

UNIVERSIDADE FEDERAL DE MINAS GERAIS

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Eduardo Luís Carneiro de Oliveira

**GEOLOGIA DA PORÇÃO NORTE DO CINTURÃO MINEIRO NO CONTATO COM O
BLOCO MINAS: uma abordagem litoestratigráfica, geoquímica, estrutural e
metalogenética**

Belo Horizonte

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BLOCO MINAS: uma abordagem litoestratigráfica, geoquímica, estrutural e
metalogenética**

Tese apresentada ao Programa de Pós-Graduação em Geologia da Universidade Federal de Minas Gerais como requisito parcial para obtenção do título de doutor em Geologia.

Orientador(a): Prof. Dr. Carlos Alberto Rosière

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“Geologia da porção norte do cinturão mineiro no contato com o bloco minas: uma abordagem litoestratigráfica, geoquímica, estrutural e metalogenética.”

Eduardo Luís Carneiro de Oliveira

Tese submetida à Banca Examinadora designada pelo Colegiado do Programa de Pós-graduação em GEOLOGIA, como requisito para obtenção do grau de Doutor em GEOLOGIA, área de concentração GEOLOGIA ECONÔMICA E APLICADA, pelo Programa de Pós-graduação em Geologia do Instituto de Geociências da Universidade Federal de Minas Gerais.

Aprovada em 29 de julho de 2024, pela banca constituída pelos membros:

Prof. Carlos Alberto Rosière - Orientador
UFMG

Profa. Lygia Maria Lobato
UFMG

Profa. Simone Cerqueira Pereira Cruz
UFBA

Prof. Atlas Vasconcelos Corrêa Neto
UFRJ

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RESUMO

Novos geocronológicos, petrológicos, estruturais e metalogenéticos são apresentados para a porção norte do Cinturão Mineiro, localizado no sul do Cráton do São Francisco. O trabalho concentrou em uma sequência vulcanossedimentar previamente definida como Formação Lafaiete (Guild, 1953; Pires, 1977), associada à uma evolução complexa para essa porção crustal. Essa sequência começa com uma rocha de composição basáltica à andesito basáltico, cristalizado em 2170 Ma e ϵ_{Hf} variando de -6,1 a -7,8. Em 2150 Ma há uma injeção de um novo pulso magmático, que gerou rochas vulcânicas ultramáficas da série picrito-boninito, relacionadas ao início do período de subdução. Com o avanço da subducção, forma-se uma suíte de rochas plutônicas em ambiente de arco magmático em 2140 Ma, associados à suíte Alto Maranhão, que é cortado por diques e veios de aplito com idade de 2130 Ma. Sobrepondo a sequência vulcânica-plutônica, existe uma sequência químico-clástica com duas fontes diferentes de sedimentos depositados entre 2100 e 2065 Ma. Em relação à compartimentação estrutural, três zonas principais de cisalhamento estabelecem os limites dos terrenos arqueano-paleoproterozóicos que compõem o Bloco de Minas com o Cinturão Mineiro na borda sudeste do proto continente arqueano do São Francisco (SFC) acomodando as fases acrescionária e colisional da orogênese Minas-Bahia, de idade riaciana. Do ponto de vista metalogenético, são apresentados dados petrográficos, geoquímicos de rocha total, além de análises in situ em magnetita em microsonda eletrônica e espectrometria de emissão óptica com plasma induzido por laser (LAICP-MS) que mostram o caráter magmático-hidrotermal de uma mineralização de Fe-(Cu) existente na área estudada.

Palavras-chave: Cinturão Mineiro; Formação Lafaiete; orogênese Minas-Bahia.

ABSTRACT

This work provides new geochronological, petrological, structural, and metallogenetic data for a volcano-sedimentary sequence, previously known as Lafaeite Formation, in the northern portion of the accretionary-to-collisional Mineiro Belt, southern of the São Francisco craton. The sequence begins with basalt to basaltic andesite formed in a forearc setting in 2170 Ma, showing a flat standard of REE, ϵ_{Hf} ranging from -6.1 to -7.8. In 2150 Ma, a new pulse of magmatism associated with a picrite-to-boninite series was registered. Intruded in the volcanic sequence, we present new geochronological ages from a quartz-diorite member of the Alto Maranhão Suite in 2140 Ma with an aplite dyke cutting the quartz-diorite in 2130 Ma. Overlying the volcanic sequence is a clastic-chemical sequence with two different sources of sediments deposited between 2100 and 2065 Ma. Three main strike-slip shear zones set the boundaries of Archean-to-Paleoproterozoic terranes that comprise the Minas Block with the Mineiro Belt at the southeastern edge of the Archean São Francisco Protocontinent (SFC) accommodating the accretionary and collisional phases of the Minas-Bahia orogenic cycle, with Rhyacian age. The related structural and shear sense indicators indicate a dominant oblique dextral movement. However, the confined puzzle-like regional geometry of the terranes suggests a complex kinematic history with recurrent activity along the suture zones. In the northern margin of the Mineiro Belt, we propose a magmatic-hydrothermal type for the Fe-(Cu) Água Limpa deposit.

Keywords: Mineiro Belt; Lafaiete Formation; Minas-Bahia orogen.

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1 INTRODUÇÃO

A história evolutiva do embasamento Cráton do São Francisco é marcada pela acreção de diferentes terrenos de idades arqueana à orosiriana, com seu embasamento sendo consolidado entre 2.1 Ga a 1.94 Ga durante a orogênese Minas-Bahia (Teixeira and Figueiredo, 1991; Teixeira et al., 2000; Barbosa and Sabaté, 2004; D'Agrella-Filho et al., 2020). Os paleocontinentes que compõem o sistema orogenético de idade riaciana a orosiriana existentes no cráton São Francisco podem ser divididos em três domínios distintos: o domínio orogenético do leste da Bahia (Barbosa and Barbosa, 2017), do oeste da Bahia (Cruz et al., 2016, 2024) e o segmento sul composto pelos arcos magmáticos e sequências intraoceânicas Mineiro, Mantiqueira e Juiz de Fora (Heilbron et al., 2010; Teixeira et al., 2015; Alkmim and Teixeira, 2017; Moreira et al., 2019).

Na porção sul do cráton São Francisco estão expostos dois terrenos distintos: 1) O Bloco Minas, composto por uma associação de granitos-greenstones belts com arcabouço em domos de granitos-gnaisses do embasamento arqueano e as rochas metavulcanossedimentares de uma sequência do tipo greenstone belt, o Supergrupo Rio das Velhas, desenvolvidas entre 3.2 Ga a 2.68 Ga. Essas rochas estão recobertas por rochas metavulcanossedimentares do Supergrupo Minas que foram depositadas em ambientes de margem continental passiva à ativa depositada entre 2.6 a 2.04 Ga (Machado and Carneiro, 1992; Alkmim and Marshak, 1998; Baltazar and Zucchetti, 2007; Romano et al., 2013; Lana et al., 2013; Farina et al., 2016; Dopico et al., 2017; Dutra et al., 2019); e 2) O Bloco Piedade, de idade arqueana (2.7 – 2.5 Ga), composto por rochas do tipo TTG e sanukitóide localizado à leste do Cinturão Mineiro. A sul do bloco Piedade está o Complexo Mantiqueira, cuja evolução se deu entre 2.2-2.0 Ga (Heilbron et al. 2017; Degler et al. 2018; Kuribara et al. 2019; Bruno et al. 2020; Cutts et al. 2020;), formado por rochas juvenis, acrescidas ao Bloco Piedade em aproximadamente 2.08 Ga ao longo da sutura Ponte Nova.

Localizado à sudeste do Complexo Mantiqueira, o arco juvenil Juiz de Fora, de idade de 2.1 Ga, foi amalgamado aos terrenos Piedade e Mantiqueira em 2.05 Ga, dando

origem zona de sutura Abre Campo (Heilbron et al., 2010; Cutts et al., 2020; Bruno et al., 2020, 2021). Nesse ambiente de amalgamação de arcos juvenis desenvolve-se o Cinturão Mineiro, constituído por microcontinentes, arcos juvenis e sequências intraoceânicas que foram diacronicamente amalgamadas entre 2.13 e 2.06 Ga (Heilbron et al., 2010; Ávila et al., 2014; Alkmim and Teixeira, 2017; Bruno et al., 2020, 2021; Cutts et al., 2020) (Fig. 1). Durante o processo de colagem do Cinturão Mineiro com o Bloco Minas ocorrido no fechamento da orogenia Minas-Bahia (Bruno et al., 2021), uma volumosa suíte de sanukitóides (Suíte Alto Maranhão) cristalizou-se em 2.13 Ga (Seixas et al., 2013; Moreira et al., 2018). Essas rochas são constituídas por rochas tonalíticas e dioríticas, compostas por plagioclásio, hornblenda e biotita, com uma porção variável de quartzo e pouco K-feldspato, derivam de um magma de composição de andesitos com alto teor de Mg gerados por fusão parcial de uma cunha mantélica metasomatizada em diferentes graus à partir da fusão de crosta oceânica (Seixas et al., 2013).

A porção norte do Cinturão Mineiro está em contato com o limite sul do embasamento arqueano do Cráton do São Francisco, que é constituído por um terreno paleo à neoarqueano composto por uma suíte TTG (tonalito, trondjemito e granodiorito), além de TTGs migmatizados, migmatito, metagranitóides com alto K agrupados no Complexo Bonfim. Essas rochas são separados das unidades do Cinturão Mineiro por um sistema de falhamentos compressivos lístricos que colocam em contato ambas unidades com o Supergrupo Minas. Para leste, o Cinturão Mineiro termina em um contato tectônico linear com as rochas do Complexo Santo Antônio do Pirapetinga, de idade neoarqueana (2.9 – 2.65 Ga; Raposo, 1991) ao longo da zona de cisalhamento Congonhas-Itaverava, que é transversal à zona de cisalhamento Lenheiros e delimita o contato entre o Cinturão Mineiro e o Complexo Mantiqueira.

A configuração final do Cráton São Francisco se dá após o evento collisional ocorrido entre 630 to 530 Ma (Heilbron et al., 2017) durante a orogênese ediacarana a cambriana, denominada de Brasiliana-Pan Africana.

Esse trabalho foi elaborado pensando contribuir para elucidar questões litoestratigráficas, gequímica e petrogenéticas na porção norte do Cinturão Mineiro no contato com o embasamento do Bloco Minas, bem como as feições estruturais

desenvolvidas no momento da colagem desses blocos durante a orogenia Minas-Bahia, e metalognéticas associadas à esse terreno crustal.

1.1 O CINTURÃO MINEIRO

O Cinturão Mineiro (Teixeira e Figueiredo, 1991) é um segmento tectônico da porção sul do Cráton São Francisco (Almeida 1977), constituído por uma série de arcos magmáticos e sequências intra-oceânicas desenvolvidos entre 2470 Ma a 2050 Ma, durante a orogenia Minas-Bahia (Bruno et al., 2021), em uma sequência acrescionária-colisional (Avila et al., 2014; Teixeira et al., 2015; Alkmim e Teixeira, 2017; Moreira et al., 2018; Bruno et al., 2021). As rochas plutônicas têm, em sua maioria, características geoquímicas de afinidade juvenil e se estendem ao longo de uma faixa de aproximadamente 200 km por 50 km, sendo que as sequências mais antigas se encontram na porção sul e as rochas mais novas estão localizadas na porção norte do Cinturão Mineiro (Noce et al. 2000; Ávila et al. 2014; Barbosa et al. 2015; Seixas et al. 2012, 2013; Teixeira et al. 2015). Uma considerável gama de trabalhos reportam granitoides intrusivos de afinidade cálcio-alcálica com idades entre 2.3 a 1.9 Ga (Noce et al. 2000; Cherman e Valença 2005; Ávila et al. 2006; Teixeira et al. 2015, Cardoso et al. 2019) (Fig.1).

Para as rochas supracrustais que compõem o Cinturão Mineiro são descritas quatro unidades informais (Alkmim e Teixeira 2017): Dores do Campo, Nazareno, Congonhas-Itaverava e Rio das Mortes. Em comum, essas unidades são compostas por rochas metavulcânicas definidas por metabasitos e anfibolitos e por rochas metassedimentares compostas por quartzitos, quartzo-xistos e gonditos.

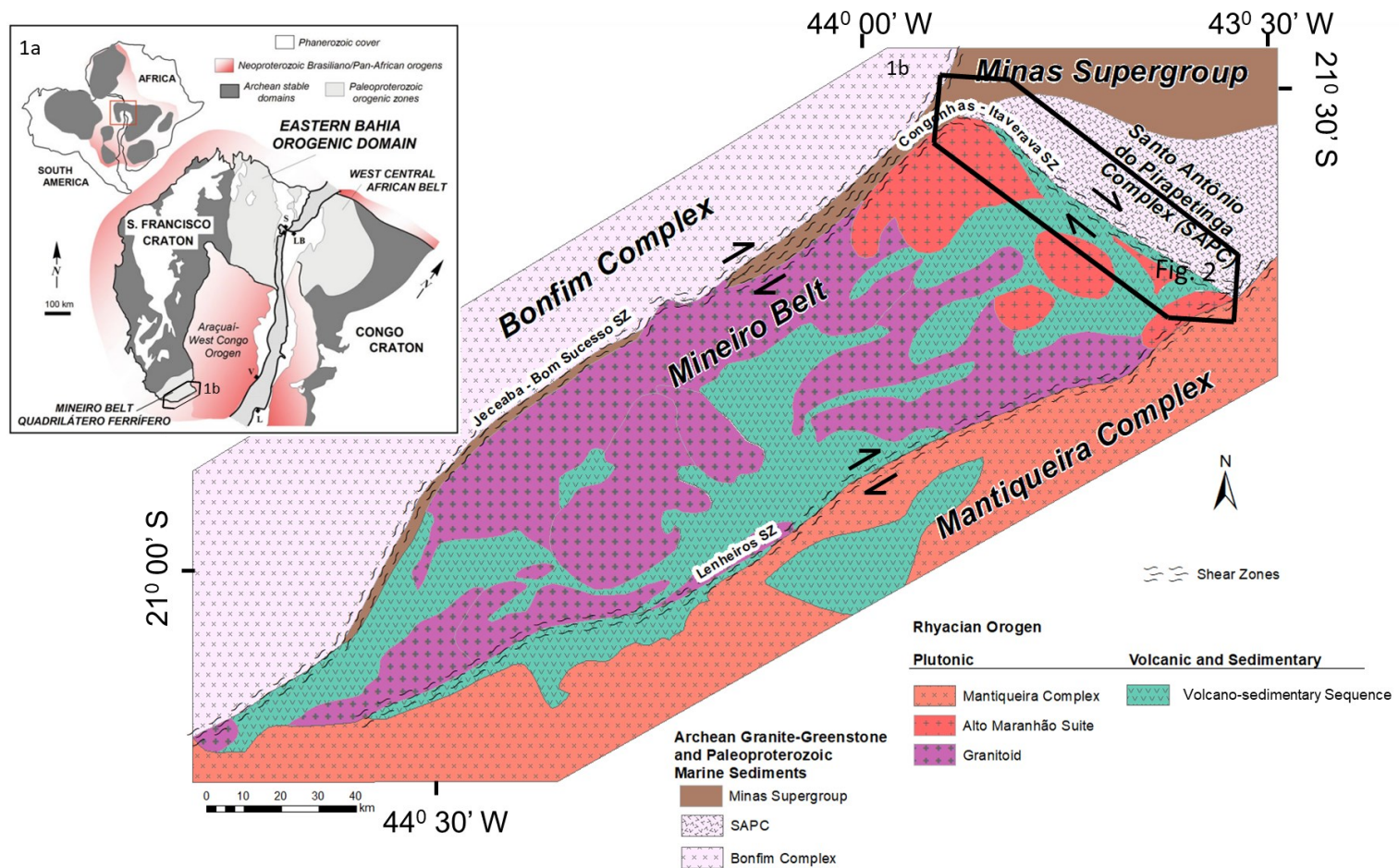


Fig. 1: Mapa geológico regional modificado de Barbosa & Barbosa (2017) e Cardoso et al. (2019). (1a) representa o contexto do Cinturão Mineiro no Cráton do São Francisco – Congo; (1b) Compartimentação geológico-estrutural do Cinturão Mineiro no contato com o Bloco Minas e o Complexo Mantiqueira, porção sul do Cráton São Francisco.

As sequências supracrustais do Cinturão Mineiro possuem como característica comum rochas com derivação vulcânica, com idades obtidas em zircões de rochas máficas entre 2.350 e 2.100 Ma (Ávila et al., 2010, 2012, 2014; Teixeira et al., 2022), bem como a presença de gnditos em todas elas.

A Sequência Congonhas-Itaverava tem aproximadamente 50 km de extensão por 3 km de largura, sendo caracterizada por uma unidade basal vulcânica máfica e ultramáfica (incluindo komatiitos), uma unidade intermediária contendo formação ferrífera bandada, marga e pelito e uma unidade superior composta por folhelho e grauvaca (Corrêa Neto et al. 2012). A sequência é intrudida pelo Corpo Congonhas de 2.195 ± 15 Ma (Seixas 2013). A idade máxima de deposição para as rochas dessa sequência foi obtida através de datação de zircões detríticos na unidade superior em 2.349 ± 14 Ma (Teixeira et al. 2015). Nas proximidades da cidade de Itaverava afloram metabasaltos que sofreram metamorfismo na fácies xisto verde baixo e apresentam características geoquímicas que sugerem ambiente de formação intra-oceânico (Teixeira et al. 2015).

A Sequência Rio das Mortes é composta essencialmente por filitos em associação com gnditos, bem como quartzitos e metabasaltos tholeíticos (anfíbolitos). Dois anfíbolitos dessa sequência foram datados por Ávila et al. (2010) em 2.231 ± 5 Ma e 2.202 ± 11 Ma.

A Sequência Nazareno ocorre na porção sul do Cinturão Mineiro e é constituída essencialmente por rochas máficas e ultramáficas, incluindo komatiitos com textura spinifex na porção basal da sequência (Toledo, 2002). As porções intermediárias e de topo incluem anfíbolitos (metabasaltos) e, subordinadamente, filitos, quartzitos e gnditos (Toledo 2002; Ávila et al. 2010). Dois anfíbolitos dessa sequência foram datados em 2.223 ± 4 Ma e 2.276 ± 14 Ma (Ávila et al. 2010, 2012).

A quarta sequência de rochas supracrustais do Cinturão Mineiro é a Dores de Campos (Ávila et al. 2004, Ávila et al. 2010) composta essencialmente por anfíbolitos, filitos, gnditos, quartzitos, com lentes de rochas metaultramáficas na base.

Entre as unidades Congonhas-Itaverava e Rio das Mortes, Ebert (1963, 1965) e Pires (1977) definiram a Formação Lafaiete associada ao Greenstone Belt Barbacena

(Barbosa, 1954; Pires, 1978; Toledo, 2002), que é composta por uma sequência vulcanossedimentar depositada entre o arqueano e paleoproterozóico, constituída por anfibolitos, metabasaltos, gouditos, formações ferríferas, grafita-xistos e quartzitos. Guild (1957) associou as rochas metavulcanossedimentares da região de Lafaiete à “Série Minas”. Candia e Girardi (1979) enfatizaram as características mineralógicas e metamórficas da Formação Lafaiete aflorante no depósito de Mn de Morro da Mina. Outros autores (Dorr II, 1969; Grossi Sad et al., 1983; Seixas, 1988; Heineck et al., 2003; Pinto e Silva, 2004) incluíram essa sequência ao Grupo Nova Lima de idade neoarqueana, enquanto Ávila et al. (2003, 2006) correlacionou as rochas vulcanossedimentares da região de Lafaiete à sequência do tipo greenstone paleoproterozóico Rio das Mortes.

O Cinturão Mineiro é balizado por três zonas de cisalhamento, desenvolvidas entre 2130 Ma a 2060 Ma (Campos, 2004; Teixeira et al., 2022; Nahas et al., 2023). Na porção NE do Cinturão Mineiro, o seu contato com o embasamento granito-greenstone arqueano do sul do craton São Francisco se dá através de um expressivo lineamento de direção NW-SE, denominado de Zona de Cisalhamento Congonhas-Itaverava, que tem mergulho forte para NE e define o limite norte dessa unidade tectônica (Fig. 1). Essa zona de cisalhamento tem deslocamento predominantemente dextral e foi parcialmente reativada, com cinemática oblíqua, reversa sinistral, durante a orogenia Pan Africana – Brasileira. O lineamento exhibe evidências de alteração hidrotermal com a presença de depósitos de ouro ao longo de todo o lineamento (Corrêa Neto et al. 2012).

Na porção sul-sudeste do Cinturão Mineiro, em contato com os arcos Mantiqueira-Juiz de Fora, existe uma zona de cisalhamento de direção NE-SW, denominada de Zona de Cisalhamento Lenheiros (ZCL) (Fig. 1), que separa as unidades plutônicas da porção norte do Cinturão Mineiro das rochas metavulcanossedimentares do arco Mantiqueira (Teixeira et al., 2008; Ávila et al., 2014). A ZCL é descrita como transcorrente dextral, com lineação de estiramento sub-horizontal desenvolvida durante a colagem riaciana e reativada na orogenia brasileira (Nepomuceno et al., 2021). Para Teixeira et al. (2022) esse cisalhamento se desenvolveu entre 2080-2020Ma, durante o período colisional do arco Mantiqueira com o Cinturão Mineiro.

No contato NW do Cinturão Mineiro com o embasamento arqueano TTG do Complexo Bonfim ocorre a Zona de Cisalhamento Jeceaba-Bom Sucesso (Fig. 1). Essa zona de cisalhamento foi responsável pela sobreposição das rochas de ambiente de margem passiva-ativa da bacia Minas, sobre as unidades do Cinturão Mineiro (Campos, 2004). Ela teria funcionado como uma zona de transferência em 2060 Ma (Campos e Carneiro, 2008).

2 OBJETIVOS

O objetivo principal desse trabalho é contribuir com uma melhor compreensão dos processos tectônicos que atuaram na fase de colagem entre os arcos magmáticos que compõem o Cinturão Mineiro e o embasamento do Bloco Minas. Os objetivos específicos foram:

(i) Caracterizar do ponto de vista litoestratigráfico e geoquímico a sequência vulcanossedimentar de Lafaiete, aflorante na porção norte do Cinturão Mineiro.

(ii) Caracterizar do ponto de vista metalogenético um depósito mineral de Fe-(Cu) existente nesse terreno crustal. Durante o trabalho foram elaborados 3 manuscritos que abordam os problemas geológicos identificados nessa porção crustal, que apresentam novos dados para elucidar os problemas, bem como discussões e conclusões a respeito dos resultados encontrados. Um desses artigos foi publicado na Journal of South American Earth Sciences e os outros dois manuscritos estão em fase de submissão e aceite para publicação.

2.1 - The Congonhas-Itaverava shear zone in the Southern São Francisco Craton (SSFC): implications for the Rhyacian amalgamation in the Minas-Bahia Orogen.

Na porção norte do Cinturão Mineiro, no contato com o embasamento arqueano do Bloco Minas, foi possível identificar um afloramento chave de quartzo-diorito associado à Suíte Alto Maranhão que apresenta uma deformação sin-tectônica associada à Zona de Cisalhamento Congonhas-Itaverava, que foi preenchida por diques de aplito. Tanto um quartzo-diorito, quanto os diques de aplito foram datados e mostraram a idade do processo colisional existente na porção sul do Cráton do São Francisco. Diante dos dados geocronológicos associados às feições geológico-estruturais dessa região, integrado com a interpretação aeromagnética da área e com os dados de campo, foi

proposto um modelo de evolução estrutural complexo para essa porção do sul do Cráton do São Francisco.

2.2 – Unveiling the Rhyacian Tholeiite-to-Boninite series in the northern portion of the Mineiro Belt, Southern São Francisco craton: Petrogenesis and Tectonic Implications

O segundo manuscrito é um trabalho de geologia regional, a partir das etapas de mapeamento geológico, coleta de amostras para descrição petrográfica e petrológica, análise química e separação de minerais pesados para datação pelo método LA-ICP-MS U-Pb e Lu-Hf, que foram realizadas no Karlsruhe Institute of Technology, Alemanha, na Universidade de Frankfurt. Ambas as análises geocronológicas e isotópicas foram realizadas em zircão.

Nesse trabalho foi possível apresentar a cartografia e caracterização, tanto do ponto de vista petrográfico, quanto químico e geocronológico, de uma sequência de rochas metavulcanossedimentares que foram descritas por Ebert (1963, 1965), Pires (1977) e Toledo (2002) como “Formação Lafaiete do Grupo Barbacena”. Essa sequência era interpretada como uma sequência arqueana. Porém, com o avançar do conhecimento científico-geológico dessa região, a mesma foi associada às sequências riacianas do Greenstone Belt Rio das Mortes (Avila et al., 2003, 2006) e posteriormente, à sequência Congonhas – Itaverava (Teixeira et al., 2015; Moreira et al., 2020), com idade variando entre 2350 a 2110Ma.

Como resultados e discussões, destacam-se as idades de cristalização magmática para um metabasalto em 2170Ma e uma rocha metaultramáfica, de composição boninítica e cristalização em 2150Ma. Rochas boniníticas são formadas nos primeiros estágios de subducção em ambientes compressivos recentes de cadeias de arcos oceânicos e tem composição de andesitos. Portanto, essas duas idades, juntamente com as duas idades de cristalização da Suíte Alto Maranhão, em 2140 Ma e 2130 Ma, de assinatura sanukitóide e é a expressão plutônica do boninito, estabelecem uma importante contribuição no entendimento dos processos relacionados à formação de uma

bacia do tipo *forearc* e o seu subsequente consumo durante o processo acrecionário-colisional no final da orogênese Minas-Bahia.

2.3 - Magnetite geochemistry and geology of the Água Limpa Fe-(Cu) Deposit: Implications for ore genesis

Propõe-se nesse trabalho uma origem do tipo magmático-hidrothermal para esse depósito, que caso se confirme, será o primeiro descrito no Cinturão Mineiro com uma importância prospectiva nesse terreno, uma vez que abre grande janela de oportunidade para descoberta de novos depósitos associados ao contexto de rochas juvenis plutônicas intrusivas em contato com rochas vulcanossedimentares, podendo gerar tanto depósitos do tipo skarn, como depósitos do tipo pórfiro Paleoproterozóico.

O depósito de Fe-(Cu) de Água Limpa possui magnetita e sulfetos de Cu e Fe no contato entre um metabasalto, formado em 2170 Ma, e um quartzo-diorito intrusivo associado à Suíte Alto Maranhão, de idade 2130Ma.

3 MATERIAIS E MÉTODOS

3.1 – Mapeamento geológico e coleta de amostras para descrições litológicas, geoquímica e geocronologia

Foram realizados trabalhos de levantamento geológico-estrutural na área estudada, a fim de cartografar as litologias aflorantes e suas relações de contato e estruturais. Do ponto de vista litológico foram identificados no mapeamento as seguintes unidades geológicas:

- uma sequência de rochas metavulcânicas de composição ultramáfica constituídas por serpentinitos, clorititos, tremolititos e hornblenditos, incluídos na unidade metaultramáfica Santana dos Montes;
- uma sequência de rochas metavulcânicas de composição máfica a intermediária, constituídas por metabasaltos a metabasaltos-andesíticos anfibolitizados denominado de unidade metavulcânica Lafaiete;
- uma sequência de rochas metassedimentares constituídas por metapelitos clásticos e carbonosos e quartzitos, por vezes com fuchcita, intercalados com as rochas metavulcânicas.
- Intrusivo na sequência vulcanossedimentar afloram granitóides de composição quartzo-diorítica a diorítica associados à Suíte Alto Maranhão de idade 2130 Ma.

Para dar subsídios químicos e petrográficos à esse empilhamento litoestratigráfico foram coletadas 42 amostras para análises químicas, 35 amostras para lâminas delgadas usadas nas descrições petrográficas e petrológicas e 07 seções polidas para análises de magnetitas e sulfetos.

3.2 – Análise geoquímica de rocha total

Foram coletadas 42 amostras de rocha que foram enviadas ao laboratório Bureau Veritas em Vancouver, Canadá, para análise de rocha total, elementos

maiores, traços e Elementos Terras Raras (ETRs) através da metodologia de fusão por lítio-borato e análise química por ICP-ES/MS e XRF.

Com os resultados analíticos foi possível classificar as litologias estudadas, seja do ponto de vista de elementos maiores, seja por elementos traços e ETR's.

3.3 – Separação de minerais pesados para datação e análises em MEV-EDS

Nessa etapa objetivou-se identificar e separar minerais que poderiam ser usados em datações radiométricas nas litologias que compõem a área de estudo.

Foram amostrados tanto rochas da sequência vulcanossedimentar para análises litoestratigráficas, quanto das rochas plutônicas existentes na área da pesquisa.

Primeiramente, as amostras de rochas metavulcânicas ultramáficas e máficas selecionadas para separação de minerais pesados que poderiam conter zircões usados na datação foram amostradas e analisadas por química multielementar. Com isso foi possível identificar as amostras que possuíam teores anômalos de Zr (>100ppm). Essas amostras foram as selecionadas para a coleta de grande volume para separação de minerais pesados.

Assim, foram coletadas 200kg de amostras de rocha metaultramáfica (boninito) e 50kg de metabasalto para britagem, moagem, concentração gravimétrica via bateia, separação magnética por ímã de mão e Frantz, e por último separação gravimétrica por líquido denso.

Em seguida foram selecionados afloramentos chave que poderiam retornar minerais usados nas análises isotópicas de U-Pb e Lu-Hf. Foram coletados em torno de 10kg de quartzito, gnaiss, granitoide e dique félsico para separação de minerais pesados. O procedimento para identificação e separação dos minerais pesados seguiu a metodologia usada no laboratório de separação de minerais pesados (SEPURA-IGC-UFMG), que é a britagem, moagem em moinhos de disco, concentração gravítica por bateia, separação magnética por ímã de mão e Frantz, separação gravimétrica por líquido denso e identificação dos minerais em lupa binocular.

As análises de MEV-EDS e imageamento por *Back Scattered Electron Imaging* (BSE) foram realizadas nos laboratórios de MEV do IGC-UFMG e da UEMG, campus João Monlevade pela Profa Flavia Silveira Braga e Dr. Joao Orestes Schneider Santos.

3.4 – Mapeamento químico em microsonda eletrônica e espectrômetro de massa LA-ICP-MS

Nessa etapa foram estudadas as magnetitas do depósito de Fe-(Cu) de Água Limpa por meio do uso de microsonda eletrônica e LA-ICP-MS de modo a poder caracteriza-lo do ponto de vista geoquímico para proposição de um modelo metalogenético.

O mapeamento químico através de microsonda eletrônica modelo JEOL – JXA 8900RL WD/ED se deu em duas lâminas polidas de formação ferrífera do depósito de Água Limpa. As análises foram feitas no Centro de Microscopia da UFMG, onde foi possível mapear os 10 elementos maiores (Mg, Al, Si, P, K, Ca, Ti, Cr, Mn e Fe) além de Ni e V.

Por sua vez, as análises de elementos traços e terras raras em magnetita do depósito de Água Limpa foram realizadas em grãos de duas seções polidas do minério de Fe-Cu, no laboratório de geoquímica do IGC-USP, através da metodologia de LA-ICP-MS utilizando-se do espectrômetro de massas com fonte de ionização por plasma do tipo quadrupolo, modelo iCAP Q – Thermo Scientific e amostrador por Laser Ablation modelo CETAC LSX-213 G2+.

3.5 – Aeromagnetometria

Foram utilizados duas etapas de processamentos para tratamento dos dados do aerolevanteamento magnetométrico da Codemig, realizado sobre a área estudada em 2001, com o objetivoda separação de domínios magnéticos, obtenção dos lineamentos magnéticos e inversão do vetor magnético.

A primeira etapa contou com processamento PPDRC (Kovesi, 2012) utilizado para ressaltar as estruturas magnéticas e traçar os lineamentos magnéticos. Na segunda etapa de processamento foi utilizado o método de MVI (Inversão do Vetor Magnético), que é utilizado para verificar a distribuição espacial da susceptibilidade magnética em 2D e 3D, considerando a natureza da anisotropia da susceptibilidade magnética em cada ambiente geológico.

