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Geochemistry, Petrography, and Genesis of the Rezvan Iron deposit, Zagros Fold–Thrust Belt, Southern Iran

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Abstract

Banded iron formations (BIFs) are chemical sedimentary deposits that record interactions among seawater, hydrothermal fluids, volcanic activity, and redox conditions in the early Earth. The Rezvan iron deposit, located in the Zagros Fold–Thrust Belt of southern Iran, is hosted by volcano–sedimentary successions comprising carbonate, volcanoclastic, and intermediate volcanic rocks. This study presents an integrated geological, petrographic, mineralogical, and geochemical investigation of the deposit to constrain its genetic evolution and depositional setting. Mineralization occurs as stratiform to lens-shaped bodies and records a multistage evolution involving early hematite ± magnetite, subsequent pyrite ± chalcopyrite mineralization, and late supergene goethite–limonite alteration. Low concentrations of Al, Ti, Zr, and Sc indicate limited terrigenous input and support a predominantly chemical marine origin. Chondrite-normalized REE patterns show low total REE concentrations, slight LREE enrichment, weak negative Ce anomalies, and locally positive Eu anomalies, suggesting seawater-dominated deposition influenced by hydrothermal fluids. Integrated evidence indicates that the deposit initially formed through chemical precipitation of iron and silica in a volcanically active, shallow-marine basin, followed by hydrothermal sulfide overprinting and supergene alteration. The Rezvan deposit has an Algoma-type affinity; however, its carbonate- and sulfate-rich character and the absence of direct geochronological constraints preclude a definitive classification. A Proterozoic age is suggested by regional geological relationships, but the depositional age remains uncertain.

Keywords: Banded iron formation (BIF); Algoma-type; Hydrothermal fluids; REE geochemistry; Zagros Fold–Thrust Belt.

Introduction

Banded iron formations (BIFs) are chemical sedimentary rocks composed of alternating iron-rich and silica-rich layers that were deposited mainly in Precambrian marine environments. They represent the largest source of iron on Earth and provide important insights into the evolution of the atmosphere, oceans, hydrothermal systems, and early biosphere (Gross, 1980; Klein, 2005; Bekker et al., 2010; Mloszewska et al., 2023). The rhythmic alternation of iron- and silica-rich bands reflects variations in seawater chemistry, hydrothermal activity, biological productivity, and redox conditions during deposition (Frei et al., 2013; Bekker et al., 2010). Most BIFs were deposited during the Neoarchean to Paleoproterozoic (ca. 2.7–2.4 Ga), when extensive hydrothermal activity and progressive oxygenation of the oceans promoted large-scale precipitation of dissolved ferrous iron. After approximately 1.8 Ga, BIF deposition became relatively rare, partly in response to increasing atmospheric oxygen and sulfate

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concentrations. A second period of iron formation occurred during the Neoproterozoic, commonly associated with global glaciation and tectonic reorganization (Klein and Beukes, 1993; Cox et al., 2013; Klein, 2005; Ilyin, 2009; Mloszewska et al., 2023).

Based on their geological setting and associated lithologies, Precambrian iron formations have commonly been discussed in terms of Algoma-, Superior-, and Rapitan-type associations (Gross, 1980; Klein, 2005; Hagemann et al., 2016). Algoma-type BIFs are generally associated with volcanic and volcanoclastic rocks in tectonically active submarine environments and commonly show evidence of hydrothermal influence. In contrast, Superior-type BIFs are typically developed in relatively stable continental shelf settings and are commonly associated with carbonate rocks, quartzites, and shales (Gross, 1980; Klein, 2005; Hagemann et al., 2016). Rapitan-type iron formations are mainly associated with Neoproterozoic glacial successions and are commonly linked to major changes in oceanic redox conditions during and following glaciation. These associations provide useful genetic frameworks; however, their diagnostic value must be evaluated together with sedimentological, mineralogical, and geochemical evidence rather than relying on a single discriminant feature.

The genetic classification of the Rezvan iron deposit requires particular consideration because the presence of carbonate- and sulfate-rich lithologies, together with the possible Neoproterozoic age of the host succession, may also be compatible with a Rapitan-type iron-formation association. In contrast, the close spatial association of the iron-rich horizons with volcanic and volcanoclastic rocks and the evidence for hydrothermal input are more consistent with an Algoma-type affinity. Therefore, the present study does not assume a priori that the deposit belongs to either type; instead, the possible Algoma- and Rapitan-type affinities are evaluated using integrated geological, petrographic, mineralogical, and geochemical evidence. In the absence of direct geochronological and definitive sedimentological evidence for a glacial depositional setting, a Rapitan-type classification cannot be regarded as conclusive.

Recent studies have provided additional insights into volcano–sedimentary iron mineralization in Iran. Ardyani et al. (2025) investigated the Rezvan iron deposit northeast of Rafsanjan, within the Poshte Badam Block of the Central Iran Zone, and described volcano–sedimentary iron mineralization hosted by the Rizou Series. Their study highlighted the association of iron mineralization with volcanic–sedimentary rocks and hydrothermal processes and provided important regional evidence for the complexity of Precambrian iron mineralization in Iran.

Table 1. Comparison of the three BIFs types

BIFs Types	Formation Age	Tectonic Settings	Rock Assemblages	Significant Deposits/Regions
Algoma	Archean and Paleoproterozoic	Intracraton rfts or volcanic Arcs settings	Greywacke, fine-grained volcanic (clastic)rocks	Archean Yilgarn And Pilbara deposits
Suprior	Neoproterozoic and Paleoproterozoic	Passive continental margin	Sedimentary formations, such as carbonates rocks, quartz sandstones, and black shales.	Hamersley and the Transvaal deposits In Australia and South Africa
Rapitan	Early Neoproterozoic	Rifted basin	Glaciogenic Lithologies such as diamictites and dropstone layers	Urucum in the Mato Grosso State, Brazil, and Rapitan in The Northwest Territories Of Canada

Although the Rezvan deposit investigated by Ardyani et al. (2025) is geographically and geologically distinct from the deposit studied here in the Zagros Fold–Thrust Belt of southern Iran, their results provide a useful regional comparison for evaluating the geological setting and genetic evolution of the Rezvan iron deposit studied in this work.

Numerous studies have investigated the geology, mineralization, and genesis of BIF-hosted iron deposits worldwide and within Iran. In southern Iran, geological and geochemical studies have been carried out on iron deposits from Hormuz Island, Tang-e-Zagh, Ak-Kahur, and several other localities in Hormozgan Province, providing valuable information on the origin and evolution of Neoproterozoic iron mineralization. However, the Rezvan iron deposit has received little scientific attention.

Previous investigations in the Rezvan area have been limited mainly to regional geological mapping by the Geological Survey of Iran, lithological descriptions, and exploration activities conducted by Kavosh Iranian Mining Company. Although these investigations confirmed the presence of economically significant iron mineralization, no integrated geological, petrographic, mineralogical, or geochemical study has previously been conducted to determine the genesis and classification of the deposit. Consequently, the origin of the Rezvan iron deposit and its relationship to the recognized BIF types have remained uncertain.

To the best of the authors' knowledge, this study represents the first comprehensive investigation aimed at determining the genesis of the Rezvan iron deposit using an integrated dataset including field observations, petrography, ore mineralogy, whole-rock geochemistry, and rare earth element (REE) geochemistry. The objectives of this study are to: (1) characterize the geological, petrographic, mineralogical, and geochemical features of the deposit; (2) evaluate the origin of the iron mineralization using major elements, trace elements, and REE geochemistry; (3) assess the role of hydrothermal fluids and depositional redox conditions during ore formation; (4) develop a multistage genetic model for the Rezvan iron deposit; and (5) evaluate its genetic affinity with recognized types of banded iron formations, particularly Algoma- and Rapitan-type associations. The results provide a genetic interpretation of the Rezvan iron deposit and contribute to a better understanding of iron formation development within the Zagros Fold–Thrust Belt of southern Iran.

Geological Setting

The Rezvan iron deposit is located in the Zagros Fold–Thrust Belt of Hormozgan Province, southern Iran. The Zagros Orogen represents one of the major tectonic provinces of the Middle East and records a complex history of sedimentation, magmatism, deformation, and hydrothermal activity related to the convergence of the Arabian and Eurasian plates (Mouthereau et al., 2012; Vergés et al., 2019).

The study area comprises a volcano–sedimentary succession consisting mainly of carbonate, volcanoclastic, and intermediate volcanic rocks, together with intrusive intermediate to felsic rocks. The sedimentary succession includes microsparitic limestone, locally dolomitic layers, and lithic arenite, whereas the volcanic–subvolcanic rocks include andesite–trachyandesite, porphyritic microquartz monzosyenite, and microgranodiorite. The iron mineralization occurs within this volcano–sedimentary succession as stratiform to lenticular bodies and is locally interlayered with carbonate and volcanoclastic units. The mineralized horizons are generally conformable with the host-rock layering, indicating a strong stratigraphic control on the initial distribution of iron mineralization.

The iron-bearing horizons are exposed at several localities within the study area and are closely associated with volcanic and volcanoclastic rocks. The close spatial relationship between iron-rich layers and volcanic units, together with the occurrence of hydrothermal sulfide minerals and structurally controlled veinlets, indicates that hydrothermal activity played an important role in

the evolution of the mineralization. However, the relative timing of iron deposition and hydrothermal activity cannot be established unequivocally from the available field data alone.

The stratigraphic position and regional age of the host volcano–sedimentary succession remain uncertain because no direct geochronological data are currently available from the Rezvan iron formation or its immediate host rocks. Therefore, a precise depositional age cannot be assigned to the mineralization on the basis of the present dataset. The possible Neoproterozoic age of the host succession is discussed in the context of regional correlations; however, this interpretation should be considered tentative until supported by direct geochronological evidence.

Regionally, the occurrence of carbonate- and sulfate-bearing lithologies together with volcano–sedimentary rocks is relevant to comparisons with Neoproterozoic iron-bearing successions of southern Iran. In particular, the possible Neoproterozoic age of the host succession and the carbonate–sulfate-rich character of the mineralization warrant consideration of Rapitan-type iron formations in addition to the Algoma-type affinity suggested by the volcanic and hydrothermal association. Rapitan-type iron formations are commonly associated with Neoproterozoic glacial successions and major changes in oceanic redox conditions; however, no direct sedimentological evidence for a glacial depositional environment has yet been established at Rezvan. Therefore, regional correlation with Neoproterozoic iron-bearing successions remains a working hypothesis rather than a definitive age or genetic assignment.

