Petrology and geochronology of the Bom Jardim de Goiás copper deposit (GO)

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Abstract The Bom Jardim de Goiás copper deposit is a remarkable example of a deposit formed in a Neoproterozoic juvenile arc in the South American Platform during the assembly of Gondwana. The deposit is inserted in the western margin of the Arenópolis Magmatic Arc, in the Tocantins Province. The host rocks comprise a volcanosedimentary sequence of intermediate to acid composition interpreted as crystal and ash tuffs. The tuffs represent calc-alkaline volcanism of rhyodacitic to dacitic composition, with lithogeochemical and biotite chemistry characteristic of magmas from intra-oceanic arcs. The tuffs are commonly cut by veins and veinlets with no preferential orientation and show $\epsilon_{\text{Nd(t)}}$ values between +3.5 and +7.4 and T_{DM} model ages ranging from 0.8 to 1.1 Ga, consistent with juvenile magmas generated in magmatic arc settings. Intrusive bodies in the volcanosedimentary sequence are represented by hornblende-biotite monzogranite, interpreted as Serra Negra Granite, and biotite syenogranite defined as Macacos Granite. The granites have geochemical characteristics of type I calc-alkaline granites, generated in a post-collision setting. Tholeiitic basalts intercalated with tuffs, interpreted as flows, show geochemical affinity with volcanic arc basalts, whereas the dikes that crosscut the granites have compositions characteristic of intra-plate tholeiitic basalts. The petrological data of the host and country rocks of the Bom Jardim de Goiás area reveal an important production of calc-alkaline magma in a magmatic arc setting in western Goiás around 750 Ma, genetically related to copper mineralization, and the generation of a non-mineralized post-collisional magmatism at 540 Ma.

Keywords: tuff; granite; copper; magmatic arc; Bom Jardim de Goiás.

Resumo *Petrologia e geocronologia do depósito de cobre de Bom Jardim de Goiás (GO).* O depósito de cobre de Bom Jardim de Goiás constitui exemplo notável de depósito formado em ambiente de arco juvenil Neoproterozoico na Plataforma Sul Americana durante a assembleia do Gondwana. O depósito está inserido na borda oeste do Arco Magmático de Arenópolis, na Província Tocantins. As rochas hospedeiras da mineralização são rochas da sequência vulcanossedimentar de composição intermediária a ácida, interpretadas como tufos do tipo cristalino e cinerítico. Os tufos representam vulcanismo cálcio-alcálinos de composição riodacítica a dacítica, com características litogeoquímicas e de química de biotitas de magmas de arcos intraoceânicos. Os tufos são comumente cortados por veios e vênulas sem orientação preferencial e mostram valores de $\epsilon_{\text{Nd(t)}}$ positivos entre + 3,5 e + 7,4 e idade-modelo TDM variando de 0,8 até 1,1 Ga, coerentes com magmas juvenis gerados em ambientes de arcos magmáticos. Corpos intrusivos na sequência vulcanossedimentar são representados por hornblenda-biotita monzogranito, atribuídos ao Granito Serra Negra; e biotita sienogranito, definido como Granito Macacos. Os granitos possuem características geoquímicas de granitos cálcio-alcálinos, do tipo I, gerados em ambiente pós-colisional. Basaltos toleíticos intercalados com os tufos, interpretados como derrames, mostram afinidade geoquímica com basaltos de arcos vulcânicos, enquanto diques que cortam os granitos têm composição de basaltos toleíticos intraplaca. Os dados de petrologia das rochas hospedeiras e encaixantes do Depósito Bom Jardim de Goiás mostram a importante produção de magma cálcio-alcálico em ambiente de arco magmático no extremo oeste de Goiás ao redor de 750 Ma, geneticamente associado a mineralizações de cobre, e a geração de magmatismo pós-colisional estéril em 540 Ma.

Palavras-chave: tufo; granito; cobre; arco magmático; Bom Jardim de Goiás.

INTRODUCTION The Bom Jardim de Goiás copper deposit is located in the westernmost part of Goiás State. Geologically speaking, it is inserted in the eastern portion of the Tocantins Province, which resulted from the convergence of the São Francisco and Amazonian cratons in the Neoproterozoic (Dardenne 2000). It is inserted in the Goiás Magmatic Arc as

proposed by Pimentel & Fuck (1992) and is defined as Neoproterozoic juvenile crust segment.

The deposit is hosted by volcano-sedimentary rocks that underwent incipient or no metamorphism. The host rocks to the mineralization are considered to be volcanoclastic rocks of the Córrego da Aldeia Formation of the Bom Jardim de Goiás Group,

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defined and described by Seer (1985). Despite the fact that the deposit has been known from a long time and the 4.5 Mt ore reserves with average 0.92% copper (Oliveira 2000) have been well constrained by the Brazilian Geological Survey (*Companhia de Pesquisa de Recursos Minerais* – CPRM), most of the existing studies derived from reconnaissance and regional mapping (Pena 1974, Fragomeni & Costa 1976, Costa *et al.* 1979, Marques *et al.* 1980). In turn, Seer (1985) studied specifically the copper mineralization in the Bom Jardim de Goiás Deposit.

The present study focuses on petrological data of the Bom Jardim de Goiás Deposit, its host rocks and rocks intrusive in the deposit, aiming at the characterization of the magmatism and the understanding of the mineralization geological setting by means of a systematic lithogeochemical and (Nd-Sr) isotopic study of these rocks. Absolute dating of host and intrusive rocks was carried out applying the U-Pb LA-ICMPS method. The well-preserved country and host rocks makes the Bom Jardim de Goiás Deposit a striking example of a copper deposit formed in the Brazilian Precambrian, and provides information for a better understanding of the deposits hosted in the Goiás Magmatic Arc and for the identification of possible targets for mineral exploration in the Goiás State.

REGIONAL GEOLOGY The Bom Jardim de Goiás Deposit is inserted in the eastern portion of the Tocantins Province, interpreted as resulting from the convergence of the São Francisco and Amazonian cratons in the Neoproterozoic (Dardenne 2000), specifically in the Goiás Magmatic Arc (Fuck 1994, Fuck *et al.* 1994) (Fig. 1).

The Goiás Magmatic Arc (AMG), as proposed by Pimentel & Fuck (1992), was defined as constituted by tonalitic and granodioritic orthogneisses associated with metavolcanic and metasedimentary rocks (Fig. 2), the typical occurrences being situated in the Arenópolis and Mara Rosa regions. The AMG rocks are crosscut by N45°-80°W- and N30°-50°E-trending transcurrent shear zones and N30°-50°E- and NS-trending thrust faults formed during the Brasiliano event. It is limited to the west by the Paraguay and Araguaia belts and to the east by the Goiás Massif (Fig. 1). It is divided in two blocks, the Arenópolis Arc in the south and the Mara Rosa Arc in the north (Araújo Filho & Kuyumjian 1996, Pimentel *et al.* 1996, Pimentel *et al.* 1997).

