Abstract

At Wadi Hamama area two types of iron ores (BIF and apatite-specularite veins) are hosted by volcanogenic metaclastics and trondhjemite, respectively. The BIF bands show conformable relationship with the enclosing volcanogenic metagreywacke and metaconglomerate of Late Proterozoic age. Field and petrographic investigations revealed the presence of some distinct types of banding in the studied BIF ore, namely macrobanding and calamina cyclothem on the megascopic scale that grade to meso- and microbanding on the microscopic scale.

Mineralogically, the Wadi Hamama BIF ore (HBIF) pertains to the oxide facies only with frequent jasper nodular clasts and microbands. Three subfacies of HBIF oxide facies are distinguished as hematite, hematite-magnetite and magnetite subfacies according to the hematite/magnetite ratio. It is evident that fine dusty hematite was formed as a precipitate and then recrystallized into platy specular hematite. This was followed by crystallization of magnetite idiomorphic porphyroblasts during diagenesis and low-grade regional metamorphism at reducing conditions. Overgrowth of idiomorphic magnetite over specular hematite suggests crystallization by incomplete sub-solidus reduction of hematite during late- and early-stage diagenesis (anadiagenesis and anchimetamorphism, respectively). Positive
correlation between $\text{Al}_2\text{O}_3$ and $\text{TiO}_2$ suggests that some clastic materials were incorporated with the chemical sediments. Although field and mineralogical constraints suggest that HBIF belongs to the Algoma-type deposits, major and trace element average compositions is geochemically intermediate between Algoma- and Lake Superior-types.

Previous works suggested that the apatite-specularite veins of Wadi hamama area resulted from the shearing of the BIF bands. The present workers contradict such hypothesis because frequent unsheared veins containing undeformed specularite are present as discordant bodies in a trondhjemite host distant from the HBIF ore. It is believed that P-rich hydrothermal fluids released from the granitoids were responsible for iron leaching from the BIF bands in the metavolcanoclastics.

**Keywords:** BIF; specularite; iron leaching; Eastern Desert; Egypt

**Introduction**

Banded iron-formation are present throughout the geological record from Middle Archean to Early Phanerozoic. James (1983) classified the distribution of the world’s iron-formation into four age groups, Middle Archean (3.5-3 Ga); Late Archean (2.9-2.6 Ga); early Proterozoic (2.5-1.9 Ga) and Late Proterozoic to Early Phanerozoic (0.75-0.45 Ga). The definition of iron-formation was the concern of many authors (James, 1954 & 1966; Gross, 1965 & 1980; Trendall and Blockley 1970; Brandt et al., 1972; Kimberly, 1978). The first definition was introduced by James (1954) as follows “...a chemical sediment, typically thin bedded or laminated, containing 15 % or more iron of sedimentary origin, commonly but not necessarily containing layers of chert”. Gross (1965) used the term “iron-formation” in a broad sense to include all stratigraphic units of layered, bedded or laminated rocks that contain 15 % or more iron in which the iron minerals are commonly interbanded with quartz, chert or carbonate and the banded structure of ferruginous rocks confirms in pattern and attitude with the bedding of the adjacent rocks. Due to this banded and/or laminated structure,
the iron-formations are called banded iron-formations (Trendall and Blockley, 1970) and the recommended abbreviations of iron-formation and banded iron-formation are “IF” and “BIF”, respectively (Gross, 1980). Thickness variations of the IF bands are arranged according to a hierarchical order of their scales into: macroband (0.5 to 15 m), “Calamina” cyclothem (80 mm to 50 cm), mesoband (2 mm to 80 mm), microband (0.2 to 2 mm) and submicroband (<200 mm) as defined by Trendall (1965 & 1973), Trendall and Blockley (1970) and Ewers and Morris (1981).

Genetically, the BIFs are divided into two main types: Algoma and Lake Superior (Gross, 1970 & 1980). Algoma-type shows close association between the iron facies and the volcanic rocks which implies a volcanic source of iron whereas the deposits of this type are commonly regarded as being exhalative in origin (Gole and Klein, 1981; Maynard, 1983). In the Lake Superior type, volcanic rocks are not always directly associated with the IFs (Morey 1983; Trendall and Morris, 1983).

In Egypt, metamorphic iron ores of Late Proterozoic age are widely distributed in the central Eastern Desert of Egypt. They were recorded in thirteen occurrences between latitudes 25º 12' and 26º 30' N (with the exclusion of Wadi Hamama occurrence that are studied for the first time in detail by the present authors). Two genetic models have been proposed for the Egyptian BIFs: a sedimentary origin during the accumulation of the Precambrian geosynclinal or shallow shelf sediments (e.g. El Manawi, 1991 & El Aref et al., 1993), and volcanogenic origin related to submarine hydrothermal activities which are developed in an island arc environment (e.g. Sims and James, 1984; Basta et al., 2000). El-Habbak and Soliman (1999) also mentioned that the Egyptian BIFs, in general, and El-Dabbah BIF in particular, have been deposited from iron brought into the basin of deposition by river water fluxes with very little contribution from hydrothermal activity. Recently, El-Habbak (2004) presented a new study on the BIF of Umm Nar area and concluded that the iron formations practiced both
regional and thermal metamorphic events. He (op.cit.) distinguished the regional signature of the BIF from the thermal one, where the latter contains skarn minerals such as grossular garnet and diopside.

In the present paper, the Wadi Hamama BIF (HBIF) will be studied in detail through their field occurrence, mega- and microscopic fabrics, in addition to the geochemical characteristics in order to clarify the nature and mode of formation. Also, the paper will try to clarify if there is any genetic linkage between this BIF ore and apatite-specularite veins in the area. Such two contrasting types of iron mineralizations are known in other localities in the world but no definite genetic linkage has not yet postulated to explain the genesis of such veins.