The regional and local geological relationships are illustrated in Figure 1. The geological map shows the distribution of the principal volcanic, volcanoclastic, sedimentary, and intrusive lithological units and their spatial relationship with the iron mineralization. The mineralized horizons occur within the volcano–sedimentary succession and are locally controlled by bedding, fractures, and faults. The map also illustrates the structural framework of the study area and the relationship between the mineralized zones and the surrounding lithological units.

Subsequent tectonic deformation, low-grade metamorphism, hydrothermal alteration, and weathering modified the primary mineral assemblages. Pyrite and minor chalcopyrite occur locally within fractures and altered zones, indicating post-depositional hydrothermal activity. Later oxidation produced martite, goethite, and limonite, particularly along fractures, pore spaces, and near-surface zones.

Petrography

Petrographic investigations of representative samples (BHC1–BHC6, REZ2, REZ12, REZ35, REZ44, and REZ45) were conducted to characterize the host rocks, iron-rich horizons, and associated alteration features.

The igneous units include porphyritic microquartz monzonite and andesite–trachyandesite (Figure 3A, B). The microquartz monzonite is characterized by a porphyritic texture, with quartz and feldspar phenocrysts set in a fine-grained quartz–feldspar groundmass. The andesite–trachyandesite displays porphyritic to trachytic and hyalomicroclitic textures. Both igneous units locally contain secondary chlorite, sericite, carbonate, and quartz veinlets, indicating variable degrees of hydrothermal alteration. The spatial association of these altered volcanic–subvolcanic rocks with the mineralized horizons suggests that at least part of the alteration may be related to the hydrothermal system associated with iron mineralization, although a later regional alteration component cannot be completely excluded.

The sedimentary succession is dominated by microsparitic limestone and dolomite-bearing microsparitic limestone, together with minor lithic arenite (Figure 4A–D). The carbonate rocks consist mainly of fine-grained calcite and dolomite, with subordinate quartz, iron hydroxides, and opaque minerals. Their fine-grained and locally recrystallized textures are consistent with deposition in a relatively low-energy marine environment followed by diagenetic modification and localized oxidation.

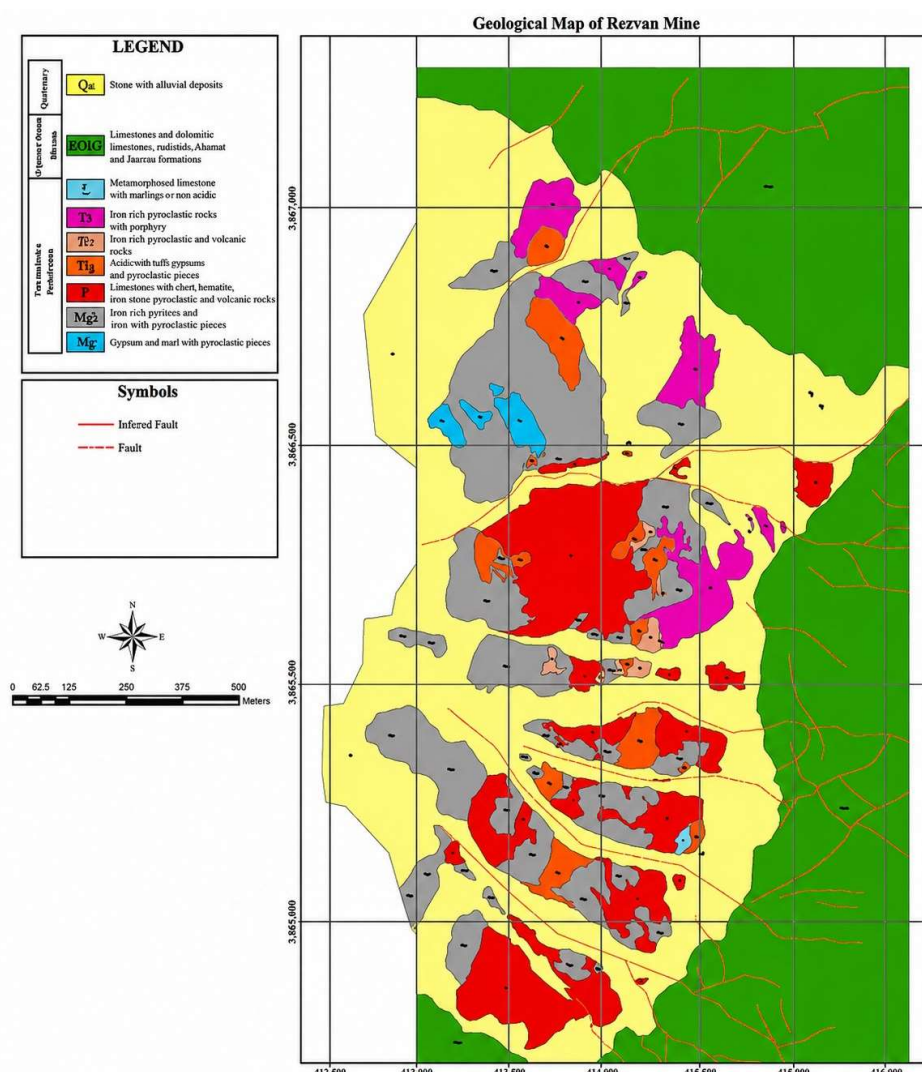


Figure 1. Simplified geological map of rezvan deposit showing major lithological units and structural features associated with iron mineralization(modified ater rezvan co,)

The iron mineralization occurs as stratiform to lenticular, laminated, and locally massive horizons interlayered with carbonate- and silica-rich layers. The iron-rich bands consist predominantly of hematite, with subordinate magnetite, martite, goethite, and limonite and minor pyrite and chalcopyrite. At the microscopic scale, hematite occurs as fine-grained aggregates as well as euhedral to subhedral and locally bladed (specular) crystals. Fine-grained hematite commonly forms continuous to discontinuous microbands that alternate with silica- and carbonate-rich bands, preserving the primary banded architecture of the iron formation.

Silica occurs mainly as fine-grained aggregates intergrown with hematite and as relatively continuous silica-rich bands. Where the silica is very fine-grained and cryptocrystalline, it is interpreted as chert-like material, whereas locally coarser and more crystalline silica occurs as quartz. The intimate intergrowth of hematite and silica, together with their rhythmic alternation, supports an original chemical sedimentary component in the formation of the iron-rich layers. Carbonate minerals locally occur between hematite- and silica-rich bands and form part of the primary to early diagenetic mineral assemblage.

The observed textures do not provide unequivocal evidence for primary microbial structures. Therefore, the possibility of biologically influenced iron precipitation cannot be established

solely from the available petrographic observations and requires additional sedimentological or microtextural evidence.

Post-depositional processes modified the primary mineral assemblage to variable degrees. Pyrite and minor chalcopyrite occur locally along fractures, veinlets, and within iron-rich horizons, and locally replace earlier iron oxides, indicating a later hydrothermal sulfide overprint. Magnetite is locally transformed to martite, reflecting oxidation of an earlier magnetite phase. Goethite and limonite occur mainly as xenomorphic phases along fractures, pore spaces, and altered zones and represent late supergene oxidation and hydration.

Overall, the petrographic relationships indicate a multistage evolution involving (1) primary chemical precipitation of iron and silica, producing laminated to banded iron-rich horizons; (2) carbonate deposition and early diagenetic modification; (3) hydrothermal alteration associated with volcanic–subvolcanic activity and subsequent sulfide mineralization; (4) oxidation and martitization of earlier iron oxides; and (5) late supergene formation of goethite and limonite. This petrographic evidence supports a chemically precipitated iron formation subsequently modified by hydrothermal and supergene processes.