The Bom Jardim Deposit is located in the western border of the Arenópolis Magmatic Arc. The volcano-sedimentary rocks that occur in this region were grouped in the Bom Jardim de Goiás Group (Seer 1985). The volcano-sedimentary sequences are constituted by metavolcanic rocks and associated subvolcanic rocks, whose compositions vary from tholeiitic basalts to rhyolites. These rocks are metamorphosed to the greenschist and amphibolite facies (Seer 1985, Pimentel & Fuck 1986). Detrital metasedimentary rocks were identified in the upper units of the volcano-sedimentary sequence.

The granitic intrusions of the study area constitute large K-rich calc-alkaline bodies, which predominantly include isotropic and equigranular biotite granites. They may contain hornblende and present porphyritic texture (Pimentel *et al.* 1999, Pimentel *et al.* 2000), such as the Macacos and Serra do Tatu granitic plutons. They also include granodiorites that are locally metaluminous to slightly peraluminous and show mylonitic foliation (Rodrigues *et al.* 1999, Pimentel *et al.* 2000). Thus, two post-Brasiliano episodes of granitic magmatism were recognized in the study area: one around 590 to 560 Ma and a younger one around 508 to 485 Ma. The older granites are considered similar to I-type granites (Caiaçó, Serra do Iran, Israelândia and Serra do Impertinente granites), whereas the younger ones are alkaline, showing A-type granite characteristics (Serra Negra and Iporá granites). Late to post-orogenic intrusive A-type granites are associated with mafic and ultramafic bodies (Pimentel *et al.* 2000).

The deposit area belongs to CPRM and comprises volcano-sedimentary rocks of the Bom Jardim de Goiás Group and granitic rocks that show incipient or no metamorphism and deformation. Rocks from the Paraná Basin (Paraná Group) occur south of the study area, and consist of Devonian sedimentary rocks of the Furnas and Vila Maria formations (Oliveira 2000). The main mapping unit recognized in the study area encompasses volcanic and volcanoclastic rocks that constitute the Bom Jardim de Goiás Group and host the copper mineralization.

GEOLOGY AND PETROGRAPHY OF THE BOM JARDIM DE GOIÁS DEPOSIT AND SURROUNDING ROCKS

Volcanic and volcanoclastic rocks The volcanic and volcanoclastic rocks surrounding the Bom

Figure 1 – Regional geologic map of the northern portion of the Tocantins Province (Dardenne 2000).

Figure 2 – Geologic map of the Goiás Magmatic Arc in southwestern Goiás (Pimentel et al. 1999, Pimentel et al. 2000).

Jardim de Goiás Deposit were divided by Guimarães (2007) in altered and unaltered rocks. The unaltered volcanic rocks are crosscut by very few veins and veinlets, their primary structures are well preserved and recognized in barren drill hole cores.

The altered volcanic rocks present a large quantity of veins and veinlets of varied compositions, contain sulfides that are either disseminated or concentrated in veinlets, and predominate in mineralized drill hole cores (Fig. 3).

The venulation characteristic of the mineralized volcanic rocks of the Bom Jardim de Goiás Deposit changes the original color of the rock,

which varies from dark gray when the rock is fresh, to brownish gray when the rock is slightly altered, to greenish gray or whitish pink, depending on the alteration type (Fig. 3).

The unaltered volcanic rocks include intermediate to acid volcanic and volcanoclastic rocks with preserved igneous structures. These rocks are slightly or not metamorphosed, unaltered or incipiently hydrothermalized. Petrographic data reveal that these volcanic rocks are predominantly pyroclastic. According to the clast-size classification proposed by Cas & Wright (1987) and modified by White & Houghton (2006), these pyroclastic rocks are

predominantly tuffs, dominantly composed of volcanic ash (particle size < 2 mm). Two sub-types are distinguished. Sub-type 1 can be classified as ash tuffs (composed of practically 100% volcanic ash), and sub-type 2 as crystal tuffs, where crystals predominate in a ash matrix. It can be also suggested that these are consolidated tuffs, due to the fact that shards (vitric fragments with a curved surface) are lacking and that the eutaxitic texture of the pumices had not evolved enough to form *fiamme* (flattened pumice lenses that generate layering during the compaction of pyroclastic rocks).

ASH tuffs The ash tuffs are the most abundant and characteristic of the study area. They are composed of a brown aphanitic matrix, in which it is possible to identify a few mineral phases under the microscope and those that compose the veinlets (Fig. 4A). Some disseminated subhedral quartz, plagioclase and magnetite crystals can be identified in a mass composed of brown biotite showing weak pleochroism to light brown. Locally, sericite- and chlorite-rich portions occur and are interpreted as alteration of brown biotite and of volcanic ash. Lithic fragments up to 1.0 mm in diameter are also present in these rocks. The fragments are composed of quartz, plagioclase and magnetite.

Random veinlets composed of brown biotite are common. Locally biotite is chloritized. Quartz veins are also present and abundant and may contain sulfides (pyrite and chalcopyrite) and brown biotite.

The brown aphanitic matrix containing some disperse quartz and feldspar grains is interpreted to be volcanic ash that underwent hydrothermal alteration, on the basis of Kano's (1990) descriptions of rhyolitic tuffs from Japan and Fritz & Stillman's (1996) study on the Country Waterford tuff in Ireland.

CRYSTAL tuffs The crystal tuffs occur intercalated with the ash tuffs. They show granular texture and are predominantly composed of angular quartz fragments (60%), brown aphanitic matrix (20%), plagioclase (15%), orthoclase (1%), lithic fragments (0.5%), and amphibole classified as ferrous hornblende (0.5%). The accessory minerals are zircon, titanite, magnetite, pyrite, chalcopyrite and rutile (4%) (Fig. 4b).

The quartz grains are angular to sub-angular, almost always microcrystalline, locally arranged in

Figure 3 – (A) Barren drill hole core of veinlet-free volcanic rock; (B) Mineralized drill hole core with quartz veins producing changes in the original color of the rock; (C) Mineralized drill hole core showing alteration (albitization) caused by the mineralized vein.

a vitriclastic texture, which results from the concentration of crypto- to microcrystalline quartz and feldspar in small lenses, interpreted as pumice relicts. The plagioclase grains are angular to sub-angular and are usually twinned according the albite law. They can also occur as very small laths and more rarely as isolated phenocrysts. Plagioclase is partially or totally altered to sericite, epidote and more rarely to carbonate. The presence of crystal fragments composed exclusively of plagioclase laths is common. The potassic feldspar is rare and occurs as angular fragments showing Carlsbad twinning. The 0.60 to 2.0 mm-sized lithic fragments are composed of quartz, plagioclase, brown biotite and more rarely magnetite and chlorite.