3.6 – Datações radiométricas

Os estudos radiométricos seguiram as metodologias de LA-ICP-MS para isótopos de U-Pb e Lu-Hf em zircão, que serão detalhadas em cada artigo. As análises foram realizadas em colaboração com o professor Armim Zeh do Karlsruhe Institute of Technology (KIT) . As análises de isótopos de U-Pb foram realizadas no KIT e análises de Lu-Hf foram realizadas na Universidade de Frankfurt.

Foram analisadas 03 amostras de rocha vulcânica e três de rocha sedimentar que compõem a sequência vulcanossedimentar na área estudada.

Também foi analisado um granitóide formado durante o período de fechamento da orogenia Minas-Bahia.

O objetivo das análises isotópicas de U-Pb e Lu-Hf nas rochas estudadas é poder elaborar um sequenciamento de formação e deposição para proposição de um empilhamento estratigráfico na área estudada.

4 ARTIGOS DESENVOLVIDOS

Artigo 1: The Congonhas-Itaverava shear zone in the Southern São Francisco Craton (SSFC): implications for the Rhyacian amalgamation in the Minas-Bahia Orogen

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Eduardo Luís Carneiro de Oliveira ^{a,d,*}, Carlos Alberto Rosière ^a, João Paulo de Souza ^b, Armin Zeh ^c, Flávia Cristina Silveira Braga ^a.

a - Programa de Pós-Graduação em Geologia - Centro de Pesquisa Professor Manoel Teixeira da Costa – CPMTC – Instituto de Geociências – IGC – Universidade Federal de Minas Gerais.

b - GSM Geoscience.

c - Karlsruhe Institut für Technologie (KIT) Institut für Angewandte Geowissenschaften – Mineralogie und Petrologie. Adenauerring 20b, Geb. 50.40 - 76131 Karlsruhe.

d - Ero Copper Corp - 625 Howe St, Vancouver BC V6C 2T6.

* Corresponding to the author. PPGG-CPMTC-IGC-UFMG. E-mail address: eduardooliveira@erocopper.com (Eduardo Oliveira).

Abstract

Three main strike-slip shear zones define the boundaries of Archean-to-Paleoproterozoic terranes comprising the Mineiro Belt and the Minas Block at the southeastern edge of the Archean São Francisco Protocontinent. These terranes accommodated the accretionary and collisional phases of the Rhyacian Minas-Bahia orogen. The related structural and shear sense indicators provide evidence of a dominant oblique dextral movement. However, the confined puzzle-like regional

geometry of the terranes suggests a complex kinematic history with recurrent activity along the suture zones that have been reactivated during a long tectonic history until the consolidation of the São Francisco Craton in the late Cambrian.

Analyzing the aeromagnetic data surveyed by the Companhia de Desenvolvimento Econômico de Minas Gerais (CODEMIG, 2001) and applying the Phase-Preserving Dynamic Range (PPDRC) technique, followed by Magnetic Vector Inversion (MVI), it was possible to individualize three main magnetic regions separated by shear zones. Six structural domains were discriminated based on the structural data, including photo-lineaments, magnetic anomalies, and field data. The age of crystallization of the quartz-diorite plutonic body of the Alto Maranhão Suite is coeval with intruding aplitic veins filling shear fractures associated with the Congonhas-Itaverava shear zone attesting its contemporary emplacement during the amalgamation of the Mineiro Belt against the Minas Block.

Keywords

Mineiro Belt; strike-slip fault system; Minas-Bahia Orogen; aeromagnetic data; structural analysis.

1. Introduction

The accretionary-collisional Minas-Bahia Orogen (Bruno et al., 2021) developed between 2400 and 2050 Ma along the eastern-southeastern São Francisco paleocontinent. Its geological and structural features have been intensively investigated by several authors in recent decades (Noce et al., 2007; Ávila et al., 2010, 2014; Heilbron et al., 2010; Teixeira et al., 2008, 2015; Alkmim and Teixeira, 2017; Moreira et al., 2018).

The boundaries of this orogenic system are delineated by the tectonic contacts between the Archean-to-Proterozoic basement of the Minas Block with Siderian-to-Rhyacian magmatic-to-volcanic arcs and intra-oceanic sequences of the Mineiro Belt, as well as the Mantiqueira and Juiz de Fora arcs. The contacts between these orogenic

terrane encompass a system of suture zones, exposed as the major Congonhas-Itaverava (Seixas, 1988; Miethke and Silva, 2007; Corrêa Neto et al., 2013; Nepomuceno et al., 2023), Lenheiros (Ávila et al., 2010, 2014; Nepomuceno et al., 2021), and Jeceaba-Bom Sucesso (Campos, 2004) Shear Zones.

All accretionary orogens have a final collisional phase that can impart a significant structural modification to the accreted terranes, causing its partial or complete obliteration due to thrusting and extensive crustal shortening (Murphy and Nance, 1991; Cawood et al., 2009; 2016) that hampers the interpretation of its tectonic evolution. In this study, aeromagnetic survey data, as well as field and geochronological data, were analyzed to help understand the processes associated with the amalgamation of the different blocks that comprise the basement of the southern São Francisco Craton (SFC) by the time of the collisional phase of the 2.1 and 1.94 Ga Minas-Bahia Orogen. We mainly focused on the kinematic analysis of the Congonhas-Itaverava Shear Zone, aiming to establish the events' structural setting and chronology.

2 Geological Setting

2.1 Regional Geology

The geological evolutionary history of the southern São Francisco Craton (SFC) is marked by the accretion of different Archean-to-Paleoproterozoic terranes. The São Francisco Protocraton is the core of the continental area consolidated during the ca. 2.1 and 1.94 Ga ~~Transamazonian~~ (Minas-Bahia) orogeny as the homonymous paleocontinent. Dates yielded by monazite and titanite point to the metamorphism peak in ca. 2.06–2.03 Ga (Machado et al., 1992; Aguilar et al., 2017) at a late orogenic stage. This paleocontinent can be subdivided into three segments: Eastern Bahia Orogenic Domain (Barbosa and Barbosa, 2017), Western Bahia Orogenic Domain (Cruz et al., 2016, 2024) and the southern segment, comprising the Mineiro, Mantiqueira, and Juiz de Fora arcs (Heilbron et al., 2010; Teixeira et al., 2015; Alkmim and Teixeira, 2017; Moreira et al., 2019).

In the southern continental border, two main terranes can be distinguished: 1) The Minas Block (MB), comprising an Archean granite-greenstone assemblage defined by the granite-gneiss basement domes and the Rio das Velhas Supergroup, developed between 3.2 to 2.68 Ga, and covered by the metasedimentary and metavolcanic rocks of the Minas Supergroup, which were deposited in the evolving passive-to-active continental margin between 2.6 to ca. 2.04 Ga (Machado and Carneiro, 1992; Alkmim and Marshak, 1998; Baltazar and Zucchetti, 2007; Romano et al., 2013; Lana et al., 2013; Farina et al., 2016; Dopico et al., 2017; Dutra et al., 2019); and 2) The Mineiro Belt, which includes Siderian-to-Rhyacian, accretionary-to-collisional, juvenile granitoids and volcano-sedimentary sequences that evolved between 2.47 to 2.12 Ga, wedged between the Minas and the Neoproterozoic (2.7 – 2.5 Ga) Piedade Blocks which consists of TTGs and sanukitoids. The ca. 2.2-2.0 Ga juvenile rocks of the Mantiqueira Complex were accreted to the Piedade block at approximately 2.08Ga with the development of the Ponte Nova Suture Zone, and the ca. 2.1 Ga juvenile Juiz de Fora arc, located further southeast, docked with the Piedade block and the Mantiqueira arc at 2.05 Ga in the Abre Campo Suture Zone (Heilbron et al., 2010; Cutts et al., 2020; Bruno et al., 2020, 2021). The consolidation of the São Francisco craton into its final configuration and assembly of the West Gondwana continent occurred after the collisional event that lasted from 630 to 530 Ma (Heilbron et al., 2017) during the Ediacaran-Cambrian Brasiliano orogeny (Fig. 1a).

The Mineiro Belt segment makes up a roughly rectangular-shaped terrane. This archetype represents different sections of microcontinents, juvenile accretionary arcs, and intra-oceanic sequences that were diachronously amalgamated between 2130 and 2060 Ma (Heilbron et al., 2010; Ávila et al., 2014; Alkmim and Teixeira, 2017; Bruno et al., 2020, 2021; Cutts et al., 2020) (Fig. 1b). During the collage process of the Mineiro Belt with the Minas Block amid the Minas-Bahia orogeny (Bruno et al., 2021), a voluminous sanukitoid plutonic suite (Alto Maranhão) crystallized at 2130Ma (Seixas et al., 2013; Moreira et al., 2018). These rocks encompass a suite of tonalitic and dioritic rocks composed of plagioclase, minor K-feldspar, hornblende, and biotite with variable proportions of quartz derived from a magma composition of high-Mg andesites, implying they originated from the partial melting of a mantle wedge, which was metasomatized to different degrees by slab-derived melts (Seixas et al., 2013).

The Mineiro Belt's northern border corresponds to the paleocontinent's southernmost limit. This terrane is composed of Paleo-to-Neoproterozoic tonalite-trondhjemite-granodiorites (TTGs), migmatites, and high-K meta-granitoids of the Bonfim Complex, separated from the Mineiro Belt units by the imbricated tectonic slices of the Minas Supergroup. To the east, the Mineiro Belt terrane ends up in a NW-SE striking linear tectonic contact with the ca. 2900–2650 Ma the Santo Antonio de Pirapetinga Complex (Raposo, 1991) along the Congonhas-Itaverava Shear Zone, transversal to the regional NE-SW Lenheiros Shear Zone.

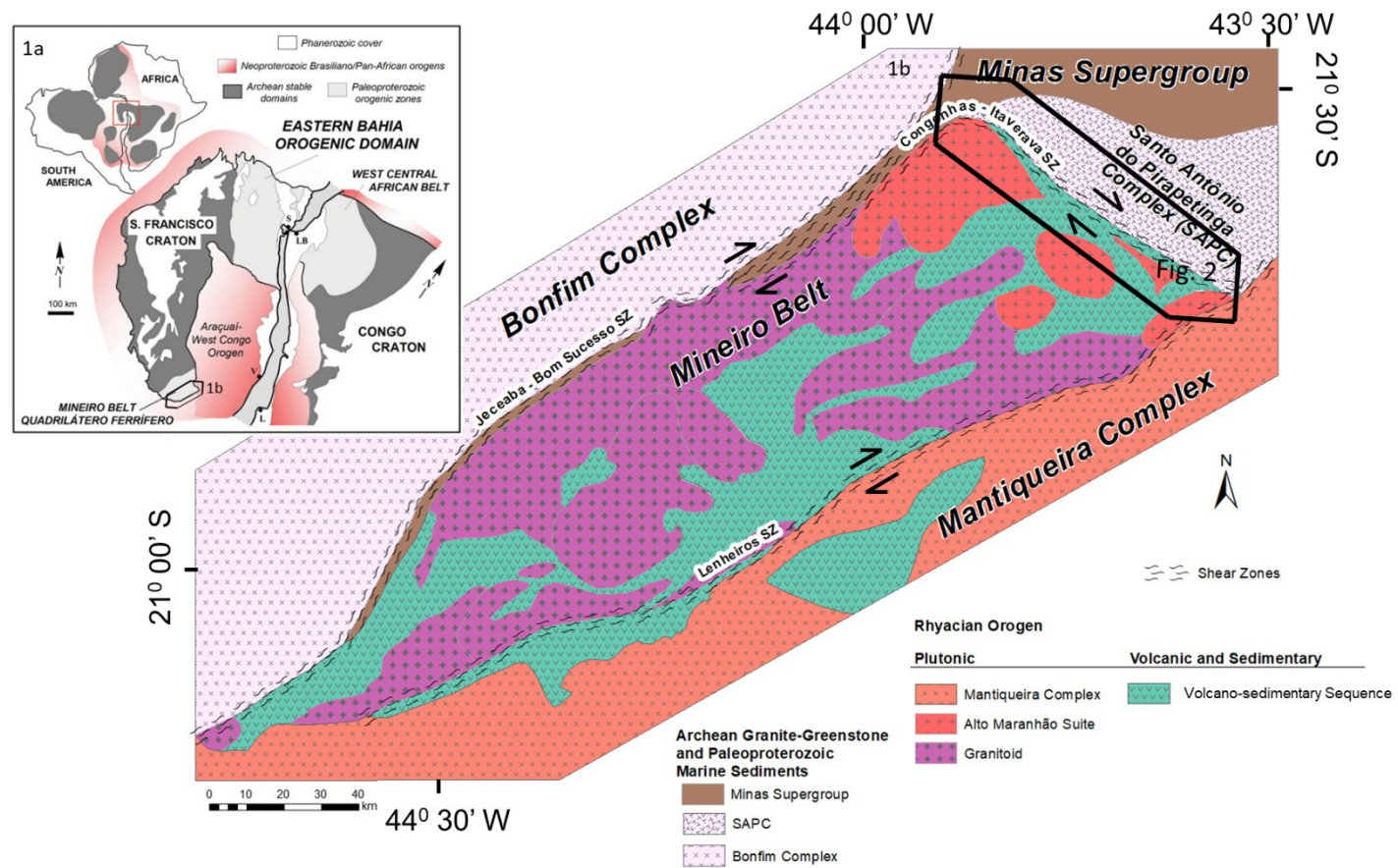


Fig. 1: Regional geological mapping, modified from Barbosa and Barbosa (2017) and Cardoso et al. (2019). (1a) represents the regional context of the São Francisco–Congo Craton; (1b) is a detail, considering the entire region.

2.2 The shear zones and the structural framework of the Mineiro Belt

Three main shear zones define the geometry of the Mineiro Belt (Fig. 1b). In the northern border, the Jeceaba-Bom Sucesso Lineament (Quéméneur et al., 2003) strikes NE-SW over 200km long, starting at the southern extremity of the Serra da Moeda Ridge near the location of Jeceaba where it ramifies in a northern branch and a southern branch that stretches eastward as the regional-scale Engenho Fault (Guild, 1957) (Fig. 2).

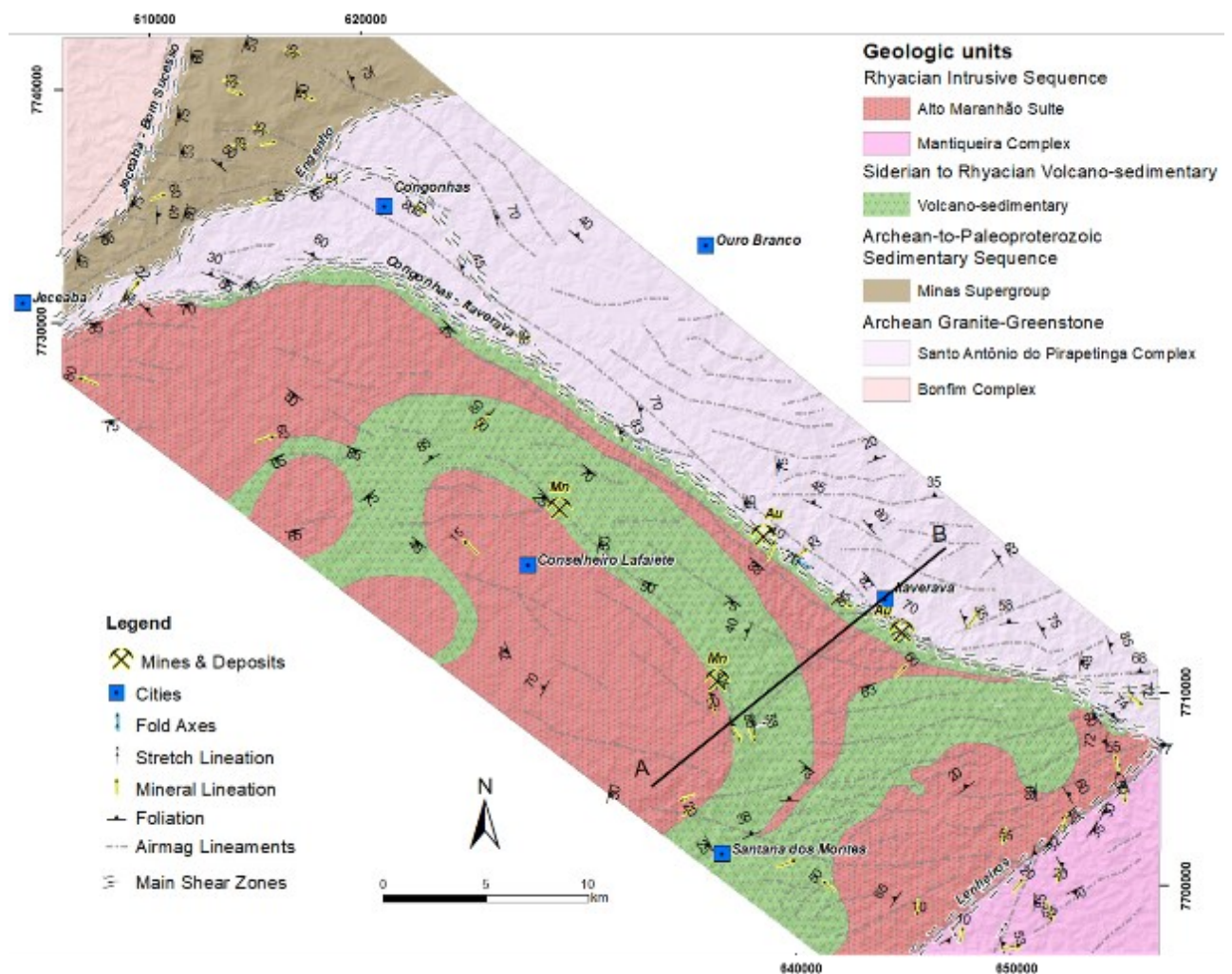


Fig. 2: Map of the local geological setting based on on-site analyses, airborne geophysical surveys (magnetic and radiometric), and the integration of regional maps taken from CPRM (Brazilian Geological Survey).

The Jeceaba-Bom Sucesso Lineament separates the Archean basement rocks and metasedimentary units of the Minas Supergroup from the Mineiro Belt, and it has been interpreted by Campos and Carneiro (2008) as a remnant of an oblique, arc-continent collision zone that culminated with a lateral displacement sub-parallel to the paleo-continental margin developed at approx. 2060 Ma (Campos, 2004).

The second major tectonic boundary is the NW-SE regional scale strike-slip Congonhas-Itaverava Shear Zone (CISZ) that coalesces in the northern extremity with the Jeceaba-Bom Sucesso Lineament and delineates the contact between the plutonic and volcano-sedimentary sequences of the Mineiro Belt to the south and southwest. The Archean Santo Antônio do Pirapetinga Complex (SAPC) is to the north and northeast. Small orogenic gold deposits, such as those documented previously by Seixas (1988), Miethke and Silva (2007), and Corrêa Neto et al. (2013), occur discontinuously along this contact zone. Nahas et al. (2023) propose a right-lateral oblique-slip kinematic during the collisional phase of the Minas-Bahia Orogen at ca. 2130 Ma, contemporaneous with the emplacement of the Alto Maranhão Suite. Corrêa Neto et al. (2013), on the other hand, proposes a reverse, sinistral, oblique-slip-type movement top-to-the-SW.

The Mineiro Belt is in fuzzy, indistinct tectonic contact with the Piedade Block at the southeast portion. It is separated from the Mantiqueira Magmatic Arc by the ill-defined Ponte Nova Suture Zone that stretches for more than 200 kilometers in the NE-SW direction. The Lenheiros Shear Zone (LSZ) is a major NE-SW-trending transpressional shear zone, where mineral (plunging between 20°- 60° to the south) and stretching (plunging 30°- 50° to the NE) lineations and shear sense criteria indicate dextral movement. In the southern Mineiro Belt, the Lenheiros Shear Zone divides two distinct metavolcano-sedimentary sequences: the Rio das Mortes (to the north) and the Nazareno-Dores do Campo greenstone belts (to the south). Nepomuceno et al. (2021) propose a dominant right-lateral movement for the structure, suggested by the sub-horizontal stretching lineation and steep SE-dipping mylonitic foliation. The Lenheiros Shear Zone was active throughout the collage between the Serrinha and Ritápolis Arcs from 2190 Ma to 2120 Ma, supporting the differential movement between distinct blocks during the accretionary-collisional transition (Ávila et al., 2010; Ávila et al., 2014; Teixeira et al., 2022).

3 Materials and methods

3.1 Aero-Magnetometry Analysis

In 2001, the Companhia de Desenvolvimento de Minas Gerais (CODEMIG) conducted an aeromagnetic survey in the study area, named “*Projeto Aerogeofísico Area 2*”. The project included a magnetometry and gamma-spectrometry survey, with lines spaced 500 meters in a NW-SE direction.

The Phase Preserving Dynamic Range Compression (PPDRC) method (Kovesi, 2012) was applied to the data to enhance the magnetic structures present in the area and determine the magnetic domains therein. The Multi-Scale Edge Detection method of identification of magnetic lineaments depends on the magnetic susceptibility contrast and the lineaments’ magnetic nature. During this method, the non-magnetic lineaments are unidentified. One way to mitigate this problem is to use monogenic filters, such as the PPDRC. This filter detects subtle features by amplifying the anomalous signal, which preserves its characteristics. Consequently, the contrast between the signatures will be enhanced.

Firstly, a high-pass filter is applied to the data to control the desired analytical scale. The 2D analytical signal data is computed to obtain a local phase and amplitude at each image point. The amplitude is attenuated by transforming the values to the logarithmic scale, and the signal is then reconstructed with the original phase values. The most significant parameter to note in this method is its cut-off wavelength during the high pass. This cut-off controls the scale of interest such that spatial frequencies in the grid with wavelengths more significant than the indicated value are suppressed, thus guaranteeing that residual features are captured more effectively (Wong et al., 2006).

In summary, it is possible to associate the obtained lineament with a cut-off wavelength using the described filtering method. This information can be applied to classify the lineaments as shallow or deep based on their cut-off values. Deeper lineaments are expected to be related to a high cut-off value, while lower values are correlated with shallow magnetic sources.

A range of wavelengths (e.g., 15, 25, 50, 100, 200, 250, and 500) could be evaluated and compared to the mapped structures and the geological interpretation for the region. Lineaments and magnetic edges were extracted automatically from the filtering results via CET Grid Analysis (Seequent®, 2019) and manually through ArcMap 10.8.2 (ArcGIS®). The extracted lineaments were then interpreted on Stereonet (11.5.1) (www.rickallmendinger.net/stereonet; Cardozo and Allmendinger, 2013).

We focused on determining the magnetic lineaments and obtaining the geometry of the Congonhas-Itaverava lineament. The Magnetic Vector Inversion (MVI) method was chosen due to the specific characteristics of the area's magnetic field. Ellis et al. (2012) proposed that the MVI technique differs from other magnetic inversion approaches because it applies the magnetic vector instead of the field's absolute value, potentially rendering a straightforward final interpretation (Macleod and Ellis, 2013).

Usually, Magnetic Vector Inversion is a solution for the spatial distribution of magnetic susceptibility in 2D or 3D, considering the anisotropic nature of magnetic susceptibility in each geological environment. Besides that, the MVI considers normal residual magnetization an essential factor when working with magnetic bodies. The results are the values of inclination, declination, and effective magnetic susceptibility (Aisengart, 2013), all of which could be very helpful in estimating structural parameters for geological exploration.

The MVI code was utilized in the Simulation and Parameter Estimation (SIMPEG) in the Geophysics database (Cockett et al., 2015). For the inversion, it was applied the original geomagnetic parameters surveyed (inclination: -25° , declination: -19° and field strength: 24256 nT), creating a 3D mesh made up of 100 m cells (measured in all directions), 1000-meter depth. To model the regional field, a padding cell of 500 meters was extended after the inversion area. The best inversion form was chosen utilizing the Tikhonov curve regularization method (Zeyen and Pous, 1991; Tsunakawa et al., 2010; Honsho and Tamaki, 2012). Results were interpreted with the Oasis Montaj® 2023 software.

3.2 Field mapping and integration data

The applied methodology consisted of geological field work and structural mapping, magnetic and geochronological data integration, satellite image analysis (satellite Aster) to generate structural lineaments, and structural interpretation using the Stereonet 11.5.1 software (Cardozo and Allmendinger, 2013). Twelve samples were taken from the plutonic and volcanic rocks in the studied area. They were analyzed to evaluate the effect of shearing in the different structural domains. Kinematic analyses were accomplished on a selected area of exceptionally good exposition of metric shear zones passing through a foliated quartz diorite with the development of a mylonitic fabric intersected by conjugated sets of shear fractures filled with sheared aplitic dykes.

3.3 Geochronology: U-Pb and Lu-Hf

Two samples were collected for isotopic analysis: one from the quartz diorite and another from aplitic dyke of the Alto Maranhão Suite. The separation of zircons followed the routine defined by the SEPURA-CPMTC-IGC-UFMG guidelines: crushing (jaw crusher), milling, pan concentration of heavy minerals, Frantz separation, and mineral identification. The zircons were mounted in epoxy and polished to expose crystal interiors. The polished grains were investigated via back-scattered electron (BSE) and secondary electron (SE) imaging using a TESCAN VEGA2 electron scanning microscope at the Mineralogy and Petrology Department of the Karlsruhe Institute of Technology (KIT), Germany.

Measurements for U–Th–Pb isotopes in zircon were performed *in situ* on polished grain mounts using a 193 nm ArF Excimer laser (Analyte Exite+, Teledyne Photon Machines) coupled to a Thermo-Scientific Element XR instrument. They were also conducted at the Karlsruhe Institute of Technology (KIT), Germany. Twenty of the zircon grains, <60 µm in length, were large enough for LA–ICP–SF–MS dating via laser spot, 20 µm in size.

Zircon grains of unknown age were dated together with the reference zircon material, BB (primary standard), Plešovice, and KA (Kaal Valley), using: a laser spot diameter of 20 µm; a laser fluence of 2.7 J/cm² at a 10-Hz repetition rate; an RF = 1255 W; a mixed, Ar–He–N₂ carrier gas consisting of Ar = 0.94 l/min, He = 0.30 cell +

0.20 cup (both l/min), and $N_2 = 12$ ml/min. Three pulses of pre-ablation, each with a 15s duration, were obtained before each analysis, with a subsequent 15s background measurement. All raw data were corrected offline for daily instrumental drift and mass offset employing an in-house MS Excel© spreadsheet program (as per Gerdes and Zeh, 2009). A Pb-common correction was also applied based on the interference- and background-corrected ^{204}Pb signal and on the model of Pb composition offered by Stacey and Kramers (1975).

Supplementary Information Table S1 shows results for the U–Pb dating of the reference zircon material and others.

The Lu-Hf isotope analyses were conducted with a Resolution M-50 193nm ArF Excimer laser system coupled to a Thermo-Scientific Multicollector—the (MC)-SF-ICP-MS (Neptune Plus) —at the FIERCE Institute, Goethe University, in Frankfurt, Germany. The employed analytical protocols are the same as those described in detail by Gerdes and Zeh (2006, including references therein). For calculation of the epsilon Hf [ϵHf], the chondritic uniform reservoir (CHUR) was used, as recommended by Bouvier et al. (2008); $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ were of 0.0336 and 0.282785, respectively; and the decay constant was of 1.867×10^{-11} (Scherer et al., 2001; Söderlund et al., 2004).

All two-stage hafnium model ages (TDM) were calculated by applying $^{176}\text{Lu}/^{177}\text{Hf} = 0.03933$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.283294$, resulting in an evolutionary line that is characteristic of the depleted mantle (DM), which ranges from zero (at 4.5 Ga) to +18 (today). Crustal evolutionary trends were modeled using $^{176}\text{Lu}/^{177}\text{Hf} = 0.0113$ for the average continental crust (values from Taylor and McLennan, 1985; Wedepohl, 1995). For all detrital zircons, initial $^{176}\text{Hf}/^{177}\text{Hf}$, ϵHf , and TDM values were calculated using the $^{207}\text{Pb}/^{206}\text{Pb}$ ages that had been previously obtained for the respective zircon domains (Table S2).

4 Results

4.1 Aero-Magnetometry

Total Magnetic Intensity (TMI) shows three magnetic domains in the study area: the southeastern domain, marked by a higher magnetic signal; the center domain, with a weak magnetic signature; and the northwestern domain, with a very strong magnetic signature (Fig. 3). The Phase Preserving Dynamic Range Compression technique was thus applied to enhance magnetic structures, better detach magnetic domains, and highlight the shear zones as well as the magnetic lineaments (Fig. 4).

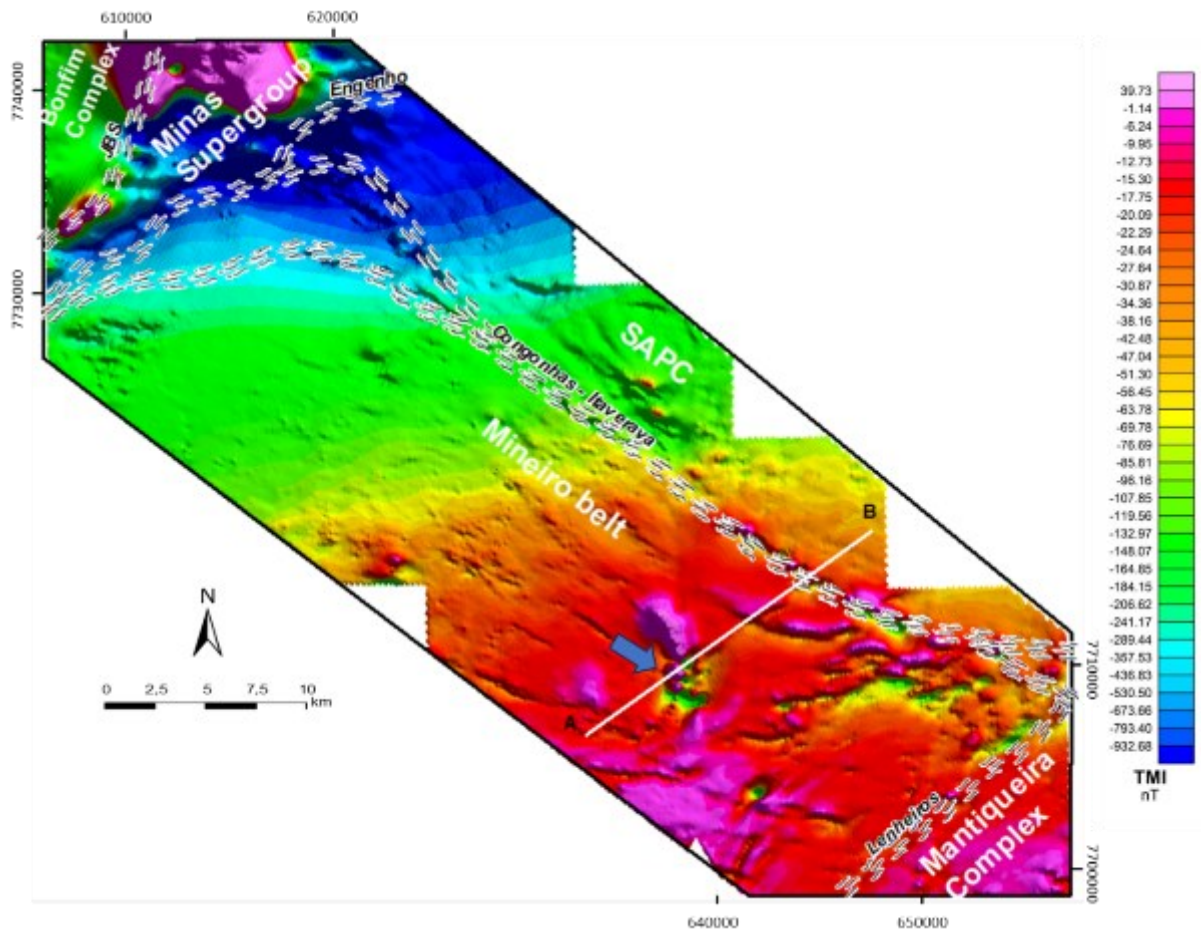


Fig. 3: Total magnetic intensity, with the main lineaments in the display. The white line indicates the section utilized for the MVI inversion. The blue arrow shows the strong magnetic anomaly chosen to be analyzed. SAPC – Santo Antônio do Pirapetinga Complex.