**Analytical Procedures**

Thirteen samples from the least deformed parts of Wadi Hamama BIF were analyzed for their major and trace element composition using the X-ray fluorescence technique (XRF). The measurements of major elements were achieved in the Kemi Lab housed at the Geological Survey of Denmark and Greenland (GEUS) in Copenhagen, Denmark. Glass discs were prepared by adding a sodium tetraborate flux to the ignited portion of the powdered sample. Loss on ignition (L.O.I) was determined by gravimetry. Only Na and Mg were analyzed by atomic absorption. Concerning the trace elements, they were analyzed directly on pressed powder pellets at the Department of Petrology, Geological Institute of Copenhagen University in Denmark by XRF using a Philips PPW 1400 spectrometer and the techniques of Norrish and Chappell (1977). Results were corrected for background interference from tube and sample spectral lines, and for matrix variation (using the major element composition). USGS standards (G-2, AGV-1, W-1, BCR-1, PCC-1, DTS-1, W-2, DNC-1, BIR-1) were used for calibration. AGV-1 was run as unknown that can be compared with the compilation of
Govindaraju (1994). Some representative electron microprobe analyses of jasper associating the BIF were carried out at the Geological Institute of Copenhagen University. The machine is a JEOL superprobe 733 equipped with four wavelength-dispersive spectrometers. Operating conditions were 15 kV acceleration voltage, 15 nA beam current and a beam diameter of 1-3 $\mu$m. A set of natural and synthetic standards (silicates and oxides) was used for calibration and the measuring time for each element was 20 seconds. The obtained data was then corrected using a ZAF correction program.

The Wadi Hamama banded iron-formation (HBIF)

Field observations

The iron-rich bands of HBIF are associated with the metavolcaniclastic rocks which make up the main outcrops of the southern side of Wadi Hamama area (Fig. 1). They also occur as isolated fragments enclosed in the mafic metavolcanics of Wadi Abu Gerida in the northern part of the area.

A simplified stratigraphic section of the metavolcaniclastics is compiled through many traverses along these rocks in different directions. The measured section elucidates that the BIF occurs within a sequence of metatuffs with some volcanogenic metagreywacke and metaconglomerate. Generally, the iron bands are concentrated mainly in two horizons, occupying the central portion of the section and are separated by metagreywacke (Fig. 2). The lower horizon and upper horizons have stratigraphic thicknesses of about 85 m and 17 m respectively. They consist of alternations of iron bands and metatuffs with subordinate intercalations of metagreywacke in the lower part of each horizon.

(Figs. 1 & 2 should be placed here)

The different varieties of metatuffs are gradationally intercalated with each other and their contacts with the alternated iron bands are sharp. The metatuffs and associated iron bands are mostly hard and highly resistant, but at the highly eroded outcrops, they are
shattered parallel to the bedding planes. The meta-greywackes are generally hard, massive, some varieties show graded bedding and they are medium-grained with dark greenish grey colour due to the presence of chlorite.

The iron ore mainly occurs as separate thin bands and less frequent in dense thick bands. Bedding is the most dominant primary structure. Generally, two scales of banding are noticed in the field, namely macrobanding and calamina cyclothem, in order of decreasing scale (Fig. 3). The macrobanding consists of alternations of iron bands of reddish colour and volcaniclastic bands of greenish colour. The iron bands range in thickness from 50 cm to 2 m while the volcaniclastics are often thicker. On the other hand, the calamina cyclothem is a cyclic alternation of different mesobands of iron and metatuffs with regular and sharp contacts (Fig. 4a). Occasionally, the iron bands are slightly folded by drag faulting (Fig. 4b). Due to their high resistance, they form standing ledges between the metatuffs in the highly weathered outcrops (Fig. 4c). The iron bands and the associated metatuffs are subjected to deformation leading to the development of some major and minor structures. At some stratigraphic levels, the metatuffs possess a well-developed foliation parallel to the lithologic contacts. Both the volcaniclastics and the associated BIF are highly intersected by fractures of different magnitudes and directions in addition to faulting in some instances. Many concordant and discordant veins and veinlets of secondary quartz and carbonates dissect some of the iron bands and their metatuffs (Fig. 4d).

(Figs. 3 & 4 should be placed here)

Megascopically, the micro- and mesobanding are the most predominant types of banding in the studied BIFs. They are attributed to the difference in colour, thickness, composition and percentage of the mineral constituents. Four main types of mesobands are distinguished, namely iron-rich band, iron-jasper band and jasper-rich band with subordinate, mostly ferruginated, tuff bands (Fig. 5). The mesobands display a thickness range from few
millimeters up to 5 cm and some mesobands show sharp contacts with the volcanogenic metaclastics (Fig. 5a). They may be massive or laminated on the millimeter scale (microbanded) or of tens of micrometers (sub-microbanded) with variable thicknesses of jasper micro- and mesobands (Fig. 5b&c). The contact between these bands is often sharp but in some few instances it appears gradational. Bedding and lamination are the most striking primary (syn-sedimentary) structures are observed in the iron mesobands such as bedding, lamination, graded bedding, minor folds and faults (Fig. 5c). Slumping, pinch and swell structures are also displayed by some specimens (Fig. 5d). Generally, the mesobands are straight or contorted, continuous or discontinuous.

(Fig. 5 should be placed here)

**Petrography of Wadi Hamama BIF (HBIF)**

Mineralogically, hematite and magnetite are the dominant constituents of the studied BIF from Wadi Hamama area. So, it belongs mainly to the oxide facies following the scheme developed by James (1954) for the classification of BIF facies. Hematite and magnetite occur either in separate bands or in various proportions in a single band. According to the relative abundance of both minerals, the oxide facies of Hamama BIF is subdivided into three subfacies: Hematite subfacies, hematite-magnetite subfacies and magnetite subfacies. These subfacies is variable in proportion from outcrop to another, but the hematite-rich subfacies is generally dominating allover the mapped area.