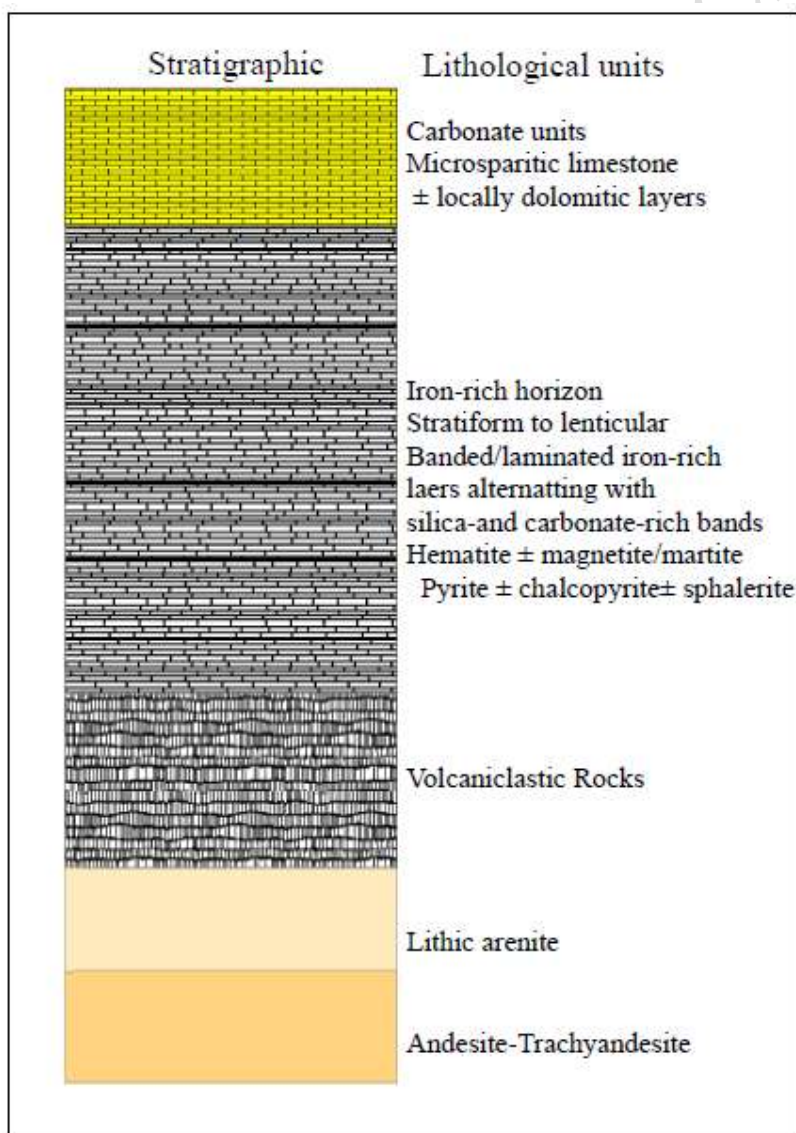


Figure 2. Schematic lithostratigraphic column of the Rezvan iron deposit

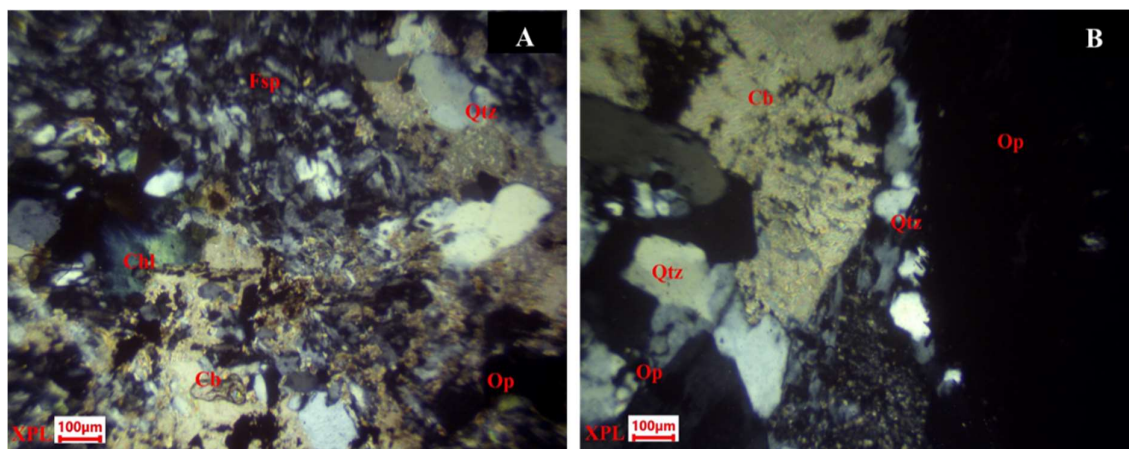


Figure 3. A. Porphyritic microquartz monzonite showing chloritic–sericitic alteration and quartz–carbonate vein filling, indicating intrusive magmatism followed by hydrothermal alteration. B. Trachytic to hyalomicrolitic porphyritic andesite–trachyandesite affected by sericitic–chloritic alteration and quartz–carbonate infilling, indicating volcanic activity and subsequent hydrothermal overprinting

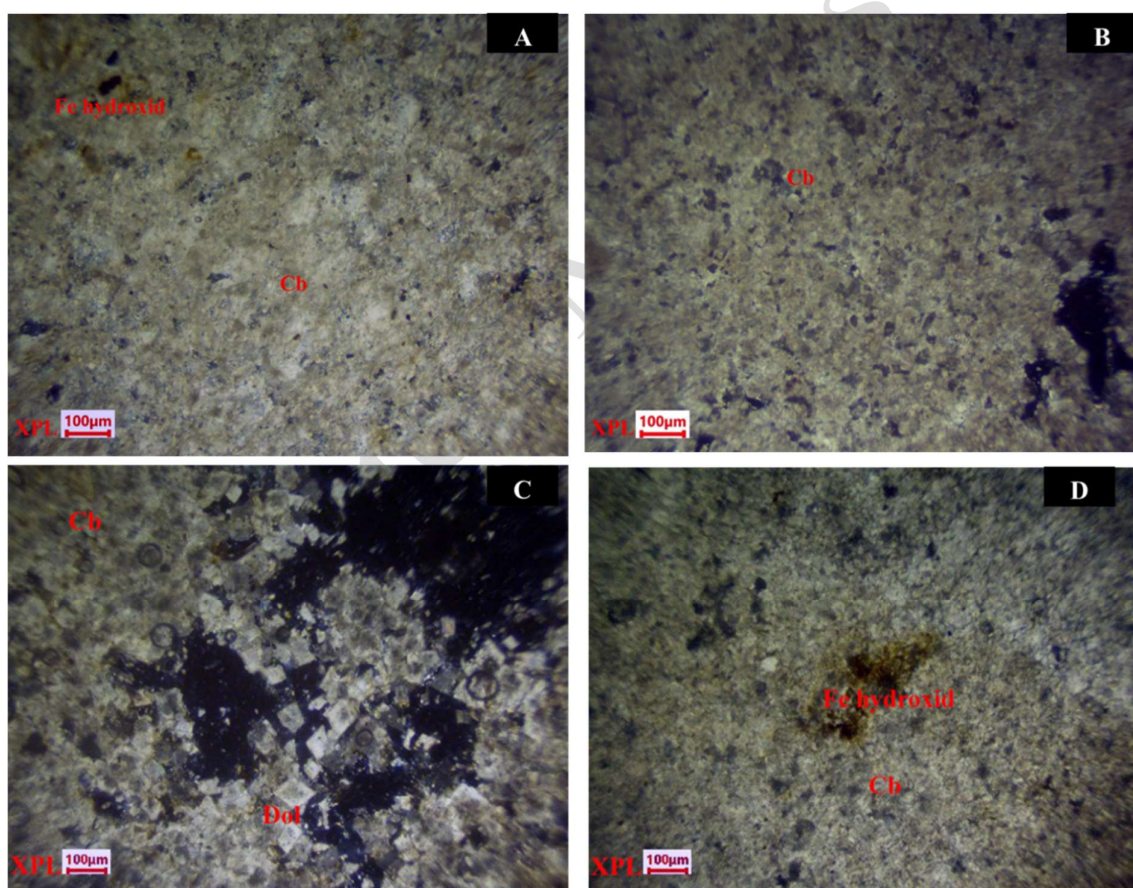


Figure 4. A. Microsparitic limestone composed predominantly of recrystallized carbonate with minor iron hydroxides and quartz, indicating low-energy carbonate deposition followed by diagenetic recrystallization and weak oxidation. B. Microsparitic limestone dominated by fine-grained crystalline carbonate with minor quartz and pore spaces. C–D. Dolomite-bearing microsparitic limestone associated with the iron-formation sequence, indicating deposition in a low-energy marine environment followed by localized diagenetic dolomitization and oxidation

Mineralogy

Mineralogical and ore-microscopic observations of the studied samples (Figure 5) reveal a complex assemblage of iron oxides, sulfides, and secondary iron hydroxides that records several stages of mineral formation and subsequent modification.

Hematite is the dominant iron oxide and occurs as fine-grained, disseminated, massive, laminated, and locally bladed or specular aggregates. Fine-grained hematite commonly forms the main constituent of iron-rich bands and is closely associated with silica and carbonate minerals. Bladed (specular) hematite occurs as relatively coarse, platy to euhedral–subhedral crystals, locally within altered and structurally controlled zones. Although its occurrence within the iron-rich horizons is compatible with an early iron-oxide generation, the petrographic evidence does not allow the bladed hematite to be unequivocally interpreted as a primary precipitate from seawater or hydrothermal fluids. Its texture may alternatively reflect recrystallization or hydrothermal modification of an earlier fine-grained iron-oxide precursor. Therefore, specular hematite is considered an early to intermediate hematite generation whose precise origin remains constrained by the available petrographic evidence (Figure 5A–C, G).

Magnetite occurs locally within the iron-rich assemblages and is partly transformed to hematite. The occurrence of martite, defined as hematite pseudomorphically replacing magnetite, provides evidence for a later oxidation stage affecting an earlier magnetite-bearing assemblage (Figure 5D, F). This relationship indicates that at least part of the hematite in the deposit formed through oxidation of pre-existing magnetite, rather than by direct primary precipitation.

Sulfide minerals are represented mainly by pyrite, with subordinate chalcopyrite and locally sphalerite. Pyrite occurs both as discrete grains and as a replacement phase associated with earlier iron oxides, indicating that sulfide mineralization postdated at least part of the oxide assemblage. The presence of chalcopyrite and sphalerite is particularly important because these minerals indicate a metal-bearing hydrothermal component and suggest that the mineralizing system was more complex than a simple oxide-facies BIF. In particular, the occurrence of sphalerite points to localized Zn-bearing hydrothermal fluids and supports a polymetallic hydrothermal overprint on the earlier iron-rich assemblage (Figure 5B, D, F, G).

Secondary iron hydroxides, mainly goethite and limonite, occur as xenomorphic phases within fractures, pore spaces, and along grain boundaries. They locally replace earlier iron oxides and sulfide minerals and are interpreted as products of late oxidation and supergene alteration. The sharp contact between goethite–limonite-rich zones and pyrite-bearing zones (Figure 5H) provides direct textural evidence for oxidation of pyrite and replacement by secondary iron hydroxides. This process may also have resulted in the release and redistribution of sulfur during weathering.

The observed textural relationships indicate that the replacement is predominantly from sulfide minerals to secondary iron hydroxides rather than the reverse. In particular, goethite–limonite locally replaces pyrite along fractures, pore spaces, and grain boundaries, whereas no clear petrographic evidence was observed for replacement of goethite–limonite by pyrite. Local replacement of chalcopyrite by chalcocite further supports supergene modification of the primary and hydrothermal sulfide assemblage (Figure 5B).

Comparable mineralogical relationships have been documented in several BIF-hosted iron deposits worldwide. In the Hamersley Province of Western Australia, BIF-hosted iron ores commonly contain martite, hematite, and goethite, and their formation has been linked to multistage hypogene hydrothermal alteration followed by supergene oxidation and hydration. In particular, magnetite-to-martite transformation and subsequent formation of goethite during supergene alteration have been documented in deposits such as Mount Whaleback, Mount Tom Price, Paraburdoo, and Hope Downs (Webb et al., 2008; Thorne, 2015).

