about 45 Ma, and it is probably no coincidence that this corresponds broadly with the estimated age of the martite-goethite event. Uplift is in fact likely to be the key factor in martite-goethite ore genesis; humid climates and related acid-generating vegetation are likely to be much more common in time and space than instances of dynamic topography overprinting a major BIF accumulation.

**Implications for exploration targeting**

A mineral system view of the martite-goethite ore system is presented in Table 6. This captures the main elements of the model, their constituent processes and the geologic proxies that can be used in exploration across a variety of scales.

Empirical limits to ore distribution around the Weeli Wolli anticline include a maximum vertical depth of high-grade mineralization of ~290 m, and a maximum down-dip extent of ~650 m. Furthermore, the vast majority of mineralization is located within 150 m of the topographic surface. This is one of the largest accumulations (~3 Bt of iron ore grading ~60% Fe) of martite-goethite ore in the Hamersley Province, and mineralization covering ~60 km in total strike length has been drilled to a high density (150 × 50 to 50 × 50 m). These empirical limits are thus likely to be a good reflection of the typical dimensions of these supergene martite-goethite ore systems.

At the camp scale, the principal components of an efficient martite-goethite ore-forming system appear to comprise the following features:

1. At the time of ore formation, Fe-rich BIF (e.g., Mount Newman Member) was cropping out or within the vadose zone.
2. Tight, mesoscale synclines with kilometer-scale strike lengths occur within Fe-rich BIF within 300 m of surface.
3. The orientation of these synclines is subparallel to major fluvialite drainages, to the regional strike of the bedrock and to the hydrologic gradient.
4. Bedrock dips are moderate to steep in these synclines and barren zones exist where bedding dips are <10°.

These targeting components can be further refined at the prospect scale to include: (1) positions where synclinal keels composed of the most Fe-rich parts of the BIF stratigraphy are <150 m deep (deeper keel positions are typically not fully mineralized); and (2) positions of maximum bedding dip—subtle variations in dip of a fold limb can have a significant effect on the local depth extent of the mineralization and small-scale dip reversals appear to terminate the ore system.

In addition, it should be noted that (1) thrusts on the steep limbs of mesoscale synclines typically act as aquitards and the prospective BIF unit beneath them may be barren if the synclinal keel is located at >150-m depth; (2) better mineralization tends to occur on one side of dolerite dikes and the dike contacts appear to act as fluid recharge zones, delivering supergene fluids to anomalous depths; and (3) in general, flat-lying structures (i.e., both fractures and bedding) tend to act as barriers to the downward propagation of the mineralization front, whereas subvertical structures facilitate deep penetration of mineralizing fluids resulting in deep pockets of ore.

**Concluding Remarks**

Martite-goethite mineralization represents the fossilized footprint of iron-rich supergene fluid plumes formed during a period of anomalous iron mobility in a region containing vast volumes of banded iron-formation, and as such the geometry of the mineralization directly reflects the flow vectors of the mineralizing fluids. The morphology of the martite-goethite orebodies at North and South Flank illustrates a strong control by bedding-parallel permeability with local contributions from structural elements. These observations, derived from three-dimension modeling and analysis of a very large database of drill hole data covering over 60 km strike length of martite-goethite mineralization, are synthesized into a physical process model of ore formation that invokes the following key stages:

1. The ongoing uplift of the BIF-rich Hamersley Province during the Cenozoic provided a mechanism for continuous delivery of unleached BIF to the vadose zone.
2. A wetter, more temperate climate in the mid-Eocene provided the optimal conditions for the generation of acidic, reduced fluids, through interaction of meteoric run-off with surficial organic matter.
3. These fluids leached Fe(II) from magnetite in the vadose zone, generating a supergene ore-fluid.
4. The ore fluid was driven by gravity and focused along bedding planes within the BIF into the body of ambient alkaline groundwater, effecting nonredox, mimetic replacement of magnetite by hematite and of the gangue minerals by goethite coupled with the release of silica into the fluid phase (porosity potentially generated by the ore-forming chemical reactions).
5. A change from silica leaching to silica deposition occurs on the downdip margins of the system before the spent ore-fluid plume is eventually diluted and becomes indistinguishable from the surrounding body of groundwater.