**i) Hematite subfacies**

On the megascopic scale, this type is fine-grained, has brownish red colour and is composed of rhythmic alternation of iron-rich and jasper micro- and mesobands together with sporadic tuff bands. The contact between these bands is generally sharp and even, especially between the iron and jasper bands. The gradational contact is developed between iron and tuff
bands. Syn-sedimentary deformational structures are quiet common, such as micro-faulting and brecciation. Pinch-and-swell structures are sometimes observed.

Microscopically, this subfacies consists essentially of hematite and jasper with or without magnetite and scarce pyrite cubes. In hematite-rich microbands, hematite occurs as very fine specularite laths (microspecularite) that are either randomly or moderately oriented. When the microspecularite laths coalesce together, they form separate sub-microbands that giving the samples the laminated and foliated appearance (Figs. 6a&b). The magnetite is variable in size and exhibits idiomorphic to subidiomorphic outlines. The finer magnetite is the most dominant one and the coarser generation is scarcely developed. Although the magnetite crystals are not common, they are variable in amount. They range from scarce disseminations in the hematite groundmass to dispersed crystals, which lead to the development of alternated specularite and specularite-magnetite micro and submicrobands. The magnetite is partially to extensively martitized mainly along the crystal boundaries and/or the octahedral (111) planes. Little pyrite cubes of variable sizes are randomly distributed in the jasper-dominated bands and rarely in the hematite-rich bands. The large crystals are almost completely oxidized while the minute disseminations are fresh (Figs. 6c&d).

Jasper occurs in separate bands or in the form of nodules. The jasper nodules are commonly enclosed in the iron-rich bands. They are mostly aligned and stretched parallel to the bedding planes. They show variation in size, amount and distribution. Jasper nodules are either randomly separated or arranged in groups. Some nodules are free of hematite or enclose scarce randomly dispersed ultra-fine hematite inclusions (hematitic dust) (Figs. 6e). Other nodules are enriched in the microspecularite lathes which are concentrated at the core of nodules or distributed allover them (Fig. 6f). Nodules with oriented microspecularite laths are rare. Some of the specularite-rich bands enclose magnetite-rich fragments with variable amount of magnetite. Magnetite is concentrated at the core of the fragment or randomly
distributed (Figs. 6g&h). The bands are dissected by secondary cracks that are filled by later silica.

(Fig. 6 should be placed here)

**ii) Hematite-magnetite subfacies**

In handspecimen, this type is medium-grained, grey in colour with reddish tint and is composed of alternated meso- and microbands, namely iron-rich, jasper and volcaniclastic bands. The contact between most of these bands is gradational.

The microscopic investigation revealed that this subfacies consists of magnetite and hematite in variable proportions associated with jasper. Some samples are highly contorted (Fig. 7a) and the magnetite occurs as idiomorphic to subidiomorphic crystals. It shows noticeable variation in grain size as ranges from very fine crystals through medium-grain size in some bands to coarse porphyroblastic crystals (Figs. 7b&c). Some of the magnetite porphyroblasts are zoned as indicated by remarkable differences in colour that may be indicate variation in Ti content (Fig. 7d). The magnetite crystals are randomly embedded in a finer microspecularite-rich band but others are concentrated in certain laminae forming separate magnetite-rich micro- and submicrobands. Some of the magnetite cut across the microspecularite laths. This crosscutting relationship is obvious between the larger generations of magnetite and the oriented microspecularite laths (Fig. 7c). The oriented magnetite crystals are not uncommon and associated mainly with the foliated microspecularite laths (Fig. 7e&f). In general, the magnetite is slightly to moderately martitized. Hematite is represented mainly by fine microspecularite laths that are mostly oriented parallel to the foliation planes and are segregated in some samples to form a monomineralic submicrobands. Some hematite is recrystalized to coarser specularite laths that are arranged in a variolitic pattern on the wall of some veinlets.
Different types of fragments are observed in the mesobands of this subfacies. The volcaniclastic rock fragments are the most dominant type. Magnetite-rich and hematite-rich fragments are present as well.

Jasper is the most predominant gangue mineral. It is coarser in size than its equivalent in the hematite subfacies. It occurs in the form of nodules which coalesced together to form jasper micro- and submicrobands. The mesobands of the hematite-magnetite subfacies are dissected by silica and carbonate veinlets which include chlorite and micas too.

(Fig. 7 should be placed here)

iii) Magnetite subfacies

This subfacies is the least in abundance comparable to the hematite-rich subfacies. The magnetite subfacies can be divided into two assemblages as follow:

a) Magnetite-jasper- (hematite) assemblage:

It is composed of magnetite, jasper and subordinate amounts of hematite. These constituents show rhythmic alternation between iron-rich, jasper-rich and iron-jasper meso and microbands. The iron-rich bands are mainly dark grey to black in colour and are strongly magnetic. The jasper bands, alternating with such iron bands of dark appearance, are characterized by their glittering red colour. Syn-sedimentary microfaulting are noticed in these bands.

Magnetite is the most dominant iron mineral in most of the bands. It shows idiomorphic to subidiomorphic outlines and exhibits variation in the crystal size. The minute magnetite crystals are coalesced in some jasper bands to form iron spherules (Fig. 8a). The growing magnetite porphyroblasts are dominant and commonly enclose jasper inclusions from the matrix. Some medium-sized magnetite crystals are aggregated around large fragments from the jasper matrix leading to the formation of coarser porphyroblastic crystals.
Magnetite overgrowth on some platy specular hematite crystals is also noticed (Fig. 8c).

Hematite is scarce, being represented by three forms, hematite dust, microspecularite laths and coarser platy specularite crystals. The hematite dust occurs as dissemination in jasper bands and fragments. The microspecularite laths are commonly segregated to form nearly separate submicrobands. They are oriented parallel to the primary bedding and are cut by larger magnetite crystals.