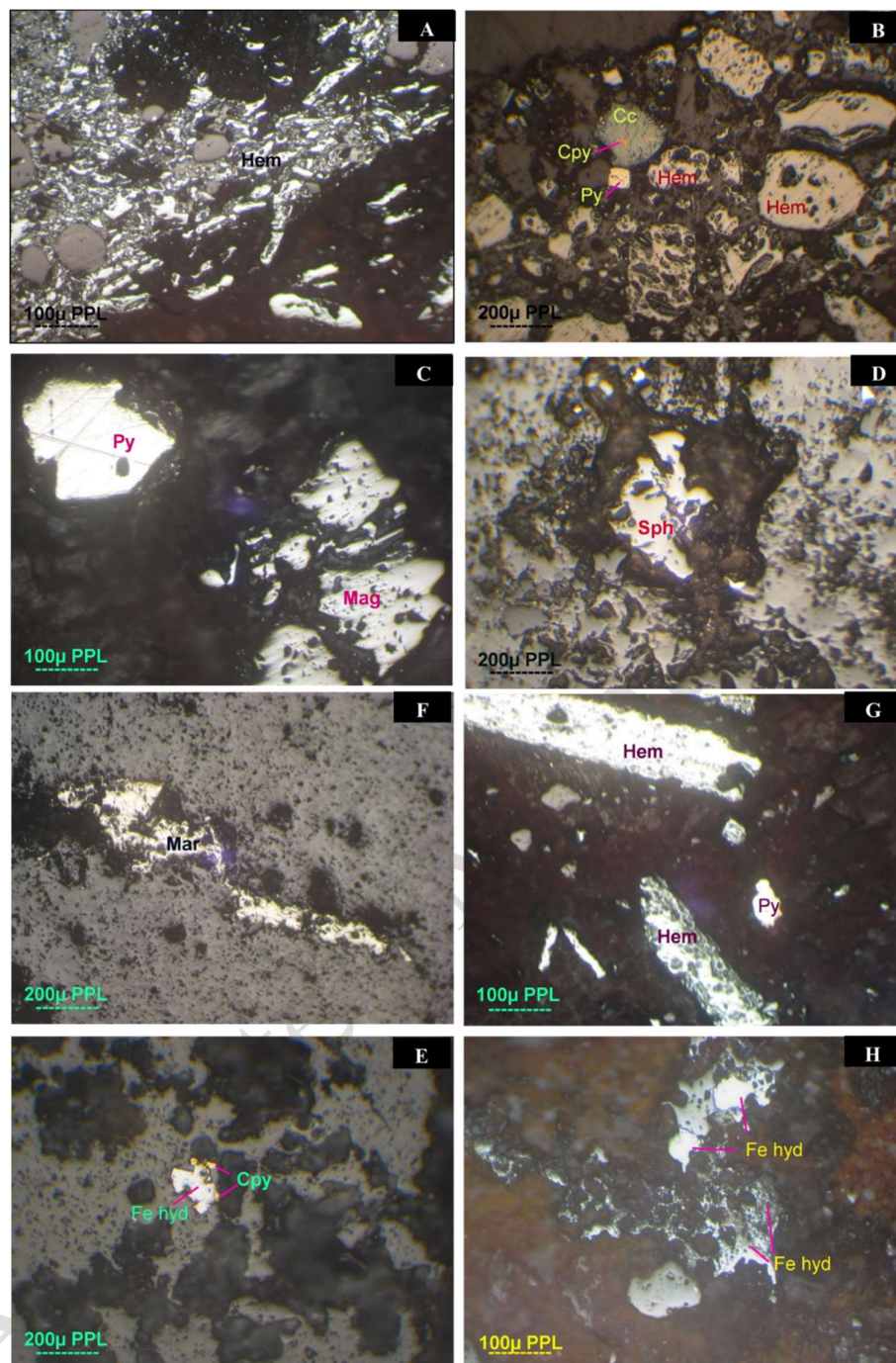


Figure 5. A. Bladed (specular) hematite occurring as relatively coarse platy crystals within the iron-rich assemblage. Its texture may reflect recrystallization or hydrothermal modification of an earlier iron-oxide precursor. B. Euhedral to subhedral hematite and pyrite, with local replacement of chalcopyrite by chalcocite, indicating supergene modification of an earlier sulfide assemblage. C. Goethite and limonite occurring as secondary xenomorphic phases associated with pyrite, hematite, and magnetite, indicating late oxidation and supergene alteration. D and F. Coexistence of martite and sphalerite. Martite records oxidation of pre-existing magnetite, whereas sphalerite indicates a localized Zn-bearing hydrothermal component. E. Goethite–limonite-dominated assemblage with minor hematite and rare chalcopyrite, indicating late supergene oxidation and secondary alteration. G. Hematite associated with pyrite $\pm$ chalcopyrite, showing a later sulfide overprint on the earlier iron-oxide assemblage. H. Sharp contact between a goethite–limonite-rich zone and a pyrite-bearing zone ($\sim 10\%$), indicating oxidation of pyrite and replacement by secondary iron hydroxides along fractures and pore spaces

Similar relationships have also been reported from BIF-hosted iron deposits of the eastern Indian Craton, where magnetite is partly transformed to martite and later hydrothermal and supergene processes produced hematite–goethite assemblages (Roy and Venkatesh, 2009). These examples demonstrate that the coexistence of hematite, martite, magnetite, and goethite may record successive hydrothermal, oxidation, and supergene processes rather than a single depositional stage. The mineralogical relationships observed in the Rezvan deposit are therefore consistent with a multistage post-depositional evolution, although the specific origin of the specular hematite and the relative timing of the hydrothermal and oxidation stages remain uncertain.

Taken together, the mineralogical and textural relationships indicate a multistage evolution rather than a simple oxide–sulfide–supergene sequence.

An early iron-oxide assemblage, dominated by fine-grained hematite with local magnetite, was subsequently affected by recrystallization and hydrothermal modification. A later hydrothermal stage introduced sulfide minerals, mainly pyrite with subordinate chalcopyrite and sphalerite, indicating locally reducing and metal-bearing fluids. Subsequent oxidation produced martite from pre-existing magnetite and finally generated goethite–limonite through supergene weathering and oxidation of iron oxides and sulfides.

These relationships support a genetic model involving an early chemical iron-formation component followed by hydrothermal overprinting, sulfide mineralization, oxidation, and supergene alteration. The volcanic–sedimentary association and hydrothermal overprint are compatible with an Algoma-type affinity; however, the occurrence of sphalerite, the carbonate–sulfate-rich character of the deposit, and the uncertainty regarding the precise origin of specular hematite indicate that the deposit should not be regarded as a simple or unequivocal Algoma-type BIF.

Ore Geology

The iron mineralization in the study area occurs as stratiform to lenticular ore bodies interlayered with volcanic–sedimentary and carbonate sequences. The mineralized horizons are generally concordant with the bedding of the host rocks, indicating a strong stratigraphic control on the distribution of the iron-rich layers. The preservation of laminated and banded textures further supports an important stratabound component to the mineralization (Gross, 1980; Klein, 2005).

The ore is dominated by hematite-rich horizons accompanied by subordinate magnetite, martite, goethite, limonite, and minor sulfide minerals, including pyrite and chalcopyrite. Iron-rich layers are commonly interbedded with carbonate- and silica-rich intervals and locally associated with volcanoclastic rocks. This association is consistent with chemical iron precipitation in a marine basin; however, the available field relationships do not by themselves establish that the volcanic rocks and iron mineralization were strictly syn-sedimentary.

In addition to the stratigraphic control, structural features played an important role in the subsequent modification of the mineralization. Bedding planes, fractures, and locally developed fault zones provided pathways for fluid circulation and locally host veinlets and sulfide–oxide assemblages. In these structurally controlled zones, the original iron-rich assemblage is locally overprinted by pyrite $\pm$ chalcopyrite and secondary iron hydroxides, indicating post-depositional hydrothermal modification and later oxidation (Figure 5G–H).

Although most mineralized horizons are approximately concordant with bedding, local structural overprinting is evident where sulfide-bearing veinlets and oxide–hydroxide assemblages occur along fractures and fault-related structures. Thus, the geometry of the deposit reflects both an original stratigraphic control and later structural modification. This distinction is important because the present distribution of mineralization cannot be attributed

exclusively to either a syn-sedimentary or a structurally controlled process.

Ore textures range from finely laminated and rhythmically banded structures to massive and locally disseminated iron-rich zones. In relatively preserved parts of the deposit, hematite-rich layers alternate with silica- and carbonate-rich bands, preserving the primary banded architecture. In strongly altered and structurally affected zones, this architecture is partly obscured by hydrothermal alteration, sulfide mineralization, oxidation, and supergene processes, resulting in heterogeneous iron enrichment and secondary mineral formation.

The close spatial association of the iron-rich horizons with volcanic–sedimentary rocks and the presence of hydrothermal mineral assemblages are consistent with a hydrothermally influenced marine depositional system. However, the available observations do not demonstrate that the volcanic activity was necessarily contemporaneous with the initial iron precipitation. The volcanic rocks may have provided a contemporaneous volcanic–hydrothermal setting, but a later hydrothermal event unrelated to the timing of volcanism cannot be excluded.

Overall, the ore geometry and textures indicate a multistage evolution involving an initial stratigraphically controlled iron-rich accumulation, followed by structural and hydrothermal modification and later supergene oxidation. The volcanic–sedimentary association and evidence of hydrothermal influence are compatible with an Algoma-type affinity, although these features alone are insufficient to establish a definitive Algoma-type classification.

Geochemistry

Geochemical investigations were conducted on representative iron ore samples collected from the Rezvan deposit. A total of 38 ore samples and 13 thin sections from seven different outcrops were selected based on minimal weathering and alteration. After pulverization using an agate mortar, whole-rock geochemical analyses were performed at ZarrAzma Laboratory using ICP–MS and XRF methods following four-acid digestion. The obtained data were used to evaluate elemental distribution patterns, depositional conditions, and mineralization processes within the studied deposit.

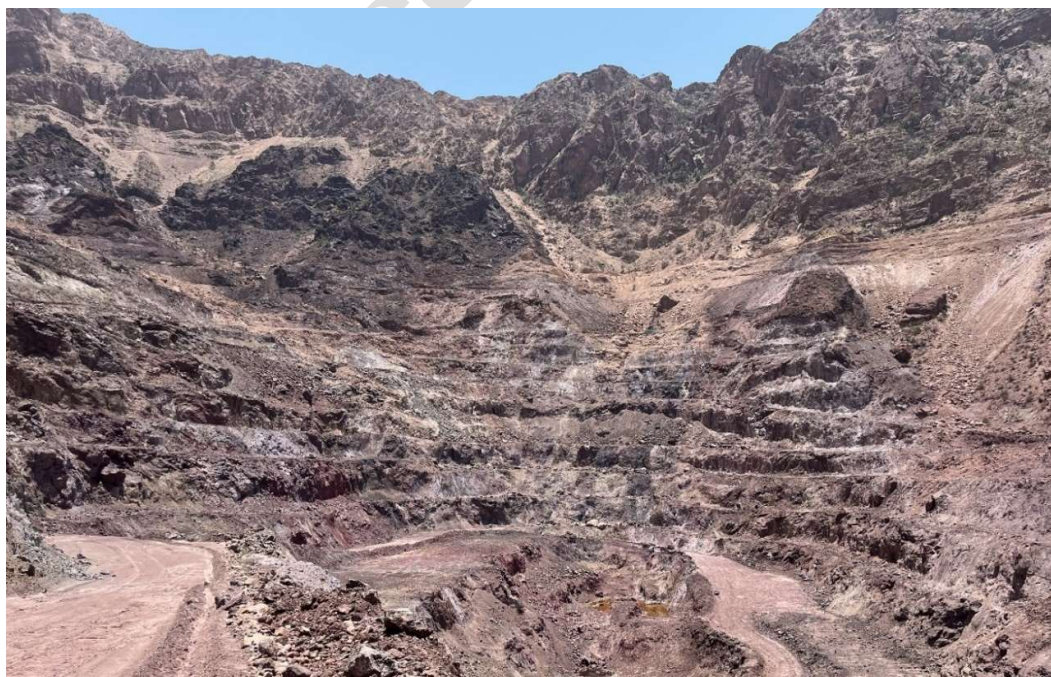


Figure 6. Overall view of the rezvan iron ore deposit. The ore body occurs as a stratiform layer between the host rock units

Chondrite-normalized REE patterns exhibit slight enrichment of light rare earth elements (LREEs) relative to heavy rare earth elements (HREEs), accompanied by overall low total REE concentrations ($<1\times$ chondrite-normalized values). Such characteristics are consistent with chemically precipitated iron formations deposited from seawater and support a sedimentary-marine affinity rather than direct magmatic derivation (Bau & Dulski, 1996).