Biotite is the main component of the aphanitic matrix. It can also occur as up to 0.5 mm-sized lamellae disseminated in the rock and in some veinlets.

Igneous amphibole showing dark to pale green pleochroism occurs as disseminated elongated grains, but is restricted to some portions of the barren drill hole cores.

Figure 4 – (A) Photomicrograph of a ash tuff; (B) Photomicrograph of a crystal tuff with quartz, feldspar and pyrite crystals in a ash matrix; (C) Photomicrograph of a crystal tuff with pumices; (D) Photomicrograph of crystal tuff with amygdaloides filled with glass and biotite aureole.

Igneous titanite, when disseminated in the rock, is approximately 0.25 mm in size. It can also occur in veins, where it reaches 1.0 mm in size. In this case, it is interpreted as a product of hydrothermalism. Reddish rutile occurs disseminated in the rock and in veinlets. Zircon, magnetite, pyrite and chalcopyrite, which can reach up to 0.10 mm in size, are disseminated in the rock.

Secondary sericite was identified in barren drill hole cores and is an alteration product after orthoclase and plagioclase (BJ 22 and BJ34). Muscovite lamellae associated with chlorite fill rock fractures. Hydrothermal actinolite predominantly occurs filling veinlets in the crystal tuffs, together with chlorite, carbonate and epidote. It is also disseminated in the rock. Locally, hydrothermal Mg-hornblende was identified in veinlets crosscutting the barren tuffs.

The presence of pumice fragments is common in the crystal tuffs and attests for their classification as pyroclastic rocks. The best occurrences of pumice

fragments are found in the barren drill hole cores, where alteration was less intense (Fig. 4C). These fragments are also observed in the rocks cropping out in the study area, and are sometimes replaced by microcrystalline quartz (vitriclastic texture). The pumices can constitute elongated bands arranged in a eutaxitic texture that results from lithostatic compaction.

Another important feature observed in these rocks is the presence of amygdaloides commonly filled with quartz. Locally the amygdaloides are filled with a yellowish, isotropic fine material, interpreted to be glass, surrounded by an aureole of biotite (Fig. 4D).

Granites In the area of the Bom Jardim de Goiás Deposit granitic bodies of different compositions crop out, such as the pink and white hornblende-biotite monzogranites named Serra Negra Granite (GSN), located east of the deposit, and the red, coarse-grained biotite syenogranite named Macacos Granite (GM), located west of the

Figure 5 – (A) Pink Serra Negra Granite with mafic enclaves; (B) White Serra Negra Granite showing mafic enclaves with corrosion gulfs; (C) Macacos Granite showing up to 5 cm-sized feldspar; (D) Outcrop showing a diabase dike crosscutting the Serra Negra Granite.

deposit. The latter can also be porphyritic, with red potassic feldspar phenocrysts of up to 5 cm in length (Fig. 5). Both granites are usually isotropic, equigranular, medium-grained and stand out from the flat relief as rounded decameter- to meter-sized blocks. They also crop out as slabs at the margins of the drainage. The white facies of the Serra Negra Granite (GSN) also occurs as dikes crosscutting the pink facies.

The pink GSN is constituted by quartz (40%), albite to oligoclase (30%), potassic feldspar (20%) and biotite (10%) as essential minerals, and titanite, zircon, allanite and magnetite as accessories. Plagioclase and potassic feldspar are very altered to sericite and epidote. Hornblende is rare and shows weak greenish pleochroism and losangular habit.

Locally the pink GSN contains rounded mafic enclaves (Fig. 5) composed of plagioclase (20%), hornblende (40%), biotite (10%), potassic feldspar (14%) and quartz (15%) as essential minerals, and magnetite and allanite as accessories (1%). In the enclaves, the quartz grains are rounded and the potassic

feldspar and plagioclase are very altered to sericite, which highlights the plagioclase zoning.

The white GSN contains potassic feldspar (40%), albite to oligoclase (28%), quartz (25%) and biotite (5%) as essential minerals, and magnetite, zircon and allanite as accessories, which together sum 2%. Quartz grains also occur as inclusions in feldspar. Plagioclase is slightly altered and occurs as twinned laths according to the albite law. Slightly altered potassic feldspar with Carlsbad twinning also occurs. The biotite lamellae show pleochroism varying from brown to pale yellow. Some lamellae are chloritized. Zircon, allanite and magnetite are disseminated in the rock. Allanite occurs as zoned yellow prisms. Zircon usually occurs as inclusions in biotite, producing pleochroic haloes in it. Primary magnetite occurs as inclusions in quartz and biotite.

Locally, the white GSN contains mafic enclaves with 92% amphibole, 5% quartz, 2% biotite and 1% magnetite (Fig. 5B). The green igneous amphibole showing weak greenish pleochroism is hornblende. Biotite is identical to that of the pink GSN, showing

pleochroism from brown to yellow. The mafic enclaves show the same characteristics of those present in the pink GSN, but are a little more rounded and show corrosion gulfs. The presence of mafic enclaves is an indication that magma mixing may have occurred.

Regarding the Macacos Granite (GM), it is composed of potassic feldspar (55%), quartz (20%), albite to oligoclase (20%), and chloritized biotite (4%). Accessories are titanite, apatite, zircon and magnetite, which together sum 1%. Potassic feldspar phenocrysts predominate in the rock and are slightly altered to sericite. They are 1 to 5 mm in size, reddish and show Carlsbad twinning. The plagioclase shows albite twinning and is 0.5 to 1.0 mm in size; quartz grains vary from 0.10 to 1.0 mm in size. Chlorite occurs as greenish pleochroic lamellae and locally shows bluish interference colors. Biotite relicts showing brown to yellow pleochroism are recognized in some chlorite lamellae, which attest that chlorite is an alteration product of brown biotite.

Basic rocks Besides the enclaves in pink and white GSN (Fig. 5B), the basic rocks identified in the study area also occur as dikes crosscutting the Serra Negra and Macacos granites, as fragments in breccia and intercalated with tuffs.

The basic dikes are dark gray and fine-grained and are composed of actinolite (58%), labradorite (25%), chlorite (5%), quartz (2%), epidote (2%), biotite (1%) and opaque minerals (pyrite, chalcopyrite and magnetite – 2%).

Actinolite is acicular and shows pleochroism varying from pale to yellowish green. Plagioclase occurs as 0.25 to 0.5 mm-long laths. Quartz grains are sub-rounded and are disseminated in the rock. Epidote, probably alteration product after plagioclase, occurs as an agglomerate of euhedral grains showing yellow pleochroism. The occurrence of biotite lamellae with pleochroism varying from brown to pale yellow is restricted. The rare opaque minerals disseminated in these rocks are pyrite, chalcopyrite and euhedral magnetite of approximately 0.10 mm in size.