Jasper nodules and magnetite-rich nests are located in the magnetite-bearing jasper bands (Fig. 8d). The rock is dissected by sub-parallel cracks that are nearly perpendicular to the primary banding.

b) Magnetite-chlorite assemblage:

It is the least common iron assemblage. It is medium to slightly coarse in size and exhibit nearly iron-rich massive mesobands that alternate mostly with the metatuffs. The contact between them is usually gradational. The iron ore mesobands are strongly magnetic and enclose chlorite-rich bundles.

(Fig.8 should be placed here)

On the microscopic basis, magnetite is the only iron mineral. It is embedded in a chlorite-rich matrix that contain also subordinate quartz. The magnetite crystals are idiomorphic to subidiomorphic nearly equant crystals. Generally, they are randomly dispersed but some crystals are coalesced to exhibit a weak alignment parallel to the primary bedding (Fig. 8e). They show different degrees of martitization. Manganese oxides and hydroxides are commonly occurring in the form of colloformed crystals in veinlets (Fig. 8f). At the gradational contact between the volcaniclastic bands and the magnetite-rich bands, magnetite
shows reduction in grain size towards the metatuffs. This could represent fining upward, which is common in cyclic sedimentation.

The tuff bands enclose some coarser porphyroblastic magnetite. Most of these crystals exhibit intense martitization at the core than at the rim, forming core-and-rim structure (Fig. 8g). Some of the magnetite porphyroblasts enclose fragments from the matrix. Some of the tuff bands also enclose magnetite-rich nests where the enclosed magnetite crystals are finer in size than those in the iron-rich bands (Fig. 8h).

**Diagenetic and metamorphic history of iron minerals**

Magnetite and hematite are the main constituents of the banded iron formation of Wadi Hamama along with the gangue minerals are represented by silica, mainly jasper with less dominant chlorite. The secondary minerals include martite, carbonates, silica, mica and Mn-oxides and hydroxides.

*i) Hematite*

Hematite is present in significant amounts in the studied BIF. It preferably oriented parallel to the foliation planes. According to the morphology of the hematite crystals, they are classified into three main types, namely hematite dust, microspecularite laths and coarse platy specularite.

The hematite dust appears to be the earliest iron oxide in the paragenetic sequence. It is ultra-fine, up to 1 µm in length with much finer width. These small needle crystals were observed as inclusions in some jasper bands and nodules or fragments. Dusty hematite of such form is interpreted in many literature to be primary depositional or diagenetic in origin (Ayres, 1972; Dimroth and Chauvel, 1973; Acharya et al., 1982; Han, 1982).

The microspecularite occurs as microplaty laths range in length from 6 to 20 µm and 2 to 10 µm in width. They show a preferred orientation parallel to the foliation in some bands,
which may indicate recrystallization of the hematite along the bedding planes during diagenesis. This primary and/or early diagenetic crystallization and recrystallization of fine-grained hematite in the BIF worldwide were recorded by many authors (e.g. Dimroth and Chauvel, 1973; Acharya et al., 1982; Lougheed, 1983). Tompkins and Cowan (2001) mentioned that primary hematite most probably forms from silica-iron hydroxide gels as suggested by Ewers and Morris (1981) during compaction and early diagenesis or the syngenesis of Tarling and Turner (1999).

Platy specularite crystals are rarely present. It is randomly distributed and the crystal size ranges from 50 to 200 µm in length and from 30 to 80 µm in width. This hematite is regarded to be of metamorphic origin (James, 1955; Dorr, 1973).

**ii) Magnetite**

Magnetite occurs in different modes, among them, the fine disseminations are the most common. They are subidiomorphic to idiomorphic crystals reaching up to 6 µm in diameter. They are dispersed within hematite or in jasper bands and are rarely aligned or coalesced to form stratiform laminae. The minute magnetite crystals in BIFs are usually described as primary detrital precipitates (e.g. Trendall, 1973a&b; Griffin, 1980). It is suggested that such primary magnetite is formed by diagenetic alteration and modification of precursor initial precipitates e.g. hydromagnetite, iron hydroxides, hematite or siderite (Dimroth and Chauvel, 1973; Klein and Bricker, 1977; Han, 1978; Lougheed, 1983).

Another coarser generation of magnetite is present, where the magnetite crystals range from 20 µm to 400 µm in size. This variety of magnetite is either randomly distributed in all bands or severely gathered together in continuous microbands with some sort of orientation. They commonly show well-developed cubic, octahedral and six-sided cross-sectional forms. As shown in the petrographic description, magnetite rims the borders of some specularite
platy crystals and jasper fragments. In this case, the magnetite shows idiomorphic terminations away from the hematite crystals or jasper fragments. So, either the hematite or jasper served as nuclei for the development of some porphyroblastic magnetite. Also, the growth of solid magnetite porphyroblastic crystals may be intermittent. This could be inferred from the difference in reflectivity or in susceptibility to oxidation viz. martitization. Such difference can be attributed to the presence of subsequent generations of magnetite in a single magnetite crystal (Han, 1978; 1982 & 1988). Morris (1980 & 1983) was in favour of magnetite formation by progressive alteration of primary hematite. Tompkins and Cowan (2001) attributed the formation of magnetite to the subsolidus reduction during late-stage diagenesis (anadiagenesis of Tarling and Turner, 1999) and anchimetamorphism. They interpreted the hematite mantled by magnetite as a result of incomplete subsolidus reduction of hematite during metamorphism. Floran and Papike (1978) agreed with them that magnetite is post-dates the formation of hematite most likely during the metamorphism. It is believed here that the coarse magnetite crystals might have been developed by recrystallization during relatively reducing metamorphic conditions.