Most samples display relatively low (La/Yb)_N ratios and nearly flat HREE patterns, suggesting that seawater was the principal source controlling REE distribution during deposition. The absence of pronounced HREE enrichment further argues against a significant direct magmatic contribution and instead supports dominant chemical sedimentation processes within a marine basin (Bau & Dulski, 1996; Planavsky et al., 2010).

Positive Eu anomalies are observed in several samples, particularly REZ-16 and REZ-41, where Eu values are significantly elevated relative to adjacent Sm and Gd concentrations. Europium enrichment is commonly associated with high-temperature hydrothermal fluids because Eu²⁺ remains stable under reducing and elevated-temperature conditions. Therefore, the positive Eu anomalies suggest a contribution from high-temperature hydrothermal fluids and are consistent with possible hydrothermal input during iron formation (Bau & Möller, 1993; Bekker et al., 2010).

In contrast, most samples show weak negative Ce anomalies characterized by slight depletion of Ce relative to neighboring La and Pr values. Cerium is readily oxidized from Ce³⁺ to insoluble Ce⁴⁺ under oxidizing marine conditions; therefore, negative Ce anomalies indicate deposition under relatively oxidizing seawater conditions. This feature is characteristic of oxide-facies BIF systems deposited in marine environments with relatively high Eh conditions (Bau, 1991; Planavsky et al., 2010).

Furthermore, relatively flat HREE patterns and low (La/Yb)_N ratios support a seawater-controlled depositional environment (Derry and Jacobsen, 1990; Muhling et al., 2023). Recent studies have demonstrated that REE and Y systematics provide robust constraints on hydrothermal activity, seawater chemistry, and redox evolution during BIF deposition (Muhling et al., 2023; Li et al., 2024).

Overall, the REE signatures suggest that the studied iron formation formed predominantly through seawater-controlled chemical precipitation in a volcanically active marine basin affected by hydrothermal fluid input. These characteristics are broadly consistent with an Algoma-type affinity, although they are not considered diagnostic of an Algoma-type classification by themselves.

Table 2. XRF Results of Iron Ore Samples from the Rezvan Iron deposit

Elements	REZ-33	REZ-1	REZ-36	REZ-29	REZ-24	REZ-6	REZ-3
SiO ₂	0.2	3	0.2	4.1	4.5	36	0.7
Al ₂ O ₃	<0.1	0.7	0.3	1.5	1.1	0.7	0.2
Fe ₂ O ₃	19.7	39.4	29.3	15.6	9.7	57.8	26.5
CaO	26.4	18	23.8	25.2	26.2	<0.1	23.3
MgO	1	0.4	0.6	0.5	0.6	0.3	0.5
Na ₂ O	0.8	1	1	0.8	0.9	1	1.1
K ₂ O	<0.1	<0.1	<0.1	0.4	0.5	0.2	<0.1
TiO ₂	<0.1	<0.1	<0.1	<0.1	<0.1	0.3	<0.1
MnO	0.1	<0.1	<0.1	0.1	0.2	<0.1	<0.1
P ₂ O ₅	0.7	<0.1	0.6	0.2	<0.1	0.4	<0.1
So ₃	37.7	26	34.1	35.9	37.4	1	33.3
L.O.I	13.1	10.7	9.8	15.2	18.6	2.3	13.9
K ₂ O	<0.1	<0.1	<0.1	0.4	0.5	0.2	<0.1

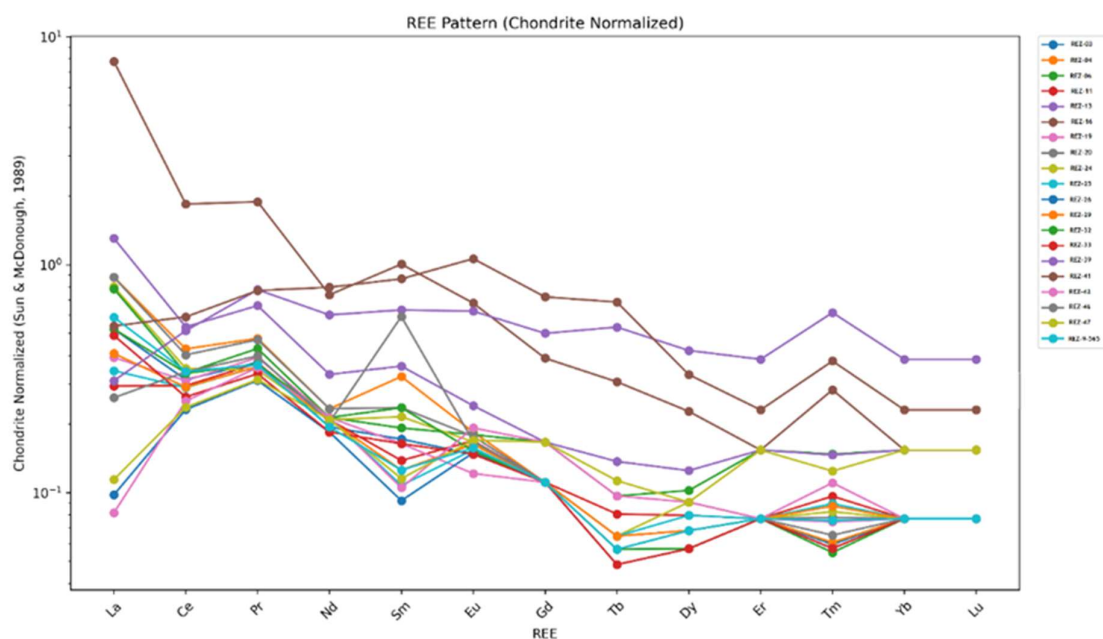


Figure 7. Chondrite-normalized REE pattern of iron ore samples from the Rezvan Iron deposit

Table 3. ICP-MS Results of Iron Ore Samples from the Rezvan Iron deposit

Elements	REZ-03	REZ-04	REZ-06	REZ-11	REZ-13	REZ-16	REZ-19	REZ-20	REZ-24	REZ-25
Al(ppm)	404	620	986	9530	29415	6043	9637	2769	4282	12494
As	24.5	6.1	97.1	13.6	91.4	133.33	8	16.3	16	22.7
Ba	191	374	293	106	3396	586	394	411	412	366
Ca	>10%	>10%	1963	7618	7915	10958	79939	92620	91255	71741
Co	4.3	3.7	3.2	51.9	105.6	100.7	37.6	20.5	33	29
Cr	32	10	146	88	185	79	53	38	23	29
Cu	7	21	138	52	204	618	57	28	91	84
Er	0.5	0.6	0.9	0.7	1.1	2	0.6	0.6	0.8	0.7
Fe	97846	36460	>10%	>10%	>10%	>10%	>10%	71702	85311	72556
K	948	855	2686	9162	22574	1936	10465	2959	3779	12253
Lu	0.1	0.1	0.2	0.1	0.2	0.2	0.1	0.1	0.1	0.1
Mg	376	279	194	11145	7549	4903	10632	1557	1931	4564
Mn	512	67	207	0.026	0.026	0.026	19067	4520	3084	7637
Mo	68.3	12.5	73.2	10.5	31.2	112	36.5	21.4	47.9	9.1
Na	133.3	133.3	552	160	429	134	334	133.3	133.3	181
Ni	1.33	2	1.33	4	13	12	2	1.33	1.33	3
P	191	537	2363	355	1487	3520	297	305	485	293
Pb	18	4	69	38	34	31	16	13	13	29
S	0.04	0.04	5153	8634	2042	13603	0.04	0.04	0.04	0.04
Sc	0.66	0.66	2.6	1.9	4.8	1.2	0.8	0.66	0.66	2
Sr	1374.9	3394.2	896.3	778.8	1183.4	2917.7	1009.1	1222.7	2752.8	6465.2
Ti	13.3	13.3	2080	306	1201	104	300	13.3	101	340
V	10	8	19	27	85	110	34	15	34	20
Y	7	7.3	9.8	10.7	11.6	17.3	7.8	7.9	7.4	8.3
Yb	0.51	0.52	1.27	0.83	1.26	2.43	0.64	0.67	0.71	0.77
Zn	14	5	30	315	122	313	67	40	22	33
Zr	6.66	6.66	33	20	53	11	18	7	10	20

Elements	REZ-26	REZ-29	REZ-32	REZ-33	REZ-39	REZ-41	REZ-42	REZ-46	REZ-47
Al	3113	10158	1334	913	56493	5487	6169	351	2090
As	51.1	26.7	133.3	38.9	34.4	39.2	7.7	8.9	11.3
Ba	191	522	382	361	375	330	123	151	1079
Ca	17939	88524	79823	>10%	7834	54503	29316	>10%	8787
Co	73.4	15.9	1.5	24.9	23.9	19.2	50.4	4	167.8
Cr	40	27	26	20	48	50	97	30	89
Cu	324	106	42	125	80	784	11	1	25
Er	0.5	0.6	0.5	0.5	3.7	2.9	0.8	0.6	0.8
Fe	>10%	78856	>10%	>10%	50452	89622	>10%	88334	>10%
K	3521	8973	852	1379	53939	2127	7867	888	2058
Lu	0.1	0.1	0.1	0.1	0.5	0.3	0.1	0.1	0.2
Mg	6529	2936	218	3036	17622	4484	7949	1634	0.026
Mn	0.026	2847	66	1474	5415	8531	0.026	1225	0.026
Mo	22.6	14.6	103.1	7.4	4.9	15.9	10.7	19	9.1
Na	193	108	414	133.3	387	169	271	137	133.3
Ni	3	1.33	1.33	10	5	10	1	1.33	87
P	181	470	942	2128	1224	238	437	933	141
Pb	88	10	11	9	2	96	28	11	116
S	28132	0.04	0.04	0.04	3879	0.04	0.04	0.04	10820
Sc	0.66	0.7	0.66	0.66	21	3.3	0.5	0.66	5.1
Sr	2269.7	1334.2	1693.2	1504.2	272.6	661.4	389.9	1007.2	475.6
Ti	13.3	235	13.3	13.3	9984	77	31	13.3	13.3
V	4	19	4	174	98	11	17	25	13
Y	7.4	8	7.1	7.1	34	32.4	9.9	7.7	9.8
Yb	0.49	0.72	0.47	0.49	5.3	3.25	0.95	0.56	1.07
Zn	4152	116	16	25	67	80	126	265	481
Zr	13	18	18	7	104	9	14	6.66	11

Major and Trace Element Geochemistry

Major element geochemistry reveals considerable variations in Fe_2O_3 (9.7–57.8 wt%) and SiO_2 (0.2–36 wt%), reflecting the characteristic alternation of iron-rich and silica-rich bands typical of BIF deposits. The inverse correlation between Fe_2O_3 and SiO_2 suggests alternating precipitation of iron oxides and silica under fluctuating physicochemical and redox conditions within the depositional basin (Gross, 1965; Klein & Beukes, 1993).