Three wavelengths were tested in the studied area: 875 m, 1875 m, and 3.125 m (Fig. 4 a, b, c). As expected, the 875 m PPDRC (Fig. 4a) exhibited more lineaments at a better resolution than the TMI result. Consequently, the 875 m grid was elected to separate the magnetic domains in the study area.

Therefore, in the PPDRC map (Fig. 4 d), the contact between the Bonfim complex and the Minas Supergroup is straightforward, following the Jeceaba-Bom Sucesso Shear Zone. Furthermore, the contact between the Mineiro Belt and the Santo Antônio do Pirapetinga Complex, along the Congonhas-Itaverava Shear Zone (Fig. 4 d), is better marked in the PPDRC than with TMI.

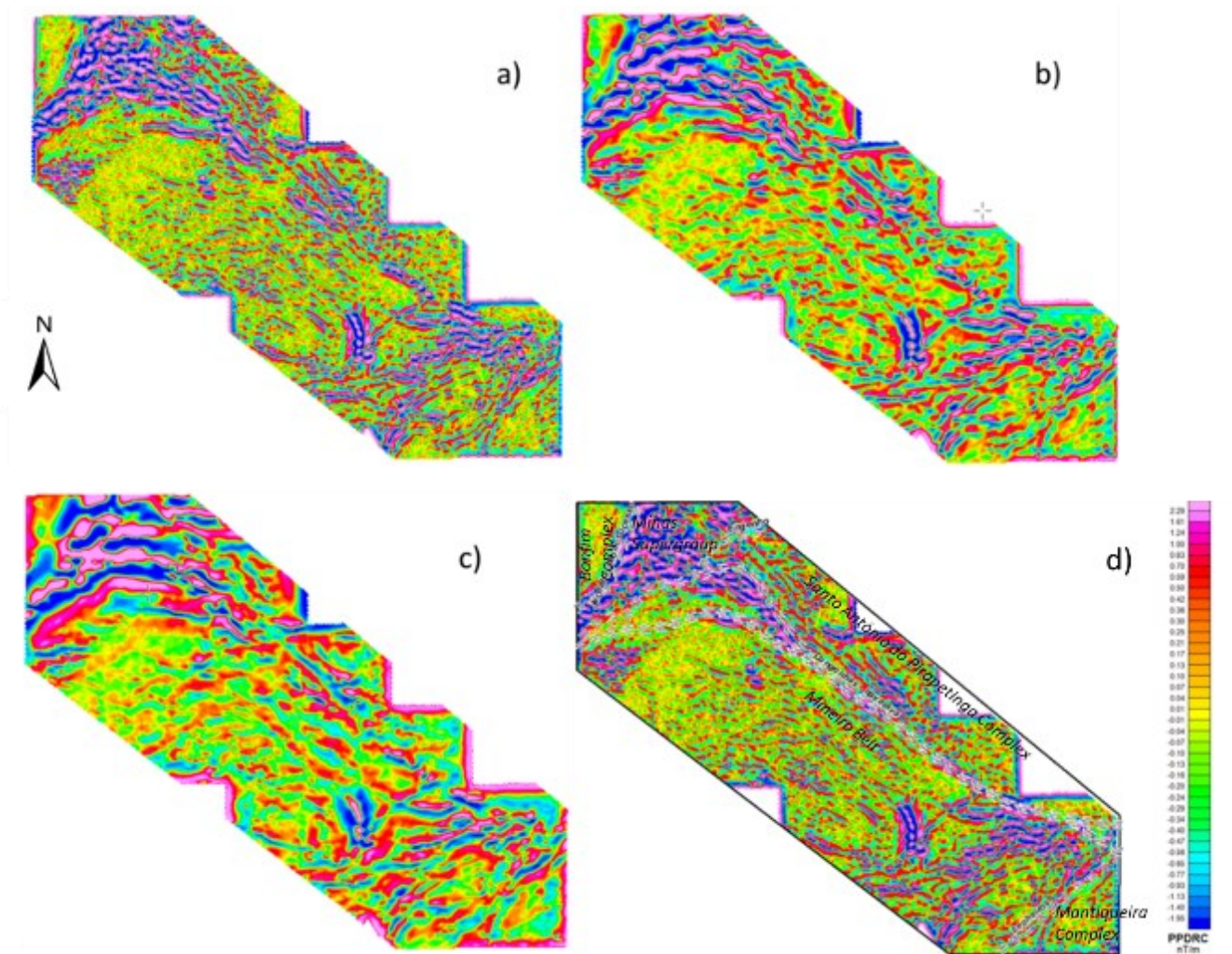


Fig. 4: Results of the PPDRC filter application: a) PDDRC for 875 m; b) 1875 m; c) 3125 m; d) Magnetic lineaments, geologic units, and the main shear zones overlaid on the PPDRC result for 875 m.

Most magnetic lineaments of the Mineiro Belt and the Santo Antônio do Pirapetinga Complex strike NW-SE (Fig. 5 and 6). They are associated with the NW-SE Congonhas-Itaverava Shear Zone (Seixas, 1988; Miethke and Silva., 2007; Corrêa Neto et al., 2011) that divide both domains. Because of the similarity of the magnetic fabric (granitic magnetic response), even for the weak magnetic zones of both

terrane, it is best to draw out the Congonhas-Itaverava Shear Zone rather than lay it onto a TMI map.

In the southeastern region, bound by the Lenheiros Shear Zone, NE-SW lineaments (Fig. 5 and 6) are the prevailing elements of the magnetic fabric. They are associated with the metavolcano-sedimentary sequence of the Mineiro Belt in contact with the granitoids of the Mantiqueira complex.

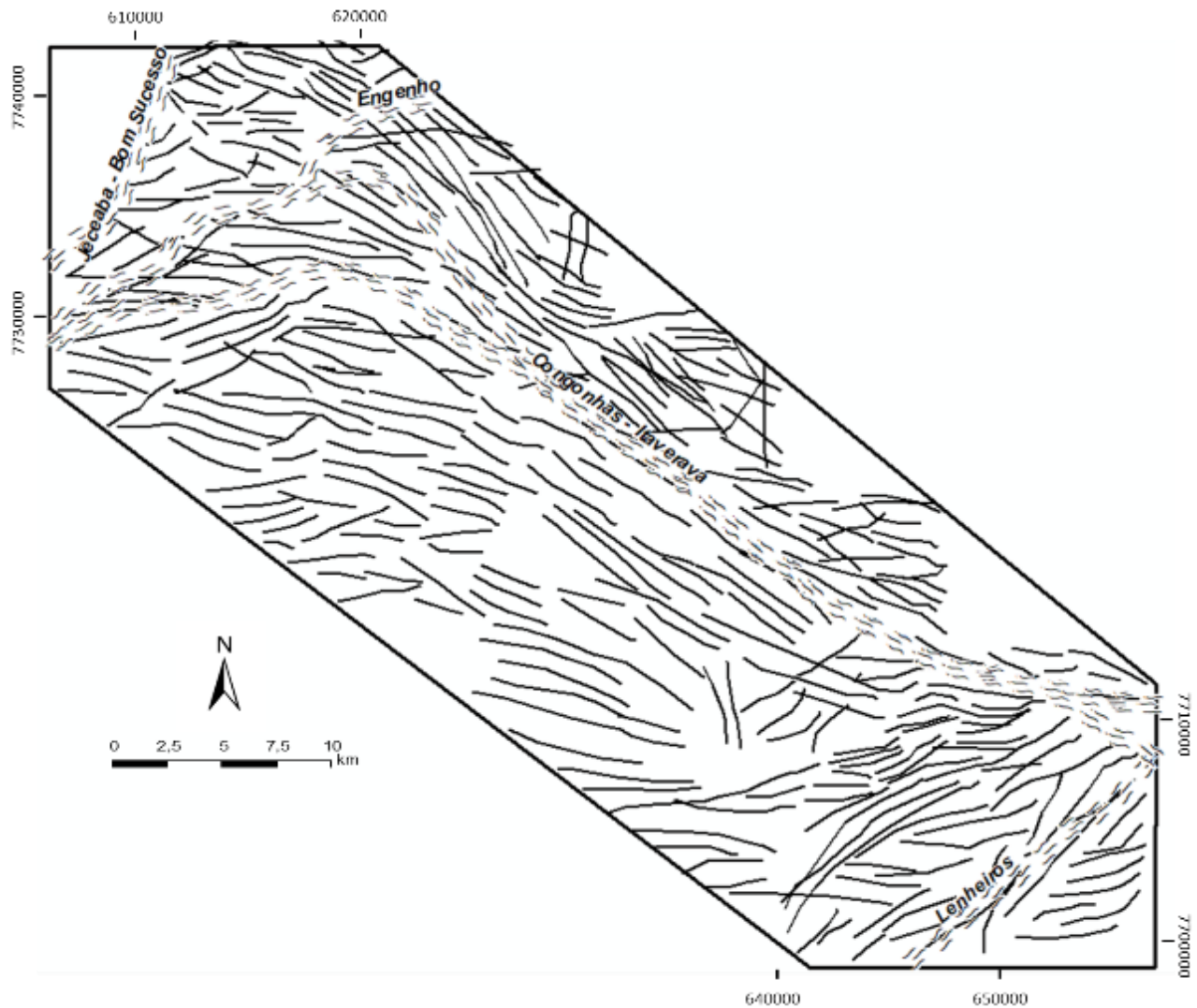


Fig. 5: Magnetic lineaments obtained for the study area.

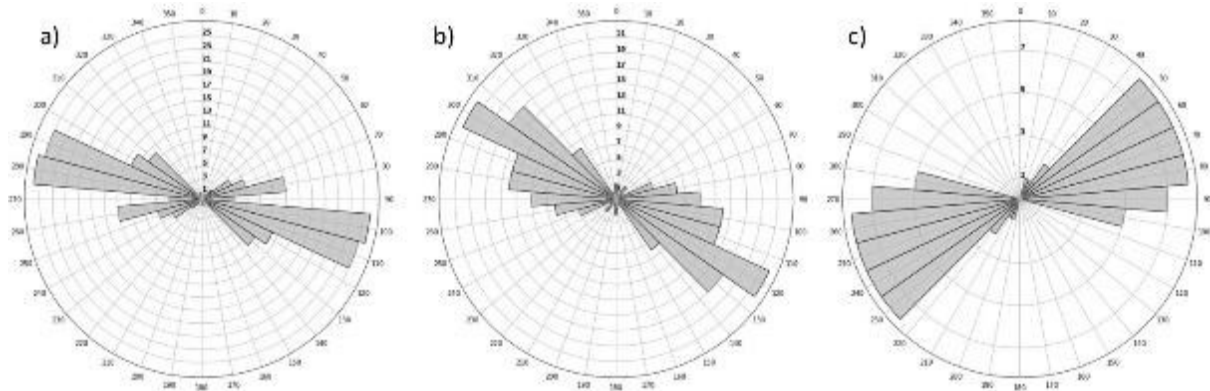


Fig. 6: Rose diagrams obtained from magnetic lineaments for: a) Mineiro Belt; b) Santo Antônio do Pirapetinga Complex; c) Mantiqueira Complex.

4.2 MVI Inversion

To determine the geometry of the Congonhas-Itaverava Shear Zone, the MVI technique was applied to a NE-SW section of the region's map (Fig. 2 and Fig. 3). The section location was chosen because two significant magnetic anomalies occur in that area, associated with a central massive magnetic anomaly.

The inversion re-enhanced the magnetic domains of the Mineiro Belt, and the Santo Antônio do Pirapetinga Complex. The magnetic susceptibility of the Congonhas-Itaverava Shear Zone is weaker than that of the metavolcano-sedimentary unit at the center of the area. However, the contact between both terranes is well marked by the shear zone (Fig. 7). It is also important to notice that the magnetic inversion dips high-angle to the NE inside the Congonhas-Itaverava Shear Zone.

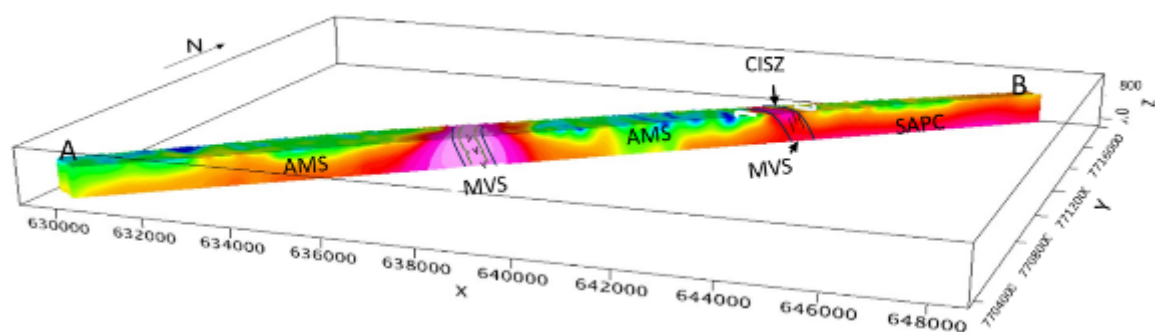


Fig. 7: MVI results for section AB including the units: MVS – Metavolcano-sedimentary sequence; AMS – Alto Maranhão Suite; SAPC – Santo Antônio do Pirapetinga Complex; CISZ – Congonhas–Itaverava Shear Zone.

4.3 Structural Domains

Six major structural domains were identified in the studied area based on the distribution and orientation of photo and magnetic lineaments and of mesoscopic planar and linear tectonic features imprinted by the regional tectonic structures (Fig. 8 and 9). Fig. 9 shows a photo-lineament integration map defined by fracture interpretation, aeromagnetic lineaments, and the structural domains defined in the geological-geophysical setting.

The *SD I* domain encompasses the southern limit of the protocraton, which includes supracrustal units of the Minas Supergroup, northwest of the Santo Antônio do Pirapetinga Complex and is bounded to the south by the NE-trending Jeceaba-Bom Sucesso and the ENE-WSW Engenho Fault. The regional foliation (S_n) strikes N-S, dipping eastward (60° - 80°), with mineral lineation plunging to E-NE (35° - 60°). The WNW-ESE is the most robust photo-lineament trend, followed by the NNW-SSE orientation. Lineaments in the NE quadrant is widely dispersed (Fig. 9). The stereoplot of the regional foliation makes up a great circle girdle comprised of two maxima with an axis parallel to the mineral lineation, suggesting the presence of recumbent folds.

The domains *SD II* and *III* south of *SD I* comprise metavolcano-sedimentary and granitoids of the Alto Maranhão Suite. On a regional scale, they differ by the inflection of the lineaments observed in *SD II* in conjunction with *SD VI*. In contrast, the prevalence of the NE-trending set of lineaments is evident in *SD III*. Besides these differences in large scale, the rose diagrams obtained for both domains are pretty homogeneous, distinguished by a robust maximum at the NW-SE direction, parallel to Congonhas-Itaverava Shear Zone, which reflects the orientation of the regional foliation (S_n), and by a second concentration NE-SW, contrasting with the main trends obtained from the data of the protocraton area (*SD I*). In domain *SD II* statistical analysis of the S_n poles with the stereogram in regional scale delivered a distribution as a diffuse small circle girdle with ill-defined axes, indicating the development of

disharmonic folds with a steeply 70° to SSW plunging axes (Fig. 8) consistent with the regional trends of the photo-lineaments (Fig. 9). The dispersion of the data is probably due 1) to the inflection observed at the concurrence of the Congonhas Itaverava Shear Zone and the Engenho Fault and 2) the interference of the Alto Maranhão plutonic body. Meso to macroscopic-scale folds in these units are not abundantly exposed because of the extensive weathering hampering an appropriate statistical evaluation of the data. Still, the field observations indicate a general dominance of the vergence towards N-NW and N-NE. In domain *SD III*, on the other hand, the foliation strikes homogeneously NE-SW, dipping NW, with a $20-60^\circ$ NW-plunging mineral lineation defining a great circle (Fig.8).

SD IV is the southernmost domain, south of the Lenheiros Shear Zone. This domain comprises granitic-gneissic rocks, and the structure is characterized by an ubiquitous NE-SW, SE-dipping regional foliation (S_n) with a mineral lineation that plunges NE to SW with variable pitch, defining a girdle distribution. The foliation is transposed to the N-S and NW-SE orientation near the Congonhas-Itaverava Shear Zone. The rose diagram plots the photo-lineaments in two main concentrations, NE-SW and NS (Fig. 9).

Domain *SD VI* is limited to the WSW by the Congonhas-Itaverava Shear Zone and encompasses the crystalline rocks of the Santo Antonio de Pirapetinga Complex. The foliation strikes NW-SE in this domain and dips $50^\circ - 60^\circ$ NE, bearing a mineral lineation that consistently plunges 45° E. *SD V* is the physical continuation of *SD VI* to the south. In this domain, the E-W S_n foliation derived from the inflection of Congonhas-Itaverava Shear Zone in contact with Lenheiros SZ is well preserved but again, partially overprinted by NW-SE structures (Fig. 8). The photo-lineaments of Domains *SD V* and *VI* plot in the synoptic rose diagram as two prominent NS and NE-SW maxima (Fig. 9) similar to the *SD IV* domain.

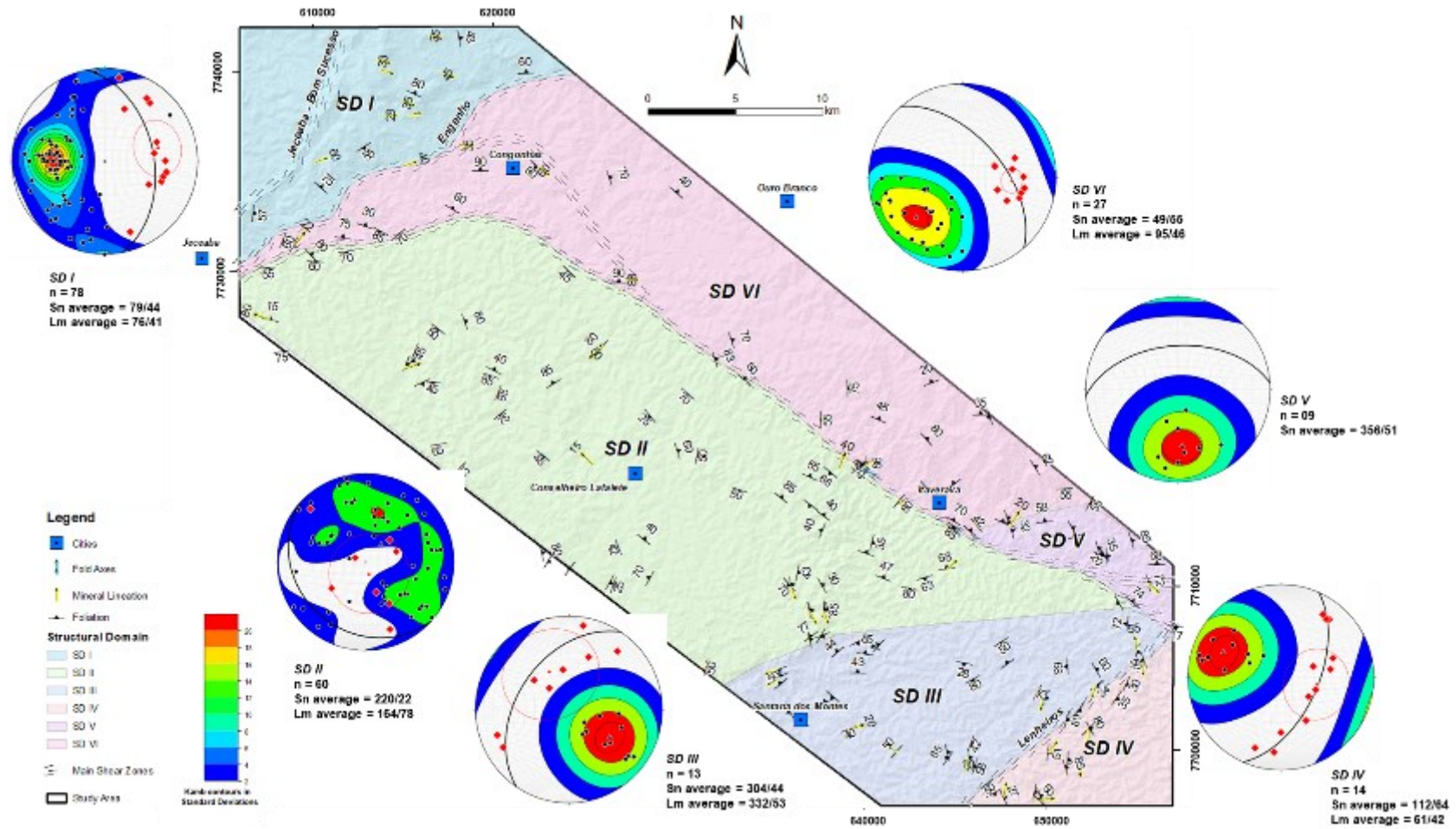
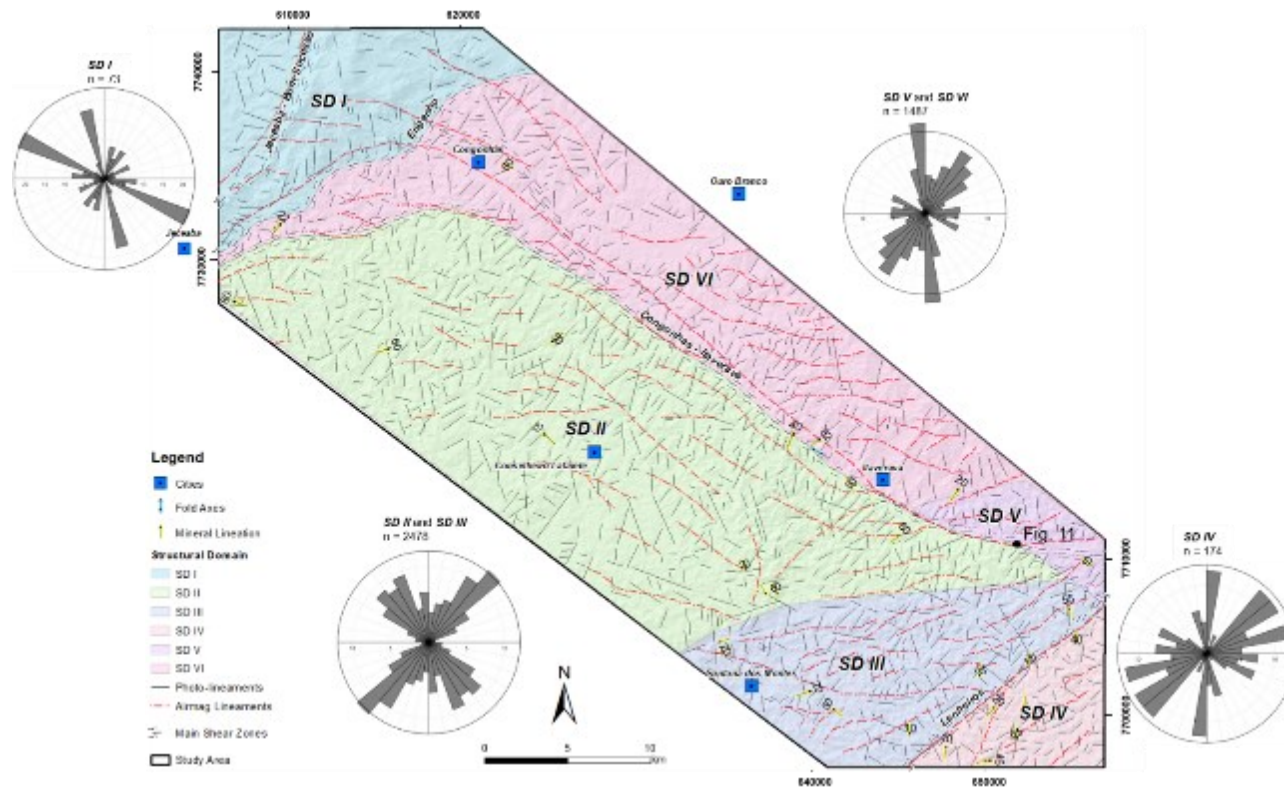


Fig. 8: Tectonic domains in the northern segment of the Mineiro Belt. *SD I* – Minas Supergroup, with an NS-striking *Sn* foliation, dipping to the east; *SD II* – NW of the Mineiro Belt, with an NW-SE-striking *Sn* foliation, dipping to the SW; *SD III* – Metavolcanic rocks of the Mineiro Belt, of a NE-SW-striking *Sn* foliation, dipping to NW; *SD IV* – Mantiqueira Complex, NE-SW-striking *Sn* foliation,

dipping to SE; *SD V* – SE of SAPC, EW-striking Sn foliation, dipping to the north; *SD VI* – SAPC, NW-SE-striking Sn foliation, dipping to the NE. The stereographic projections (equal area, lower hemisphere) show the best-fit great circle to the Sn foliations, poles to the foliation (black points), Kamb statistic analyses to the poles of foliation, and mineral lineations (red points).



4.4 The Congonhas-Itaverava Shear Zone (CISZ)

The Congonhas-Itaverava Shear Zone is a major tectonic discontinuity between the Paleoproterozoic Mineiro Belt and the Archean Santo Antonio de Pirapetinga Complex. In map view, it stretches as a straight line along 65 km in the NW-SE direction, cutting across the Alto Maranhão Pluton and metavolcanic sedimentary sequences with the development of a mylonitic foliation that overprints the regional NE-SW schistosity.

The regional fabric of the Alto Maranhão pluton varies gradually from a homogeneous massive igneous texture (Fig. 10 a) in the central parts of the intrusion to a NE-SW-striking-foliation (Fig. 10 b) parallel to the regional schistosity of the metasedimentary units. The intensity of the fabric is increasing at the proximity of major structures such as the Lenheiros Shear Zone with the development of an anastomosed foliation comprised of partially altered oriented hornblende and biotite and recrystallized quartz deflecting around cataclastic deformed feldspar grains. Close to the Congonhas-Itaverava Shear Zone, however, the NE-SW foliation is deflected, and the fabric is overprinted by a steeply dipping, well-developed, pervasive NW-SE mylonitic foliation comprising a planar shape fabric (Fig. 10 c) with a mineral lineation with variable angles of plunge from sub-horizontal to moderate values (40-50°).

To analyze the kinematics of the structure, we selected a broad area of exposition of a quartz diorite from the Alto Maranhão Suite. The microscopic protomylonitic structure is defined by fine-grained, polygonal quartz foam microstructure (Fig. 10 d, e, f), elongated mica flakes that embrace asymmetric feldspar and hornblende, epidote porphyroclasts with biotite and quartz in the strain shadows (Fig. 10 g, h, i). The local development of S-C structures and the asymmetric core and mantle fabric indicate dextral as the dominant shear movement.

The protomylonitic fabric is intersected by a set of shear fractures, partially filled by aplitic dikes. These fractures are consistent with the sheared fabric of the quartz diorite and indicate a right-lateral displacement (Fig. 11). They represent synthetic R and P arrays and the left-lateral antithetic R' shear (Riedel et al., 1929; Tchalenko and Ambraseys, 1970; Wilcox et al., 1973; Logan et al., 1979; Yu et al., 2022) and make

angles, respectively, of $\phi/2$ (R), $-\phi/2$ (P), and $90^\circ - \phi/2$ (R') with the strike of the shear zone, where ϕ is related to the angle of internal friction of the rock matrix (Ahlgren, 2001; Davis et al., 2000; Tchalenko, 1970).

The aplitic dykes have undergone deformation exhibiting cataclastic feldspar crystals and quartz grains displaying sutured boundaries and undulose extinction, with the formation of subgrains and asymmetric aggregates of new grains comprising a fine, incipient but elongated protomylonitic fabric, similar to the host plutonic rock. Sets of C-type shear bands intersect the fabric at a low angle, indicating shear senses consistent with the right-lateral dislocation of the Congonhas-Itaverava Shear Zone described by Nahas et al. (2023).

The above data, however, contrast with observations of specific outcrops (Fig. 12). The main shear planes associated with the Congonhas-Itaverava Shear Zone are NW-SE-striking, showing moderate to high dips towards the NE (Fig. 12e), which coincides with the information of the profile obtained from the aeromagnetic interpretation (Fig. 7). The mineral and stretching lineations tend to show high pitches (Fig. 12e), suggesting predominantly dip-slip displacements. However, low plunge lineations are also observed (Fig. 12d). The strike-slip motion component is often sinistral (Fig. 12a, c). The few outcrops in which it is possible to observe the dip-slip displacement component indicate reverse kinematics for the structure (Fig. 12b). All this agrees with the results obtained by Correa-Neto et al. (2013), who suggested the upthrusting of the Santo Antônio do Pirapetinga Complex against the Mineiro Belt. A possible explanation of these contrasts in the kinematics of the Congonhas-Itaverava Shear Zone, in terms of the tectonic evolution of the entire region, will be given in the discussion section.

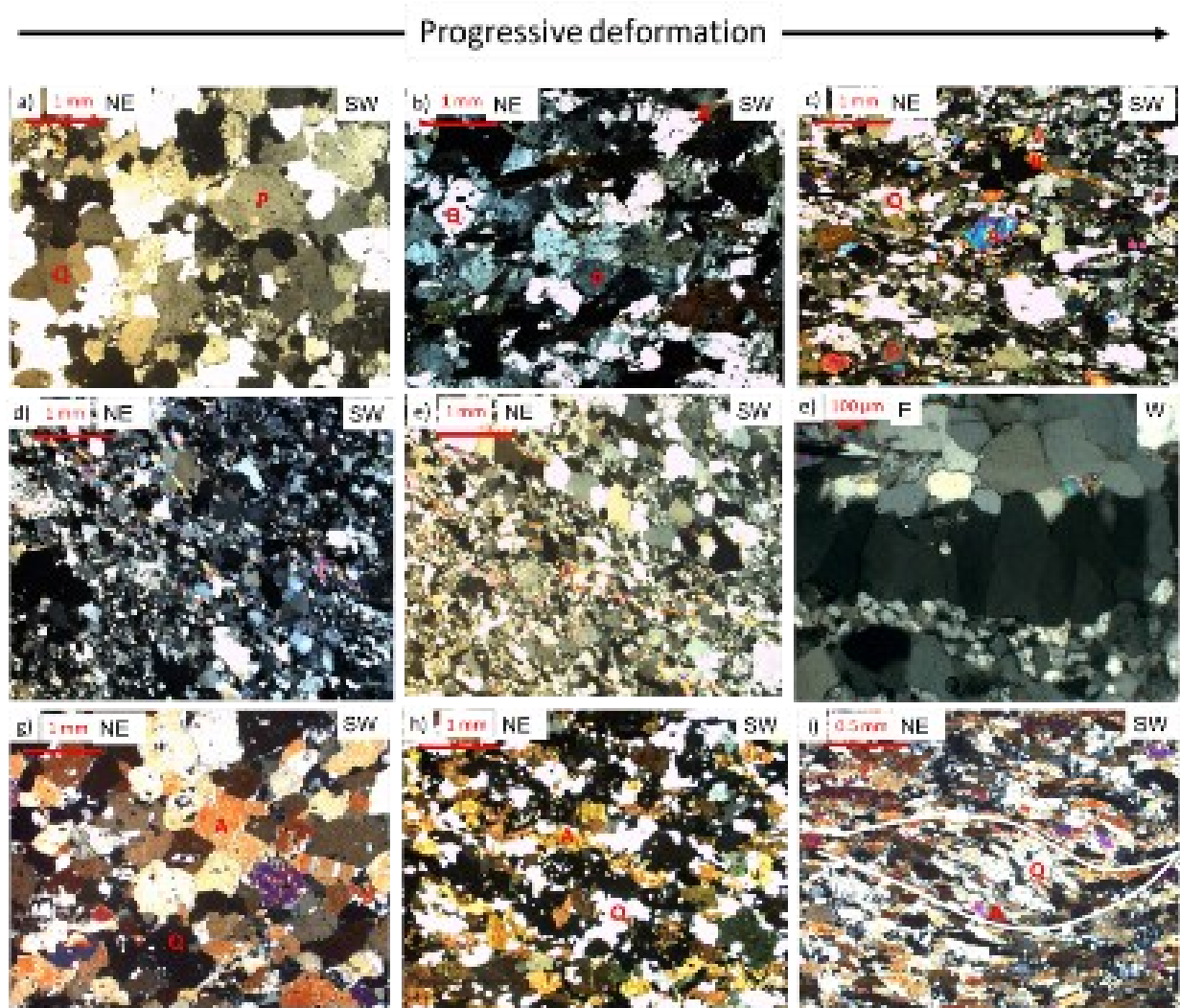


Fig. 10: a) Granoblastic quartz-plagioclase fabric of weakly metamorphic quartz-diorite of the Alto Maranhão Suite with relictic igneous texture; b) Foliated quartz-diorite from in the border of the intrusion (Alto Maranhão Suite); c) Protomylonitic fabric of sheared quartz-diorite of the Alto Maranhão Suite; d, e, f) Foam microstructure of recrystallized quartz in aplitic dyke; g) Quartz-hornblende granoblastic fabric of metavolcanic rock; h) Granolepidoblastic biotite-quartz-hornblende fabric of foliated metavolcanic rock; i) Crenulated foliation in the biotite-quartz-hornblende metavolcanic. A – Amphibole (hornblende-actinolite); E – Epidote; B – Biotite; Q – Quartz; P – Plagioclase.