Martitization is common in the different generations of magnetite. It ranges from incipient to extensive replacement to form secondary hematite or martite, leaving just small relics of the parental magnetite. The observed martitization displays many forms, but the oxidation of magnetite into hematite along the octahedral planes (111) and along the borders of the magnetite crystals (peripheral or rim replacement) are the most dominant forms.

**iii) Gangue components**

Silica represents the essential gangue components with less common chlorite. In the studied BIF from Wadi Hamama, silica is present in many forms, namely Jasper, chert and rarely as mega-quartz. Jasper is the most dominant form of silica. It commonly occurs as
meso- to microbands alternating with the iron bands or as nodules embedded in a matrix rich in iron minerals. Most of the nodules are arranged in discontinuous array, lenticular, jasper lumps or pods. This could be attributed to the lateral flow of silica during distortion under load (Ewers and Morris, 1981; Ewers, 1983). Trendall and Blockley (1970) supposed that the whole jasper-rich bands, either pods or not, result from gross compaction and diagenesis of uniform gel with vertical escape of silica. On the other hand, Mel’nik (1982) stated that the red jasper nodules found in BIF, are believed to be colloidal accretions or lumps of silica that have been reworked by water. The megacrysts of quartz, when present, either fills fracture or occupies nodules, representing mobilized or recrystalized chert and jasper, respectively.

As indicated by some representative electron microprobe analyses of Wadi Hamama jasper (Table 1), the amount of Fe$_2$O$_3$ varies greatly from low (4.01-6.22) to high (30.42-62.81) according to the modal percentage of accumulated fine hematite dusty inclusions. Remarkable enrichment of the hematitic chert (jasper) containing up to 3.28 wt% MgO is an indication of deposition in marine conditions, more specifically in back-arc basins. Source of Ti in the jasper (containing up to 0.77 wt% TiO$_2$) is regarded to the volcanic output of mafic to intermediate composition.

(Table 1 should be placed here)

**Geochemistry of HBIF ore**

The BIF of Wadi Hamama area was analyzed for both major and trace elements. Data in Table 2 are the analytical results for 13 representative samples of the studied BIF. Samples number 1&13 are mainly from the jasper mesobands. Where the FeO is not indicated in the table, the Fe$_2$O$_3$ values represent the total iron oxide (Fe$_2$O$_3$).
i) Major oxides composition

The analyzed samples are dominated by silica and iron oxide, where they vary from 16 to 60 wt% and 24 to 78 wt%, respectively. The SiO$_2$ and Fe$_2$O$_3$ vary markedly, whereas SiO$_2$ decreases as the total Fe increases (Fig. 9a). This is a general feature of all Precambrian BIFs (James, 1966). The other components of the analyzed BIF are present in small but highly variable amounts. CaO, Al$_2$O$_3$ and K$_2$O contents range from 0.36 to 4.3 wt%, from 1.4 to 11.41 wt% and from 0.02 to 2.78 wt% respectively. The higher content of CaO in some samples can be attributed to the presence of secondary calcite and epidote. The larger amount of Al$_2$O$_3$ and K$_2$O can be attributed to the contamination by volcanioclastic input (Klein and Beukes, 1989). This is also supported by the positive correlation between Al$_2$O$_3$ and TiO$_2$ (Fig. 9b), which reflects incorporation of clastic materials to these chemical sediments (Ewers and Morris, 1981; Davy, 1983; Dymek and Klein, 1988).

(Fig. 9 should be placed here)

For comparison, the average major element composition of the studied BIF of Wadi Hamama area is presented with the average major element of other Egyptian and worldwide BIFs (Table 3). From this table and the graphical presentation shown in Fig. 10, higher Al$_2$O$_3$, TiO$_2$ and K$_2$O values characterize Wadi Hamama BIF, but CaO is lower. All these values reveal a close similarity to the BIFs of Algoma-type. As compared to two BIF examples from Egypt, Abu Marawat BIF (Basta et al., 2000) and Umm Nar BIF (El Manawi 1991; El-Aref et al., 1993), Wadi Hamama BIF exhibits higher Al$_2$O$_3$, TiO$_2$, Na$_2$O and K$_2$O values. The other elements are intermediate in their values between Abu Marawat and Um Nar BIFs, except for moderately lower CaO.

(Table 3 should be placed here)

(Fig. 10 should be placed here)
**ii) Trace element composition**

The concentrations of 19 trace elements were determined and their abundances are listed in Table 2. Most iron-formations have low values for all trace elements (Davy, 1983). Like most iron-formations, the studied BIF has low trace element values. Nearly all the determined trace elements values fall within the range of common values recorded in the oxide and silicate facies BIF of Davy (1983) as shown by the comparison given in Table 4.

In the studied BIF samples from Wadi Hamama area, most of the transitional metal concentrations are generally low, especially Ni which is extremely low and does not exceed 12 ppm and many samples have abundances lower than the detection limit (<2 ppm). The concentrations of V and Cr vary from 58 to 141 ppm and 6 to 48 ppm, respectively. The Cu and Sc vary from <2 to 35 ppm and 7 to 18 ppm, respectively. Remarkably high V content can be assigned to the metal capturing in the magnetite structure.

Rb concentration is generally low. The maximum Rb concentration (65 ppm) is found in a sample containing the highest K$_2$O (2.8 wt%). Sr concentration ranges from 8.4 to 91 ppm. There is a clear good positive correlation between Sr and CaO (Fig. 9c). Zr ranges from 12 to 121 ppm and shows a good positive correlation between Zr and TiO$_2$ in the studied BIF (Fig. 9d). Such positive correlation is a strong indication that the source of the material could be pyroclastics from intermediate igneous rocks (Ewers and Morris, 1981).