Locally, some dikes contain preserved vesicles of 1 to 3 mm in diameter. These dikes are composed of actinolite (45%), potassic feldspar (20%), plagioclase (10%), chlorite (10%), quartz (5%), epidote (5%) and magnetite (5%). Actinolite is acicular and shows pleochroism varying from pale green to pale yellow. The plagioclase laths are 0.10 to 0.25 mm

long. Chlorite lamellae showing weak greenish pleochroism are 0.25 mm in diameter. Quartz grains occur disseminated in the rock and are 0.05 mm in size. Magnetite is euhedral to subhedral and occurs disseminated in the rock. The vesicles are composed of quartz $\pm$ actinolite $\pm$ magnetite. A fine, brown material thought to be volcanic ash is concentrated at the rims of the vesicles.

The basic rock associated with GM occurs as centimeter-sized blocks. It is composed of plagioclase (50%), magnetite (30%), chlorite (15%), sericite (5%), quartz (1%), zircon (< 1%), and epidote (< 1%). Plagioclase occurs as laths and is usually sericitized. Chlorite lamellae showing pleochroism that varies from green to pale green and bluish interference color are considered alteration products; however, it was not possible to identify under the microscope the mineral from which they were formed. The rare, disseminated quartz grains are angular. Magnetite is also disseminated, which darkens the original color of the rock. Zircon is colorless and prismatic, and epidote is an alteration product after plagioclase.

The basic rocks intercalated in the volcanoclastic rocks are predominantly basaltic in composition. The basalt intercalated in the barren volcanoclastic rocks is very fine grained to aphanitic, greenish, and is composed of chlorite and biotite (87%), quartz ($\pm$ 10%), magnetite (3%) and sericite, alteration product after biotite. Chlorite is a transformation product after biotite, which is recognized as a few disperse lamellae and as relict in chlorite lamellae.

The basalt associated with the mineralized volcanoclastic rocks is composed of plagioclase (40%), actinolite (33%), quartz (10%) and chlorite (10%), and zircon, titanite (1%), magnetite and chalcopyrite (5%) as accessory minerals. Plagioclase laths are 0.10 to 1.0 mm in size. Actinolite is acicular and 0.25 to 1.5 mm in length. Chlorite lamellae showing greenish pleochroism are approximately 0.10 mm long. Quartz is not a rock-forming mineral; rather, it characterizes the silicification that affected the rock. Zircon, titanite, magnetite and chalcopyrite are disseminated. Carbonate, thought to be alteration product after preexisting calcic amphibole, occurs at the rims of actinolite and chlorite.

Evidences of hydrothermal alteration in the tuffs and implications to mineralization A systematic study of the veins and veinlets identified in the

tool in the study of the nature and physical-chemical conditions of the magmas it crystallizes from.

When preserved, the GSN biotite occurs as millimeter-sized brown lamellae (Fig. 19A). However, it is often partially transformed to chlorite, which changes its color to green (Fig. 19B).

Only the biotite from the crystal tuffs was analyzed, because in these rocks it occurs as brown millimeter-sized lamellae (Fig. 19C). The grain-size of the biotite from the ash tuffs makes the chemical analysis impossible. In the ash tuffs, biotite is locally transformed in chlorite, its color changing to green (Fig. 19D).

Figure 6 – (A) Photomicrograph of a mineralized drill hole core of crystal tuff showing a-type veins composed of biotite; (B) Photomicrograph of an ash tuff with crosscutting b-type veinlets; (C) Photomicrograph of a mineralized drill hole core of ash tuff showing a d-type vein composed of pyrite (py) and chalcopyrite (ccp); (D) Photomicrograph of a barren drill hole core of ash tuff showing an e-type vein composed of chlorite (chl) with some titanite (ttn) grains at the rims; (E) Photomicrograph of a barren drill hole core of ash tuff showing a f-type vein with actinolite (act) in comb texture, quartz and pyrite (black); (F) Photomicrograph of a g-type vein composed of carbonate only crosscutting an ash tuff.

volcanic rocks was carried out in order to better understand their relationship with mineralization.

The most characteristic alteration that affected the volcanic rocks is silicification. It advances from the veinlet walls towards the host rocks as microcrystalline quartz. It modifies the original color of the rock, which changes from brownish, to greenish gray, and to whitish pink. Chloritization is also common, and in many cases it is restricted to the veinlet walls, making the color of the rock greenish. Epidotization was observed only in drill hole sample BJ 22. It advances from the rims of an epidote agglomerate, along a vein composed of actinolite (Fig. 6).

The veins have no preferential orientation. They may be parallel, but more frequently they cross-cut one another. Regarding their mineral composition, Guimarães (2007) identified eight types of mineral associations in these veins and veinlets (Fig. 6):

- a) usually chloritized brown biotite;
- b) quartz $\pm$ chlorite $\pm$ carbonate;
- c) quartz + biotite + chlorite + pyrite + chalcopyrite $\pm$ carbonate $\pm$ titanite $\pm$ actinolite $\pm$ epidote $\pm$ albite $\pm$ Mg-hornblende;
- d) pyrite + chalcopyrite $\pm$ quartz $\pm$ chlorite;
- e) chlorite $\pm$ titanite $\pm$ magnetite $\pm$ carbonate;
- f) actinolite $\pm$ quartz $\pm$ pyrite;
- g) albite + quartz;
- h) carbonate.

It is difficult to establish a chronological order for the veins/veinlets because some are parallel, such as the c- and e-type veins, or crosscut one another several times, as the a- and b-type veins, or do not occur in the same sample.

The late b-type veins/veinlets are the most abundant and occur along all the barren and mineralized drill hole cores. Concomitant or later, there are the a-type veins, which are also common and abundant.

The c- and d-type veins are directly related to mineralization. Veins composed of only pyrite + chalcopyrite also occur. The casual presence of carbonate, titanite, actinolite, epidote and Mg-hornblende in these types of veins may be due to the alteration of certain minerals and to fluid-rock reactions that originated hydrothermal Ca- and Mg-rich minerals.

Carbonate-rich, h-type veinlets are considered to be the youngest, as carbonate also occurs filling rock fractures. The actinolite-rich, f-type

veinlets do not show any clear relationship with the other types of veinlets.

The characteristics observed in the study area, such as the succession of tuffs; the eutaxitic texture; the presence of pumice fragments; the presence of nodules filled with quartz and/or phyllosilicates, interpreted to be altered volcanic glass, and the presence of pillow lavas, as described by Seer (1985), open the possibility that these tuffs result from shallow submarine eruptions or from subaerial eruptions where the pyroclastic material flowed into water. Cas & Wright (1987) suggest that the presence of quartz nodules, such as those observed in the Bom Jardim de Goiás area, indicate welding influenced by an aqueous environment.