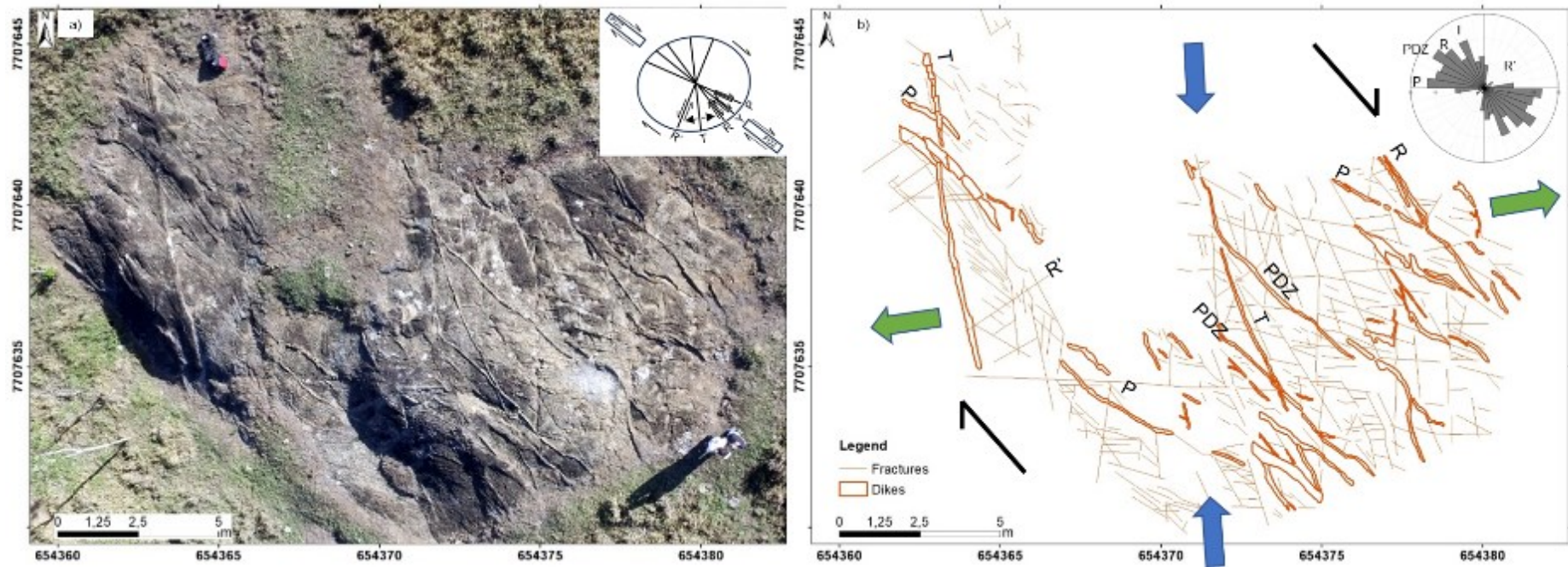


Fig. 11: a) Drone image showing a sheared syntectonic quartz-diorite of the Alto Maranhão Suite with protomylonitic fabric, intersected by the Congonhas-Itaverava Shear Zone, with fractures filled by syntectonic acid dykes. Microfabric and fracture geometry suggest a dextral displacement. b) Interpretation of the field photograph. The dikes were emplaced in open fractures of a Riedel shear system (P, T, R, and R' fractures). PDZ – Principal Displacement Zone. Blue arrows represent the σ_1 principal stress axis. Green arrows represent the σ_3 principal stress axis. Coordinates: UTM X – 654375E/UTM Y – 7707736N – DATUM Sirgas 2000, Zone 23S.

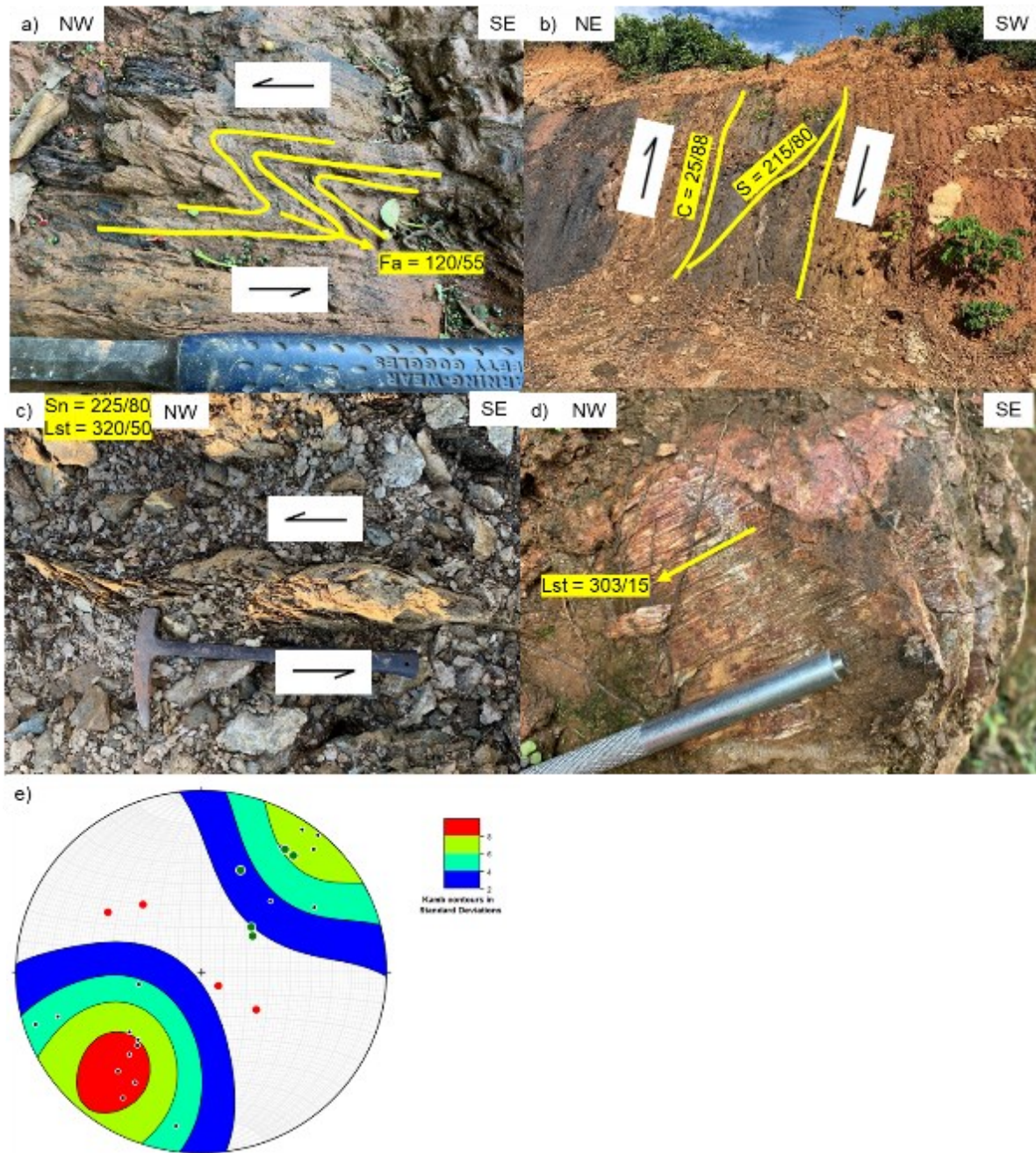


Fig. 12: a) Intrafolial NW-vergent S-fold in Banded Iron Formation (BIF) Domain *SD II*. Coordinates: UTM X 639863E/UTM Y 7715214N – DATUM Sirgas 2000, Zone 23S. Fa – Fold axis; b) Exposition of a reverse shear zone in the proximities of the Congonhas-Itaverava SZ. The structure intersects a metasedimentary sequence near the town of Itaverava. The outcrop dips toward NE, exposing an oblique section of the shear zone. Coordinates: UTM X 644934E/UTM Y 7711802N – DATUM Sirgas 2000, Zone 23S; c) Sheared, a foliated pod of chlorite schist indicating left lateral movement in chlorite-schist of SAPC, outcropped near the Congonhas town. Coordinates: UTM X 621119E/UTM Y 7732595N – DATUM Sirgas 2000, Zone 23S; d) Stretching

lineation defined by elongated mineral aggregates plunging NW (303/15) in Congonhas-Itaverava Lineament; indicating left lateral movement. Coordinates: UTM X 645040E/UTM Y 7712029N – DATUM Sirgas 2000, Zone 23S; e) Synoptic stereogram of the main structural elements of the Congonhas-Itaverava Shear Zone. The contour diagram represents the orientation of shear planes. The red dots depict the stretching lineation with variable plunges from NW to SE, and the green dots represent the mineral lineation-plunging to NE.

4.5 Age and nature of the magmatic rocks

4.5.1 U-Pb isotope analyses

One of the analyzed samples had 10 zircon grains separated from a quartz diorite of the Alto Maranhão Suite. Other 05 zircon crystals were concentrated from the aplitic dike. Zircons from both samples show similar morphologies: euhedral and elongated, ranging from 80 μm to 220 μm in length (Fig. 13). The grains exhibit oscillatory zoning, with fractures and reaction borders. In both cases, analyses were performed in the growth areas of different grains to obtain the igneous crystallization age. Results are shown in Supplementary Table S1.

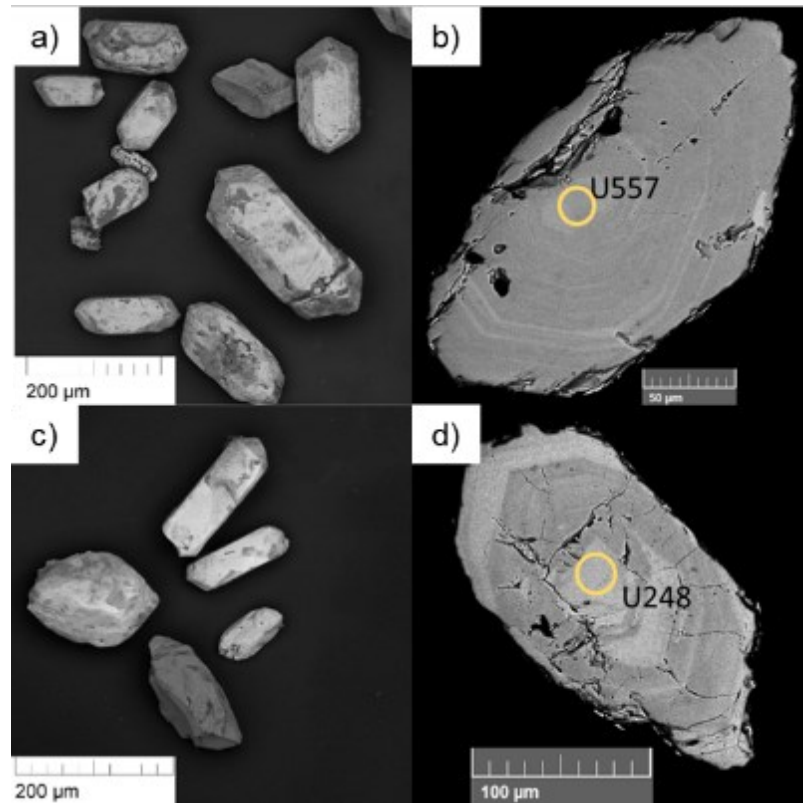


Fig. 13: Zircons analyzed in this study: a) BSE images from the group of zircons identified in the diorite sample; b) BSE images from one zircon analyzed in the diorite; c) BSE images from a group of zircons identified in the aplitic sample; d) BSE image from the zircon.

Nine of the analyzed zircons from the quartz-diorite revealed concordance values over 90% and were used to determine the crystallization age of that rock (Fig. 14 a). The age of the dyke was obtained by analyzing 06 spots exhibiting a concordance higher than 90%. The $^{206}\text{Pb}/^{238}\text{U}$ analysis of the quartz-diorite led to a Concordia age of 2140.7 ± 15 Ma (MSWD = 3.2, 2σ), which overlaps the age obtained to the aplitic dyke (Fig. 14 b) in 2131.4 ± 9.5 Ma (MSWD = 0.0099, 2σ), considering the error within the analyses.

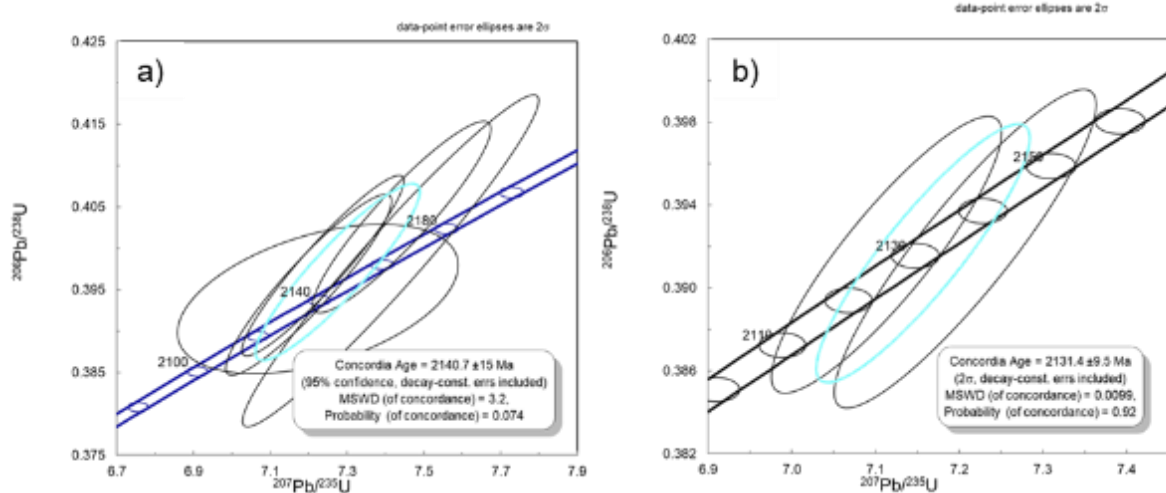


Fig. 14: Concordia diagrams (2σ ellipses error) for zircon U-Pb LA-ICP-MS analyses: a) Diorite; b) Aplitic dyke.

4.5.2 Lu-Hf isotope analyses

Analyses of zircons from quartz-diorite showed two distinct clusters. The ϵHf values obtained from one group vary from +3.4 to -1.3, which attests to the juvenile origin of these rocks with a Paleoproterozoic T_{DM} age of 2300 Ma. The values of the other group range from -4.9 to -6.9, correlated with the cluster yielded from the zircon grains collected from the dyke, which range from -6 to -8, indicating crustal contamination. The difference in the ϵHf values is possibly due to infracrustal reworking during collision (Moreira et al., 2018), when juvenile magmas were likely contaminated by mixing with the pre-existing crust of the Mineiro Belt or older crustal-derived segments at the Archean margins like the Santo Antônio do Pirapetinga Complex (Barbosa et al., 2015) (Fig. 15).

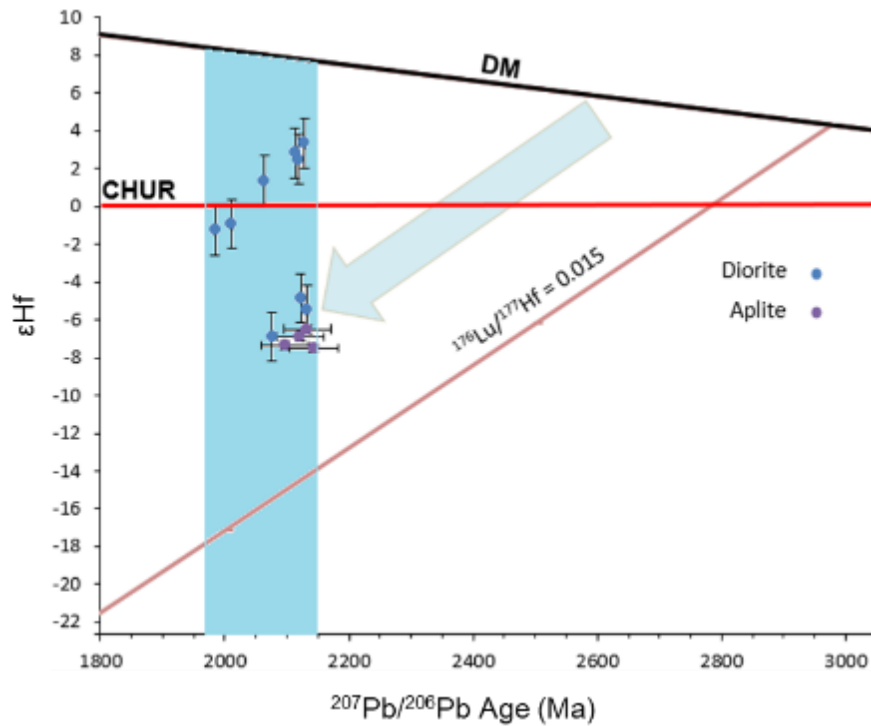


Fig. 15: Lu-Hf results for the analyzed samples. DM – Depleted Mantle model from DePaolo (1981); CHUR – Chondrite Uniform Reservoir. The light blue arrow shows a crustal evolution of Archaean-to-Proterozoic zircons calculated using a crustal $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$ (Dhuime et al., 2011; Goodge and Vervoort, 2006; Griffin et al., 2002).

5 Discussion

5.1 Aero-Magnetometry

The aeromagnetic survey pointed to a strong magnetic anomaly in the NW portion of the studied region, over the Minas Supergroup and between the Jeceaba-Bom Sucesso and Engenho lineaments. The sources of this magnetic anomaly in *SD I* can be associated with the iron formation of the Minas Supergroup. In the domain *SD II*, the stronger magnetic field could represent the metavolcano-sedimentary sequence. On the other hand, the region with a weaker magnetic signature can potentially correspond to the Alto Maranhão Suite.

The *SD III* shows a distinctive magnetic signature. An elongated magnetic anomaly, 5 km long and striking in the NW-SE direction with 200 nT of amplitude, can be observed in this region. Such an anomaly can be correlated with the metavolcanic units associated with the intra-oceanic sequences from the northern part of the Mineiro Belt.

The high magnetic amplitude signals caused by the iron formation from the northern part of the Mineiro Belt mask the low magnetic amplitude anomalies caused by magnetic lineaments. Therefore, we cannot assume that the central portion of the area does not show any significant magnetic anomaly.

The magnetic anomalous domain in the SE portion of the studied area is parallel to the LSZ. The second direction observed in this area can be associated with the collision of the Mantiqueira Arc, which was reactivated during the Neoproterozoic during the Araçuaí Orogen (Noce et al., 2007; Kuribara et al., 2019).

5.2 Structural setting and consequences

The tectonic structures in the northern portion of the Mineiro Belt are consequential to the accretionary-collisional history of the Minas-Bahia Orogen, which occurred approximately in the period 2130 to 2070 Ma (Aguilar et al., 2017 and Moreira et al., 2019). The framework of the continental crust at the southern border of the São Francisco protocraton most likely played an important role.

The development of this orogen is well documented by many authors (e.g., Noce et al., 2000; Teixeira et al., 2000, 2015, 2022; Ávila et al., 2010, 2014; Barbosa et al., 2015; Moreira et al., 2018), representing a classic evolution of an arc sequence with supracrustal domains, and juvenile plutonic rocks emplaced as a product of oceanic subduction and partial mantle melting (Teixeira and Figueiredo, 1991; Noce et al., 2000; Ávila et al., 2010; Ávila et al., 2014; Alkmim and Teixeira, 2017; Cardoso et al., 2019). The collisional system of this orogen, however, is characterized by three main transcurrent to transpressive shear zone that delineate the contact of the allochthonous volcano-sedimentary sequences of the Mineiro Belt with the paraautochthonous and autochthonous foreland terranes: the northeastern portion is

sharply delimited by the Congonhas-Itaverava Shear Zone. At the same time, the Jeceaba-Bom Sucesso Shear Zone stretches along the north-northwestern limit. On the other hand, the southeastern boundary of the Mineiro Belt units with the continental Piedade block is less evident, but the Lenheiros Shear Zone broadly controls the regional structure.

The Congonhas-Itaverava Shear Zone is a long-living structure that reflects the primary boundary conditions of the basin. The first record of compressive tectonic activity associated with this structure is very early in the history of the Minas Bahia orogeny in the southern São Francisco Craton, as indicated by the age obtained from the syntectonic aplitic dykes (2131.4 ± 9.5 Ma), which is coeval with the crystallization of plutonic quartz-diorite of the Alto Maranhão Suite. The analyses of shear sense criteria in different scales indicate a prevailing right-lateral movement for the structure. However, different shear senses have been locally observed and reported (Correia Neto et al., 2013), suggesting multiple recurrent lateral displacements of the Mineiro Belt's terranes against the continent along the shear zone. This is illustrated in Fig. 12 as a consequence of its reactivation along the tectonic history of the southern border of the protocraton in successive events from the Paleo to Neoproterozoic.

The tectonic activity along the Congonhas-Itaverava SZ has probably compensated for the frontal collision of the allochthonous terranes against the southern continental border. This was counterbalanced by the evolution of Jeceaba-Bom Sucesso Shear Zones, which guided the colliding units towards a lateral scape in a similar manner to a sliding puzzle, with the development of an anastomosing strike-slip dextral system of faults. According to Campos (2004), the Jeceaba-Bom Sucesso Lineament developed around 2060 Ma probably represents a late stage of orogenic history.

The transpressional Lenheiros Shear Zone has probably developed in a late collisional stage associated with the docking of the Mantiqueira and, later, of the Juiz de Fora Complexes against the Mineiro Belt (Noce et al., 2007; Heilbron et al., 2010; Aguilar et al., 2017; Kuribara et al., 2019). Teixeira et al. (2022) propose that the Lenheiros SZ is contemporaneous with the development of the Jeceaba-Bom Sucesso Lineament between 2080 and 2060 Ma and with the syntectonic granitoids of the

colliding terranes (Silva et al., 2002; Noce et al., 2007; Heilbron et al., 2010; Aguilar et al., 2017).

The structure of the terranes affected by the collision of the Mantiqueira Complex in the southern border of the Mineiro Belt is characterized by a SE-dipping, NE-SW striking foliation, a pervasive mineral lineation plunging 20°- 60° to the south, and a second stretching lineation defined by elongated porphyroclasts that plunge 50° to NE indicating W-NW-upwards movement and coinciding with the structures associated with the Neoproterozoic collage of the Brasiliano-Pan African, West Congo-Araçuaí Orogen, which took place between 620-520 Ma (Pedrosa-Soares et al., 2001; Noce et al., 2007; Kuribara et al., 2019). In this area, it is also evident that the recurrent tectonic activity reactivated the NE-SW and NS-trending structures associated with lineaments in the domains *SD IV*, *V*, and *VI*.

In summary, according to the available geochronological data and analyzed tectonic structures, the tectonic evolution can be described in the following two main stages (Fig. 16):

1.

Subduction of oceanic crust and magmatic activity with crystallization of the sanukitoids that compose Alto Maranhão Suite dated from ca 2130 Ma. The magmatic activity was contemporaneous with the docking of juvenile terranes and lateral displacement on a transfer zone defined by the Congonhas – Itaverava dextral shear zone.

2.

The period of 2.08-2.06 Ga records the accretion of the Mantiqueira Complex against the Mineiro Belt terranes and the evolution of the Lenheiros dextral shear zone and the Jeceaba-Bom Sucesso SZ (2060 Ma - Campos, 2004) at the suture zone between the Mineiro Belt and the foreland basement of the Minas Block.

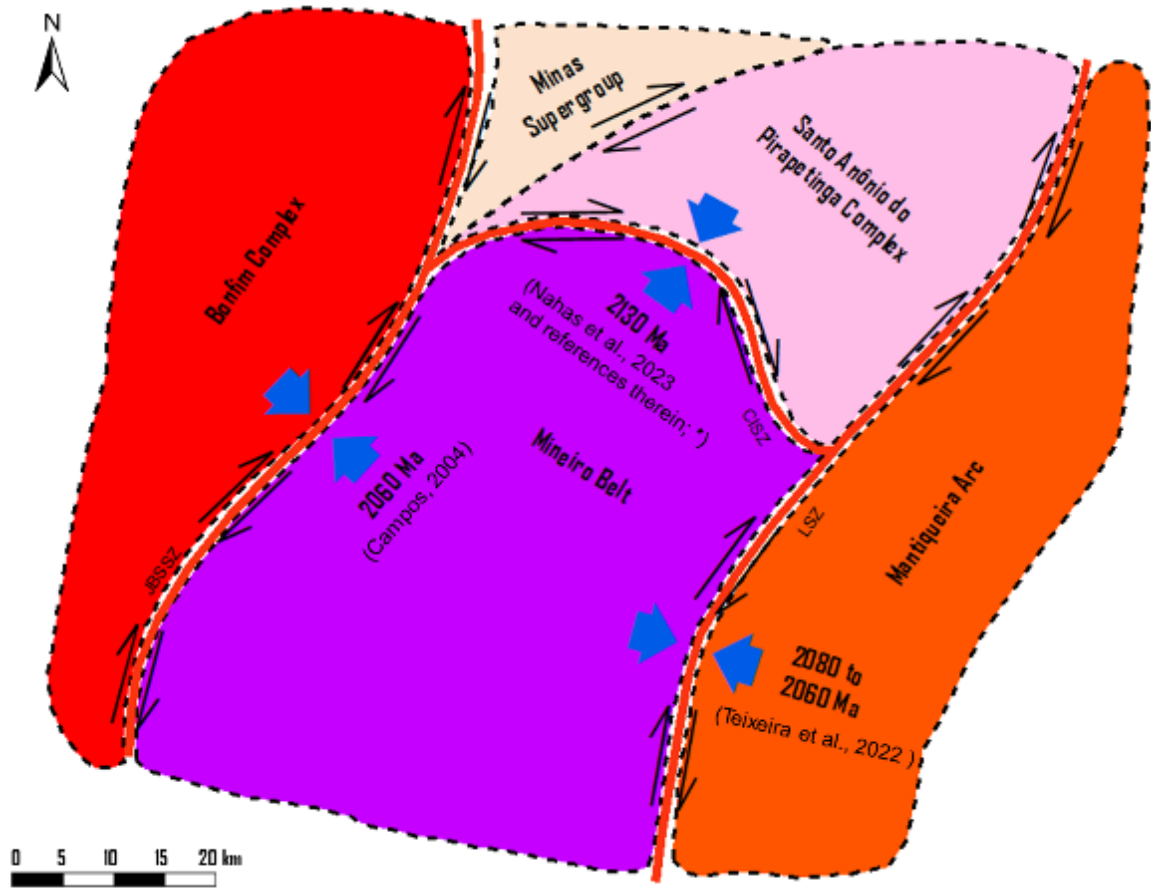


Figure 16: Collisional sketch of the SFC with the ages of the collisional process defined by the shears on the borders of the blocks. * This study. CISZ – Congonhas – Itaverava Shear Zone; JBSSZ – Jeceaba – Bom Sucesso Shear Zone; LSZ – Lenheiros Shear Zone.

6 Conclusions

The mapped structures in the northern portion of the Mineiro Belt reveal an accretionary-collisional system dominated by transcurrent to transpressive zones that partially shape the geometry of that sector of the southern São Francisco protocontinent.

Six structural domains were defined in the studied area. Regional-scale shear zones and associated structural features compose the geometry of the southeastern border of the São Francisco craton. Three main transcurrent faults delineate the

contact of the Mineiro Belt with the Archean-to-Paleoproterozoic basement of the southern São Francisco craton and the general geometry of the Belt. They were activated between 2130 and 2060 Ma during the Rhyacian amalgamation related to the Minas-Bahia Orogen in the southern São Francisco craton.

The Jeceaba-Bom Sucesso SZ defines the northern boundary of the Mineiro Belt at the contact zone with the Minas Block along the NE-SW direction approx. parallel to the regional strike of the metasediments of the Minas Supergroup, while the NW-SE Congonhas-Itaverava SZ delineates the northeastern boundary abutting against the Jeceaba-Bom Sucesso Lineament and defines the contact with Santo Antonio de Pirapetinga Complex.

The Jeceaba-Bom Success and the Congonhas-Itaverava are long-lived dominantly strike-slip to transpressive dextral shear zones. Associated with these major structures developed second-order shears that were successively reactivated, in many cases in the opposite sense, hampering a clear kinematic interpretation.

The structural framework of the south-southeastern border of the Mineiro Belt is less conspicuous at large scale, with the development of reverse faults and low-angle thrusts, but displays the unmistakable imprint of the Lenheiros SZ, such as a steep foliation, developed during the collision between the Mantiqueira and Juiz de Fora Complexes and the Mineiro Belt.

Structures associated with a late reverse movement component, such as discrete cleavage planes and lineation, are also recognizable in the Congonhas-Itaverava and Lenheiros Shear Zones, probably associated with the Neoproterozoic collage of the São Francisco and Congo cratons between 620-520 Ma (Pedrosa-Soares et al., 2001; Noce et al., 2007; Kuribara et al., 2019).

7 Author statement

Eduardo Oliveira: Conceptualization, Methodology, Formal analysis, Investigation, Writing – Original Draft. Carlos Rosière: Conceptualization, Methodology, Validation, Investigation, Writing - Review and Editing, Supervision, Funding acquisition. João Paulo de Souza: Methodology, Formal analysis, Validation,

Writing - Review and Editing. Armin Zeh: Methodology, Formal analysis, Validation, Writing - Review and Editing. Flávia C. S. Braga: Formal analysis, Writing - Review and Editing.

8 Declaration of competing interest

The authors declare that they have no interests, competing financial interests, or personal relationships that could have appeared to influence the work reported in this paper.

9 Data availability

Data will be made available on request.

10 Acknowledgments

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Artigo 2: Unveiling the Rhyacian Tholeiite-to-Boninite series in the northern portion of the Mineiro Belt, Southern São Francisco craton: Petrogenesis and Tectonic Implications

Manuscrito submetido.