The very low concentrations of Al_2O_3 and TiO_2 suggest limited terrigenous clastic input and support predominantly chemical precipitation in a marine setting (Klein & Beukes, 1993; Murray, 1994). Similarly, low concentrations of detrital-sensitive trace elements such as Zr, Sc, and Ti are consistent with limited siliciclastic input into the depositional basin.

The elevated CaO (18–26 wt%) and SO_3 (26–38 wt%) concentrations indicate substantial carbonate- and sulfate-bearing components within the mineralized horizons. Petrographic observations should be used to determine whether these sulfate phases occur as primary sedimentary components or as secondary vein- and fracture-filling minerals such as gypsum and/or anhydrite. If the sulfate minerals are predominantly fracture- and vein-controlled, their abundance is interpreted as a post-depositional feature rather than a primary component of BIF deposition.

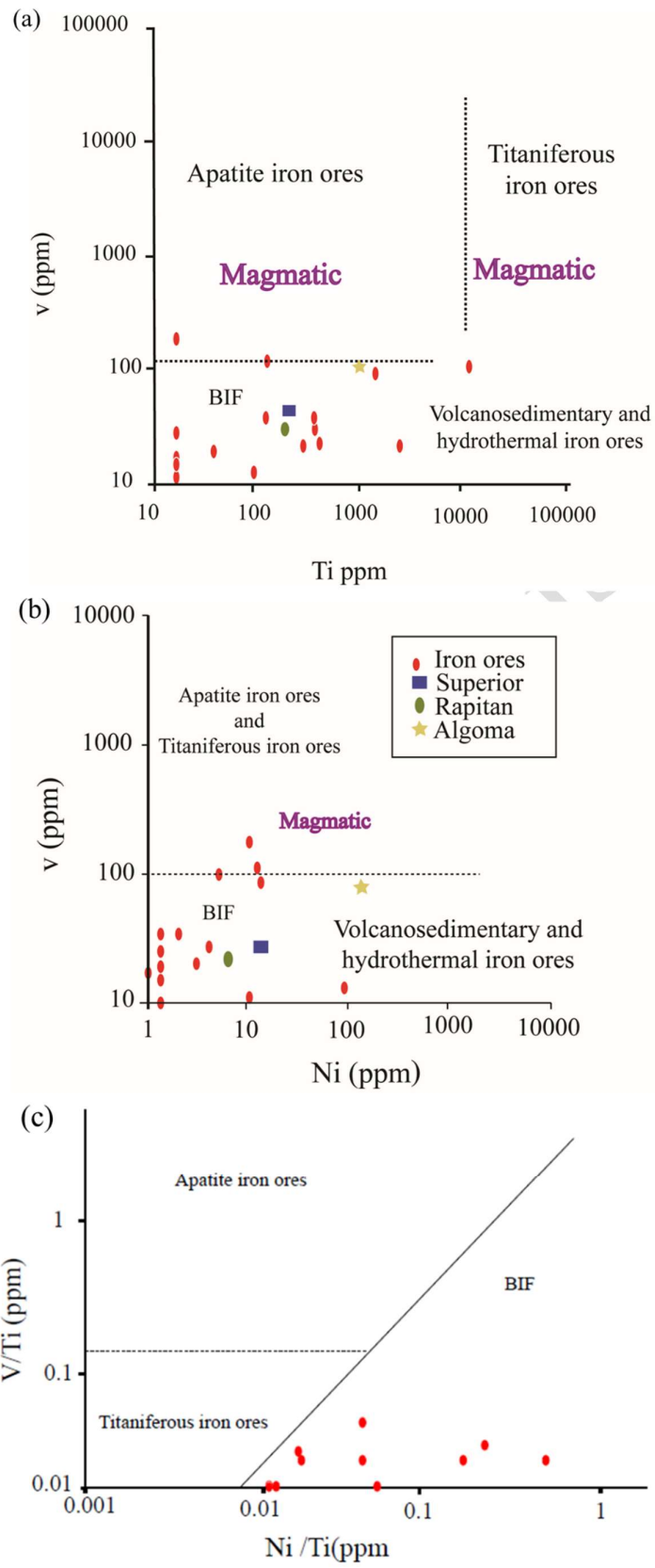


Figure 8. The plots of iron ores from rezvan iron ores on geochemical diagrams. a) V vs Ti, b) V s Ni c) V/Ti vs Ni/Ti(after Loberg and Horndahl, 1983)

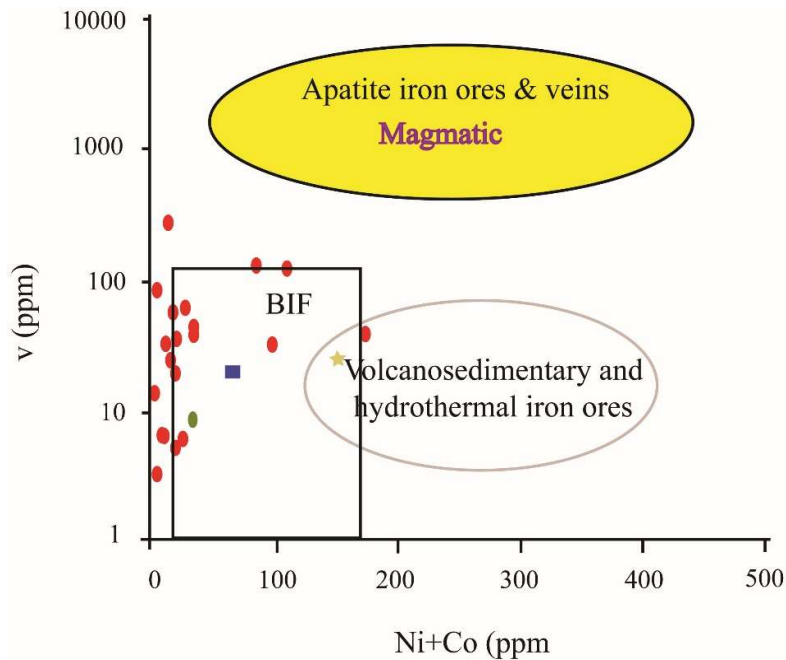


Figure 9. V versus (Ni + Co) discrimination diagram for the Rezvan iron ore deposits. The diagram indicates a predominantly hydrothermal influence on the studied samples; however, this diagram is not considered diagnostic of BIF type and is therefore used only as supporting geochemical evidence. (Løberg and Hørndahl, 1983)

Low MnO contents further support deposition under predominantly oxidizing marine conditions because manganese enrichment is generally associated with more reducing depositional environments. In addition, enrichment of sulfur in some samples, together with the occurrence of pyrite and minor chalcopyrite, reflects post-depositional hydrothermal sulfide mineralization and localized reducing conditions during later stages of mineral evolution. Transition-metal distributions also support hydrothermal influence on mineralization. Elevated concentrations of Pb, Zn, Mo, and locally Cu suggest circulation of metal-bearing hydrothermal fluids through permeable structural pathways including fractures and bedding planes. These patterns are consistent with a contribution from hydrothermal processes to metal redistribution and enrichment, whereas hydrogenous input appears to have been comparatively limited.

Collectively, the major-, trace-, and rare earth element data indicate chemical precipitation of iron in a hydrothermally influenced marine basin associated with volcanic activity, followed by post-depositional hydrothermal modification and supergene alteration. These characteristics support an Algoma-type affinity, but the classification should be considered provisional and evaluated together with the petrographic, sedimentological, and mineralogical evidence.

Discussion

Origin of the Iron Formation

The integrated geological, petrographic, mineralogical, and geochemical evidence indicates that the Rezvan iron deposit formed through the interaction of chemical sedimentation and hydrothermal processes within a volcanically active marine basin. Several characteristics, including the stratiform geometry of the mineralized horizons, their close association with volcanic and volcanoclastic rocks, limited terrigenous input, and evidence of hydrothermal

influence, support an Algoma-type affinity. However, the available evidence does not permit a definitive classification of the deposit as an Algoma-type BIF.

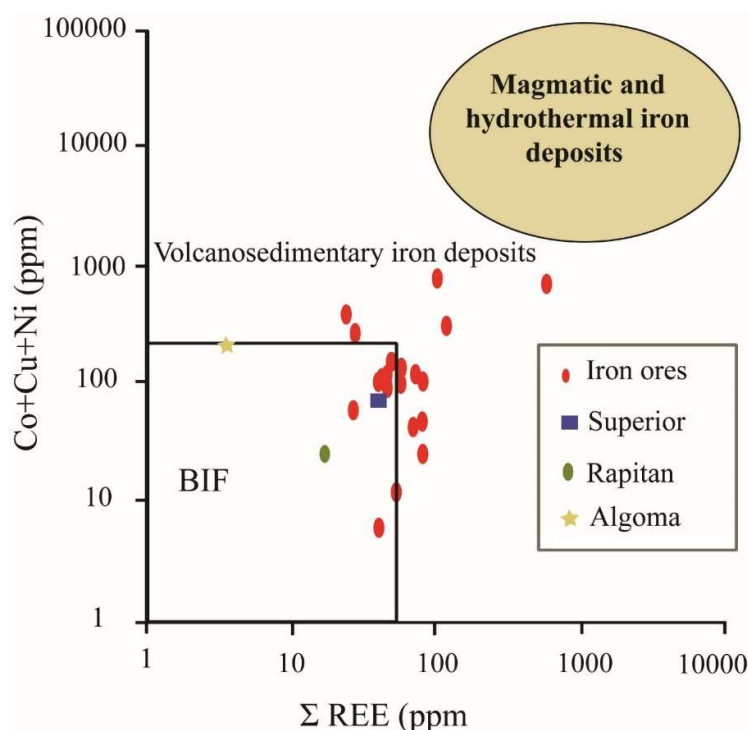
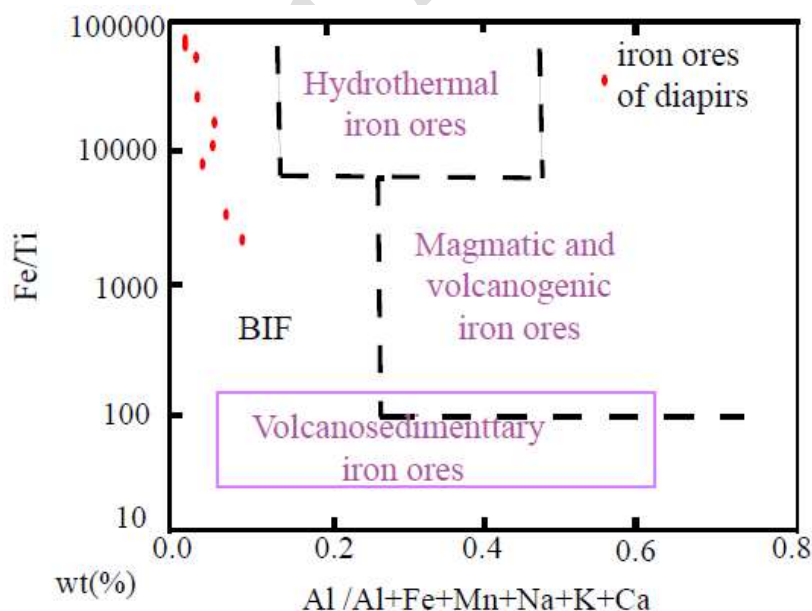