LITHOGEOCHEMISTRY

Representative chemical analyses of the rocks of the study area are listed in Tab. 1.

Tuffs Rock samples collected from drill hole cores and outcrops of the study area were analyzed for major and trace elements. Three analyses correspond to the ash tuffs (BJ34-278.74m; BJ35-110.31m; BJP8a), and the other ones to the crystal tuffs (Tab. 1).

The SiO₂ contents in the crystal tuffs are higher than those obtained for the ash tuffs. Harker-type major-element variation diagrams show that there is a good negative correlation between SiO₂ and MgO and between SiO₂ and Al₂O₃, and a weak negative correlation between SiO₂ and Fe₂O₃(t). CaO, P₂O₅ and K₂O do not show correlation with SiO₂ (Fig. 7).

When plotted in the AFM diagram (Fig. 8), the Bom Jardim de Goiás intermediate volcanic rocks follow a calc-alkaline trend. The SiO₂ x Zr/TiO₂ diagram of Winchester & Floyd (1977) shows that the Bom Jardim de Goiás acid volcanic rocks tend to be subalkaline. The Zr/TiO₂ ratio can be used as an alkalinity index, because Zr occurs in large concentrations in alkaline rocks. In the diagram of Fig. 9, the samples fall in the andesite and rhyodacite-dacite fields. The low SiO₂ contents in some samples are due to hydrothermalism.

The Ni concentrations, as well as the Cu and Au concentrations in the ash tuffs, are higher than those in the crystal tuffs. According to Guimarães (2007), anomalous Cu contents were obtained for three samples: BJ35-110.31m, BJ34-73.80m and BJ34-70.10m (1132.5 ppm, 343 ppm and 1100 ppm,

Table 1 – Representative chemical analyses of the pink Serra Negra Granite (samples BJP4a and BJP11b1); white Serra Negra Granite (sample BJP11f); Macacos Granite (sample BJP13a); basalt intercalated with volcanic rocks (drill hole core sample BJ34-143.95); basaltic dike crosscutting pink Serra Negra Granite (sample BJP11b2); ash and crystal tuffs from barren (BJ22 and BJ 34) and mineralized (BJ 35) drill hole cores, and crystal tuff sampled in the field (BJP8a)

Sample	Granites					Basalts					Tuffs									
	BJP4a	BJP11b1	BJP11f	BJP13a	BJ34 143.95m	BJP11b2	BJ22 20.00m	BJ22 42.19m	BJ22 201.45m	BJ34 218.36m	BJ34 278.74m	BJ35 19.06m	BJ35 110.31m	BJ35 130.55m	BJP8a					
SiO ₂	72.23	75.74	75.20	73.23	51.48	45.54	63.42	66.60	70.11	64.34	65.90	63.56	62.64	67.16	63.94					
TiO ₂	0.45	0.17	0.23	0.30	0.78	2.12	0.85	0.55	0.86	0.85	0.68	0.84	0.91	0.90	0.82					
Al ₂ O ₃	13.36	12.55	12.55	13.93	14.02	14.36	17.36	15.89	14.27	15.62	15.16	16.35	15.31	14.66	16.48					
Fe ₂ O ₃	3.08	1.70	2.27	1.66	19.60	16.34	5.88	4.10	4.62	5.79	5.15	3.13	5.32	6.18	6.69					
MnO	0.06	0.03	0.02	0.02	0.29	0.19	0.14	0.07	0.06	0.03	0.05	0.10	0.08	0.10	0.16					
MgO	0.80	0.16	0.23	0.46	4.77	8.45	2.24	1.88	1.38	2.44	2.62	0.78	2.23	2.08	3.03					
CaO	1.44	0.60	1.17	0.76	1.47	11.34	1.58	3.26	1.77	2.01	2.23	3.87	2.89	1.61	2.13					
Na ₂ O	3.65	3.62	3.31	3.07	0.10	0.99	2.29	5.05	2.42	5.63	3.02	4.43	5.78	2.14	3.24					
K ₂ O	4.32	4.91	4.45	6.18	1.53	0.10	4.18	1.58	2.60	1.86	3.63	2.46	0.77	3.09	2.28					
P ₂ O ₅	0.13	0.03	0.06	0.08	0.18	0.18	0.18	0.14	0.14	0.22	0.19	0.20	0.23	0.18	0.27					
Cr ₂ O ₃	0.002	0.001	0.001	0.001	0.007	0.027	0.005	0.004	0.018	0.009	0.011	0.010	0.013	0.013	0.006					
Total	100.02	100.11	100.08	100.09	100.04	99.96	100.02	99.93	99.85	100.01	99.94	100.03	99.88	100.01	99.96					
Ba	538.8	205.7	289.7	625.9	75.3	21.4	674.1	660.1	785.0	336.5	1164.9	176.1	127.4	480.8	597.9					
Co	5.5	1.3	2.9	2.2	45.7	55.2	13.5	9.1	10.1	12.1	13	17.1	12.8	16.7	19.9					
Hf	8.1	7.9	9.7	4.6	3.9	3.7	5.3	6.6	5.1	4.2	4.8	4.2	4.2	6.8	5.3					
Nb	29.9	32.2	15	13.4	4.3	8.1	5.6	12.2	11.3	4.3	7.4	4.7	4.7	13	7					
Sn	4	4	0	1	0	1	1	2	2	0	1	0	0	1	0					
Ta	3	2.1	0.9	0.9	0.3	0.6	0.4	0.9	0.8	0.3	0.6	0.3	0.4	1	0.4					
Th	23.6	17.4	13.3	23.7	3	1	3.1	8.5	6.8	3.5	5.2	3.2	3.2	8.6	5.1					
U	9.5	4.6	2	2	0.9	0.3	18.0	12.0	15.0	17.0	15.0	16.0	19.0	15.0	19.0					
V	31	6	13	15	102	458	102	66	90	125	86	98	146	99	99					
Zr	245.3	204.6	296.5	185.1	139.5	123.8	185.6	215.3	191.3	142.9	174.3	153.3	160.7	236	181.3					
Y	99.7	54	28.5	22.3	18.5	28.9	31.1	45.1	25.4	25	32.6	23.3	26.9	29.2	37.1					
La	78.2	48.6	90.6	105.3	12.4	9.4	15.7	28.5	31.3	15.7	20.7	18.8	17.9	36.9	21.7					
Gd	12	8.76	6.61	5.06	3.57	6.05	5.18	7.53	5.21	5.16	5.66	4.35	4.68	5.38	6.84					
Tb	2.01	1.48	0.99	0.7	0.52	0.96	0.9	1.15	0.77	0.79	0.96	0.7	0.73	0.91	1.06					
Dy	12.31	9.73	5.3	3.77	3.42	5.68	5.68	7.54	4.79	4.69	5.71	4.24	4.75	5.35	6.45					
Mo	0.6	2.4	1.3	0.4	0.6	0.6	0.2	0.3	0.2	0.4	0.4	0.4	0.6	0.9	0.9					
Cu	2.1	1	9.2	2.1	12.4	220.4	21.1	12.3	2.8	31.2	37.5	342.9	1132.5	28	129.1					
Pb	21	30.5	16.3	37.2	2.6	19.1	11	8.8	9.6	112.5	6.4	3.8	10.2	8.6	17.3					
Zn	34	30	20	16	53	13	56	24	14	5	27	12	20	52	35					
Ni	3.4	0.7	2.3	2	65.1	16.3	20.1	15.1	25.8	16.6	42.9	13.9	23.5	40.7	43.1					
As	0.5	2.1	1.1	0.5	18.8	0.5	0.5	0.8	0.5	0.6	1.4	8.1	2.1	1	0.6					
Ag	0	0	0	0	0.2	0	0	0	0	0	0	0.4	0.3	0	0					
Au (ppb)	0.5	0.5	1.6	0.5	4.6	3.6	0.6	0.5	0.5	1.6	1.1	0.7	3.5	1.8	6.3