Average of trace elements in Wadi Hamama BIFs and the averages of some Egyptian and world BIFs are also compared (Table 4). From the comparison with the Egyptian BIFs, the studied BIF has an extremely higher Co and Zn values, slightly higher V and Sc and slightly lower Cr and Sr values. Other trace elements show intermediate values. As compared to the average values of both the Algoma and Superior-types, Wadi Hamama BIF shows lower values of Cr, Y and Zr and higher Co value and the rest of elements show intermediate values (Fig. 11).
Specularite mineralization

Field observations

The specularite mine consists of several shafts that are more or less parallel to each other and generally has the NNE-SSW trend. All shafts were dug in the trondhjemite-tonalite masses at Wadi Hamama (Fig. 12a). The main shaft expands from the contact of the trondhjemite and the mafic metavolcanics inside the granitic body.

The specularite veins and lodes are tabular in shape, but some of the shafts exhibit shortage in one of the proposed long dimensions and hence some lodes display subtubular appearance. All of the shafts are steeply dipping, nearly vertical in most places (Fig. 12b), and show some sort of regularities in their orientations suggesting their development on one of the fracture systems in the trondhjemite. The occurrence of slickensides along the wall of some shafts emphasizes the restriction of the specularite veins along such fractures. The thickness of the specularite veins seems to be variable from a shaft to another. The maximum measured width of the mined shafts is 5 meters. Bifurcation of some veins is quite common and is recorded on different scales (Fig. 12b).

The filling material of the veins consists either of quartz-specularite or quartz alone. The quartz-specularite veins are the main vein type and the barren quartz veins are so scarce. Despite of their infrequency, the most primitive form of the quartz-specularite veins can be observed at Wadi Abu Gerida. These veins are limited in their thickness, only up to 50 centimeters thick. They consist of massive undeformed specularite selvages (Fig. 12c) surrounding vugy specularite-bearing quartz. Minor fresh pyrite cubes are observed in this quartz. The hematite along the selvage is finer in size and more massive than the hematite associated with quartz. The latter is characterized by its large specular form and some of
them line up the cavities in the quartz. The predominance of vugs in such veins, even in the specularite selvages, suggests gas emanation during the development of these veins that are sometimes lined by fine drusy quartz crystals. Some shafts of the highly mineralized area at Wadi Hamama are characterized by standing specularite-specularite ledges with exploited borders. These specularite-quartz ledges are brecciated and this can be attributed to the reactivation of the movements along the infilled fractures. All the trondhjemitic walls of the shafts are almost characterized by the presence of the massive specularite offshoots and less common quartz-specularite veinlets. The specularite offshoots vary in their densities and in their thicknesses, from few centimeters up to 25 centimeters. The quartz-specularite offshoots are less common and smaller in thicknesses than the specularite offshoots. Quartz in these veinlets is the early-formed phase, lining the fracture, and it is followed by a later generation of specular hematite. No mining activities are recorded in the pure quartz veins.

Small quartz-specularite veins, up to 25 centimeter thick, were injected in the metabasalt, in proximity to the contact with the trondhjemite close to the mining area. These veins range in composition from massive specularite, quartz-specularite, to ferruginated quartz. The veining led to the silicification of the metabasalts, which gives the rocks their creamy to brownish yellow appearance. Large flaky crystals forming rosette of specular hematite encrust the surfaces of some fractures in the metabasalts.

The wall-rocks of some shafts are broken up into angular pieces along the joint planes which can be attributed to the forceful injections of the hydrothermal solutions along the major fractures. In addition to silicification and chloritization of the host rocks that enclose the shafts, other surficial minerals are developed on the wall rock of the shafts, such as anhydrite and iron oxides & hydroxides that give the walls their white, yellow and red colours.
Petrography of the specularite veins

The tonalite-trondhjemite rocks are generally suffered from chloritization, epidotization and sericitization in a decreasing order. Along the walls of some shafts, the tonalite-trondhjemite host rocks are subjected to intense chloritization giving them a greenish colour. In this case, the chlorite is a secondary product after Na-feldspars of the felsic host that is why in most cases it psuedomorphically replaces euhedral prisms. In other cases, a reddish colour characterizes some of the shafts’ wall rocks. The development of such colour is attributed to the staining of the sodic feldspars crystals of the trondhjemitic rocks by iron oxides and/or hydroxides. The iron oxides and/or hydroxides weakly corrode the rims of some quartz crystals. The iron-rich solution invaded the feldspar crystals along the cleavage planes as well as the tiny fractures.

Quartz, in general, is the main infill for most of the veinlets that dissect the granitic host rocks adjacent to the mined shafts. These quartz veinlets grade into pyrite-quartz assembly in some cases. The formation of pyrite crystals follows the formation of the quartz whereas the former occupies the core of the veinlets only. The pyrite cubes are more or less entirely altered to goethite that forms a colloform texture in most cases. The secondary goethite also formed microveinlets at the expense of aligned pyrite crystals.

Hematite, which is the main extracted ore from the shafts, generally has the specular or the micaceous form which refers to specularite especially in the old literature. The specular hematite is present microscopically as flaky crystals, which aggregate in bundles that crosscut each other. Other specular crystals of hematite are radiating from a common center, forming either wavy or rosette shape (Fig. 13a). Some of the specularite crystals are folded in an irregular manner (Fig. 13b). Psuedomorphic goethite (after pyrite cubes) includes laths of specularite of different sizes either partially or completely (Figs. 13c&d). It appears that goethite was subjected to remobilization and then engulfs the hematite crystals.
and partly corrodes them (Figs. 13e). It was also noticed that some growing quartz crystals in
the veins corrode specularite that appears as relics in such a case (Fig. 13f).

In the intense chloritized walls, the quartz is generally sizable and also highly
corroded by chlorite. Few quartz crystals enclose minute fresh pyrite crystals and some of
them altered to goethite.