While this model provides a robust framework to guide exploration and resource definition there are significant pieces of further research required to complete our understanding of this economically significant category on mineralization. These include (1) refining the precise timing of the martite-goethite event (or events) and developing a better understanding of the geodynamic and paleoclimatic context; and (2) development of an integrated chemical process model, which explains the observed textural and mineralogical relationships, to complement our physical process model.

**Acknowledgments**

BHP is thanked for permission to publish this paper. We acknowledge and recognize the many BHP geoscientists involved in the discovery and evaluation of the North and South Flank deposits whose work has made this analysis possible. The manuscript was greatly improved as a result of constructive reviews by Matthew Cross, Paul Duuring, and an
<table>
<thead>
<tr>
<th>Model elements</th>
<th>Geologic proxies</th>
<th>Scale</th>
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<tbody>
<tr>
<td>Rainwater becomes reduced and acidic after passage through a surficial layer of rotting organic matter</td>
<td>Temperate climate, seasonally wet</td>
<td>Regional</td>
</tr>
<tr>
<td>Reduced acidic fluid leaches Fe from BIF in the vadose zone; possibly enhanced by microbial reduction of Fe$^{3+}$ to Fe$^{2+}$</td>
<td>Evidence of regional uplift and erosion at the time of mineralization</td>
<td>Camp</td>
</tr>
<tr>
<td>Acidic, Fe-rich supergene fluid plume is driven by gravity along bedding planes in BIF and down steeply dipping structures and displaces the alkaline, Fe-poor groundwater, causing goethetization of gangue layers and release of silica, martitization of magnetite layers, stabilization of Al as kaolinite, and formation of martite-goethite ore</td>
<td>Steeply dipping structures (faults, dike contacts, joints, cleavage planes)</td>
<td>Prospect</td>
</tr>
<tr>
<td>The supergene fluid plume changes chemistry with depth, losing Fe and becoming more silica rich, the mineralizing potential of the system declines</td>
<td>Siliceous mineralization (e.g., Fe &lt;48% but SiO$_2$ &gt;5%)</td>
<td></td>
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<tr>
<td>The spent supergene fluids, now Fe poor and silica rich, cause silica cementation on the outflow path Mixing with and dilution by ambient groundwater eventually destroys all trace of the supergene fluid plume</td>
<td>Silica precipitation takes the place of leaching</td>
<td></td>
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Subsurface topology of BIF bedding planes continues to drive the supergene fluid plume downward

Subsurface topology of BIF bedding planes focuses supergene fluid flow

Folding produces zones of structurally enhanced permeability

Folding produces zones of structural thickening of the BIF protolith

Bedding planes within BIF dip at >10° without dip reversals

Synchonal keels of mesoscale folds, especially favorable within 150 m of surface and if plunge is subparallel to regional strike and regional hydrologic gradient

Axial-planar foliation, small-scale parasitic folding

Fold-induced thrusting on the steeply overturned limbs of mesoscale folds; note: these thrust planes can act as aquitards, particularly where BIF is thrust over shale, shielding steeply dipping BIF stratigraphy beneath the thrust plane from the continued downward passage of the supergene fluid plume

Declining ore quality indicates the fringes of the ore system have been reached

With increasing depth, the vertical, gravity-driven component of fluid flow is gradually overwhelmed by lateral flow driven by the hydrologic gradient

Siliceous mineralization (e.g., Fe <48% but SiO$_2$ >5%)

Maximum vertical depth of high-grade mineralization ~290 m (empirical observation)

Maximum downdip extent of high-grade mineralization ~650 m (empirical observation)

Vast majority of mineralization is located within 150 m of surface (empirical observation)

Mineralization is most persistent with depth along BIF-shale contacts (at the member or macroband scale)

Dense, silica-cemented BIF preserved fringing the downdip, outflow portion of the ore system
unknown reviewer. We acknowledge and thank the Banjima people—the Traditional Owners of Mining Area C.

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