Figure 10. Plot of iron ores from rezvan iron ore deposits. On geochemical diagram of Co+Cu+Ni vs. Σ REE diagram (after Klein and beukes, 1993)



Interpretation of Figure11. indicates that the studied samples display very high Fe/Ti ratios and low $Al/(Al+Fe+Mn+Na+K+Ca)$ values, reflecting the dominance of chemical precipitation processes and the negligible contribution of detrital materials. The clustering of the samples within the hydrothermal iron deposit field suggests that the iron-rich units were formed under the influence of submarine hydrothermal fluids. These geochemical characteristics are consistent with the genetic model of Algoma-type banded iron formations (BIFs) developed in Archean volcano-sedimentary environments. (after Bostrom, 1973)

The criteria used to evaluate the genetic affinity of the Rezvan iron formation include its stratigraphic and lithological setting, association with volcanic–volcaniclastic rocks, depositional geometry, degree of terrigenous input, mineralogical assemblage, and major- and trace-element as well as REE characteristics. The stratiform to lenticular geometry of the iron-rich horizons, their close association with volcanic and volcaniclastic lithologies, low concentrations of detrital-sensitive elements such as Al, Ti, and Zr, and evidence for hydrothermal input represented by positive Eu anomalies collectively support an Algoma-type affinity. However, the relatively high carbonate and sulfate contents, together with the occurrence of sphalerite and the absence of direct geochronological and definitive sedimentological evidence for either an Archean–Paleoproterozoic or Neoproterozoic depositional age, prevent a definitive classification. Accordingly, these criteria are used to assess genetic affinity rather than to assign the deposit unequivocally to a specific BIF type. The stratiform geometry of the ore bodies, preservation of primary banding, and close spatial association with volcanic and volcaniclastic rocks are consistent with characteristics commonly described for Algoma-type BIFs, which typically occur in volcano–sedimentary successions and may be associated with submarine hydrothermal systems (Gross, 1980; Klein, 2005; Bekker et al., 2010). Unlike Superior-type BIFs, which commonly developed on relatively stable continental shelves and may display extensive lateral continuity, the studied mineralization occurs as localized iron-rich horizons closely associated with volcanic units and structural pathways (Gross, 1980; Hagemann et al., 2016).

Petrographic observations further support an early chemical-sedimentary component in the formation of the iron-rich horizons. Euhedral hematite, preserved banded textures, and the rhythmic alternation of iron-rich and silica-rich layers are consistent with chemical precipitation under fluctuating physicochemical and redox conditions. The coexistence of iron oxides, silica, and carbonate minerals further suggests deposition within a chemically heterogeneous marine basin influenced by hydrothermal activity (Klein, 2005; Bekker et al., 2010).

No direct geochronological data are currently available for the Rezvan iron formation; therefore, the age of iron deposition and the temporal relationship between iron formation, volcanic activity, and subsequent hydrothermal events remain uncertain. This age uncertainty limits a definitive correlation of the Rezvan iron formation with a specific BIF type, particularly Algoma- or Rapitan-type iron formations. Consequently, the proposed genetic model is based primarily on field relationships, petrographic and mineralogical characteristics, and geochemical evidence, and should be considered a working model until direct geochronological and additional basin-scale sedimentological constraints become available.

Hydrothermal Contribution

Geochemical evidence suggests a significant hydrothermal contribution to the mineralization. Positive Eu anomalies observed in several samples, particularly Z-16 and Z-41, are consistent with an input from high-temperature hydrothermal fluids, as Eu^{2+} can remain mobile under reducing and elevated-temperature conditions (Bau & Möller, 1993; Danielson et al., 1992; Bekker et al., 2010).

The generally low concentrations of Al_2O_3 , TiO_2 , Zr, and Sc indicate limited terrigenous clastic input and support precipitation from seawater rather than detrital sedimentation (Klein & Beukes, 1993; Murray, 1994). Furthermore, low total REE abundances and relatively flat HREE patterns are typical characteristics of hydrothermally influenced marine iron formations (Derry & Jacobsen, 1990; Bau & Dulski, 1996).

The close association between iron mineralization and altered volcanic rocks suggests that volcanic centers acted as conduits for hydrothermal fluid circulation.

Table 4. REE Analysis Results of Iron Ore Samples from the Rezvan Iron deposit

REE	REZ-03	REZ-04	REZ-06	REZ-11	REZ-13	REZ-16	REZ-19	REZ-20	REZ-24	REZ-25
La	24	40	20	12	55	312	15	32	32	13
Ce	32	54	32	18	80	478	24	54	49	21
Pr	2.88	3.93	3.09	2.71	4.91	16.97	2.85	3.7	3.23	2.66
Nd	12	15.3	13.8	11.9	21.3	60.8	12.6	15.1	12.7	11.5
Sm	2	2.4	2.2	2.1	3.4	7.6	2.2	2.4	2.1	2.2
Eu	0.67	1.26	0.75	0.54	1.4	3.92	0.33	0.92	0.84	0.42
Gd	1.52	1.9	1.85	1.74	2.48	6.99	1.67	1.82	1.72	1.59
Tb	0.2	0.2	0.3	0.2	0.3	0.7	0.2	0.2	0.2	0.2
Dy	0.7	0.7	1.2	1	1.7	3.8	0.8	0.8	0.7	0.8
Er	0.5	0.6	0.9	0.7	1.1	2	0.6	0.6	0.8	0.7
Tm	0.1	0.1	0.2	0.1	0.2	0.2	0.1	0.1	0.1	0.1
Yb	0.51	0.52	1.27	0.83	1.26	2.43	0.64	0.67	0.71	0.77

REE	REZ-26	REZ-29	REZ-32	REZ-33	REZ-39	REZ-41	REZ-42	REZ-46	REZ-47
La	3	17	45	23	11	17	2	12	2
Ce	6	25	48	30	19	33	5	16	7
Pr	2.13	2.67	3.07	2.42	4.72	5.43	2.32	3.12	2.18
Nd	10	11.5	11.8	10.7	25	24.8	11.4	12.8	10.1
Sm	1.9	2.1	2.2	1.9	6.2	8.2	2.2	2.1	2.1
Eu	0.36	0.49	0.92	0.64	2.47	3.38	0.41	2.31	0.45
Gd	1.54	1.68	1.57	1.52	6.44	10.94	1.98	1.73	1.75
Tb	0.2	0.2	0.2	0.2	0.9	1.3	0.3	0.2	0.3
Dy	0.6	0.8	0.7	0.6	6.6	8.5	1.2	0.7	1.4
Er	0.5	0.6	0.5	0.5	3.7	2.9	0.8	0.6	0.8
Tm	0.1	0.1	0.1	0.1	0.5	0.3	0.1	0.1	0.2
Yb	0.49	0.72	0.47	0.49	5.3	3.25	0.95	0.56	1.07

These fluids may have contributed dissolved Fe and Si to the basin, where mixing with ambient seawater could have promoted precipitation in response to changes in temperature, pH, and oxidation state (Isley, 1995; Bekker et al., 2010).

Redox Conditions and Basin Evolution

The REE signatures and mineral assemblages provide important constraints on the redox evolution of the depositional system. Weak negative Ce anomalies together with the abundance of hematite are consistent with relatively oxidizing marine conditions during the early stages of iron deposition (Bau, 1991; Planavsky et al., 2010). Under such conditions, dissolved Fe^{2+} supplied by hydrothermal fluids could have been oxidized and precipitated as ferric iron minerals.

The redox conditions can be further evaluated using the normalized REE ratios and anomalies. The Ce/Ce ratio, calculated as $\text{CeN}/(\text{LaN} \times \text{PrN})^{1/2}$, is used to evaluate the behavior of Ce relative to neighboring REEs. The weak negative Ce anomalies observed in the studied samples are consistent with preferential removal of Ce from the dissolved marine reservoir under relatively oxidizing conditions. In addition, Eu/Eu, calculated as $\text{EuN}/(\text{SmN} \times \text{GdN})^{1/2}$, provides an indicator of hydrothermal influence. The positive Eu anomalies observed in selected samples, particularly REZ-16 and REZ-41, are consistent with an additional contribution from high-temperature hydrothermal fluids. Therefore, the combined behavior of

Ce/Ce* and Eu/Eu*, together with the normalized REE patterns, suggests that the Rezvan iron formation developed under predominantly oxidizing marine conditions with a significant hydrothermal contribution.

However, mineralogical evidence demonstrates that the system subsequently evolved toward more reducing conditions. Pyrite locally replaces earlier iron oxides, indicating circulation of reduced sulfur-bearing hydrothermal fluids during post-depositional stages. The occurrence of pyrite $\pm$ chalcopyrite therefore records hydrothermal sulfide mineralization superimposed upon the primary oxide-facies assemblage.

In addition, the presence of martite (hematitized magnetite) indicates oxidation of an earlier magnetite phase, whereas secondary goethite and limonite formed through later supergene weathering processes. These mineralogical relationships suggest a complex evolutionary history involving repeated fluctuations in redox conditions (Hagemann et al., 2016).

