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BIF-hosted iron mineral system: A review

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ABSTRACT

The BIF-hosted iron ore system represents the world's largest and highest grade iron ore districts and deposits. BIF, the precursor to low- and high-grade BIF hosted iron ore, consists of Archean and Paleoproterozoic Algoma-type BIF (e.g., Serra Norte iron ore district in the Carajás Mineral Province), Proterozoic Lake Superior-type BIF (e.g., deposits in the Hamersley Province and craton), and Neoproterozoic Rapitan-type BIF (e.g., the Urucum iron ore district).

The BIF-hosted iron ore system is structurally controlled, mostly via km-scale normal and strike-slips fault systems, which allow large volumes of ascending and descending hydrothermal fluids to circulate during Archean or Proterozoic deformation or early extensional events. Structures are also (passively) accessed via downward flowing supergene fluids during Cenozoic times.

At the depositional site the transformation of BIF to low- and high-grade iron ore is controlled by: (1) structural permeability, (2) hypogene alteration caused by ascending deep fluids (largely magmatic or basinal brines), and descending ancient meteoric water, and (3) supergene enrichment via weathering processes. Hematite- and magnetite-based iron ores include a combination of microplaty hematite–martite, microplaty hematite with little or no goethite, martite–goethite, granoblastic hematite, specular hematite and magnetite, magnetite–martite, magnetite-specular hematite and magnetite–amphibole, respectively. Goethite ores with variable amounts of hematite and magnetite are mainly encountered in the weathering zone.

In most large deposits, three major hypogene and one supergene ore stages are observed: (1) silica leaching and formation of magnetite and locally carbonate, (2) oxidation of magnetite to hematite (martitisation), further dissolution of quartz and formation of carbonate, (3) further martitisation, replacement of Fe silicates by hematite, new microplaty hematite and specular hematite formation and dissolution of carbonates, and (4) replacement of magnetite and any remaining carbonate by goethite and magnetite and formation of fibrous quartz and clay minerals.

Hypogene alteration of BIF and surrounding country rocks is characterised by: (1) changes in the oxide mineralogy and textures, (2) development of distinct vertical and lateral distal, intermediate and proximal alteration zones defined by distinct oxide–silicate–carbonate assemblages, and (3) mass negative reactions such as de-silicification and de-carbonatisation, which significantly increase the porosity of high-grade iron ore, or lead to volume reduction by textural collapse or layer-compaction. Supergene alteration, up to depths of 200 m, is characterised by leaching of hypogene silica and carbonates, and dissolution precipitation of the iron oxyhydroxides.

Carbonates in ore stages 2 and 3 are sourced from external fluids with respect to BIF. In the case of basin-related deposits, carbon is interpreted to be derived from deposits underlying carbonate sequences, whereas in the case of greenstone belt deposits carbonate is interpreted to be of magmatic origin. There is only limited mass balance analyses conducted, but those provide evidence for variable mobilization of Fe and depletion of SiO_2. In the high-grade ore zone a volume reduction of up to 25% is observed.

Mass balance calculations for proximal alteration zones in mafic wall rocks relative to least altered examples at Beebyn display enrichment in LOI, F, MgO, Ni, Fe_2O_{3total}, C, Zn, Cr and P_2O_5 and depletions of CaO, S, K_2O, Rb, Ba, Sr and Na_2O. The Y/Ho and Sm/Yb ratios of mineralised BIF at Windarling and Koolyanobbing reflect distinct carbonate generations derived from substantial fluid–rock reactions between hydrothermal fluids and igneous country rocks, and a chemical carbonate-inheritance preserved in supergene goethite.

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Hypogene and supergene fluids are paramount for the formation of high-grade BIF-hosted iron ore because of the enormous amount of: (1) warm (100–200 °C) silica-undersaturated alkaline fluids necessary to dissolve quartz in BIF, (2) oxidized fluids that cause the oxidation of magnetite to hematite, (3) weakly acid (with moderate CO_2 content) to alkaline fluids that are necessary to form widespread metasomatic carbonate, (4) carbonate-undersaturated fluids that dissolve the diagenetic and metasomatic carbonates, and (5) oxidized fluids to form hematite species in the hypogene- and supergene-enriched zone and hydroxides in the supergene zone.
Four discrete end-member models for Archean and Proterozoic hypogene and supergene-only BIF hosted iron ore are proposed: (1) granite–greenstone belt hosted, strike-slip fault zone controlled Carajás-type model, sourced by early magmatic (± metamorphic) fluids and ancient "warm" meteoric water; (2) sedimentary basin, normal fault zone controlled Hamersley-type model, sourced by early basinal (± evaporitic) brines and ancient "warm" meteoric water. A variation of the latter is the metamorphosed basin model, where BIF (ore) is significantly metamorphosed and deformed during distinct orogenic events (e.g., deposits in the Quadrilátero Ferrífero and Simandou Range). It is during the orogenic event that the upgrade of BIF to medium- and high-grade hypogene iron took place; (3) sedimentary basin hosted, early graben structure controlled Urucum-type model, where glaciomarine BIF and subsequent diagenesis to very low-grade metamorphism is responsible for variable gangue leaching and hematite mineralisation. All of these hypogene iron ore models do not preclude a stage of supergene modification, including iron hydroxide mineralisation, phosphorous, and additional gangue leaching during substantial weathering in ancient or Recent times; and (4) supergene enriched BIF Capanema-type model, which comprises goethitic iron ore deposits with no evidence for deep hypogene roots. A variation of this model is ancient supergene iron ores of the Sishen-type, where blocks of BIF slumped into underlying karstic carbonate units and subsequently experienced Fe upgrade during deep lateritic weathering.