Eduardo Luís Carneiro de Oliveira^{1*}, Carlos Alberto Rosière¹, Armin Zeh², and Flávia Cristina Silveira Braga¹

1: Programa de Pós-graduação em Geologia e Centro de Pesquisas Manoel Teixeira da Costa, Instituto de Geociências, Universidade Federal de Minas Gerais (CPMTC-IGC-UFMG), Campus Pampulha, Av. Antônio Carlos 6627, CEP 31270-901 Belo Horizonte, MG, Brazil

2: Karlsruhe Institut für Technologie (KIT) Institut für Angewandte Geowissenschaften–Mineralogie und Petrologie. Adenauerring 20b, Geb. 50.40 - 76131 Karlsruhe

* Corresponding to the author. PPGG-CPMTC-IGC-UFMG. E-mail address: oliveira.eduardoluiscarneiro@gmail.com (Eduardo Oliveira).

Abstract

New petrological, geochemical, and geochronological data are presented to enhance our understanding of the tectonic and petrogenetic evolution of the southern São Francisco craton, Brazil. A systematic sampling campaign, whole-rock rock geochemistry, and combined U-Pb and Lu-Hf LA-MC-ICP-MS dating of igneous volcanic and plutonic bodies overlaid by sedimentary rocks revealed the occurrence of volcanic rocks of the tholeiite-to-boninite series in the accretionary-to-collisional Minas-Bahia Rhyacian orogen in the contact of the Mineiro Belt with the basement of Minas Block. The tholeiite series comprises basalt to basaltic andesite, characterized by hornblende and aggregates of fine crystals of plagioclase and quartz, displaying metamorphic foliation. Picrite-to-boninite forms the boninite series,

composed of Mg-chlorite and zircon as an accessory mineral. The geochemistry signature of volcanic rocks ranges from silica-high magnesium basalt and basalt-andesite to magnesium-high, silica-high, and low titanium boninite. Two magmatic ages in this volcanic sequence were obtained, showing ages of 2170.8 ± 5.9 Ma to the basalt and 2149.8 ± 5.9 Ma for the boninite, which is consistent with the early subduction stage responsible for the crystallization of the Alto Maranhão sanukitoid suite. Boninite is a rare magnesium-high, silica-high, and low-titanium rock formed in the early stage of subduction, melting a metasomatized mantle in a forearc setting. The enrichment in incompatible elements, together with negative ϵ_{Hf} and Archean T_{DM} ages, reveal that the influence of Archean crustal matter during the emplacement of melting has significant implications for our understanding of the Earth's tectonic processes in the Rhyacian period. With the evolution of the subduction responsible for forming the boninite series, a voluminous sanukitoid plutonic suite (Alto Maranhão) crystallized from 2140 to 2130 Ma. These rocks were previously described as having a magma composition of high-Mg andesites in modern subduction zones. Overlying the volcanic-plutonic rocks, we present three new detrital maximum deposition ages consistent with the changes in the deposition source and tectonic extensional setting from this crustal segment of the southern São Francisco craton.

Keywords

Orogenic system, Rhyacian subduction, tholeiite-boninite series.

1. Introduction

The tholeiite-to-boninite series is a crucial geological feature in a forearc setting, such as an Izu-Bonin-Mariana arc type associated with the early stage of the subduction process in the modern arc environment. This dominant setting has only been observed as far back as approximately 2000 Ma (Pearce and Reagan, 2019). Boninite is a rare, high-silica, high-magnesium, low-titanium volcanic rock with significant tectonic implications, considered to be the product of lithospheric extension and asthenospheric upwelling generated in the suprasubduction zone (SSZ) tectonic

environments as primary melts from the refractory mantle (e.g., Hickey and Frey, 1982; Beccaluva and Serri, 1988; Crawford et al., 1989; Stern and Bloomer, 1992).

In Proterozoic to Phanerozoic volcano-sedimentary sequences, the boninite series have been associated with several ophiolites and ophiolite-like terranes such as the Troodos ophiolite (e.g., Cameron, 1985; Flower and Levine, 1987), the Koh ophiolite (Sameshima et al., 1983; Meffre et al., 1996), the Betts Cove ophiolite (Bédard et al., 1998; Bédard, 1999), and Dachadaban ophiolite (Xia et al., 2012).

The accretionary-to-collisional terrane of the Mineiro Belt (Teixeira et al., 2000; Avila et al., 2014; Moreira et al., 2018), located at the southern edge of the São Francisco craton, comprises reworked Rhyacian to Orosirian rocks, magmatic arcs, and intra-oceanic sequences amalgamated against the Minas Block during the evolution of the Minas-Bahia orogeny in the Siderian-to-Rhyacian transition from 2400 to 2050 Ma (Bruno et al., 2021 a, b).

In our current research, the discovery of new petrological, geochemical, and isotopic U-Pb and Lu-Hf data is a significant milestone. These findings offer valuable insights into the lithostratigraphy, petrogenesis, geochronology, and tectonic setting to the northern Mineiro Belt. The volcano-sedimentary sequence studied is proposed to be individualized as the Lafaiete Sequence instead of the uncertain Lafaiete "Formation" (Ebert, 1963, 1965; Pires, 1977; Toledo, 2002) or Rio das Mortes Sequence (Avila et al., 2003 and 2006).

The tholeiite-boninite series that was solidified around 2170 Ma and 2150 Ma is in contact with a sanukitoid plutonic body of the Alto Maranhão Suite, dated at 2140 Ma, and filled by aplitic dykes crystallized in 2130 Ma. These ages provide insights into a crucial period of evolution of the Minas Block, occurring chrono-correlated as the extension of the continental crust and sedimentation of the upper Minas Supergroup (Sabará Group) in the proto-continent. The beginning of tholeiitic magmatism can be related to the end of a decreased volcanic activity known as "magmatic lull" (Spencer et al., 2018). The subsequent sedimentation between 2100 and 2060 Ma represents a new stage in the belt's tectonic development associated with an extensional event (Machado et al., 1996; Alkmim and Martins-Neto, 2012).

2. Geological setting

The southern portion of the São Francisco Craton comprises the Paleoproterozoic metasedimentary sequences of the Minas Supergroup covering Archean terranes (Marshak et al., 1997; Cutts et al., 2019) in tectonic contact with the accretionary-collisional arc systems developed from 2400 to 2100 Ma comprising the Mineiro Belt (Teixeira and Figueiredo, 1991) and Mantiqueira Complex (Heilbron et al., 2010) (Figs. 1a and 1b). The Archean terranes in the study area comprise the Granite-Greenstone Complex of Santo Antônio do Pirapetinga (Raposo et al., 1991) and the Bonfim Complex (Carneiro, 1992). The Santo Antônio do Pirapetinga Complex (SAPC) is located between the Minas Block in the north and the Mineiro Belt in the south, separated from these units by the Engenho and Congonhas-Itaverava Lineaments., respectively (Figs. 1b and 1c).

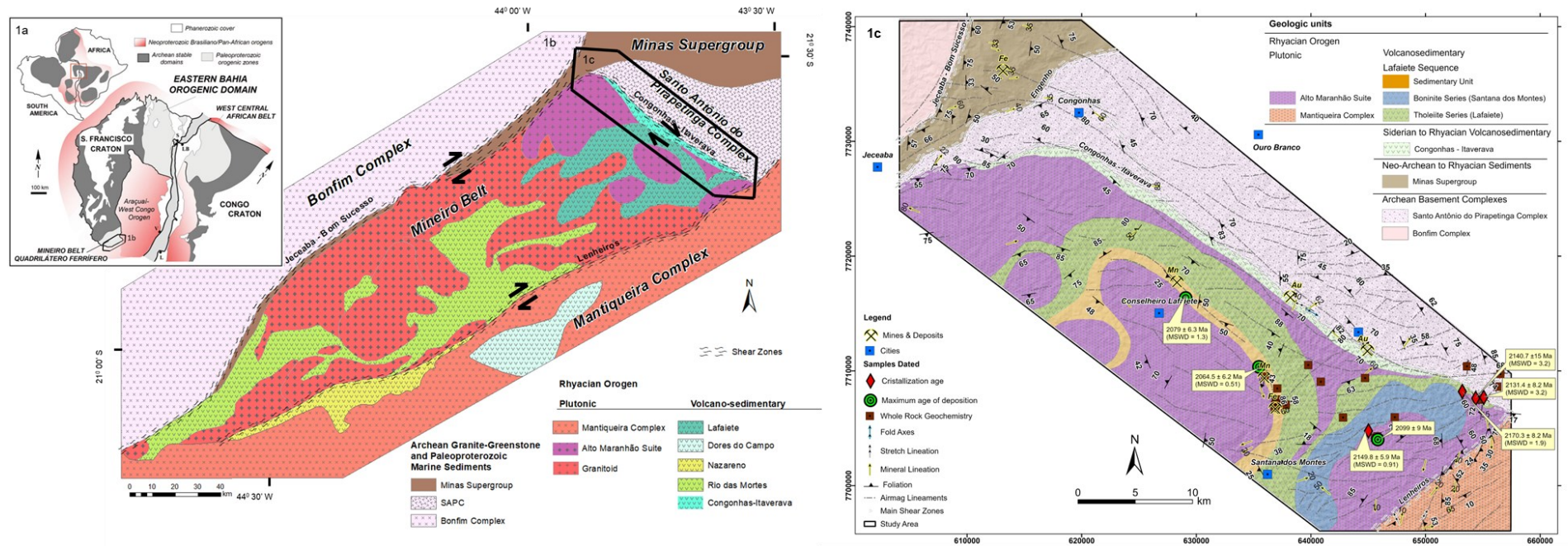


Fig. 1: Regional geological mapping, modified from Barbosa & Barbosa (2017) and Cardoso et al. (2019). (1a) represents the regional context of the São Francisco–Congo Craton; (1b) is a detail, considering the entire region; (1c) a detailed geologic map showing the samples of whole-rock geochemistry and geochronology ages obtained from the study area.

2.1 Minas Supergroup

The Minas Supergroup comprises a 6000 m thick sedimentary sequence deposited in an initial continental rift that evolved to a passive margin (Alkmim and Marshak, 1998; Hartmann et al., 2006; Farina et al., 2016). The basal Caraça Group is constituted by the Moeda and Batatal formations comprising quartz-rich metasandstones and conglomerates that grade to gray and black shales and minor carbonate rocks (Dorr II, 1969; Renger et al., 1994; Koglin et al., 2014; Dopico et al., 2017). Deposited over the Caraça Group, there is a sequence of platformal (Lake Superior-type) Cauê banded iron formations covered by limestones and dolomites of the Gandarela formations that comprise the Itabira Group (Alkmim and Martins-Neto, 2012; Babinski et al., 1995; Dorr II, 1969; Renger et al., 1994). A major erosional surface and sedimentary gap with the development of a low-angle, angular unconformity (Simmons, 1968) separates the Itabira from the quartzites and phyllites of the Piracicaba Group (Pomerene, 1964; Dorr II, 1969).

The Piracicaba Group comprises deltaic and shallow-marine deposits overlain by deep-water sediments. The depositional environment is contentious as some authors propose a passive margin (Alkmim and Martins-Neto, 2012 and references therein), whereas others consider it a foreland sequence (Machado et al., 1996; Bekker et al., 2003; Rossignol et al., 2020). Rossignol et al. (2020) proposed a maximum depositional age of 2333 ± 11 Ma, although Cabral et al. (2022) advocated an approximate age of 2140 Ma for the onset of sedimentation of the sequence. In any case, the deposition of the Piracicaba Group started during the late Siderian or was already in the Rhyacian period, indicating a stratigraphic gap that lasted more than 60 Ma.

The overlying Sabará Group in the Minas Basin comprises immature, clastic sedimentary rocks interbedded with pelitic rocks and rare felsic tuffs (Dorr II, 1969; Reis et al., 2002; Almeida, 2005), indicating volcanogenic sources from south-southeast source (Alkmim and Noce, 2006; Dutra et al., 2019), which represents a change in the sedimentation regime from an extensional environment to a foreland basin. The maximum depositional age is usually taken as ca. 2120 Ma (Farina et al.,

2016; Dopico et al., 2017) during the inversion of the Minas Basin, and recent data suggest that sedimentation continued until ca. 2040 Ma (Dutra et al., 2019).

The post-orogenic collapse resulted in the deposition of the Itacolomi Group, suggesting that most of the sediments were derived from terranes generated during the Rhyacian orogeny, with sources located to the south and southeast (Farina et al., 2016; see also reference therein).

2.2 Mineiro Belt

In tectonic contact with the basement of the Minas Block, the Mineiro Belt comprehends a series of Siderian to Rhyacian magmatic arcs amalgamated in an accretionary-collisional system (Noce et al., 2000, 2007; Campos et al., 2008; Ávila et al., 2010; Corrêa Neto et al., 2012; Seixas et al. 2012, 2013; Teixeira et al., 2015; Barbosa et al., 2015; Barbosa et al., 2019; Moreira et al., 2018, 2019). This terrane is comprised of juvenile granitic bodies of calcium-alkaline affinity, grouped into plutons, suites, or batholiths with ages that spread from Siderian to Rhyacian between stripes of volcano-sedimentary sequences (Seixas et al., 2012; Ávila et al., 2014; Teixeira et al., 2015; Moreira et al., 2018; Barbosa et al., 2019). Granitic bodies crystallized from magmas formed by the subduction of the oceanic plate with continental crust contamination (Noce et al., 2000; Cherman and Valença, 2005; Ávila et al., 2006; Seixas et al., 2012; Teixeira et al., 2015). The most representative, from a petrogenetic point of view, are the Lagoa Dourada-Resende Costa (2350 Ma), Serrinha-Tiradentes (2230 - 2210 Ma), and Alto Maranhão (2140 - 2130 Ma) suites, as well as the Ritápolis batholith (2150 - 2120 Ma), (Alkmim and Teixeira, 2017).

The Mineiro Belt is bound to the north by the NE-SW Bom Sucesso - Congonhas Lineament (Seixas et al., 1988) and to the south by the Lenheiros strike-slip Shear Zone (Nepomuceno et al., 2021). The southern limit of this tectonic domain is not yet well established (Alkmim and Teixeira, 2017).

Four volcano-sedimentary sequences exposed in the Mineiro Belt are developed in an oceanic arc environment, defined by Toledo (2002), Teixeira et al. (2008), Corrêa Neto et al. (2012) and Ávila et al. (2014) as the Congonhas-Itaverava,

Rio das Mortes, Nazareno, and Dores do Campo sequences, respectively (Fig. 1). They comprise basal mafic-ultramafic volcanic rocks, followed by detrital sediments with some spots of banded iron formation and gondite.

The Congonhas-Itaverava Sequence forms the eastern boundary of the Mineiro Belt, covering an area approximately 50 km long and 3 km wide, in tectonic contact with the Archean basement of the Minas Block, along the strike-slip Congonhas Shear Zone (Chemale Jr. et al., 1994; Endo, 1997; Campos, 2004, Corrêa Neto et al., 2012; Teixeira et al., 2015) (Fig. 1). The sequence is characterized by a mafic and ultramafic volcanic basal unit, including komatiites, an intermediate unit containing banded iron formation, marl, and pelitic layers, and a top unit composed of shale and graywacke (Corrêa Neto et al., 2012), with a maximum depositional age of 2349 ± 14 Ma (Teixeira et al., 2015).

The Rio das Mortes Sequence comprises phyllites associated with gondite, quartzite, and tholeiitic metabasalt (amphibolites). Two amphibolites of this sequence were dated by Ávila et al. (2010) at 2231 ± 5 Ma and 2202 ± 11 Ma.

The Nazareno Sequence occurs in the southern portion of the Mineiro Belt. It consists essentially of mafic and ultramafic rocks, including komatiites with a spinifex texture characteristic of the basal portion of the sequence. The intermediate and upper units include amphibolites (metabasalts) and, subordinately, phyllites, quartzites, and gondites (Toledo, 2002; Ávila et al., 2010). Two samples of amphibolites from this sequence were dated at 2223 ± 4 Ma and 2276 ± 14 Ma (Ávila et al. 2010, 2012).

The Dores do Campos Sequence is composed of ortho-amphibolites, phyllites, gondites, quartzites, and lenses of metaultramafic rocks at the base (Avila et al. 2004, Avila et al. 2010). The composition of the ortho-amphibolite is basaltic and basalt-andesitic, with magmatic ages of 2255 ± 51 Ma and TDM between 2500 and 2200 Ma, ϵ Nd (2000 Ma) between +1.2 to +3.6, indicating a mantellic source. A T_{DM} age of 2500 Ma and ϵ Nd (2200 Ma) -2,5 suggest crustal contamination.

The Lenheiros suture (Ávila et al., 2010, 2014; Nepomuceno et al., 2021) is the eastern boundary of the Mineiro Belt, in tectonic contact with the Mantiqueira Magmatic Arc (ca. 2200 - 2000 Ma), that comprises juvenile-to-crustal-contaminated TTGs, sanukitoids, alkaline rocks, intraplate tholeiitic basic rocks, and peraluminous

granitoid rocks (e.g., Heilbron et al., 2010; Ávila et al., 2014; Alkmim and Teixeira, 2017; Teixeira et al., 2017; Moreira et al., 2018; Bruno et al., 2020; Cutts et al., 2020).

2.3 The Lafaiete Sequence

In this work, we individualize the Lafaiete Sequence, the unit defined as Lafaiete Formation by Ebert (1965, 1963), who described a volcano-sedimentary sequence following a northwest-trending belt between the localities of Congonhas and Itaverava (Figs. 1b and 1c) comprising chlorite and talc shists, metagraywackes, banded iron formations BIFs, and gondites. Guild (1957) inferred that the chlorite schists and metasedimentary rocks were part of the “Minas Series”, while the metaultramafic rocks were intrusive bodies. Ebert (1963), Pires (1977), Candia and Girardi (1979), and Barbosa (1985) included all these rocks in the Barbacena greenstone belt, purported as of Archean age. Other authors (e.g., Dorr II, 1969; Grossi Sad et al., 1983; Seixas, 1988; Heineck et al., 2003; Pinto and Silva, 2014) included this sequence in the Neoproterozoic Nova Lima Group, while Ávila et al. (2003, 2006) correlated it with the Paleoproterozoic Rio das Mortes greenstone belt.

2.3.1. Volcanic units

Volcanic rocks, comprising mafic to ultramafic volcanic rocks, are the main lithologies associated with the Lafaiete Sequence (Fig. 1c).

The Tholeiite Series (Água Limpa Unit)

The tholeiite series is composed of basalt and basaltic andesite. These rocks are the main lithology of the Lafaiete formation, exposed between the locations of Conselheiro Lafaiete and Itaverava. The metamorphism is of amphibolite facies with greenschist retrograde overprint. The mineralogy comprises hornblende, plagioclase, and quartz as main minerals and titanite, apatite, zircon, and epidote as accessories. (Fig. 2 a, b). Some portions of the volcanic mafic present a saussuritization in the plagioclase-forming fine grains of clinozoisite-zoisite, epidote, and sericite.

Covering the volcanic rocks follows a clastic-chemical sedimentary sequence with greywackes, black shales, quartzite, and Fe- and Mn-bearing rocks. That can be mapped along the Morro da Mina Mn Deposit to the Água Limpa Fe Deposit.

The Boninite Series (Santana dos Montes Unit)

An ultramafic volcanic unit composed of picrite to boninite series comprises a monomineralic hydrated metamorphic suite as Mg-chlorite, serpentinite, and tremolite. This unit outcrops between the cities of Santana dos Montes and Itaverava. It is here described as the Santana dos Montes Unit. The ultramafic enriches in some incompatible elements, forming zircon as an accessory mineral. The presence of zircons in an ultramafic unit may be related to crustal contamination during the early stage of the subduction period that forms the picrite-boninite series.

This sequence outcrops in the southern-eastern part of the area, defined by chloritite, soapstone, tremolitite, and hornblendite outcrop in gradational contact with the basaltic andesite volcanic unit. It crops out between Santana dos Montes and Itaverava, forming a NE-SW trend affected by the deformation in the southeastern border of Southern São Francisco Craton (Fig. 2 c, d, e). This sequence is younger than the mafic volcanic (Água Limpa Unit), crystalized at 2150 Ma (U-Pb dating, presented in this work).

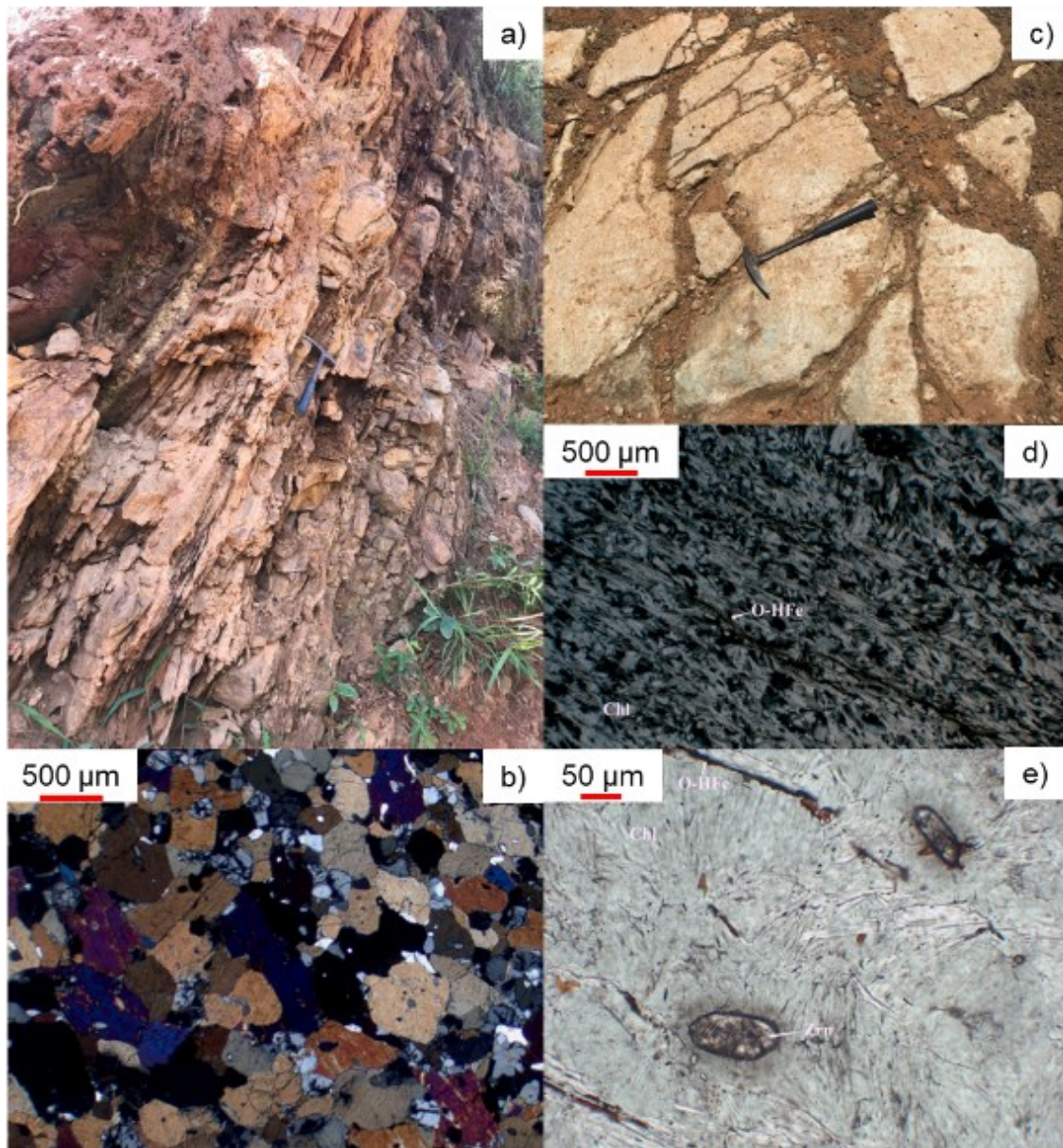


Fig. 2: a) Amphibolite that represents the metabasalt showing a foliated hornblende-plagioclase-quartz composition; b) f volcanic mafic composed by foliated hornblende, plagioclase, and quartz; c) Boninite outcrop, represented by chloritite; d) thin section of chloritite associated with the boninite volcanic rock; e) volcanic ultramafic-ultrabasic showing magmatic zircons in a matrix of the rock that attests the crustal contamination with enrichment of incompatible elements.

2.3.2 Sediments (The Morro da Mina Unit)

A chemical-clastic sequence composed of a fuchsite-quartzite with rutile, monazite, and zircons, grades laterally to greywacke, graphite schist, and iron-manganese rich rocks. The sequence is exposed in the surroundings of Conselheiro Lafaiete, unconformably overlies the basalt/basaltic-andesite unit, and is intruded by granites of the Alto Maranhão Suite. The mineralogy of the manganese-rich member comprises spessartine, rhodonite, and rhodochrosite and hosts the Mn-oxide Morro da Mina deposit (Candia and Girardi, 1979). The Fe-rich member comprises medium to coarse-grained layers of magnetite and hornblende with biotite, chlorite, and minor spessartine (Fig. 3 a, b, c, d, e).

Cabral et al. (2019) obtained a maximum depositional age of ca. 2075 Ma (U-Pb LA-ICP-MS method) for detrital zircons from the Morro da Mina Mn Deposit, much younger than for the units comprising the Congonhas-Itaverava metavolcano-sedimentary belt as well other greenstone belts exposed in the Mineiro belt.

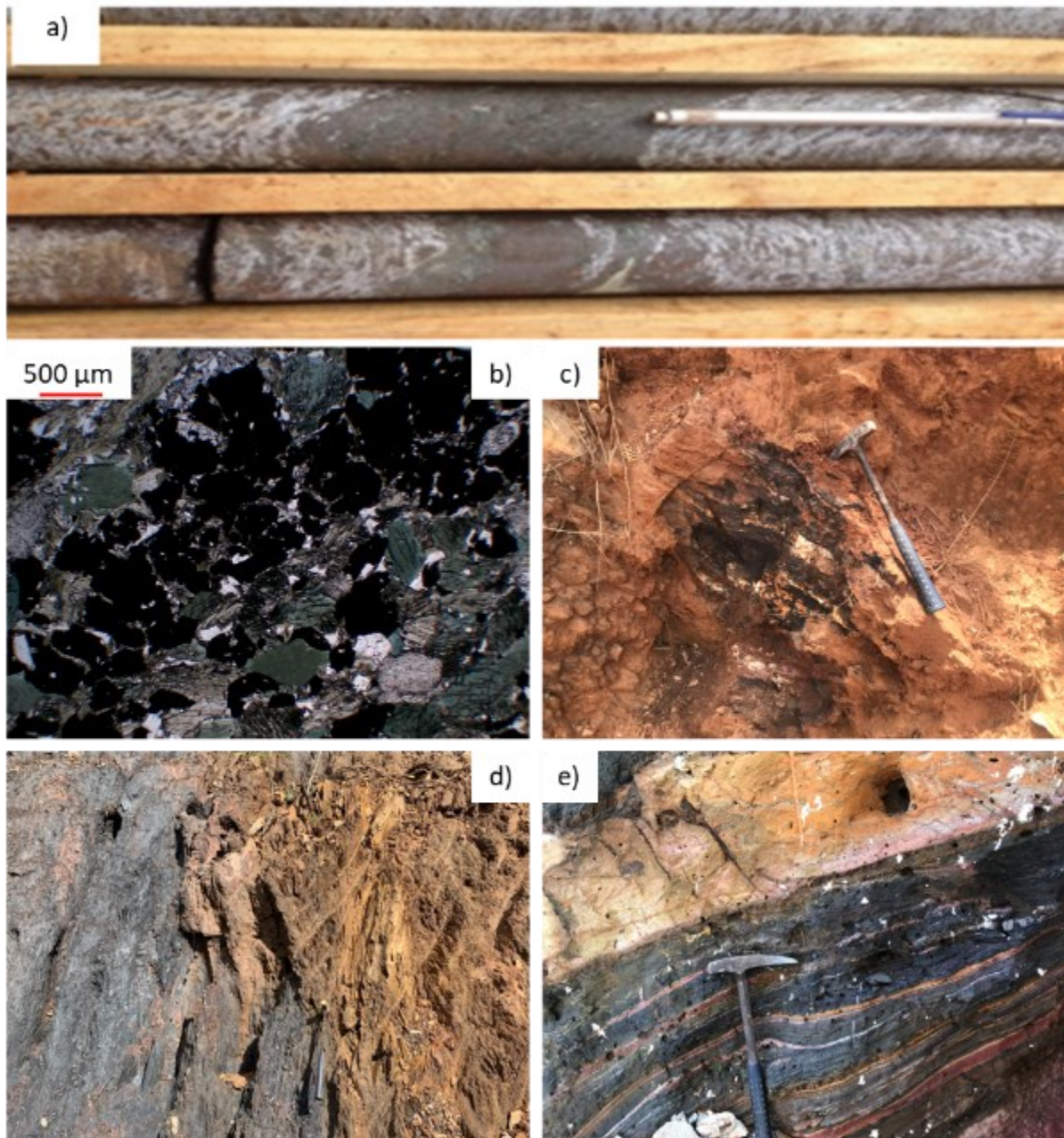


Fig. 3: Sedimentary sequence in the study area: a) hornblende-quartz-magnetite banded bearing rock; b) thin section of the iron-bearing rock; c) weathered gondite (Mn-bearing rock); d) gradation of the shale (graphite schist) to pelite; e) greywacke outcropping in the town of Lafaiete.

3. Methodology

3.1. Whole rock geochemistry

Twenty-three samples were collected in outcrop and drill cores for whole rock analyses of primary, trace, and rare earth elements (REE). The total results are in Table 1. The studies were conducted in the ACME Vancouver Labs, Vancouver, Canada, using the ICP-MS (Inductively Coupled Plasma Mass Spectroscopy) method. ICP-MS analyzed eleven principal oxides following a lithium fusion and dilute acid digestion of a 0.2 g sample pulp. Forty-six trace elements were analyzed using aqua regia digestion to precious and base metals. The standard SO-19 from USGS, with an achieved accuracy of $\pm 3\%$ SD, was used to determine the accuracy of the analytical assay. A duplicate sample was introduced in the analysis batch to measure the assay precision. The accepted values have a maximum error of $\pm 4\%$ RSD.

3.2. U-Pb and Lu-Hf isotopes

3.2.1. Sample preparation and imaging

Seven samples (02 volcanic, 03 sedimentary, 02 plutonic-volcanic rocks) were processed with conventional crushing, grinding, and screening methods at Centro de Pesquisa Professor Manoel Teixeira da Costa, UFMG. The volcanic had 200 kg of mass. The others followed a standard size to concentrate enough zircon grains for analysis. After the screening, the heavy fraction concentration was pre-concentrated by panning. After concentration, samples were sieved and washed to remove clay and silt size. The pre-concentrate was purified using heavy liquid (bromoform) to remove all light minerals. Then, a Frantz LB1 magnetic separator was used to separate the less magnetic minerals. Zircon was handpicked and organized in an epoxy mount, which was polished and carbon-coated for SEM (Scanning Electron Microscope) study using back-scattered electron (BSE) and secondary electron (SE) images that were taken using a TESCAN VEGA2 scanning electron microscope at the Department of Mineralogy and Petrology of Karlsruhe Institute of Technology (KIT).

3.2.2. U-Th-Pb isotope analysis

Measurements for U–Th–Pb isotopes in zircon were performed on polished grain mounts using a 193nm ArF Excimer laser (Analyte Exite+, Teledyne Photon Machines), coupled to a Thermo-Scientific Element XR instrument at the Karlsruhe Institute of Technology (KIT), Germany. Twenty zircon grains, <60 µm in length, were large enough for LA–ICP–SF–MS dating using a laser spot size of 20 µm.