Figure 7 – Harker-type major-element variation diagrams for the volcanic rocks of the Bom Jardim de Goiás region.

Figure 8 – AFM diagram (A: $\text{Na}_2\text{O} + \text{K}_2\text{O}$, F: total FeO, M: MgO) for the intermediate volcanic rocks of the Bom Jardim de Goiás region. Line separating calc-alkaline and tholeiitic trends from Irvine & Baragar (1971).

respectively). The samples with anomalous Cu contents also show high Au contents. The anomaly can be explained by the fact that the crystal tuff represented by sample BJ34-70.10m is hydrothermalized, and shows a significant abundance of disseminated chalcopyrite. Sample BJ35-110.31m is an ash tuff that contains pyrite, chalcopyrite and magnetite, both disseminated and in veins. Sample BJ34-73.80m is a crystal tuff with disseminated pyrite and chalcopyrite (Guimarães 2007).

Hydrothermal alteration is also attested by geochemistry (Fig. 10). The trace-element variation diagrams show that for the less altered rocks there is a positive correlation between SiO_2 and Ta (Fig. 10C) and negative correlation between SiO_2 and V (Fig. 10D). Zr, Rb, Ni and Co do not show correlation with SiO_2 (Fig. 10E-H).

The rare-earth element patterns normalized to chondrite (Nakamura 1974) obtained for the Bom Jardim de Goiás volcanic rocks are very similar (Fig. 11). The patterns show high depletion in LREE and enrichment in HREE in relation to the chondrite. The patterns show weak fractionation ($\text{La}_n/\text{Yb}_n \approx 5$) and negative Eu anomalies.

The tectonic setting discrimination diagrams of Pearce *et al.* (1984) show that the Bom Jardim de Goiás acid to intermediate volcanic rocks plot in the field of rocks generated in volcanic arc settings (VAG) (Fig. 12).

Figure 9 – $\text{SiO}_2 \times \text{Zr}/\text{TiO}_2$ diagram of Winchester & Floyd (1977) for the volcanic rocks of the Bom Jardim de Goiás region.

Granites The granites cropping out in the Bom Jardim de Goiás area are classified as peraluminous to slightly metaluminous, yield MgO/TiO_2 ratios between 0.9 and 1.7, $\text{K}_2\text{O}/\text{Na}_2\text{O} > 1$ and Al_2O_3 contents from 12 to 16 wt.%, and show a calc-alkaline character.

The rare-earth element patterns normalized to chondrite (Nakamura 1974) obtained for the granites are very similar (Fig. 13). The patterns show a weak depletion in LREE and enrichment in HREE in relation to the chondrite and pronounced negative Eu anomalies for some samples. Pearce's (1996) discrimination diagrams show that the granite samples from the Bom Jardim de Goiás area can be classified as post-collisional granites (Fig. 14). Similar result is obtained when plotting these data in the triangular $\text{Rb}/100 \times \text{Y}/44 \times \text{Nb}/16$ and $\text{Rb}/100 \times \text{Tb} \times \text{Ta}$ diagrams (Fig. 15).

Basalts Table 1 presents two analyses for the basic rocks of the study area. The first analysis corresponds to the drill hole sample BJ34, in which basalt is intercalated with volcanic rocks (BJ34-143.95). The other analysis corresponds to a mafic dike that cross-cuts the pink Serra Negra Granite (BJP11b2).

The Bom Jardim de Goiás basalts are of tholeiitic character. The chondrite-normalized REE data (Nakamura 1974) yield a weakly fractionated pattern, with no negative Eu anomaly, and weak

Figure 10 – Trace-element variation diagrams for the volcanic rocks of the Bom Jardim de Goiás region.

Figure 11 – Rare-earth element patterns normalized to chondrite (Nakamura 1974) for the intermediate volcanic rocks of the Bom Jardim de Goiás region.

depletion in LREE (Fig. 16). The basalts associated with the acid to intermediate volcanic rocks show geochemical characteristics of calc-alkaline rocks generated in a volcanic arc setting, whereas the mafic dikes have intraplate basalt compositions (Figs. 17 and 18).

BIOTITE CHEMISTRY Chemical analyses of biotite from the granitic rocks and ash and crystal tuffs were carried out using the CAMECA SX-50 electronic microprobe of Universidade de Brasília. Operation conditions were 15 kV and 15 nA and beam aperture varying from 1 to 7 μm . Biotite plays an important role in petrogenesis, being a potential

Figure 12 – Tectonic setting discrimination diagrams (Pearce et al. 1984) for the acid and intermediate volcanic rocks of the Bom Jardim de Goiás region.

Figure 13 – Rare-earth element patterns normalized to chondrite (Nakamura 1974) for the granitic rocks of the Bom Jardim de Goiás region.

Figure 14 – Tectonic setting discrimination diagrams (Pearce et al. 1984) for the granitic rocks of the Bom Jardim de Goiás region.

Figure 15 – Plotting of granitic rock samples from the Bom Jardim de Goiás region in geotectonic discrimination diagrams proposed by Thiéblemont & Cabanis (1990) for syn-collisional (syn-C), post-collisional + syn-subduction (P), anorogenic non-hyperalkaline (A) and hyperalkaline (H) magmatic rocks.

Figure 16 – Rare-earth element patterns normalized to chondrite (Nakamura 1974) for the basalts of the Bom Jardim de Goiás region.

Figure 18 – Plotting of basic rocks samples from the Bom Jardim de Goiás region in the Th – Hf/3 – Ta diagram proposed by Wood (1980) for tectonic setting classification.