Accordingly, the Rezvan iron deposit is interpreted to record a multistage evolution comprising: (1) primary marine chemical precipitation of iron and silica; (2) syn-sedimentary or early hydrothermal input associated with volcanic activity; (3) post-depositional hydrothermal sulfide overprinting, represented mainly by pyrite $\pm$ chalcopyrite; (4) oxidation and partial replacement of magnetite by hematite, producing martite; and (5) late supergene alteration, resulting in the formation of goethite and limonite.

Collectively, these geological, mineralogical, and geochemical characteristics support an Algoma-type affinity for the Rezvan iron formation, particularly because of its volcanic–hydrothermal association. However, the absence of direct geochronological and definitive sedimentological constraints prevents a conclusive classification.

The high SO₃ concentrations may reflect either primary sulfate-bearing sedimentation or secondary sulfate formation associated with post-depositional alteration and oxidation of sulfide minerals. However, the available dataset does not permit an unequivocal distinction between these alternatives. Therefore, the sulfate-rich character of the deposit should be considered when evaluating its genetic affinity and should not be used alone as evidence for an Algoma-type classification.

Genetic Model

Based on the integrated geological, petrographic, mineralogical, and geochemical evidence, a multistage genetic model is proposed for the Rezvan iron deposit.

Stage 1—Primary chemical precipitation: Iron and silica were initially introduced into a shallow-marine basin, where chemical precipitation produced stratiform, iron-rich and silica-rich layers. The preservation of primary banding, low concentrations of detrital-sensitive elements, and the predominantly oxide-rich mineral assemblage support a chemical-sedimentary component in the formation of the iron-rich horizons.

Stage 2—Hydrothermal input and early modification: Contemporaneous volcanic activity was accompanied by hydrothermal fluid circulation. Hydrothermal fluids may have contributed Fe and Si and modified the chemistry of the depositional system. Positive Eu anomalies in selected samples are consistent with a hydrothermal contribution. Mixing between hydrothermal fluids and seawater may have promoted additional precipitation and modification of the primary iron-rich layers.

Stage 3—Post-depositional hydrothermal overprint: Subsequent tectonic deformation provided pathways for hydrothermal fluids along fractures, bedding planes, and fault zones. This stage resulted in partial remobilization and redistribution of iron and introduced sulfide mineralization, mainly pyrite $\pm$ chalcopyrite. The local replacement of earlier iron oxides by pyrite indicates that at least part of the sulfide mineralization postdates the primary oxide assemblage.

Stage 4—Oxidation and martitization: During later oxidation, pre-existing magnetite was partly converted to hematite, producing martite. This process modified the original oxide assemblage while locally preserving the primary banded architecture.

Stage 5—Supergene alteration: Subsequent uplift and surface weathering promoted oxidation and hydration of primary and hydrothermal minerals, resulting in the formation of secondary goethite and limonite and contributing to localized secondary iron enrichment.

This multistage history indicates that the present mineral assemblage of the Rezvan deposit cannot be attributed to a single depositional event. Rather, the deposit records an initial chemical-sedimentary iron-formation stage followed by hydrothermal modification, sulfide mineralization, oxidation, and supergene alteration.

The possible relationship of the Rezvan iron formation to Rapitan-type iron formations should also be considered. Rapitan-type iron formations are generally associated with Neoproterozoic glacial successions and major changes in oceanic redox conditions. The carbonate- and sulfate-rich character of the Rezvan mineralization and the regional occurrence of Neoproterozoic successions provide a basis for considering a possible Rapitan-type affinity. However, the absence of direct geochronological constraints and definitive sedimentological evidence for a glacial depositional setting prevents a firm assignment to the Rapitan type. Therefore, an Algoma-type affinity is currently supported by the volcanic–hydrothermal association, but a Rapitan-type or other Neoproterozoic iron-formation affinity cannot be completely excluded.

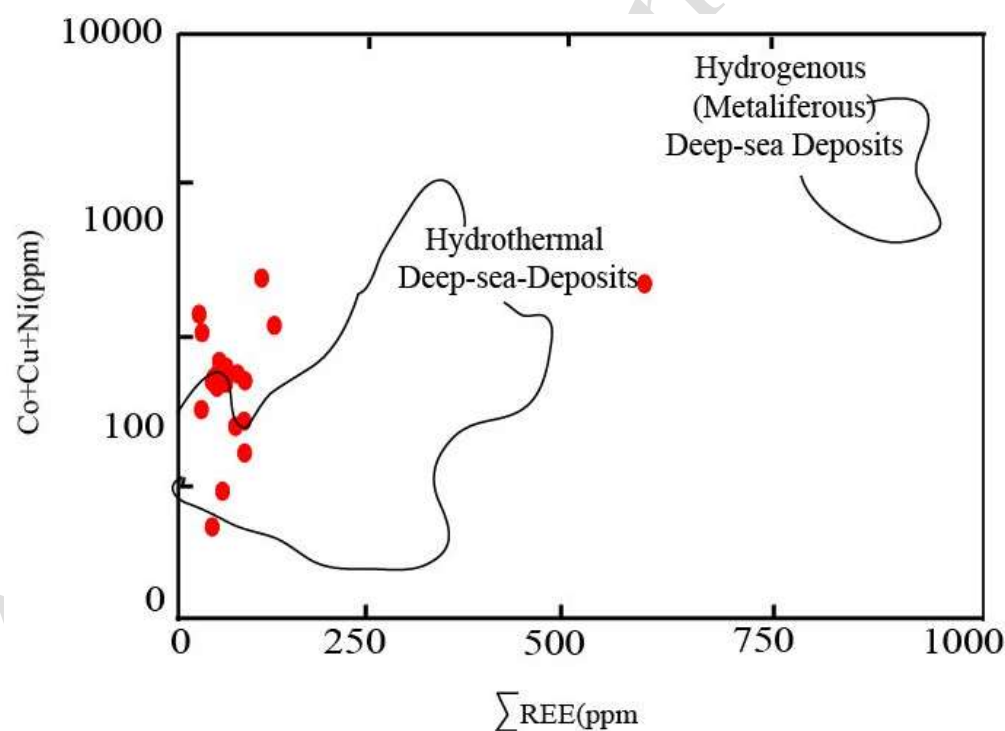


Figure 12. shows the distribution of the samples on a plot of ΣREE versus $(\text{Co} + \text{Cu} + \text{Ni})$. Most samples plot within or close to the field of deep-sea hydrothermal sediments, suggesting a significant hydrothermal contribution to the geochemical characteristics of the iron-rich units. However, this diagram alone does not uniquely constrain the depositional environment or genetic type of the deposit. The relatively low REE concentrations compared with $\text{Co} + \text{Cu} + \text{Ni}$ may indicate a greater contribution of hydrothermal fluids relative to hydrogenous processes. These results are therefore consistent with, but do not independently demonstrate, a hydrothermally influenced depositional system and support the interpretation of an Algoma-type affinity when considered together with the geological, petrographic, mineralogical, and other geochemical evidence

Conclusions

The Rezvan iron deposit, located within the Zagros Fold–Thrust Belt of southern Iran, is hosted by volcano–sedimentary successions composed of carbonate, volcanoclastic, and intermediate volcanic rocks.

Petrographic, ore-microscopic, and mineralogical investigations reveal a multistage evolution involving primary chemical precipitation of iron oxides, subsequent hydrothermal alteration and sulfide mineralization (pyrite $\pm$ chalcopyrite), followed by oxidation and supergene alteration that produced goethite, limonite, and martite.

The preservation of banded iron-rich layers, low abundances of detrital-sensitive elements (Al, Ti, and Zr), and the close spatial association with volcanic rocks are consistent with chemical iron deposition in a hydrothermally influenced marine basin with limited clastic input.

Petrographic characteristics of the associated volcanic and intrusive lithounits, including andesite–trachyandesite and porphyritic microquartz monzonite, together with their hydrothermal alteration features, indicate a close association with the hydrothermal system. However, the available data do not allow these rocks to be conclusively identified as the direct source of Fe, Si, or other elements. Their spatial association and alteration characteristics suggest that they may have acted as conduits and/or contributed to hydrothermal fluids during the evolution of the iron mineralization.

Chondrite-normalized REE patterns of the iron-rich samples are characterized by slight LREE enrichment, relatively flat HREE patterns, positive Eu anomalies in selected samples, and weak negative Ce anomalies. These features are consistent with chemical precipitation influenced by hydrothermal input and suggest relatively oxidizing conditions during the early stages of iron deposition. However, a direct comparative evaluation of REE patterns between the iron mineralization and the associated volcanic lithounits was not possible because REE data for the host volcanic rocks were not available in the present dataset.

Positive Eu anomalies suggest a contribution from high-temperature hydrothermal fluids, whereas weak negative Ce anomalies and the dominance of hematite are consistent with relatively oxidizing conditions during the early stages of mineralization. The occurrence of pyrite $\pm$ chalcopyrite records a subsequent hydrothermal stage involving relatively reducing fluids.

Fractures, bedding planes, and fault zones acted as pathways for hydrothermal fluid circulation and contributed to the post-depositional redistribution of iron and sulfide minerals. The high SO_3 concentrations may reflect either primary sulfate-bearing sedimentation or secondary sulfate formation associated with post-depositional alteration and oxidation of sulfides. However, the available dataset does not permit an unequivocal distinction between these possibilities.

The integrated geological, petrographic, mineralogical, and geochemical evidence indicates that the Rezvan iron formation shares several characteristics with Algoma-type BIFs, particularly its close association with volcanic–sedimentary rocks and evidence of hydrothermal influence. However, this affinity should not be regarded as definitive. The carbonate- and sulfate-rich character of the deposit, together with the absence of direct geochronological and basin-scale sedimentological constraints, leaves open the possibility of a Rapitan-type or other Neoproterozoic iron-formation affinity. Supergene modification of a pre-existing Fe-rich sediment cannot also be completely excluded.

Overall, the Rezvan iron deposit is interpreted to record a complex multistage history involving primary chemical precipitation of iron and silica, subsequent hydrothermal modification and sulfide mineralization, oxidation and martitization, and late supergene alteration. This interpretation indicates that the present mineral assemblage and geochemical characteristics reflect several superimposed processes rather than a single depositional event.

Authors' Contributions

Alireza Jafari Rad: Conceived and designed the study, collected data, analyzed and interpreted the data, and drafted the manuscript.

Melika Aminolroaya: Conceived and designed the study, collected data, analyzed and interpreted the data, and drafted the manuscript.

Gholamreza Ghadami: Contributed to the study design, analyzed the data, and critically revised the manuscript for important intellectual content.

Seyyed Mohammad Mirhosseini: Collected and interpreted the data, and revised and edited the final manuscript.

Ali Solgi: Collected and interpreted the data, and revised and edited the final.

All authors have read and approved the final version of the manuscript and agree with its submission to this journal.

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