The calibration of Pb/U ratios was conducted using the zircon standards BB (primary standard), Plešovice, and KA (KaapValley). The laser spot diameter used was 20 µm, with a laser fluence of 2.7 J/cm², at 10-Hz repetition rate, RF = 1255 W, a mixed Ar–He–N₂ carrier gas consisting of Ar = 0.94 l/min, He = 0.30 cell + 0.20 cup (both l/min), and N₂ = 12 ml/min. Three pulses of pre-ablation were obtained before each 15-s duration analysis, following 15-s background measurement. All raw data were corrected offline for daily instrumental drift and mass offset employing an in-house MS Excel© spreadsheet program Gerdes and Zeh (2009). We applied a standard Pb correction based on the interference- and background-corrected ²⁰⁴Pb signal and a model Pb composition Stacey and Kramers (1975). U–Pb dating results of reference zircon and unknowns are shown in Supplementary Table S1. Multiple measurements of the reference zircon BB and KA yielded Concordia ages of 561.6 ± 3.2 Ma (MSWD = 0.1, Probability = 0.98, n = 10), 338.9 ± 1.4 Ma (MSWD = 0.1, probability = 0.99, n = 10), and 3225.8 ± 7.9 Ma (MSWD = 5.3, probability = 0.02, n = 10), respectively, in agreement with published TIMS ages (Santos et al., 2017; Sláma et al., 2008; Schoene et al., 2006).

3.2.3. Lu-Hf isotope analysis

Lu-Hf isotope analysis was carried out with a Resolution M-50 193nm ArF Excimer laser system coupled to a Thermo-Scientific multi-collector (MC)-SF-ICP-MS (Neptune Plus) at FIERCE Goethe University of Frankfurt, Germany. The analytical protocols used are described in detail by Gerdes and Zeh (2006; and reference therein). For calculation of the epsilon Hf [εHf], the chondritic uniform reservoir (CHUR) was used as recommended by Bouvier et al. (2008); ¹⁷⁶Lu/¹⁷⁷Hf and ¹⁷⁶Hf/¹⁷⁷Hf of 0.0336 and 0.282785, respectively, and a decay constant of 1.867 x 10⁻¹¹ (Scherer et al., 2001; Söderlund et al., 2004). All two-stage hafnium model ages (T_{DM}) were

calculated by applying $^{176}\text{Lu}/^{177}\text{Hf} = 0.03933$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.283294$, resulting in a depleted mantle (DM) evolutionary line ranging from zero (at 4.5 Ga) to +18 (today). Crustal evolutionary trends were modeled by applying $^{176}\text{Lu}/^{177}\text{Hf} = 0.0113$ for the average continental crust (average of Taylor and McLennan, 1985 and Wedepohl, 1995). For all detrital zircons, initial $^{176}\text{Hf}/^{177}\text{Hf}$, $\epsilon\text{Hf}(t)$, and T_{DM} were calculated using the $^{207}\text{Pb}/^{206}\text{Pb}$ ages obtained for the respective zircon domains (Supplementary Table S2).

4. Results

4.1 Whole-rock geochemistry: major and trace elements

The analytical data for the volcanic, plutonic, and sedimentary rocks that compose the region studied are depicted in Table 1.

	Picrite 1	Picrite 2	Picrite 3	Boninite 1	Boninite 2	Boninite 3	Boninite 4	Boninite 5	Boninite 6
SiO2 (wt%)	42.96	46.59	44.96	27.78	53.34	52.23	55.17	57.48	52.88
TiO2	1.15	0.21	0.27	0.77	0.60	0.23	0.41	0.04	0.42
Al2O3	9.20	7.59	6.61	20.02	4.36	3.63	4.86	2.64	5.85
Fe2O3	14.17	15.17	20.80	15.19	9.99	8.00	10.74	7.94	12.70
MnO	0.22	0.25	0.37	0.17	0.12	0.20	0.06	0.09	0.04
MgO	18.20	17.82	18.83	24.21	22.81	19.74	21.70	25.04	20.51
CaO	6.43	6.49	0.31	0.02	1.88	11.26	0.02	0.17	<0.01
Na2O	0.42	0.72	<0.01	<0.01	0.08	0.38	0.03	0.02	<0.01
K2O	0.04	0.06	<0.01	0.01	0.03	0.07	<0.01	<0.01	<0.01
P2O5	0.14	0.02	0.02	0.04	0.04	<0.01	0.06	0.04	0.03
LOI	6.00	4.30	6.80	11.00	5.50	3.50	6.20	5.60	6.60
Rb (ppm)	0.50	0.70	1.20	<0.1	<0.1	0.60	0.50	0.40	0.10
Ba	55.00	12.00	41.00	55.00	8.00	23.00	11.00	16.00	3.00
Sr	40.00	13.20	2.10	9.50	10.10	43.30	2.50	1.80	<0.5
Th	0.30	0.20	0.70	32.90	0.40	0.30	0.60	<0.2	<0.2
Nb	3.30	1.20	1.40	9.00	2.30	0.40	2.70	0.60	1.70
Zr	61.20	20.20	35.10	502.20	41.30	10.60	24.90	2.70	25.30
Hf	1.60	0.70	0.90	24.70	0.90	0.30	0.70	<0.1	0.70
Y	294.60	85.90	25.50	22.70	120.40	25.00	12.50	56.30	2.60
V	246.00	127.00	128.00	124.00	104.00	126.00	52.00	52.00	41.00
La	400.00	67.00	8.90	106.50	413.10	28.40	7.60	132.90	1.70
Ce	44.60	4.70	70.90	71.90	31.00	8.10	11.00	97.30	9.40
Pr	153.00	14.89	2.68	26.00	137.51	11.10	1.36	35.97	0.51
Nd	566.10	57.70	11.20	97.20	530.80	48.10	6.50	139.00	2.40
Sm	137.84	11.91	3.66	16.56	98.82	11.87	1.82	23.90	0.79
Eu	42.23	4.10	1.41	3.72	26.47	3.13	0.46	6.23	0.21
Gd	107.18	13.19	4.64	11.01	61.63	8.61	2.45	17.51	0.72
Tb	18.17	2.33	1.07	1.27	8.83	1.21	0.43	2.32	0.16
Dy	92.83	14.80	7.50	5.98	43.19	6.52	2.87	12.00	1.11
Ho	15.22	3.20	1.61	0.91	6.46	1.08	0.66	2.24	0.20
Er	37.79	8.84	5.19	2.24	15.09	2.80	2.00	5.68	0.63
Tm	5.03	1.29	0.90	0.31	2.02	0.39	0.29	0.79	0.10
Yb	31.49	8.27	6.92	2.05	12.60	2.47	1.94	4.85	0.90
Lu	4.18	1.23	1.01	0.39	1.57	0.32	0.31	0.68	0.13
ΣREE	1655.66	213.45	127.59	346.04	1389.09	134.10	39.69	481.37	18.96
Cr ppm	1854.18	2066.28	3092.58	684.20	3414.16	2641.01	1087.88	1806.29	2408.38
Cr2O3 (wt%)	0.27	0.30	0.45	0.10	0.50	0.39	0.16	0.26	0.35
Ni (ppm)	1094.00	809.00	1766.00	974.00	1063.00	317.00	1292.00	1335.00	1822.00
Sc	28.00	25.00	18.00	9.00	20.00	43.00	11.00	7.00	9.00
Sum (wt%)	99.38	99.62	99.60	99.44	99.40	99.62	99.61	99.50	99.63
Be (ppm)	<1	<1	<1	<1	1.00	1.00	<1	<1	<1
Co	126.70	80.70	205.20	102.90	87.70	81.20	70.90	95.40	92.10
Cs	0.20	<0.1	0.10	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1
Ga	12.60	6.80	6.00	27.10	6.10	4.70	5.90	3.50	3.70
Sn	1.00	2.00	<1	<1	<1	<1	<1	<1	<1
Ta	0.20	<0.1	<0.1	0.70	0.10	<0.1	0.10	<0.1	0.10
U	0.30	0.20	0.20	7.40	0.20	<0.1	0.30	0.40	0.60
W	<0.5	<0.5	0.50	<0.5	1.00	<0.5	<0.5	<0.5	1.10
Zr/Y	0.21	0.24	1.38	22.12	0.34	0.42	1.99	0.05	9.73
Nb/Y	0.01	0.01	0.05	0.40	0.02	0.02	0.22	0.01	0.65
Zr/Ti	0.01	0.02	0.02	0.11	0.01	0.01	0.01	0.01	0.01
Th/Yb	0.01	0.02	0.10	16.05	0.03	0.12	0.31	0.04	0.22
Nb/Yb	0.10	0.15	0.20	4.39	0.18	0.16	1.39	0.12	1.89
Mg#	0.53	0.50	0.44	0.58	0.66	0.68	0.64	0.73	0.58
Ti	6894.37	1258.97	1618.68	4616.23	3597.06	1378.87	2457.99	239.80	2517.94
				0.02					
Sr/Y	0.14	0.15	0.08	0.42	0.08	1.73	0.20	0.03	0.00
V/Sc	8.79	5.08	7.11	13.78	5.20	2.93	4.73	7.43	4.56

	Basalt 1	Basalt 2	Basalt 3	Basalt 4	Basalt 5	Basalt 6	Basalt 7	B-Andesite 1	B-Andesite 2	B-Andesite 3	B-Andesite 4	B-Andesite 5	B-Andesite 6	B-Andesite 7
SiO2 (wt%)	47.76	49.08	46.31	48.23	45.93	45.49	49.91	55.67	54.56	52.32	54.02	51.68	53.94	55.46
TiO2	2.80	0.85	1.60	1.75	2.82	1.47	1.29	0.83	0.77	0.96	0.93	0.96	2.01	0.03
Al2O3	12.30	14.36	15.35	14.15	11.87	12.80	13.98	14.80	14.74	14.10	14.22	14.12	8.97	0.72
Fe2O3	20.11	13.09	16.48	17.68	22.00	21.20	14.37	9.38	10.04	12.39	11.51	13.02	24.78	34.80
MnO	0.25	0.19	0.19	0.22	0.26	0.20	0.20	0.12	0.12	0.13	0.12	0.13	0.46	0.26
MgO	4.32	8.29	6.06	5.24	4.54	6.20	6.23	5.74	6.38	6.74	6.51	6.96	0.81	5.83
CaO	8.52	11.14	8.88	7.76	8.82	7.95	10.26	6.79	6.48	6.58	5.78	6.49	7.00	3.12
Na2O	2.38	1.64	2.53	1.05	2.12	1.33	2.23	3.94	3.94	3.06	3.01	2.85	0.74	0.07
K2O	0.70	0.20	0.33	0.26	0.71	1.84	0.33	1.24	1.53	1.96	2.23	2.10	0.28	0.05
P2O5	0.17	0.04	0.12	0.14	0.23	0.24	0.12	0.11	0.10	0.12	0.11	0.12	0.70	0.04
LOI	0.40	0.80	1.80	3.20	0.40	0.90	0.80	1.00	1.00	1.30	1.20	1.20	0.20	-0.50
Rb (ppm)	5.30	1.20	2.80	2.50	5.00	34.80	1.80	33.30	39.20	54.00	65.40	55.60	3.40	<0.1
Ba	181.00	25.00	103.00	198.00	158.00	780.00	48.00	455.00	431.00	590.00	686.00	599.00	138.00	32.00
Sr	228.80	111.40	152.60	30.70	195.40	157.40	139.90	489.30	505.70	463.40	474.30	484.10	64.50	8.40
Th	2.40	<0.2	0.80	1.00	1.40	5.70	1.00	26.90	19.30	3.40	3.30	3.50	0.70	0.50
Nb	9.80	1.90	5.20	5.80	9.80	9.70	4.10	3.80	3.10	4.70	4.50	4.70	9.20	0.40
Zr	145.30	37.20	100.40	123.60	169.80	151.60	82.50	197.00	138.70	63.70	79.70	68.50	145.50	9.00
Hf	3.90	1.00	2.70	3.40	4.80	4.40	2.20	4.40	3.50	2.00	2.40	1.80	4.10	0.20
Y	44.50	20.50	35.30	195.90	45.40	44.30	30.90	13.60	11.70	14.30	15.00	15.40	55.00	7.60
V	592.00	297.00	610.00	369.00	600.00	280.00	348.00	123.00	123.00	134.00	128.00	128.00	<8	<8
La	17.30	3.60	13.90	37.20	16.60	22.70	8.30	97.50	68.50	13.40	10.70	10.90	11.60	4.00
Ce	38.00	5.80	18.20	18.60	39.10	44.40	17.80	152.40	110.80	25.70	22.90	23.90	27.50	6.70
Pr	4.76	0.99	3.76	7.42	5.11	5.23	2.65	13.85	10.48	2.99	2.82	2.90	4.61	0.73
Nd	21.60	4.90	17.00	35.70	22.20	21.20	12.30	43.10	32.10	11.80	11.90	12.30	24.60	3.20
Sm	5.96	1.62	4.35	8.64	5.81	5.51	3.53	5.06	4.20	2.73	2.67	2.68	7.50	0.65
Eu	1.96	0.69	1.43	3.11	1.99	1.47	1.21	1.13	0.99	0.89	0.87	0.93	2.20	0.45
Gd	7.32	2.65	5.37	16.31	7.32	6.65	4.61	3.83	3.27	2.92	2.97	3.11	9.45	0.84
Tb	1.28	0.45	0.91	2.50	1.23	1.14	0.79	0.48	0.43	0.46	0.43	0.49	1.58	0.13
Dy	8.06	3.20	5.99	16.72	7.52	7.45	5.17	2.51	2.28	2.78	2.77	2.77	9.83	0.94
Ho	1.75	0.70	1.33	4.04	1.71	1.70	1.11	0.47	0.42	0.51	0.53	0.54	2.09	0.21
Er	5.16	2.01	3.98	11.83	5.08	5.10	3.32	1.37	1.32	1.45	1.50	1.73	6.14	0.72
Tm	0.70	0.30	0.57	1.40	0.74	0.73	0.44	0.19	0.16	0.20	0.20	0.21	0.88	0.10
Yb	4.48	1.83	3.65	7.56	4.84	4.73	2.91	1.12	1.03	1.25	1.28	1.50	5.53	0.71
Lu	0.67	0.29	0.56	1.18	0.73	0.63	0.43	0.16	0.17	0.19	0.19	0.23	0.85	0.11
ΣREE	119.00	29.03	81.00	172.21	119.98	128.64	64.57	323.17	236.15	67.27	61.73	64.19	114.36	19.49
Cr ppm	13.68	273.68	157.37	130.00	13.68	136.84	123.16	198.42	260.00	294.21	260.00	287.36	20.53	20.53
Cr2O3 (wt%)	0.00	0.04	0.02	0.02	0.00	0.02	0.02	0.03	0.04	0.04	0.04	0.04	0.00	0.00
Ni (ppm)	<20	146.00	107.00	71.00	<20	45.00	246.00	135.00	160.00	169.00	159.00	168.00	<20	<20
Sc	43.00	40.00	41.00	44.00	44.00	34.00	37.00	13.00	14.00	16.00	15.00	16.00	47.00	2.00
Sum (wt%)	99.74	99.76	99.72	99.75	99.75	99.77	99.78	99.75	99.76	99.78	99.79	99.78	99.92	99.89
Be (ppm)	<1	<1	<1	<1	3.00	4.00	<1	1.00	<1	1.00	1.00	1.00	<1	<1
Co	51.00	60.50	58.60	66.30	51.90	28.20	124.40	35.50	37.40	37.60	34.90	36.70	26.70	4.20
Cs	<0.1	<0.1	0.10	0.10	0.10	0.90	<0.1	0.80	0.90	1.40	1.70	1.70	<0.1	<0.1
Ga	20.60	13.70	19.00	18.20	20.90	23.40	14.70	14.90	15.30	13.40	14.10	13.60	17.30	0.60
Sn	2.00	<1	<1	<1	1.00	4.00	3.00	<1	<1	<1	<1	<1	<1	<1
Ta	0.70	0.10	0.20	0.40	0.70	0.40	0.30	0.10	0.20	0.30	0.30	0.40	0.60	<0.1
U	0.30	<0.1	0.20	0.20	0.30	0.70	0.20	2.10	1.60	1.60	2.40	2.20	0.20	0.20
W	0.70	0.50	0.50	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	0.60	<0.5
Zr/Y	3.27	1.81	2.84	0.63	3.74	3.42	2.67	14.49	11.85	4.45	5.31	4.45	2.65	1.18
Nb/Y	0.22	0.09	0.15	0.03	0.22	0.22	0.13	0.28	0.26	0.33	0.30	0.31	0.17	0.05
Zr/Ti	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.04	0.03	0.01	0.01	0.01	0.01	0.05
Th/Yb	0.54	0.11	0.22	0.13	0.29	1.21	0.34	24.02	18.74	2.72	2.58	2.33	0.13	0.70
Nb/Yb	2.19	1.04	1.42	0.77	2.02	2.05	1.41	3.39	3.01	3.76	3.52	3.13	1.66	0.56
Mg#	0.16	0.35	0.24	0.20	0.15	0.20	0.27	0.35	0.35	0.32	0.33	0.32	0.03	0.13
Ti	16786.28	5095.84	9592.16	10491.43	16906.18	8812.80	7733.68	4975.93	4616.23	5755.30	5575.44	5755.30	12050.15	179.85
Sr/Y	5.14	5.43	4.32	0.16	4.30	3.55	4.53	35.98	43.22	32.41	31.62	31.44	1.17	1.11
V/Sc	13.77	7.43	14.88	8.39	13.64	8.24	9.41	9.46	8.79	8.38	8.53	8.00	0.02	0.50

	AMS 1	AMS 2	AMS 3	AMS 4	AMS 5	APT 1	IBR 1	IBR 2	IBR 3	MBR 1	MBR 2
SiO₂ (wt%)	60.56	61.87	58.53	76.63	65.62	76.69	43.45	42.11	41.48	24.18	49.95
TiO₂	0.26	0.27	0.26	0.11	0.23	0.04	0.05	0.02	0.03	0.59	0.25
Al₂O₃	16.82	19.13	20.82	13.77	14.62	13.68	0.64	0.51	0.67	13.95	14.85
Fe₂O₃	7.77	4.28	4.16	1.30	6.16	0.93	54.22	55.78	56.16	10.21	4.34
MnO	0.06	0.03	0.05	0.01	0.06	0.02	0.06	0.07	0.07	25.53	21.41
MgO	1.15	1.36	1.52	0.17	1.16	0.15	1.46	1.48	1.41	0.16	0.22
CaO	6.15	4.92	5.02	0.92	6.92	2.32	1.20	1.17	1.21	1.30	2.57
Na₂O	4.48	5.83	7.01	4.92	1.21	5.20	0.05	0.04	0.05	0.02	0.03
K₂O	0.96	1.15	1.29	1.14	2.11	0.14	0.01	0.01	0.01	0.26	0.06
P₂O₅	0.06	0.13	0.12	0.02	0.09	<0.01	0.04	0.03	0.02	0.13	0.02
LOI	1.50	0.70	1.00	0.80	1.40	0.70	-1.20	-1.20	-1.10	13.50	6.20
Rb (ppm)	21.20	27.30	32.00	43.70	44.10	6.80	0.30	<0.1	0.10	12.40	1.10
Ba	808.00	873.00	675.00	1376.00	1222.00	99.00	7.00	2.00	1.00	1286.00	361.00
Sr	1043.70	1066.20	990.40	233.00	1679.20	536.30	4.00	4.00	15.90	339.70	113.10
Th	1.30	14.60	4.00	20.30	6.40	6.70	<0.2	<0.2	<0.2	4.90	0.20
Nb	3.00	2.30	2.50	5.60	1.70	65.20	0.50	0.40	0.50	6.50	0.40
Zr	74.80	117.90	112.30	92.20	92.80	16.50	2.50	1.80	1.90	86.50	102.20
Hf	1.80	2.80	2.70	3.00	2.40	2.20	<0.1	<0.1	<0.1	2.50	2.40
Y	5.00	3.80	7.60	6.80	5.00	16.40	2.40	2.10	1.70	92.40	2.60
V	54.00	38.00	33.00	15.00	49.00	12.00	45.00	43.00	43.00	244.00	89.00
La	14.30	34.00	30.70	33.90	21.20	9.40	0.40	0.20	0.50	39.20	4.20
Ce	24.70	59.80	52.30	62.10	35.80	35.90	1.00	0.60	1.00	186.70	50.80
Pr	2.89	6.39	5.59	5.28	3.82	2.98	0.11	0.05	0.10	10.31	0.67
Nd	11.30	22.50	21.00	16.30	13.60	12.20	0.40	<0.3	0.50	45.60	2.90
Sm	1.90	3.37	3.58	2.19	2.03	4.28	0.08	0.09	<0.05	11.15	0.53
Eu	0.72	0.71	1.01	0.40	0.60	0.08	0.03	0.02	<0.02	3.36	0.68
Gd	1.53	2.04	2.62	1.68	1.47	3.81	0.23	0.17	0.21	12.16	0.76
Tb	0.19	0.23	0.32	0.21	0.18	0.59	0.03	0.02	0.03	1.92	0.07
Dy	0.94	0.96	1.52	1.26	0.98	2.83	0.24	0.21	0.22	11.62	0.48
Ho	0.17	0.14	0.27	0.23	0.19	0.42	0.07	0.05	0.05	2.37	0.09
Er	0.47	0.36	0.75	0.90	0.51	0.97	0.25	0.17	0.15	6.90	0.31
Tm	0.06	0.05	0.11	0.12	0.08	0.14	0.02	0.02	0.02	0.94	0.04
Yb	0.37	0.29	0.66	0.94	0.45	0.95	0.15	0.14	0.15	5.86	0.32
Lu	0.06	0.04	0.09	0.16	0.08	0.13	0.02	0.02	0.02	0.90	0.04
ZREE	59.60	130.88	120.52	125.67	80.99	74.68	3.03	1.76	2.95	338.99	61.89
Cr ppm	20.53	13.68	20.53	13.68	20.53	13.68	6.84	34.21	6.84	47.89	27.37
Cr₂O₃ (wt%)	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00
Ni (ppm)	<20	<20	<20	<20	25.00	<20	7.10	2.10	3.50	670.00	37.00
Sc	4.00	3.00	5.00	1.00	3.00	11.00	<1	<1	<1	28.00	15.00
Sum (wt%)	99.83	99.81	99.82	99.95	99.76	99.91	99.97	99.98	99.98	90.08	99.94
Be (ppm)	3.00	1.00	1.00	<1	1.00	2.00	<1	<1	<1	<1	2.00
Co	8.60	6.10	8.60	1.20	8.80	2.70	7.70	9.20	8.70	2588.90	87.90
Cs	0.70	0.70	0.30	0.20	0.10	0.20	<0.1	<0.1	<0.1	1.40	0.10
Ga	17.40	17.40	19.30	13.50	15.60	25.40	5.00	4.90	6.20	21.40	20.90
Sn	<1	<1	1.00	<1	<1	4.00	<1	<1	<1	3.00	<1
Ta	<0.1	<0.1	0.20	0.80	0.10	9.00	<0.1	<0.1	<0.1	0.40	<0.1
U	0.20	0.60	0.40	2.30	0.70	2.80	<0.1	<0.1	<0.1	2.10	0.60
W	<0.5	<0.5	0.80	0.60	0.70	1.10	<0.5	<0.5	<0.5	2.00	1.90
Zr/Y	14.96	31.03	14.78	13.56	18.56	1.01	1.04	0.86	1.12	0.94	39.31
Nb/Y	0.60	0.61	0.33	0.82	0.34	3.98	0.21	0.19	0.29	0.07	0.15
Zr/Ti	0.05	0.07	0.07	0.14	0.07	0.07	0.01	0.02	0.01	0.02	0.07
Th/Yb	3.51	50.34	6.06	21.60	14.22	7.05	0.07	0.07	0.07	0.84	0.63
Nb/Yb	8.11	7.93	3.79	5.96	3.78	68.63	3.33	2.86	3.33	1.11	1.25
Mg#	0.11	0.22	0.24	0.10	0.14	0.12	0.02	0.02	0.02	0.01	0.04
Ti	1558.73	1618.68	1558.73	659.46	1378.87	239.80	299.76	119.90	179.85	3537.11	1498.78
Sr/Y	208.74	280.58	130.32	34.26	335.84	32.70	1.67	1.90	9.35	3.68	43.50
V/Sc	13.50	12.67	6.60	15.00	16.33	1.09	45.00	43.00	43.00	8.71	5.93

Table 1: Whole-rock geochemical data from the volcanic rock sampled. IBR – Iron-bearing rock; MBR – Manganese-bearing rock; AMS – Alto Maranhão Suite; APT – Aplite.

4.1.1 The Agua Limpa Unit

The Agua Limpa Unit comprises basaltic rocks with SiO₂ content ranging from 45 and 50 wt%. MgO from 4 to 8 wt%, TiO₂ between 1 and 3 wt%, Na₂O + K₂O between 1 and 3 wt% and Fe₂O₃ between 13 and 22 wt%. The Ni values vary from <20 ppm to 250 ppm, and Cr from 10 ppm to 270 ppm. In the basaltic-andesites SiO₂ varies from 52 to 57 wt%, MgO is low, around 6 wt%, TiO₂ range from 1 to 2 wt%, Na₂O + K₂O is around 5 and 6 wt% and Fe₂O₃ exhibit values around 10 to 15 wt%. The Ni content is <20 ppm to 170 ppm, while the Cr varies from 20 ppm to 300 ppm. The samples on the TAS diagram are located within the subalkaline field (Fig. 4a). According to the classification from Le Bass (2000), the tholeiite series samples plots in the basalt and basaltic andesite (Fig. 4c). Fenner's diagrams exhibit compositional trends with negative correlations for Fe₂O₃, MnO, CaO, Al₂O₃, TiO₂, and V (Fig. 5).

The Chondrite-normalized REE diagram (Sun and McDonough, 1989) in the basalts displays a flat standard of crystallization, indicating a juvenile magma association, while the basaltic andesite has an enrichment in LREE associated with arc environmental of volcanism (Fig. 6 a). The spider diagram (McDonough et al., 1992) of the trace elements shows a negative anomaly to Rb, Ta, and P, with positive anomalies to Th, La, and Nd, characterizing the volcanic arc environment (Fig. 6 b).

4.1.2 Santana dos Montes Unit

The analytical results of the Santana dos Montes volcanic rocks, crucial for our understanding of volcanic processes, fall, according to the classification from Le Bass (2000), in the fields corresponding to the picrite and boninite series (Fig. 4 c). Following the classification of Pearce and Reagan (2019) (Fig. 4 c, d, e, and f), the picrite is classified as LOTI (Low Ti Basalt), and basaltic andesite is SHMB (Silica High Magnesium Basalt).

The picrite samples are characterized by high MgO, 18 wt%, low SiO₂, 45 wt%, while the boninite display high silica, 52 and 57 wt%, and high magnesium from 18 and 25 wt%. Both rocks have the TiO₂ < 1 wt% and Na₂O + K₂O < 1 wt%. The picrite has

Fe_2O_3 ranging between 15 and 20 wt%, and the boninite's Fe_2O_3 content remains constant at approx. 10 wt%. The Ni and Cr in the picrite vary from 800 ppm to 1800 ppm and 1850 ppm to 3100 ppm, respectively. The boninite range of the Ni is 300 ppm to 1800 ppm, and the Cr varies from 700 ppm to 3400 ppm. In the TAS diagram (Fig. 4 a), the points plot the picrite samples in the basalt and boninite in basaltic andesite and andesite.

Fenner's diagram reveals significant compositional trends. Notably, it demonstrates a negative correlation to Fe_2O_3 , MnO, CaO, and Al_2O_3 , and a positive correlation to TiO_2 and SiO_2 (Fig. 5).

The normalized REE diagram (Sun and McDonough, 1989) in the picrite and boninite shows a slight LREE enrichment with a strong negative Ce anomaly (Fig. 6 c). The spider diagram (McDonough et al., 1992) of trace elements shows a negative anomaly to U, K, Ta, Sr, and Ti in the boninite, while the picrite shows a negative anomaly to K, Ta, Sr, Hf, Zr, and Ti (Fig. 6 d).

4.1.3 The Alto Maranhão Suite

The rocks from the Alto Maranhão Suite exhibit a wide range of SiO_2 contents, from 58.53% to 76.63% (total). In the TAS diagram (Fig. 4 b), we observe that intermediate samples are plotted in the field of diorite and monzonite, and acid samples in the field of granodiorite and granite. The suite of rocks also shows lower TiO_2 , Na_2O , and K_2O content. In addition to these, we also analyzed an associated aplite rock with a high SiO_2 content, 76 wt% (Fig. 4 a).

The Alto Maranhão Suite is enriched in LREE, Rb, Ba, K, La, Sr, and Tb. The aplite displays a strong negative Eu anomaly with Ta and Nb positive anomaly (Figs. 6 e and f).

4.1.4 The Morro da Mina Unit

The Fe_2O_3 content in the iron-bearing member is 54 - 56 wt%, and the SiO_2 is 41- 43 wt%. The MnO content in the manganese-rich member is 21 - 26 wt%, and SiO_2 is 24 - 50 wt%. Both rock types exhibit flat REE Chondrite normalized plots. The Mn-rich member is Ce-rich (see Cabral et al. 2019) (Fig. 6 g). The spidergrams of the trace elements show an enrichment in Ba, U, Nd, and Sm, whereas the iron-bearing rock shows an enrichment in U and Ta (Fig. 6 h).

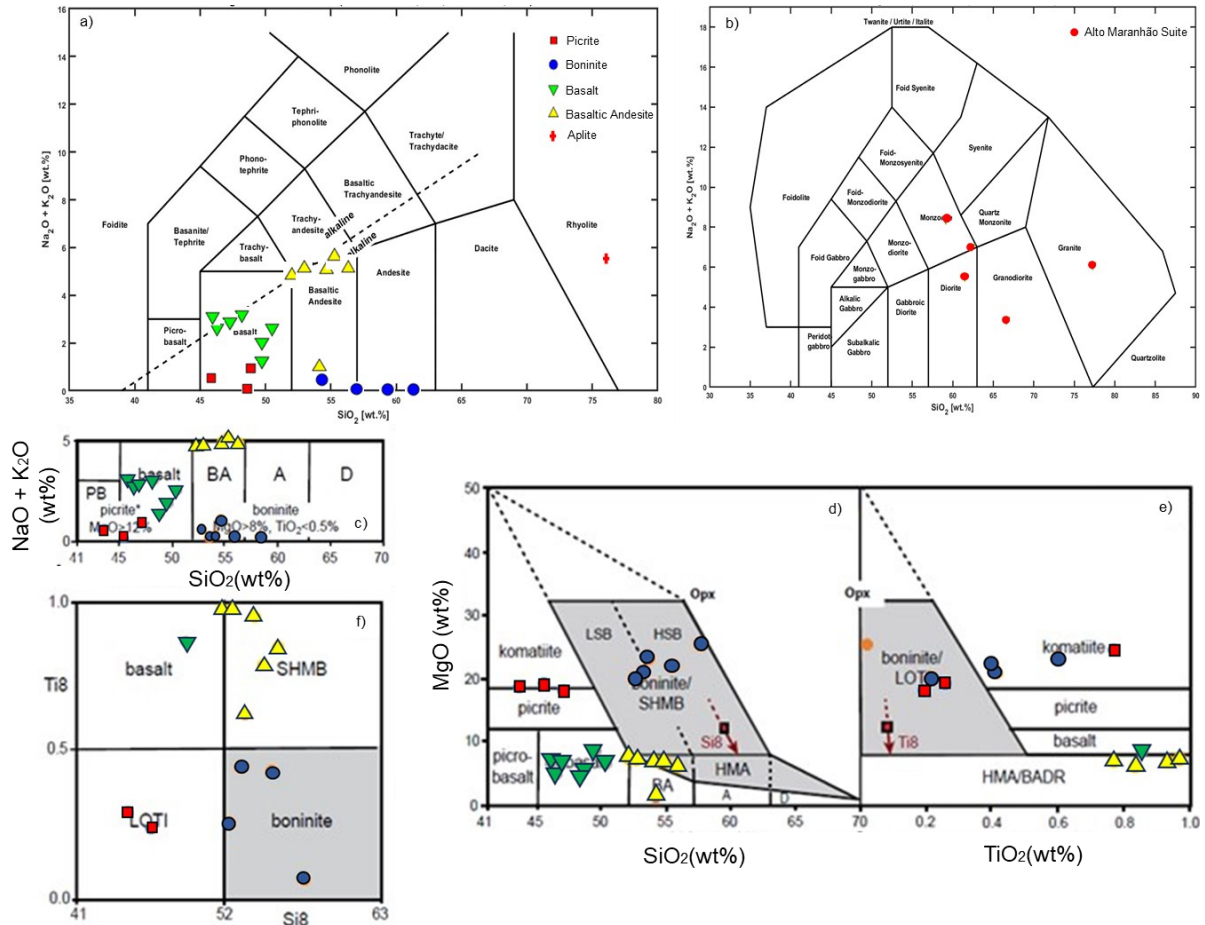


Fig. 4: a) TAS diagram for volcanic rocks after McDonald (1968) and Le Bas et al. (1986); b) TAS diagram for the plutonic rocks modified from Middlemost (1994); c), d), e), f) Major elements binary diagrams defined by Le Bas (2000) and Pearce and Reagan (2019) to classify high magnesium volcanic rocks.