Figure 17 – Plotting of basic rocks samples from the Bom Jardim de Goiás region in the Zr – Ti/100 – Y*3 diagram proposed by Pearce & Cann (1973) for tectonic setting classification.

The GSN biotite is characterized by moderate FeO(t), MgO and Al₂O₃ contents, approximately 24, 8 and 12 wt.%, respectively. The FeO(t), MgO and Al₂O₃ contents in the biotite from crystal tuffs are respectively 20, 11 and 16 wt.%.

In the Mg/(Mg+Fe) x Al diagram, the GSN biotite clusters in the intermediate portion between annite and phlogopite, with Al < 2.5 (Fig. 20). In the same diagram, the biotite from crystal tuffs clusters in the field corresponding to Al around 2.5, and intermediate composition between annite and phlogopite, according to Deer *et al.* (1992).

The compositions obtained for the GSN biotite are very homogeneous, which makes it possible to define a mean structural formula representative of the biotite analyses: (K_{1.91}Ca_{0.02})_{1.93}(Fe_{3.22}Mg_{1.97}Ti_{0.43}Mn_{0.05})_{5.67}(Si_{5.75}Al_{2.25})₈O₂₀(OH,F)₄.

Figure 19 – (A) Preserved brown biotite from the Serra Negra Granite (BJP11d); (B) Brown biotite partially transformed in chlorite from the Serra Negra Granite (BJP4a); (C) Brown biotite in crystal tuff (BJ31-78.60m); (D) Chloritized brown biotite in crystal tuff (BJ22-23,87m).

No matter if from barren or mineralized drill hole cores, the biotite from crystal tuffs has an homogeneous composition, defined by the mean structural formula: $(K_{1.8}Na_{0.01}Ca_{0.01})_{1.82}(Fe_{2.54}Mg_{2.3}Al_{0.65}Ti_{0.21})_{5.71}(Si_{5.54}Al_{2.55})_8O_{20}(OH,F)_4$.

In the triangular $FeO+MnO - 10*TiO_2 - MgO$ diagram of Nachit (1986), the GSN biotite falls in the field corresponding to primary biotites, whereas the biotite from crystal tuffs falls in the field corresponding to re-equilibrated primary biotites (Fig. 21).

In the $Mg \times Al^{3+}$ diagram of Nachit (1986), which distinguishes different magmatic biotite families based on their composition, the GSN biotite falls in the field corresponding to the alkaline family, whereas the biotite from crystal tuffs falls in the field corresponding to the calc-alkaline family (Fig. 22).

According to Abdel-Rahman (1994), the biotite compositions depend on the nature of the magmas it crystallizes from. The author observed that mean

Figure 20 – Compositions of biotites from the Serra Negra Granite and from (mineralized and barren) crystal tuffs of the Bom Jardim de Goiás Deposit represented in the $Mg/(Mg+Fe) \times Al$ diagram, in which the main end members of Li-free trioctahedral micas are represented. The area delimited in the graph represents the field of the majority of natural biotites (Deer et al. 1992).

Figure 21 – Compositions of biotites from the Serra Negra Granite and from (mineralized and barren) crystal tuffs plotted in the diagram proposed by Nachit (1986) for biotites from different rock types. (A) field of primary biotites; (B) domain of re-equilibrated primary biotites; (C) domain of secondary biotites.

Figure 22 – Compositions of biotites from the Serra Negra Granite and from (mineralized and barren) crystal tuffs plotted in the diagram proposed by Nachit (1986) to distinguish different magmatic families on the basis of their Mg/Al ratios.

Figure 23 – Compositions of biotites from the Serra Negra Granite and from (mineralized and barren) crystal tuffs plotted in the $\text{FeO} - \text{MgO} - \text{Al}_2\text{O}_3$ diagram proposed by Abdel-Rahman (1994) to discriminate biotites from anorogenic alkaline suites (A); peraluminous suites, including S-type suites (P), and orogenic calc-alkaline suites (C).

FeO(t) , MgO and Al_2O_3 contents in biotites are approximately 30.6, 4.4 and 11.2 wt.% in alkaline suites; 22.1, 6.3 and 18.9 wt.%, in peraluminous suites, and 19.7, 11.2 and 14.9 wt.% in calc-alkaline suites. The GSN biotite yields mean FeO (t) , MgO and Al_2O_3 contents of 24, 8 and 12 wt.%, respectively, which match with the chemical characteristics of biotites from alkaline suites. The biotite from crystal tuffs shows chemical characteristics of biotites from calc-alkaline suite, with FeO(t) , MgO and Al_2O_3 contents of 20, 11 and 16 wt.%, respectively. This is confirmed by the triangular $\text{FeO} - \text{MgO} - \text{Al}_2\text{O}_3$ diagram, in which the analyses of the GSN biotite fall in the alkaline suite field, and those of the biotite from crystal tuffs between the calc-alkaline and peraluminous fields (Fig. 23).

GEOCHRONOLOGY AND ISOTOPIC GEOLOGY

Sm, Nd and Sr isotopic analyses were carried out using representative samples of the study area. The results are listed in Table 2. The analytical procedures follow Gioia & Pimentel (2000) for the Sm-Nd method and Buhn *et al.* (2009) for the LA-ICPMS absolute dating.

U-Pb zircon dating of an acid volcanic rock sample from the study area (BJP7b) reveals a complex pattern. Ages around 2.6, 2.2 and 1.9 Ga are interpreted as evidence of isotopic heritage, whereas the concordant age at 749 ± 6 Ma is interpreted as the crystallization age of the volcanic rock (Fig. 25).

Sample BJP4a from the Serra Negra Granite (GSN) presents a concordant population at 538 ± 17 Ma, interpreted as the granite crystallization age.

Dispersion of Nd and Sr isotopic ratios, when plotted in the $^{87}\text{Sr}/^{86}\text{Sr} \times \epsilon_{\text{Nd(t)}}$ diagram, occurs in all quadrants (Fig. 24). The tuffs concentrate in quadrant II, which correspond to rocks generated from sources of high Sm/Nd ratios. According to Faure (1986), island arc volcanic rocks have Nd and Sm isotopic compositions in this quadrant. The GSN samples fall in quadrants III and IV, characteristic of sources of low Sm/Nd ratios. The single basalt sample (dike) falls in quadrant I of the graph.

Associated tuffs and basalts yield positive $\epsilon_{\text{Nd(t)}}$ values. The tuffs present Neoproterozoic TDM model ages, which indicate juvenile sources. The basalt that occurs intercalated with tuffs (BJ34-143.95m) presents TDM model ages very close to those of the tuffs, around 0.8–1.0 Ga.