(Fig. 13 should be placed here)

Discussion and conclusions

The Wadi Hamama banded iron-formation (HBIF) is concentrated principally in two
horizons within the metavolcaniclastic sequence. The lower horizon (~85m) and the upper
horizon (~17m) are separated by metagreywacke. The iron content in the HBIF ranges from
24 to 41%. The iron bands alternate mainly with metatuffs and minor volcanogenic
metagreywacke. The contacts between the iron bands and their metatuffaceous host are
commonly sharp but some gradational relationship was also identified occasionally. HBIF
shows various scales of primary bedding and banding (either primary depositional or
metamorphic). The iron ore bands are mainly of oxide facies, whereas the main iron
constituents are hematite and magnetite associating with jasper. This facies can be further
divided into three subfacies, namely hematite, hematite-magnetite and magnetite subfacies
depending on the hematite/magnetite ratio. The mineral assemblages of HBIF are
characterized by generally fine-grained iron oxide and jasper, indicating that these rocks
were merely affected by diagenetic to low-grade metamorphic processes. It is evident that
growing idiomorphic magnetite porphyroblats enclose jasper microclasts in addition to clear
overgrowth of magnetite over recrystallized hematite dust (specular hematite). It is suggested
that the first deposition of hematite was in the form of fine chemically precipitated dust
intermittent with the precipitation of hematitic chert or jasper. The hematite dust was then recrystallized into specular hematite during diagenesis and low-grade regional metamorphism in reducing conditions. Then the specular hematite was followed by magnetite overgrowth as a result of incomplete sub-solidus reduction of hematite during early-stage diagenesis (anadiagenesis) and late-stage diagenesis and beginning of metamorphism (anchimetamorphism). Formation of magnetite at the expense of hematite by sub-solidus reduction was accepted for similar world examples of BIF especially during the last five years (e.g. Tarling and Turner, 1999; Tompkins and Cowan, 2001&2002).

The limited extension and the close association with the arc metavolcaniclastics, supported by some geochemical evidence, suggest the idea that Hamama banded iron-formation akin to the Algoma-type similar to many BIFs in the Eastern Desert of Egypt. The present paper suggests that the studied HBIF ore was originated as chemical sediments of exhalative fumarolic sources developed in intra-arc basins. It is here recommended that the chemical composition of BIF should not be a decisive tool for the ore genesis because the averages of major and trace elements compositions of HBIF suggest an intermediate example between the Algoma- and Lake Superior-types despite of being clearly affiliated to the former category as suggested by the field and mineralogical constraints materialized in the present paper. The suggested origin is in harmony with the general one proposed by Sims and James (1984) and Hussein and El Sharkawi (1990) for other Egyptian BIFs. Despite of being misleading in parts, the whole-rock chemical composition of the HIBF ore, some positive sides are here presented. The positively correlated Al₂O₃ with TiO₂ suggest incorporation of clastic materials with the chemical precipitates in the basin of deposition. Remarkable enrichment of the hematitic chert (jasper) containing up to 3.28 wt% MgO is an indication of deposition in marine conditions in the back-arc basins. Source of Ti in the jasper (containing up to 0.77 wt% TiO₂) is regarded to the volcanic output of mafic to intermediate
composition. Actually, chert bands associating the BIF ores were deposited in back-arc basins but the source of silica might be the weathering of highlands on the continent during the Precambrian as indicated by their Ge/Si ratio (Hamade et al., 2003). In a recent study, Bolhar et al. (2005) concluded that jasper and iron minerals in the BIF ores are precipitated as alternating bands in response of episodic changes in ambient water chemistry. They (op. cit.) added that chert is deposited from silica-saturated seawater by evaporation during quiescence of volcanic exhalations. Pickard et al. (2004) attributed the formation of chert in BIF macrobands to provence-scale seafloor silicification where the source of silica would be terruginous. The present authors are in favour that Wadi Hamama hematitic chert was precipitated when the volcanic exhalations were quiet or even when there was supply of iron from volcanicity as indicated by the presence of jasper bands with dense aggregated hematite dust wher $\text{Fe}_2\text{O}_3$ reaches up to 62.81 wt%.

The specularite ore of Wadi Hamama area was known since Antiquity especially during the rule of Romans. Specularite from the apatite-specularite veins was exploited from the veins, which are located within the trondhjemitic host in particular (Hume, 1937). The concentration of specularite veins along the contact between the trondhjemites and the metavolcanics led the present authors to conclude that residual P-rich hydrothermal fluids, released from the calc-alkaline magma, was responsible for the leaching of iron from the metavolcanics and the metavolcaniclastics that host BIF strata. These mineralized fluids followed the fractures in the trondhjemites forming the specularite veins with common apatite crystals and the gases of the fluids resulted in the formation of vugs that are sometimes lined by drusy quartz crystals. Such conclusion opposed the hypothesis of El-Mansi (1994) that Wadi Hamama specularite originated from the shearing of the BIF bands in the metavolcaniclastics. The present authors contradicts his conclusion because field evidence suggest that specularite occurrences are far from the BIF, in addition to the
presence of discordant specularite veins in the granites themselves with no signs of shearing. Also, apatite is neither an essential nor accessory constituent of the BIF bands. In similar world examples, veins rich in Fe-rich apatite and hematite (preumbaly Kiruna-type mineralization) are connected to volcanic arc calc-alkaline granitoids like those of the north Chilean Andes (Grocott and Taylor, 2002).

Acknowledgements

The analytical works were carried out by the second author during a scientific leave to Copenhagen University, Denmark in 2002. Sincere helps by John Bailey, who generously facilitated the XRF analyses, are greatly acknowledged. Thanks are also due to Berit Winzell for her kind assistance in the electron microprobe laboratory at the Geological Institute.

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**Figure Captions:**

Fig. 1: Geological map of Wadi Hamama area

Fig. 2: Graphic presentation of the different types of banding recorded in Wadi Hamama BIF (HBIF) from macro- to microbanding.

Fig. 3: Compiled stratigraphic sections showing the BIF-bearing horizons and their details.