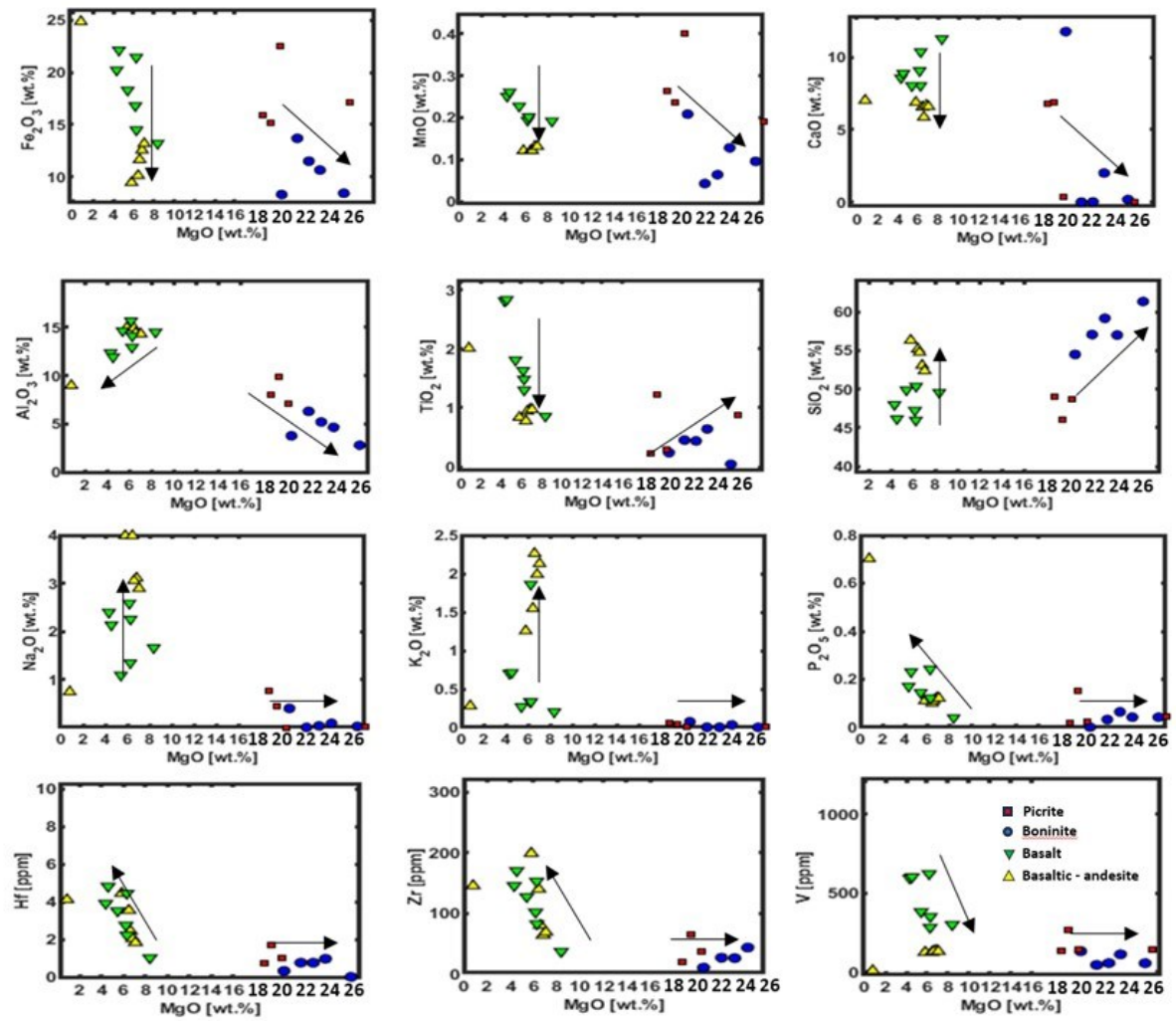


Fig. 5: Fenner's binary diagram showing different fractionation trends for selected major and trace elements versus MgO%.

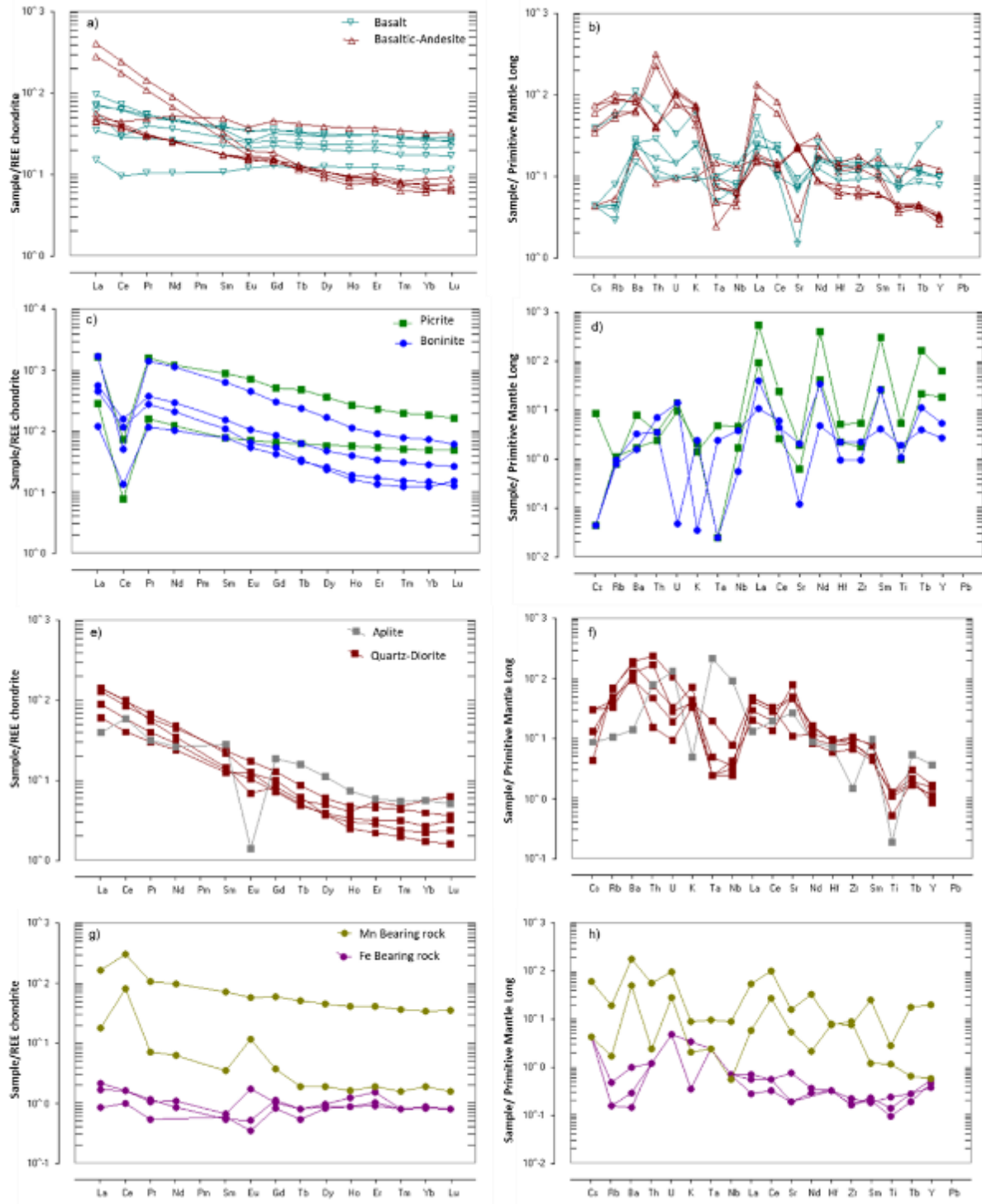


Fig. 6: a) Normalised chondrite patterns (Sun and McDonough, 1989) of analyzed basalt and basaltic andesite samples. b) Spidergrams normalized to the primitive mantle of McDonough et al. (1992) for basalt and basaltic andesite analysis. c) Normalised chondrite patterns (Sun and McDonough, 1989) of analyzed samples of picrite and boninite. d) Spidergrams normalized to the primitive mantle of McDonough et al. (1992) for picrite and boninite analysis. e) Normalised chondrite patterns (Sun

and McDonough, 1989) of analyzed samples of picrite and boninite. f) Spidergrams normalized to the primitive mantle of McDonough et al. (1992) for picrite and boninite analysis. g) Normalised chondrite patterns (Sun and McDonough, 1989) of analyzed samples of picrite and boninite. h) Spidergrams normalized to the primitive mantle of McDonough et al. (1992) for picrite and boninite analysis.

4.2 Geochronology

4.2.1 U-Pb geochronology

To establish a chronology for the geological events associated with the deposition of the Lafaiete Sequence were concentrated 70 magmatic zircons from 02 volcanic rocks, 150 detrital zircons from 03 sedimentary rocks, and 15 magmatic zircons from a quartz-diorite and aplite of the Alto Maranhão Suite. Only the pristine zircons with levels of concordance of 90-110% in the volcanic rocks and 95-105% in sedimentary rocks were considered for interpretation. The results are summarized in Table 2.

Sample	Lithology	Coordinate	U-Pb	Interpreted age	Lu-Hf	
			U-Pb age (Ma) $\pm 2\sigma$		$\varepsilon_{\text{Hf}}(t)$	T_{DM2} (Ga)
956	Basalt	43°30'38.700" 20°43'26.220"	2170.3 $\pm$ 8.2 (MSWD = 1.9) n=10	Crystallization	(-6.1 to -7.8) n=3	2.99 - 2.91
99	Boninite	43°36'05.103" 20°45'25.207"	2149.8 $\pm$ 5.9 (MSWD = 0.91) n=9	Crystallization	(-4.8 to -16.1) n=24	3.41 - 2.84
955a	Quartz-Diorite	43°31'04.323" 20°43'26.912"	2140.7 $\pm$ 15 (MSWD = 3.2) n=6	Crystallization	(3.4 to -6.9) n=9	2.90 - 2.39
955b	Aplite	43°30'36.911" 20°43'26.912"	2131.4 $\pm$ 9.5 (MSWD = 1.8) n=8	Crystallization	(-6.5 to -7.5) n=4	2.98 - 2.92
73	Quartzite	43°35'56.913" 20°45'28.802"	2099 $\pm$ 9	Maximum age of deposition	(-1.1 to -13.3) n=25	3.302 - 2.052
987	Greywacke	43°45'42.068" 20°38'52.327"	2079 $\pm$ 6.3 (MSWD = 1.3) n=19	Maximum age of deposition	---	---
130	Mn bearing rock	43°45'42.087" 20°42'03.589"	2064.5 $\pm$ 6.2 (MSWD = 0.51) n=8	Maximum age of deposition	(-10.5 to 0.0) n=21	2.417 - 2.046

Table 2: U-Pb and Lu-Hf data obtained for the studied volcanic and sedimentary samples.

4.2.1.1 Volcanic rocks

The zircon grains of all volcanic rocks are highly fractured and hydrothermally altered, with inclusions of chlorite, preserving only some points of non-metamict core available for suitable analysis. Martins et al. (2022) proposed that grains from the same populations present similar internal structures, indicating multiple zircon growth and alteration stages. Only the zircons from the basalt show fewer signs of resorption and recrystallization (Fig. 7 a). Generally, the grains display large, fractured metamict domains that are entirely heterogeneous, structureless, and porous (Fig. 7 b).

The zircons from the basalt display a euhedral and anhedral form with a size between 100 to 200 μm (Fig. 7 a). The Th/U varies from 0.02 to 1.35, and values <0.2 suggest the influence of crustal contamination (see Supplementary Table S1). Twelve spots were considered in the basalt sample. A regression through these points defines a Pb-loss Discordia with an upper intercept age of 2170 ± 8.2 Ma (MSWD=1.9), interpreted as crystallization age (Fig. 7 a).

The zircon grains from boninite are euhedral, forming a prismatic structure with a straight border. They are intense fractures preserving only minor points with magmatic signatures, showing on the surface a dissolution metamorphic process “imprinted” by Fe-chlorite from the matrix of the ultramafic (Fig. 7 b). 37 spots in 09 grains were measured in the boninite sample. A regression through these points defines a Pb-loss Discordia with an upper intercept age of 2149 ± 5.9 Ma (MSWD=0.026), interpreted as crystallization age (Fig. 7 b). A metamorphic overprint was defined in 2098.8 ± 6.2 Ma (MSWD=1.5), which is interpreted as the age of the amalgamation of the arc into the Minas block.

4.2.1.2 Plutonic rocks and aplite dykes

Two distinct zircon sample batches were analyzed in the Alto Maranhão Suite: 10 zircons from quartz-diorite and 05 zircons from aplite dikes that fill the fabric of the diorite. The zircons from both samples present similar morphologies, with euhedral and elongated characteristics, ranging from 80 μm to 220 μm in length (Figs. 7 c and d).

The grains exhibit oscillatory zoning, with fractures and reaction borders. In both cases, analyses were performed in the growth areas of different grains to obtain the igneous crystallization age (Carneiro de Oliveira et al., 2024) (Figs. 7 c and d).

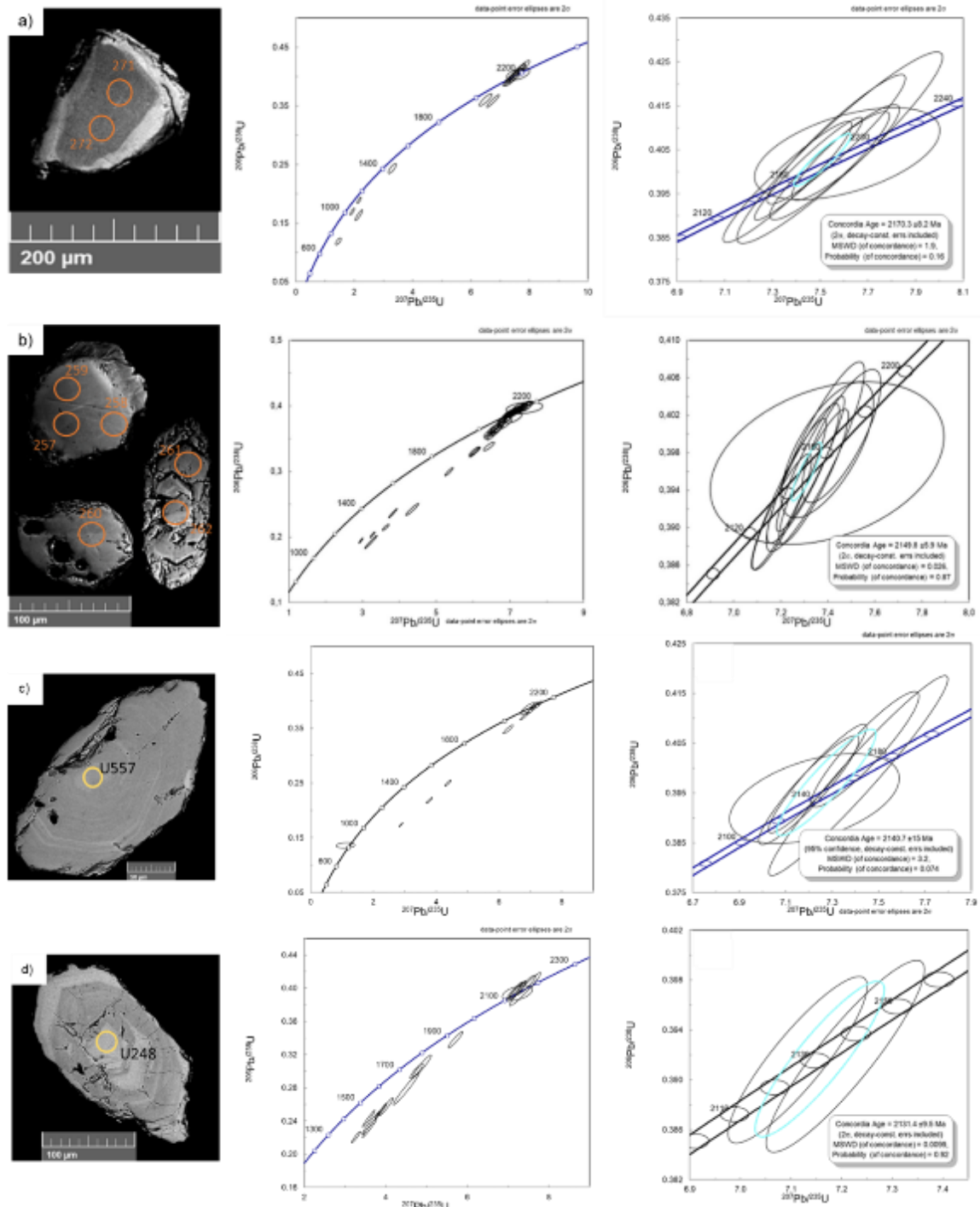


Fig. 7: Zircons analyzed in the volcanic rocks: (a) basalt zircons and (b) boninite zircons. Fig. 13: Concordia diagrams (2σ ellipses error) for zircon U-Pb LA-ICP-MS analyses: a) quartz-diorite; b) aplite.

Nine grains of zircon from the quartz-diorite revealed concordance values over 90% and were used to determine the crystallization age of that rock (Fig. 7 c). The $^{206}\text{Pb}/^{238}\text{U}$ analysis led to a Concordia age of 2140.7 ± 15 Ma (MSWD = 3.2, 2σ), which is considered in the error margin of the age obtained to the aplite (Fig. 7 d) in 2131.4 ± 9.5 Ma (MSWD = 0.0099, 2σ). The age of the aplite, in turn, was obtained by analyzing 06 spots with a concordance higher than 90%.

4.2.1.3 The Morro da Mina Unit

The zircons of sedimentary rocks are predominantly euhedral, with size varying from 150 to 250 μm long and 50-80 μm wide and with the border abrasive (Fig. 8). The oscillatory zoning in the CL images and low Th/U ratios (0.1 to 0.2, Supplementary Table S2). A total of 24 spots were analyzed in the grains from quartzite, 42 spots were analyzed in the grains from greywacke, and 29 spots were analyzed in the grains from Mn-bearing rock.

The quartzite sample presents a distinct population of euhedral grains with different sizes and grades of alteration. The analyses in the quartzite show a predominant Archean source of the sediments, with a significant concentration at Mesoarchean. This sample also presents some grains with Rhyacian age (Fig. 8 a and d).

The greywacke sample presents a significant concentration of zircons of Rhyacian age, with some grains of Siderian and Mesoarchean age. (Fig. 8 b and e).

The Mn-bearing rock shows the concentration of detrital zircons in the Paleoproterozoic source, with a great concentration in the Rhyacian (Fig. 8 c and f).

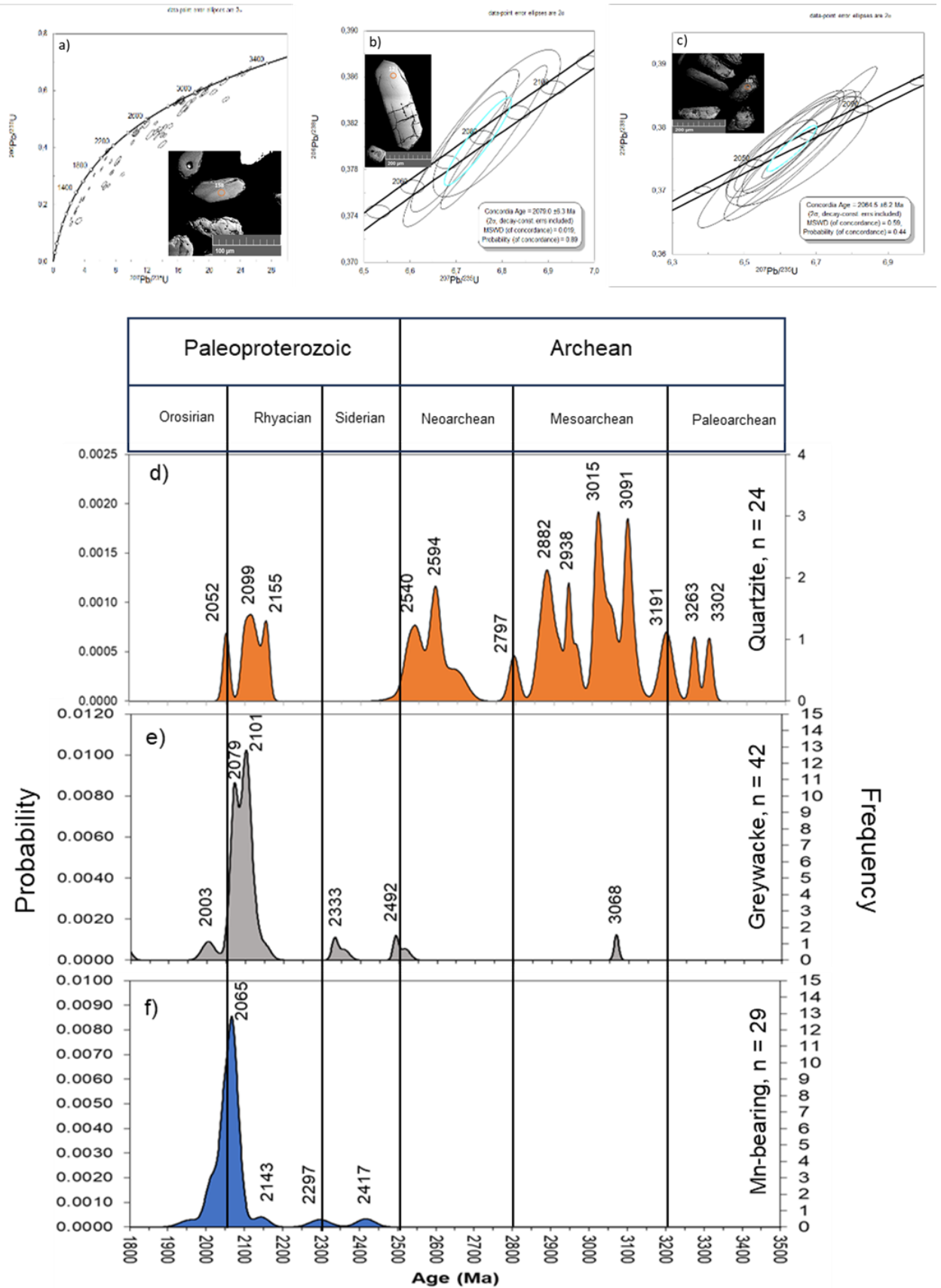


Fig. 8: Detrital zircons age data presented in relative Probability vs. Age (Ma) diagrams (only data with 95% – 105% concordance). (a, d) Quartzite showing a

maximum age of deposition in $2099 \pm 9\text{Ma}$; (b, e) Greywacke showing a maximum age of deposition in $2079.0 \pm 6.3\text{Ma}$; (c, f) Mn-bearing rock showing a maximum age of deposition in $2064.5 \pm 6.2\text{Ma}$. See details for the detrital analyses in Supplementary Table S2.

4.2.2 Lu-Hf signature

Eighty-six (86) Lu-Hf isotopic analyses were performed in 6 samples that have been already documented in terms of U-Pb isotopic analysis: two (02) from the volcanic rocks (basalt and boninite); two (02) from the quartz-diorite and aplite of the Alto Maranhão suite; two (02) from the sedimentary unit (quartzite and Mn-bearing rock). The spots analyzed show 95% to 105% concordance in U-Pb ages.

The initial $^{176}\text{Hf}/^{177}\text{Hf}$ values of magmatic spots from the basalt sample (ca. 2170 Ma) varied from 0.281195 to 0.281233. The Hf T_{DM} model ages yield values between ca. 2.99 and 2.91 Ga with negative $\epsilon\text{Hf}(t)$ values between - 7.8 and -6.1 (Fig. 9 and Supplementary Table S2).

The initial $^{176}\text{Hf}/^{177}\text{Hf}$ values of magmatic spots from the boninite sample (ca. 2150 Ma) varied from 0.280990 to 0.281276. The Hf T_{DM} model ages yield values between ca. 3.41 and 2.89 Ga with negative $\epsilon\text{Hf}(t)$ values between - 16.1 and - 4.8 (most data around - 16.1; Fig. 9 and Supplementary Table S2).

The initial $^{176}\text{Hf}/^{177}\text{Hf}$ values of magmatic spots from the quartz-diorite from Alto Maranhão suite (ca. 2140 Ma) varied from 0.281263 to 0.281517. The Hf T_{DM} model ages yield values between ca. 2.90 and 2.37 Ga with positive to negative $\epsilon\text{Hf}(t)$ values between - 6.9 and +3.4 (Fig. 9 and Supplementary Table S2).

The initial $^{176}\text{Hf}/^{177}\text{Hf}$ values of magmatic spots from the aplite that fill the Alto Maranhão suite sample (ca. 2130 Ma) varied from 0.281202 to 0.281238. The Hf T_{DM} model ages yield values between ca. 2.98 and 2.92 Ga with positive to negative $\epsilon\text{Hf}(t)$ values between - 7.5 and + 6.5 (Fig. 9 and Supplementary Table S2).

The initial $^{176}\text{Hf}/^{177}\text{Hf}$ values of detrital spots from the quartzite sample (ca. 2100 Ma) varied from 0.280374 to 0.281502. The Hf T_{DM} model ages yield values between

ca. 4.02 and 2.44 Ga with positive to negative $\epsilon\text{Hf}(t)$ values between - 13.3 and + 0.9 (Fig. 9 and Supplementary Table S2).

The initial $^{176}\text{Hf}/^{177}\text{Hf}$ values of detrital spots from the Mn-bearing rock sample (ca. 2065 Ma) varied from 0.281143 to 0.281436. The Hf T_{DM} model ages yield values between ca. 3.07 and 2.61 Ga with positive to negative $\epsilon\text{Hf}(t)$ values between - 10.5 and +0.0 (Fig. 9 and Supplementary Table S2).

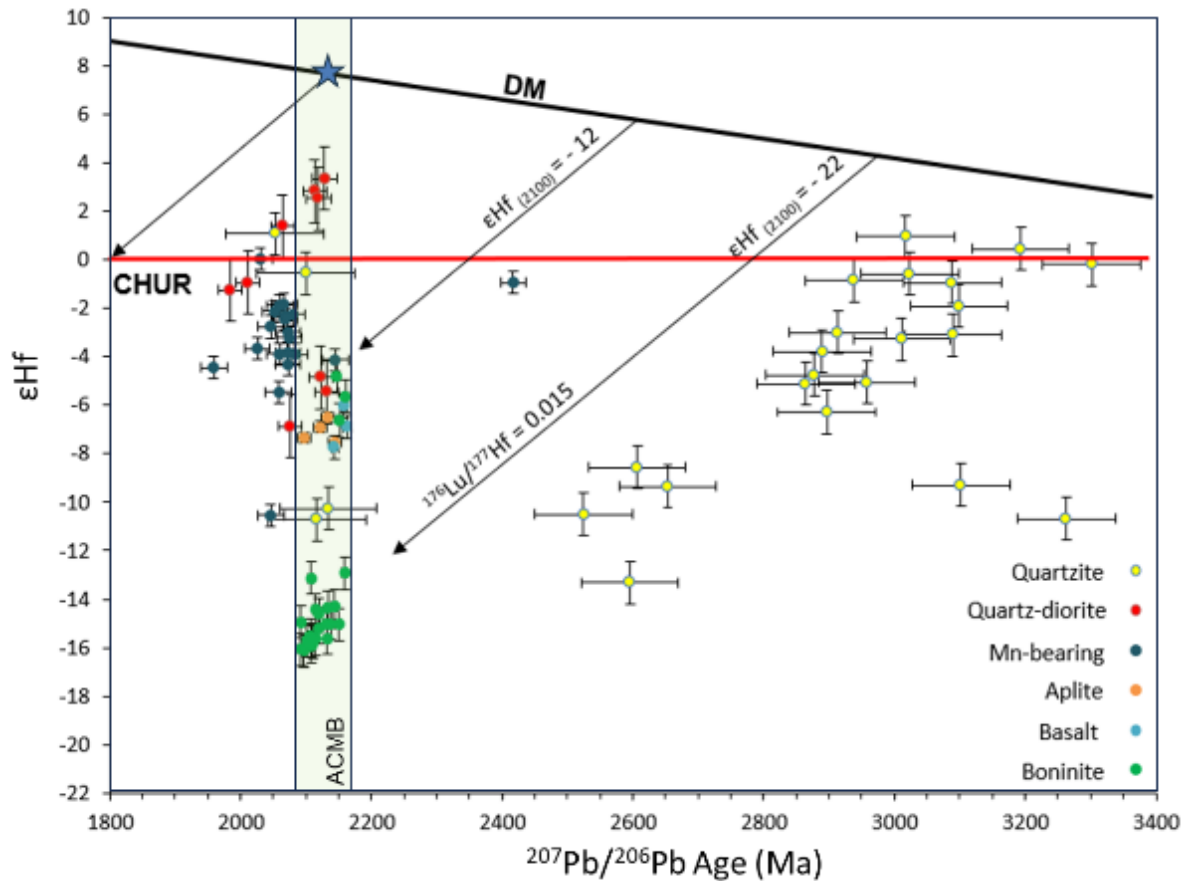


Fig. 9: Hafnium isotopic composition of zircons from volcanic and sedimentary rocks of the Mineiro Belt, Brazil, represented for the diagram $^{207}\text{Pb}/^{206}\text{Pb}$ Age (Ma) vs. ϵHf . The blue star represents the Primitive Mantle source in ~2125 Ma. Error bars indicate 2 σ confidence level. ACMB – Accretionary-to-collisional period of the Mineiro Belt terranes against the Minas Block from ~2170 Ma to 2100 Ma. See details for DM and CHUR parameters in Supplementary Table S1.

5.1 Volcanic magma sources and emplacement time

The comprehensive whole rock geochemistry and petrography analysis have unveiled a significant geological phenomenon with three distinct volcanic magma pulses in the northern Mineiro Belt. These pulses occurred during the amalgamation of the Alto Maranhão arc against the Minas Block after a period of magmatic lull in the southern São Francisco craton between ca. 2195 Ma and 2170 Ma.

Geochronological and whole rock geochemistry supported by extensive fieldwork leads to the conclusion that the Lafaiete Sequence comprises a volcano-sedimentary succession formed by two volcanic units overlaid by a sedimentary at the top in an unconformity contact. The volcanics begin as a tholeiite series forming basalt-to-basaltic andesite. This sequence is followed by an intrusive ultramafic volcanic sequence composed of dunitic-to-harzburgitic magma contaminated with crustal sediments that form a sequence of picrite-to-boninite series enriched in Zr. The volcanic units are unconformably overlain by a clastic-chemical sequence comprising quartzite, Mn-bearing rock, Fe-bearing rock, black shale, graphite schist, and greywacke (Fig. 1c).

Intrusive in the volcanic sequence, formed during the collisional process of the Mineiro Belt with the Minas Block amid the Minas-Bahia orogeny (Bruno et al., 2021 a), a voluminous sanukitoid plutonic suite (Alto Maranhão) crystallized at 2130Ma (Seixas et al., 2013; Moreira et al., 2018). These rocks encompass a suite of tonalitic and dioritic rocks composed of plagioclase, minor K-feldspar, hornblende, and biotite with variable proportions of quartz derived from a magma composition of high-Mg andesites, implying they originated from partial melting of a mantle wedge, which was metasomatized to different degrees by slab-derived melts (Seixas et al., 2013).