The granites yield negative $\epsilon_{\text{Nd(t)}}$ values and low Sm/Nd ratios, which characterizes melting from

a reworked continental crust. The TDM model ages for the pink Serra Negra Granite samples BJP4a and BJP11B, respectively 1.62 and 1.91 Ga, indicate melting from a dominantly Paleoproterozoic source, whereas the source for the white Serra Negra (1.13 Ga) and Macacos (1.21 Ga) granites has a dominant component derived from melting of the arc itself. These results can be interpreted as product of an older source or product of mixing of sources of different ages.

The tholeiitic basalts associated with the granites can represent crustal contamination of these magmas, with negative $\epsilon_{\text{Nd}(t)}$ values. The basalt associated with the pink Serra Negra Granite (BJP11B2) yields a TDM model age of 1.23 Ga, which means that the basaltic magma was generated after the Serra Negra granitic magma formation.

DISCUSSIONS AND CONCLUSIONS

The Bom Jardim de Goiás Deposit is hosted by a succession of ash tuffs intercalated with crystal tuffs. These tuffs possibly resulted from shallow submarine eruptions or subaerial eruption where pyroclastic material flowed into water. The ash tuffs are classified as dacites, whereas the crystal tuffs fall in the andesite and dacite fields. These tuffs present a calc-alkaline lithogeochemical character, confirmed by biotite chemical analyses. The tuffs of the study area present geochemical and isotopic characteristics of a volcanic arc setting. They are similar to the rocks from the Córrego da Aldeia Formation of the Bom Jardim de Goiás Group described by Seer (1985).

The granitic bodies that occur adjacent to the Bom Jardim de Goiás Deposit are known as the Serra Negra and Macacos granites. These calc-alkaline granites are metaluminous to peraluminous and show strong negative Eu anomalies. Tectonic setting discrimination diagrams indicate that these granites have compositional characteristics of post-collisional granites. Analyses of biotite from these granites reveal intermediate compositions between annite and phlogopite and $\text{Al} < 2.5$, which are characteristic of biotites from alkaline suites.

There are dikes of basaltic composition cross-cutting the granites that show intraplate tholeiitic basalt characteristics. Basalts intercalated with ash and crystal tuffs, interpreted as lava flows, present calc-alkaline character typical of volcanic arcs. $\epsilon_{\text{Nd}(t)}$ values for these basalts indicate that they are less contaminated than the granites.

It is possible that the igneous rocks (granites and basalts) geochemically similar to those typical

Table 2 – Sm, Nd and Sr isotopic analyses of granitic and basaltic rocks from the Bom Jardim de Goiás area

Rock/Sample	Sm (ppm)	Nd (ppm)	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr} \pm 2\sigma$	$^{87}\text{Sr}/^{86}\text{Sr}(t)$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd} \pm 2\sigma$	$\epsilon_{\text{Nd}(0)}$	TDM (Ga)	$\epsilon_{\text{Nd}(t)}$	t (Ga)
Pink Serra Negra Gr. (BJP4A)	12.11	58.70	194.30	125.90	4.481	0.743213 ± 1	0.706	0.124	0.512086 ± 6	-10.73	1.62	-5.81	0.54
Pink Serra Negra Gr. (BJP 11 B1)	9.52	38.61	181.20	33.00	14.787			0.149	0.512223 ± 12	-8.06	1.91	-4.81	0.54
White Serra Negra Gr. (BJP 11F)	9.29	60.96	97.30	57.20	4.939	0.743259 ± 2	0.702	0.092	0.512135 ± 7	-9.77	1.13	-2.6	0.54
Macacos Gr. (BJP13A)	7.16	52.38	165.00	152.50	3.131	0.70852 ± 1	0.683	0.083	0.512001 ± 21	-12.39	1.21	-4.56	0.54
Basalt assoc. with volcanic rock (BJ34-143.95B)	3.63	16.77	51.30	16.40	9.112	0.77785 ± 2	0.657	0.131	0.512439 ± 15	-3.84	1.1	2.45	0.75
Basalt assoc. with granite (BJP 11 B2)	5.13	17.72	2.60	40.20	0.187	0.707914 ± 1	0.706	0.175	0.512724 ± 7	1.72	1.23	3.19	0.54
Crystal tuff (BJ 22-20.00)	4.00	16.78	94.90	164.10	1.557			0.144	0.512607 ± 17	-0.57	0.95	4.45	0.75
Crystal tuff (BJ22-139.06)	7.29	33.65	55.00	421.50	0.378	0.708957 ± 2	0.704	0.131	0.512415 ± 15	-4.31	1.15	1.96	0.75
Crystal tuff (BJ34-73.80)	4.39	18.49	68.30	309.80	0.638	0.71022 ± 4	0.702	0.144	0.512625 ± 8	-0.21	0.91	4.86	0.75
Crystal tuff (BJ34-218.36)	4.58	19.77	46.00	274.00	0.486	0.708562 ± 2	0.702	0.140	0.512669 ± 17	0.64	0.78	6.04	0.75
Crystal tuff (BJ35-130.55)	6.48	33.86	87.80	121.40	2.096	0.72616 ± 2	0.698	0.116	0.511953 ± 15	-13.32	1.68	-5.6	0.75
Crystal tuff (BJP 7B)	5.92	25.32	35.80	895.90	0.116	0.705242 ± 1	0.704	0.141	0.512671 ± 6	0.68	0.79	5.97	0.75

Gr.: granite; assoc.: associated with

of volcanic arcs be products of magmatism generated in a setting similar to the modern island arcs and be related to the Arenópolis Magmatic Arc evolution, from a primitive island arc system that evolved to a post-collisional continental setting marked by post-tectonic, alkali-rich granitic intrusions.

The mineralization, characterized by the paragenesis pyrite + chalcopyrite $\pm$ electrum $\pm$ pyrrhotite $\pm$ magnetite $\pm$ sphalerite $\pm$ ilmenite $\pm$ hematite, is disseminated and confined to a vein system similar to the stockwork zone of volcanogenic deposits.

On the other hand, the deposit does not show alteration haloes typical of volcanogenic deposits. Only a zone of intense silicification associated with

mineralized veinlets and a more external chloritization zone with local epidotization are recognized. Even so, it is possible to affirm that the Bom Jardim de Goiás Deposit is a volcanogenic deposit localized in a stockwork zone surrounded by a silicification zone (Guimarães 2007).

The characteristics of the volcanic rocks that host the Bom Jardim de Goiás mineralization, such as successions of tuffs of intermediate composition; calc-alkaline character; chemical and isotopic characteristics of volcanic arc, and intense silicification associated with pyrite- and chalcopyrite-rich veinlets are important exploration guides for the Bom Jardim de Goiás region.

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Figure 24 – Graph of the distribution of Nd and Sr isotopic ratios for the rocks of the Bom Jardim de Goiás region.

Figure 25 – U-Pb concordia diagram for the acid volcanic rocks of the Bom Jardim de Goiás region.

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