Fig. 4: Field occurrence of HBIF:
   a. Megascopic or macrobands of BIF intercalated with the volcanogenic metaclastics.
   b. BIF band exhibiting drag folding.
   c. Resistant ledges of BIF in their host volcanogenic metaclastics.
   d. Fine calcite and quartz veinlets (Cc-Qz) injecting BIF hosted by the volcanogenic metaclastics (Vol).

Fig. 5: Polished slabs of HBIF showing some primary structure and association of jasper:
   a. Volcanogenic metaclastics sandwiched by BIF mesobands.
   b. Thin BIF mesobands intercalated with thick jasper bands.
   c. Jasper-poor BIF showing syn-depositional micro-faulting.
   d. Slab showing slumped surfaces, in addition to pinch and swell structure.

Fig. 6: Ore mineralogy of HBIF, hematite subfacies (all microphotographs are in Polarized Reflected Light):
   a. Sub-microbands formed by the coalescence of fine specularite laths. Notice the presence of flattened jasper.
   b. Sub-microbands of specularite lacking jasper.
   c. Pyrite cube pseudomorphed by ill-developed goethite-hematite.
   d. Very fine, but fresh, pyrite disseminations.
   e. Aggregation of fine hematite into dense lensoidal bodies.
   f. Confinement of fine hematite (Hem) to the core of jasper (Js)
   g. Aggregated fine magnetite crystals.
   h. Jasper nodule with some fine magnetite crystals.

Fig. 7: Ore mineralogy of HBIF, hematite-magnetite subfacies (all microphotographs are in Polarized Reflected Light except for a, in Plane-Polarized Light):
   a. Contorted magnetite-rich sample with common light silica bands.
   b. Fine magnetite crystals scattered in silicate matrix.
   c. Coarse magnetite crystals showing slight martitization.
   d. Zoned idiomorphic magnetite porphyroblast with dark core.
   e. Oriented idiomorphic magnetite set in fine specularite-rich matrix. Notice that magnetite is slightly martitized especially along its peripheries.
   f. Same as e, but with less frequent magnetite.

Fig. 8: Ore mineralogy of HBIF, magnetite subfacies (all microphotographs are in Polarized Reflected Light):
   b. Growing sub-idiomorphic magnetite porphyroblasts in a matrix rich in finer idiomorphic magnetite generation, magnetite-hematite-jasper assemblage.
c. Magnetite overgrowing platy specular hematite, magnetite-hematite-jasper assemblage.
d. Magnetite “nests” in jasper-rich microbands, magnetite-hematite-jasper assemblage.
e. Weak alignment of idiomorphic magnetite, magnetite-chlorite assemblage.
f. Mn-oxides & hydroxide colloformed crystals (Mn) in veinlets, magnetite-chlorite assemblage.
g. Highly martitized core of magnetite (Mt) partly enveloped by goethite (G) after hydrated pyrite, magnetite-chlorite assemblage.
h. Aggregated fine magnetite crystals in the metatuffs.

Fig. 9: Binary relationships between some major and trace elements in the analyzed HBIF samples.

Fig. 10: Average major oxides composition of HBIF compared to other Egyptian and world examples. For number notation, refer to Table 3.

Fig. 11: Average trace elements composition of HBIF compared to other Egyptian and world examples. For number notation, refer to Table 4.

Fig. 12: Field characteristics of the specularite veins in the granitoids:
   a. Ancient mined vertical shafts dug along specularite veins in trondhjemite.
   b. Steeply dipping bifurcated shaft at the vicinity of excavated specularite veins.
   c. Undeformed specularite (spec)-quartz (Qz) selvage in trondhjemite (a fine metal pin as a scale).

Fig. 13: Ore mineralogy of the specularite veins (all microphotographs are in Reflected Light):
   a. Radiating specularite crystals cut perpendicular to their platy axes.
   b. Folded specularite aggregates.
   c. Hydrated pyrite cubes, now goethite (G) interstitial to randomly oriented specularite crystals.
   d. Goethite (G) after pyrite enclosing and corroding specularite (Spec).
   e. Specularite crystals filled by mobilized goethite (G).
   f. Relict specularite (Spec) corroded by growing quartz (Qz).
Fig. 1:
Fig. 2

Alternating iron bands with banded metatuffs and subordinate volcanogenic metagreywacke

Alternating iron microbands with jasper
Fig. 4
Fig. 5

- **a**: Volcaniclastics
- **b**: Jasper, iron, jasper, iron
- **c**: Micro-faulting
- **d**: Slumping, pinch and swell structure

**2 cm** (for all samples)
Fig. 7
Fig. 9
Fig. 10
Fig. 11

Graph showing the concentration of various trace elements in ppm.
openings of vertical shafts

Fig. 12
Table 1: Electron microprobe analyses of hematitic chert (jasper) from Wadi Hamama BIF

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### Tab 2: Major and trace elements composition of Wadi Hamama BIF determined by the XRF technique

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Table 3: Comparison of average major oxides of the studied BIF with other Egyptian and world BIF examples.

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1- Wadi Hamama oxide facies BIF.
2- Abu Marawat oxide facies BIF (Basta et al., 2000).
3- Um Nar BIF (El-Aref et al., 1993).
4- Algoma (all facies) BIF (Graf, 1978; Gross & McLeod, 1980).
5- Superior (all facies) BIF (Graf, 1978; Gross & McLeod, 1980).

Table 4: Comparison of average trace elements (ppm) in Wadi Hamama BIF and some Egyptian and World BIF examples.

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1- Wadi Hamama oxide facies BIF.
2- Abu Marawat oxide facies BIF (Basta et al., 2000).
3- Um Nar BIF (El-Aref et al., 1993).
4- Algoma (all facies) BIF (Graf, 1978; Gross & McLeod, 1980).
5- Superior (all facies) BIF (Graf, 1978; Gross & McLeod, 1980).