1. Introduction

Banded-iron-formation-(BIF) hosted iron deposits, particularly the high-grade (>60% Fe) version of those, are the most important Fe metal system worldwide as far as production, reserves and employment of geologists are concerned. This is reflected in the significant amount of scientific papers published and conferences and short courses held over the past 15 years (see Hagemann et al., 2008 in Economic Geology Reviews vol. 15 for a recent review). The BIF hosted iron (ore) system is defined as iron ore that is located in, and in large parts derived from, BIF which is defined as "a variety of iron formation that contains distinct chert layers, known as bands; iron formation is a finely laminated to thinly bedded chert-bearing chemical sedimentary rock containing at least 15 wt.% iron of sedimentary origin" (Beukes and Gutzmer, 2008; Hagemann et al., 2008 for references and further detailed subdivisions). Unlike most other metal systems the host rock of iron ore (i.e., BIF, Fig. 1) and thus protolith is already significantly rich in Fe, containing on average about 30 to 35 wt.% Fe (Klein, 2005), therefore pointing to a local source for Fe. Banded Iron Formation alone, in most rational economic cases, does not represent iron ore, instead BIF is upgraded via complex geochemical processes into iron ore. These processes can be rapid, <5 million years as established for many hydrothermal systems (Chiaradia et al., 2014), or may span a total of 100's of millions of years if hypogene iron ore is further upgraded by weathering and/or supergene enrichment that often took place during the last 20 million years (Angerer et al., 2014a).

Unlike many other metal systems, such as orogenic gold, porphyry copper, or MVT lead and zinc, the BIF-hosted iron system contains several significant ore types defined by various proportions of oxides and hydroxide mineral species and with characteristic contaminants (e.g., P, Si, Al) and physical properties (e.g., hard-consistent, friable-inconsistent, blue dust).

There are presently only a limited number of major BIF-hosted iron ore districts in the world (Fig. 2, Table 1). They are hosted in Archean cratons and Proterozoic mobile belts constrained to the Precambrian Era. In the Hamersley Province of Western Australia the three largest world class deposits are the Mt. Whaleback, Mt. Tom Price and Paraburdoo–Channar deposits; other significant deposits include Hope Downs, Mining Area C and Cloud Break (Thorne et al., 2008a). In the Archean Yilgarn and Pilbara cratons of Western Australia the Koolyanobbing camp in the Yilgarn (Angerer et al., 2012) and the Abydos, Wodgina, and Corunna Downs camps in the Pilbara, are presently the only significant producing deposits (Teitler et al., 2014). Brazil contains three major iron ore districts: the Quadrilátero Ferrífero (QF) in the São Francisco craton (Rosière et al., 2008), Serra Norte at Carajás in the Amazon craton (Figueiredo e Silva et al., 2008), and Urucum near Corumbá in the Amazon craton-Rio Apa block (Urban et al., 1992). In West Africa, the Pic de Fon deposit in the Simandou Range of the West African craton (Republic of Guinea) is planned to be in production within the next five years (Cope et al., 2008). In South Africa the Thabazimbi and Sishen deposits in the Kapvaall craton are significant producers (Beukes et al., 2002b). In India, major districts include the Noamundi-Koira Valley in the Singhbhum craton, the Bailadila–Dalli–Rajhara deposits in the Bastar craton, the Donimalai–Hospet deposits in the Eastern Dharwar craton and the Goa deposits in the Western Dharwar craton (Mukhopadhyay et al., 2008). In the Ukraine, the Krivoy Rog deposit is located in the Ukrainian Shield (Dalstra and Guedes, 2004). In China, BIF-hosted deposits are the main source of iron, representing more than 58% of known iron ore resources (Zhang et al., 2012); however, these deposits are mostly lower-grade, with ores containing more than 50 wt.% Fe accounting for just 1% of total iron ore resources (Zhang et al., 2014).

This paper provides a mineral system analysis of the BIF-hosted iron ore system with emphasis on: (1) geotectonic controls, (2) fluid pathways from the source of hydrothermal fluids to the depositional sites, (3) depositional sites in terms of their architecture (structural framework and geometry), footprint of ore and alteration minerals, hydrothermal fluid characteristics and source(s), and physico-chemical processes necessary to upgrade the BIF protoliths to high-grade iron ore, and (4) ore preservation, uplift and supergene modification.

2. Geotectonic controls of BIF deposition

There are three main types of geotectonic settings in which BIF is located (Fig. 3):

(1) Algoma-type banded iron formations are stratigraphically linked, or interlayered with, submarine-emplaced volcanic rocks in convergent margin settings of Archean and Paleoproterozoic granite–greenstone belts (Gross, 1980, 1993, Fig. 3). Significant deposits include the Archean Yilgarn- and Pilbara-type deposits (Angerer et al., 2014a; Teitler et al., 2014) and Carajás deposits (Figueiredo e Silva et al., 2008).
(2) Lake Superior-type banded iron formations (Gross, 1980, 1993) are developed in passive-margin Proterozoic sedimentary rock successions. These contain some of the largest known BIF-hosted