The first volcanic unit in the Lafaiete Sequence, a Fe-rich tholeiite series, crystallized in 2170 Ma, is a testament to the intricate nature of geological processes. It is related to the early stages of subduction in a forearc setting and is characterized by relative enrichment in LREE and LILE (Fig. 8 a, b), with a positive correlation between SiO₂, Na₂O, and K₂O about MgO in a basalt-to-basaltic-andesite volcanic series (Fenner's diagram, Fig. 7). The ϵ_{Hf} present a sub-chondritic source and

Archean T_{DM} ages (Fig. 9), further adding to the complexity of the geological history of the Mineiro Belt.

The second magma pulse in the Lafaiete Sequence is characterized by the picrite-to-boninite series, a high-magnesium olivine basalt derived from the early stages of the subduction of tholeiitic basalt setting (Fig. 6 a, b). The beginning of the subduction of the tholeiitic magma induces a partial melting of a metasomatized mantle, forming a boninite magma crystallized in 2150 Ma. Boninite is a rare Mg- and Si- high and Ti-low rock (Pearce and Arculus, 2021). The magma originated from a dunitic-to-harzburgitic composition (Fig. 11 a), with olivine and orthopyroxene as the primary liquidus phases of boninite in modern subduction settings and in many ophiolite complexes (Shen et al., 2022). According to Cameron et al. (1979), melting a mantle segment that has been depleted in incompatible elements by previous melting episodes would require very high temperatures to produce such magnesian liquids. The pristine magma is contaminated with crustal material, crystallizing a picrite-boninite series enriched in LREE and incompatible elements and depleted in LILE, with a strong Ce negative anomaly (Fig. 6 c). The Ce negative anomaly can be related to the introduction of the sediments in the early stage of the subduction (Bellot et al., 2018). The enrichment in incompatible elements (Fig. 9 b), together with negative ϵ_{Hf} and Archean T_{DM} ages (Fig. 9), reveal the influence of Archean continental crust during the melt emplacement.

The third sequence of volcanic magma, which crystallized an aplite volcanic rock in 2130 Ma (Carneiro de Oliveira et al., 2024), can be attributed to the fractionation process of mafic-ultramafic magma. The enrichment of LREE, as indicated by the analysis of the plutonic of Alto Maranhão Suite, to which this volcanic is related, is a significant finding. The aplite, with its negative Eu anomaly associated with the previous crystallization of plagioclase in the plutonic rocks (Fig. 6 e, f), provides valuable insights. The ϵ_{Hf} , with its sub-chondritic source and Archean T_{DM} ages (Fig. 9), demonstrate the juvenile character of this magma source.

5.2 Alto Maranhão Suite

According to the model presented by Moreira et al. (2020), the sanukitoid suite of the Alto Maranhão formed as a product of the interaction of magma proveniencies from the oceanic crust subducting after a quiescence period known as “magmatic lull” that interacted with weathering and erosional materials developed in an arc setting. This process recycled sediments back into the mantle, retaining a relatively juvenile Hf signature. This process can be described as self-feeding of the arc (Moreira et al., 2020) formed between ca. 2140 Ma to 2130 Ma.

5.3 Sedimentary deposition and provenance

U-Pb ages and Hf isotope data were carried out in detrital zircon grains from three sedimentary rocks comprising the Morro da Mina Unit: a quartzite, greywacke, and Mn-bearing rock.

Quartzite contains abundant detrital zircon grains with ages > 2.5 Ga and some grains in the Rhyacian age (Fig. 15 a). The initial $^{176}\text{Hf}/^{177}\text{Hf} < 0.281502$. This significant dispersion in terms of source of sedimentation indicates that the quartzite was deposited at different times and supplied from distinct sources, concentrated in the Archean proto craton to the north, but with also a proximal source originating from the erosion of the rocks that were formed in the Rhyacian orogen system. The youngest ages of detrital zircon grains and the timing of metamorphism in the Mineiro Belt's northern portion constrain quartzite maximum age of deposition in 2099 Ma (Figs. 8 a and d).

Greywacke concentrates detrital zircon grains in the Rhyacian age, but some grains also show the Siderian and Mesoarchean ages (Figs. 8 b and e). The initial $^{176}\text{Hf}/^{177}\text{Hf} < 0.281502$. This concentration of the grains in the Rhyacian age, associated with the morphology of the grains, indicates that the main source of the sedimentation is the proximal sanukitoid suite eroded in a post-orogenic setting.

Mn-bearing rock also concentrates the detrital zircon grains in the Rhyacian period, with few grains showing the Siderian age (Figs. 8 c and f). The $^{176}\text{Hf}/^{177}\text{Hf} < 0.281436$. This concentration of the grains in the Rhyacian age follows the greywacke distribution of the grains and is associated with the morphology of the grains, which

can indicate that the main source of the sedimentation is the proximal sanukitoid suite eroded in a post-orogenic setting.

5.4 Lithostratigraphic sequence

The proposed stratigraphic evolution for the northern portion of the Mineiro Belt is based on the published data and new geochronological, petrological, and geochemistry data presented in this paper (Fig. 10). The magmatic arcs were compiled from Moreira et al. (2018) and Teixeira et al. (2022).

According to Barbosa et al. (2015), the history of the Mineiro Belt starts with the formation of the Cassiterita magmatic arc at ca. 2470 - 2410 Ma.

The second magmatic arc is Lagoa Dourada – Resende Costa, which comprises a voluminous intra-oceanic sequence intruded by granitoids at ca. 2350 – 2317 Ma (Seixas et al., 2012; Teixeira et al., 2015).

Following the evolution of the Mineiro Belt, the next magmatic arc formed is the Serrinha – Tiradentes, composed of different volcano-sedimentary sequences intruded by juvenile granitoid at ca. 2230 to 2200 Ma (Ávila et al., 2010, 2012, 2014).

The Alto Maranhão Arc is composed of a voluminous sanukitoid suite crystallized in ca. 2140 – 2130 Ma (Seixas et al., 2013; Moreira et al., 2019; Carneiro de Oliveira et al., 2024) that derived from the subduction of the tholeiite-to-boninite series of the volcanic rocks formed in ca. 2170 - 2150 Ma.

Overlying the volcanic sequence and the sanukitoid suite is a sedimentary chemical-clastic unit with three maximum deposition ages related to the break-up of the orogen system: 2.099 ± 9 Ma in a fuchsite-quartzite, 2079 ± 6.3 Ma in greywacke, and 2065 ± 6.2 Ma in an Mn-bearing rock.

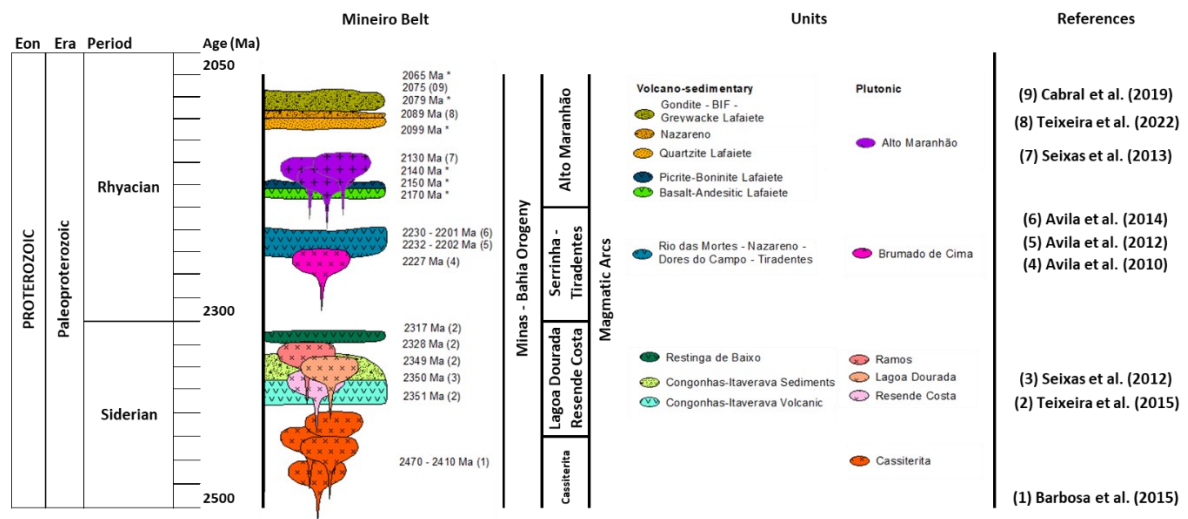


Fig. 10: Stratigraphy column based on (1) Barbosa et al., 2015: 2470 to 2410 Ma; (2) Teixeira et al., 2015: 2351 Ma to 2317 Ma; (3) Seixas et al., 2012: 2350 Ma; (4) Avila et al., 2010: 2345 ± 14 Ma; (5) Avila et al., 2012: 2232 Ma to 2202 Ma; (6) Avila et al., 2014: 2230 Ma to 2201 Ma; (7) Seixas et al., 2013: 2130 Ma; (8) Teixeira et al., 2022: 2089 ± 16 Ma; (9) Cabral et al., 2019: 2075 ± 12 Ma. (*) present the results from this study.

5.5 Tectonic setting

The same chemical characteristics of low-Ti-K tholeiitic basalts and boninites, derived from high refractory mantle sources (Duncan and Green, 1987), that occur in the modern style of forearc basins associated with early stages of subduction in the Izu-Bonin-Mariana (IBM) system (Li et al., 2022), can be observed in the Lafaiete volcanic units. The tholeiite series has an initial geochemistry composition from orthopyroxene + olivine associated with the basalt to basaltic andesite derived from a forearc basin setting (Fig. 11 a). With the beginning of subduction, an increase in partial melting from the metasomatized mantle contaminates the magma, forming a picrite-boninite series of rock (Fig. 6 a). Analyzing the major elements (MgO , $\text{Fe}_2\text{O}_3 + \text{TiO}_2$, Al_2O_3) as proposed by Jensen (1976) to the mafic-ultramafic volcanic rocks, it is possible to distinguish two classes of magmas associated with the volcanic rocks from Lafaiete Sequence: the first, associated to Fe-rich tholeiite basalt crystalizing basalt to basaltic - andesite. The second is related to high MgO fractionated from a komatiite to

composition following a trend of crustal contamination. On the other hand, the boninite series follows the subduction trend in a modern arc array (Fig. 11 d).

In Fig. 12, we propose a tectonic setting evolution to the study area. According to Moreira et al. (2019), the proto-subduction began around 2200 Ma as part of the mantle wedge hypothesis for sanukitoid derivation. At this time, the incremental density of eclogite residue at the base of the crust reached a 'tipping point', causing this dense material to sink into the mantle (e.g., Bédard, 2018). The subduction started with north vergence, developing a forearc basin as a compression response to the subduction. Beginning at 2170 Ma, a tholeiite series of magma is formed, followed by a second pulse of boninite series at 2150 Ma. The boninite has a magma composition of high-Mg andesites in modern subduction zones, which is the exact composition of the sanukitoid from the Alto Maranhão suite, crystallized between 2140 and 2130 Ma, suggesting that this suite originated from partial melting from a mantle wedge, which was metasomatized to different degrees by slab-derived (Seixas et al., 2013). In 2100 Ma, the collisional dynamics may have involved a soft collision with plate overriding along the Jeceaba-Bom Sucesso, Congonhas-Itaverava, and Lenheiros Lineaments, juxtaposing two different blocks along its trace. After the collision, an extensional period was responsible for the erosion and deposition of sediments, with most zircons seeming to come from the same proximal granitic source (euhedral zircons with the same typologies) between 2100 and 2065 Ma.

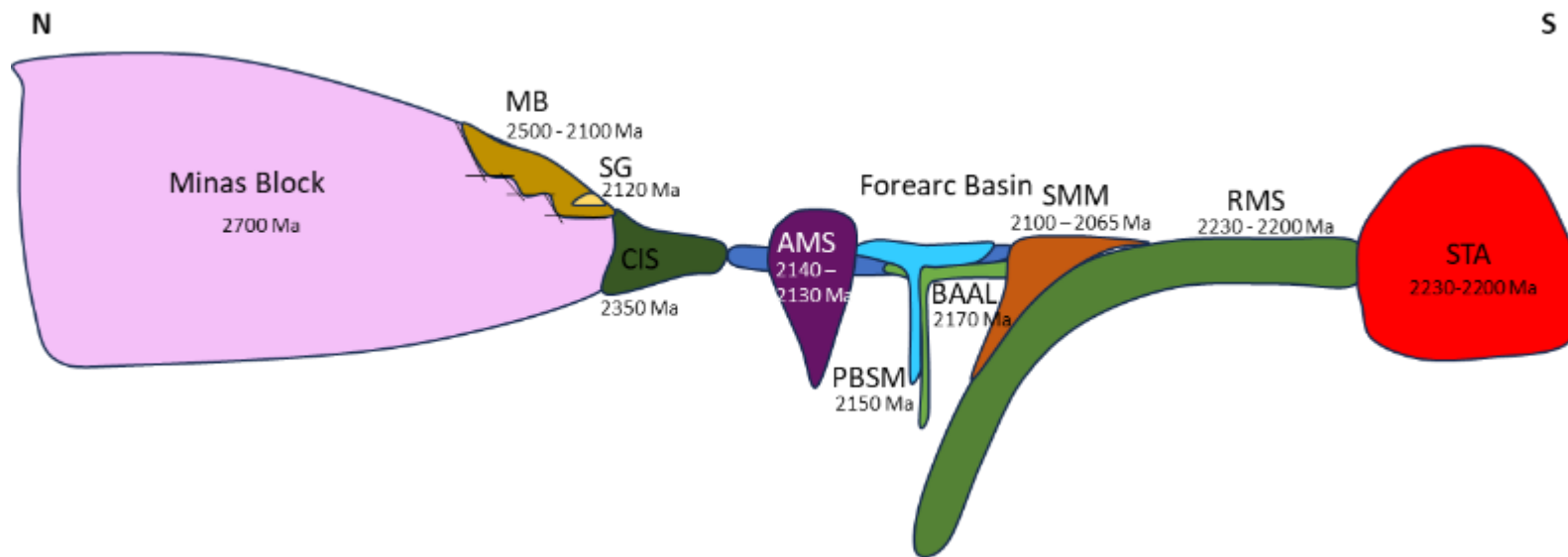


Fig. 12: Schematic model illustrating the tectonic setting developed in the Rhyacian orogen between the Minas Block in the north and the sequences that comprise the Mineiro Belt in the south. MB – Minas Basin; SG – Sabará Group (Martinez Dopico, 2017); CIS – Congonhas-Itaverava Sequence (Teixeira et al., 2015); RMS – Rio das Mortes Sequence (Avila et al., 2010); STA – Serrinha-Tiradentes Arc (Ávila et al., 2010); AMS – Alto Maranhão Suite (Seixas et al., 2013); Lafaeite Sequence: BAAL – Água Limpa Tholeiite Series; PBSM – Santana dos Montes Boninite Series; SMM – Morro da Mina Sediments.

6 Conclusions

The geochronological and petrological data presented to the Lafaiete formation reveal a volcano-sedimentary sequence in the north of Mineiro Belt that has a crustal accretion implication:

1. The Lafaiete sequence can be individualized compared to Congonhas or Rio das Mortes Sequences.
2. The U-Pb dating presents a long-lived history of 100 Myr in the geodynamic evolution of this critical crustal segment of the Earth. The age of 2170.3 ± 8.2 Ma obtained in the basalt shows the beginning of magmatism in a forearc setting after tectonic quiescence represented by a magmatic lull period. It is followed by an ultra-magnesian-low Ti magmatism associated with picrite-boninite series, crystalized in an early stage of subduction in 2149.8 ± 5.9 Ma derived from dunite-to-peridotite magma. The evolution of the subduction process is responsible for the crystallization of the Alto Maranhão sanukitoid suite from 2140 to 2130 Ma. Overlying the volcanic sequence, a sedimentary unit showing the first peak of maximum deposition ages in 2099 ± 9 Ma, where the sediments express the provenance from the granite-greenstone terrain in the proto craton as well as erosion of rocks formed in the Rhyacian orogen. To the top of the sequence, the sedimentation is related to proximal rocks from the Rhyacian orogen system in 2079 ± 6.3 Ma and 2064.5 ± 6.2 Ma. The metamorphism is marked in the boninite unit in 2098.9 ± 6.2 Ma and can be related to the amalgamation of the Mineiro Belt against the Minas Block.
3. The volcanic sequence displays the forearc environment in the Mineiro Belt chrono-correlated with the sediments of the Sabará Group in the Minas basin. A subduction process with vergence to the proto craton in the north forms a forearc setting with sediment apport from the rocks formed in the orogen system.
4. The enrichment in incompatible elements in the boninite volcanic rock, LREE, and negative ϵ_{Hf} and Archean T_{DM} ages reveal the influence of the Archean continental crust during the emplacement of the melt.
5. Two sources of sediments are shown in the Hf isotope study of the sedimentary rocks: one associated with the Archean basement of the São Francisco

craton responsible for the maximum age of deposition in 2099 ± 9 Ma to the quartzite. The other samples show 2079.0 ± 6.3 Ma and 2064.5 ± 6.2 Ma to the Mn-bearing rock and greywacke, respectively. Both ages are related to the maximum age of deposition presented by Cabral et al. (2019) to the Mn protolith of Morro da Mina Mn deposit in 2075 ± 12 Ma.

7 Author statement

Eduardo Oliveira: Conceptualization, Methodology, Formal analysis, Investigation, Writing – Original Draft. Carlos Rosière: Conceptualization, Methodology, Validation, Investigation, Writing - Review and Editing, Supervision, Funding acquisition. João Paulo de Souza: Methodology, Formal analysis, Validation, Writing - Review and Editing. Armin Zeh: Methodology, Formal analysis, Validation, Writing - Review and Editing. Flávia C. S. Braga: Formal analysis, Writing - Review and Editing.

8 Declaration of competing interest

The authors declare that they have no interests, competing financial interests, or personal relationships that could have appeared to influence the work reported in this paper.

9 Data availability

Data will be made available on request.

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Artigo 3: Magnetite geochemistry and geology of the Água Limpa Fe-(Cu) Deposit: Implications for ore genesis

Eduardo Luís Carneiro de Oliveira^{1*}, Carlos Alberto Rosière¹

¹: Programa de Pós-graduação em Geologia e Centro de Pesquisas Manoel Teixeira da Costa, Instituto de Geociências, Universidade Federal de Minas Gerais (CPMTC-IGC-UFMG), Campus Pampulha, Av. Antônio Carlos 6627, CEP 31270-901 Belo Horizonte, MG, Brazil

* Corresponding e-mail: oliveira.eduardoluiscarneiro@gmail.com

Esse trabalho encontra-se em desenvolvimento.

Abstract

A combination of field geology, petrographic observations, whole-rock geochemistry, EMPA, and laser ablation inductively coupled plasma mass spectrometry (LAICP-MS) was applied in this study to distinguish the origin and determine compositional variations in igneous and hydrothermal magnetite from the Água Limpa Fe-(Cu) magmatic-hydrothermal deposit hosted by a volcanic unit of the Paleoproterozoic Lafaiete Formation (2170 Ma) in the accretionary terranes of the Mineiro Belt located at the southern Sao Francisco Craton. The deposit comprises discontinuous lenses and layers of magnetite-hornblende-epidote-quartz layered with pyrite-chalcopyrite-pyrrhotite as a sulfide phase along 5 km in the strike and 30 m in thickness, hosted in a metabasalt in contact with the Água Limpa pluton from the Alto Maranhão Suite (2130 Ma). The texture and mineral chemistry indicate the crystallization of three generations of magnetite: one dense-granular, hypidiomorphic to idiomorphic, suggesting a magmatic origin by direct crystallization associated with hornblende and biotite, enriched in Ti, Ca, Mg, and Al (Mag 1). A second magnetite (Mag 2) is porous-granular, hypidiomorphic to idiomorphic, suggesting a dissolution and re-precipitation phase caused by the introduction of high-temperature Ca-rich fluid with high oxygen fugacity altering the previous compact magnetite, which is depleted in Ti, Ca, Mg, and Al concerning the Mag 1, and has the RIM depleted in Fe. A third

phase (Mag 3), platy-shaped mushketovite, is associated with late-stage hydrothermal mineralization, related to a change in oxidized to a more reduced environment, thereby creating conditions favorable for the precipitation of pyrite-chalcopyrite-pyrrhotite that occurs as disseminated and in small veins cross-cutting the magnetite-rich rock. Electric microprobe analysis (EMPA) geochemistry map shows a chemical variation from the center to the border of magnetite, hornblende, and biotite. The results of LAICP-MS trace elements analysis in magnetite from two different samples of magnetite-rich rock attest to the magmatic-hydrothermal characteristic of the Fe-(Cu) mineralization generated from metasomatic fluids responsible for forming a magmatic-hydrothermal deposit hosted in a metabasalt.

Keywords

Fe-(Cu) deposit, magnetite chemistry, Mineiro Belt, São Francisco craton

1. Introduction

The Agua Limpa Fe-(Cu) deposit in the northeastern portion of the Mineiro Belt (Teixeira and Figueiredo, 1991), close to the Congonhas-Itaverava and Lenheiros lineaments (Fig. 1 a, b), is composed of granoblastic magnetite hosted in discontinuous lenses associated with hornblende, epidote, and quartz intersected by late stages of the sulfide veins comprised of pyrite, pyrrhotite, and chalcopyrite. The orebody straddles the 5-km contact zone between the metabasalt ascribed to the volcano-sedimentary Lafaiete Formation (Pires, 1977) and an intrusive plutonic body of the Alto Maranhão Suite (Seixas et al., 2013) (Fig. 1 c). Investigation of the ore fabric and the mineral chemistry of the magnetite suggests a primary magmatic origin for the crystallization of iron oxides altered by the circulation of hydrothermal fluids that evolved from oxic-to-sulfidic conditions and affected the external walls of the orebodies. The mineralization developed during the collision of the accreted Siderian to Rhyacian terranes that comprise the Mineiro Belt with the São Francisco Protocontinent in the course of the Minas-Bahia orogeny (Bruno et al., 2020)

In 2001, the Companhia de Desenvolvimento de Minas Gerais (CODEMIG) conducted an aeromagnetic survey in the study area, named “Projeto Aerogeofísico Area 2”. The project included a magnetometry and gamma-spectrometry survey, with lines spaced 500 meters in a NW-SE direction. This survey revealed a robust aeromagnetic anomaly occurring along a 5 km NS trend associated with the intraoceanic sequence of the Alto Maranhão arc (Moreira et al., 2019). During the field check of the aeromagnetic anomaly, a banded magnetite-quartz-hornblende epigenetic rock was identified, starting thus a mineral exploration program responsible for the discovery and estimating 60Mt@28%Fe +(Cu). The deposit was discovered after investigating a 5km NS trend aeromagnetic anomaly identified in the CODEMIG Airmag survey conducted in 2001 (Fig. 1 d).

This study presents data from field geology, petrography analysis, whole rock geochemistry, EMPA, and LAICP-MS analysis in magnetite to support a genetic proposition for this deposit.

2. Geologic Setting

2.1 Regional Geology

In the Mineiro Belt at the southern border of the Sao Francisco paleocontinent, foreland/active continental margin units, Siderian-to-Rhyacian magmatic and volcanic arcs, volcano-sedimentary sequences, and juvenile-to-crust-contaminated magmatic-arc granitoid rocks, including high Ba– Sr rocks, TTGs, sanukitoids, and hybrid granitoids (Teixeira et al., 2015; Moreira et al., 2018), make up an approximately rectangular-shaped terrane with 200 km X 60 km following a NE-SE trend (Fig. 1a). They represent different sections of microcontinents, juvenile accretionary arcs, and intra-oceanic sequences that were diachronously combined between 2130 and 2060 Ma (Heilbron et al., 2010; Ávila et al., 2014; Alkmim and Teixeira, 2017; Bruno et al., 2020, 2021; Cutts et al., 2020). (Fig. 1b). The intraoceanic sequence is composed of the units: Congonhas-Itaverava (Corrêa Neto et al., 2012), Rio das Mortes (Ávila et al., 2010), Nazareno (Avila et al., 2010), Dores do Campo (Avila et al., 2014), and Lafaiete Formation (Guild, 1957; Pires, 1977).

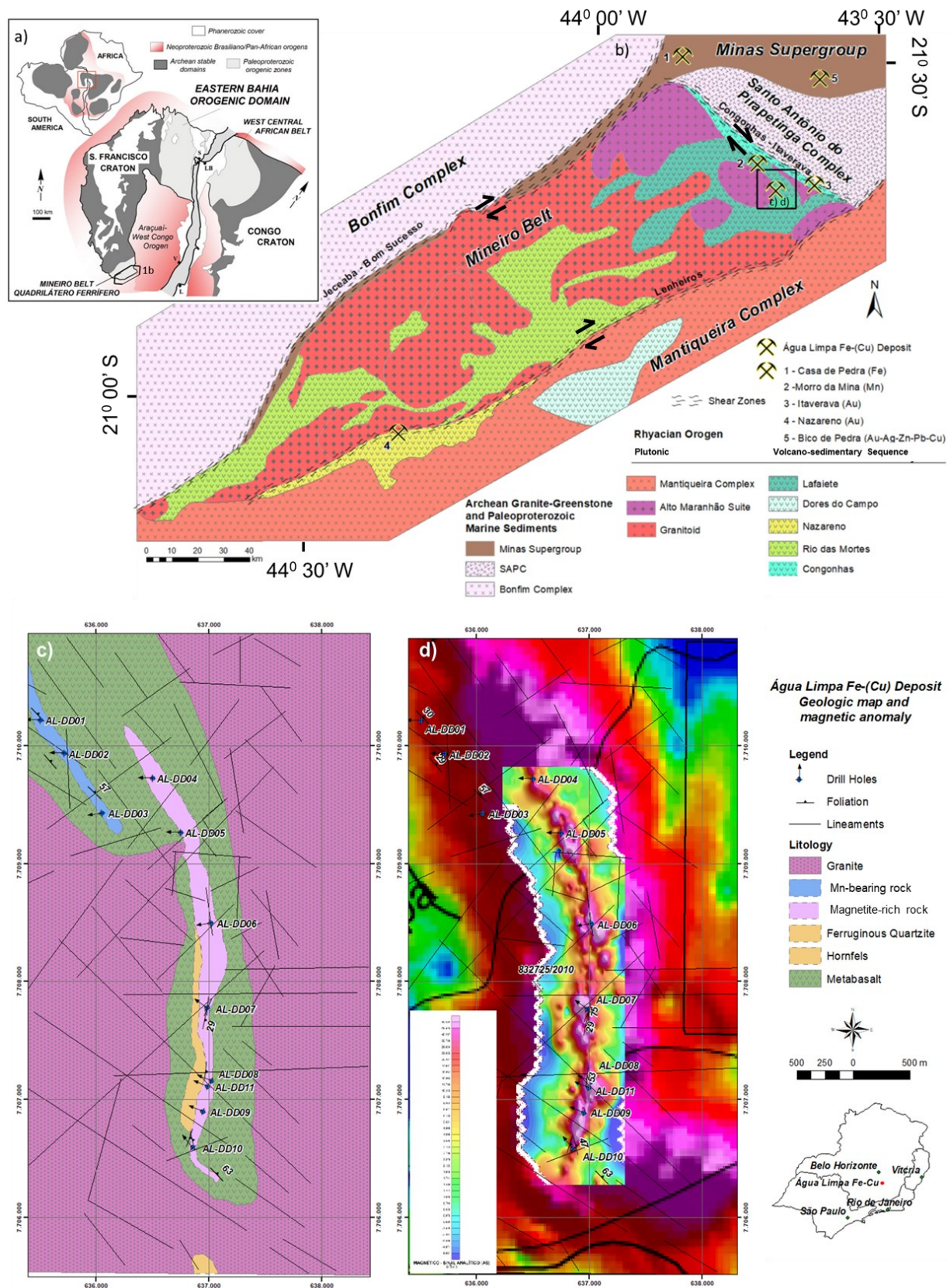


Fig. 1: Regional geological mapping, modified from Barbosa & Barbosa (2017) and Cardoso et al. (2019). (1a) represents the regional context of the São Francisco–

Congo Craton; (1b) is a detail, considering the entire region of the study. c) Geologic map of the Água Limpa Fe-(Cu) deposit; d) Aeromagnetic and ground magnetic detailed signature of the deposit.

2.2 Deposit Geology

The Água Limpa Fe-(Cu) deposit is located in the northern portion of the Mineiro Belt. It is characterized by elongated hornblende-quartz-magnetite±pyrite-chalcopyrite-pyrrhotite lenses up to 30m measured along the contact between metavolcanic rock and the Água Limpa granitic body (Fig. 1c).

The whole-rock geochemistry signature (Fig. 6a) shows that metavolcanic rock comprises metabasalt and metabasalt-andesitic rock. It is foliated and composed of fine bands of oriented hornblende and plagioclase, with apatite and zircon as trace minerals. A magnetic anomaly is not present in this unit.

The Água Limpa pluton is part of the Alto Maranhão Sanukitoid suite. It stands out for its unique composition of plagioclase, lesser K-feldspar, hornblende, biotite, and variable proportions of quartz (Fig. 2a). The fabric of the pluton transitions from a homogeneous massive igneous texture in the central parts of the intrusion to a NE-SW foliated structure parallel to the regional schistosity. Breccia zones in the Água Limpa pluton are enriched in epidote veins, biotite-chlorite-sericite, and pyrite. (Fig. 2c)

The contact zone between the Água Limpa pluton and the metavolcanic rock is characterized by a pervasive hydrothermal alteration in both rocks, forming zones of the massive fine-grained hornfels, silicified foliated metavolcanic rock, with epidote development. Late strings of massive and disseminated pyrite, pyrrhotite, and chalcopyrite, are further evidence of this alteration (Fig. 2b).

The Magnetite-rich rock is marked by lens-shaped strings (5-10 cm thick) of magnetite, quartz, hornblende, and epidote that grade from finely layered to massive layers of up to 30 meters, with an average of ca. 20m (Fig. 2d). These layers are made up of partially oxidized aggregates of granoblastic magnetite (Pires 1977) and minor sulfides in a very fine quartz matrix with hornblende (Fig 3 f). In contact with the

metavolcanic rock, slender to massive-elongated veins of chalcopyrite increase with the dissemination of pyrite and pyrrhotite (Fig. 2e).

Magnetite occurs as euhedral 0.1–10 mm crystals, immersed in a fine-grained matrix of hornblende and quartz, as 1 to 5 mm layers, separated by fine-grained quartz layers (Fig. 2f, g), and also compact massive (Fig. 3 h). Platy-shaped magnetite (mushketovite) is associated with pyrite, chalcopyrite, and pyrrhotite (Fig. 7 a, b, c). The thickness of the magnetite layers varies from 10 m to 35 m, with an average of 20 m (Fig. 3). The weathering surface in the deposit is estimated at 30 m depth.

Three paragenesis phases have been recognized locally within the Água Limpa pluton and between the intrusion and surrounding metabasalt: (1) The first phase is dominated by the hornblende, magnetite, titanite, and epidote assemblage. The second phase is defined by epidote, allanite, carbonate, apatite, chlorite, biotite, and sericite. The third phase is related to hydrothermal alteration, which comprises mushketovite (magnetite platy shape), pyrrhotite, chalcopyrite, and pyrite.



Fig. 2: a) Contact between metavolcanic rock and quartz-diorite; b) Intercalation between quartz-diorite, metavolcanic rock, and hornfels; c) Brecciated granite with veins of pyrite-epidote-chlorite; d) Magnetite-hornblende massive and disseminated

quartz-magnetite in magnetite-rich rock; e) Chalcopyrite (Ccp) vein in contact with hornblende (Hb) -magnetite (Mag); f) Hornblende-magnetite-quartz (Qtz) layered with disseminated traces of fine chalcopyrite; g) Magnetite-hornblende-quartz layered; h) Compact massive magnetite.

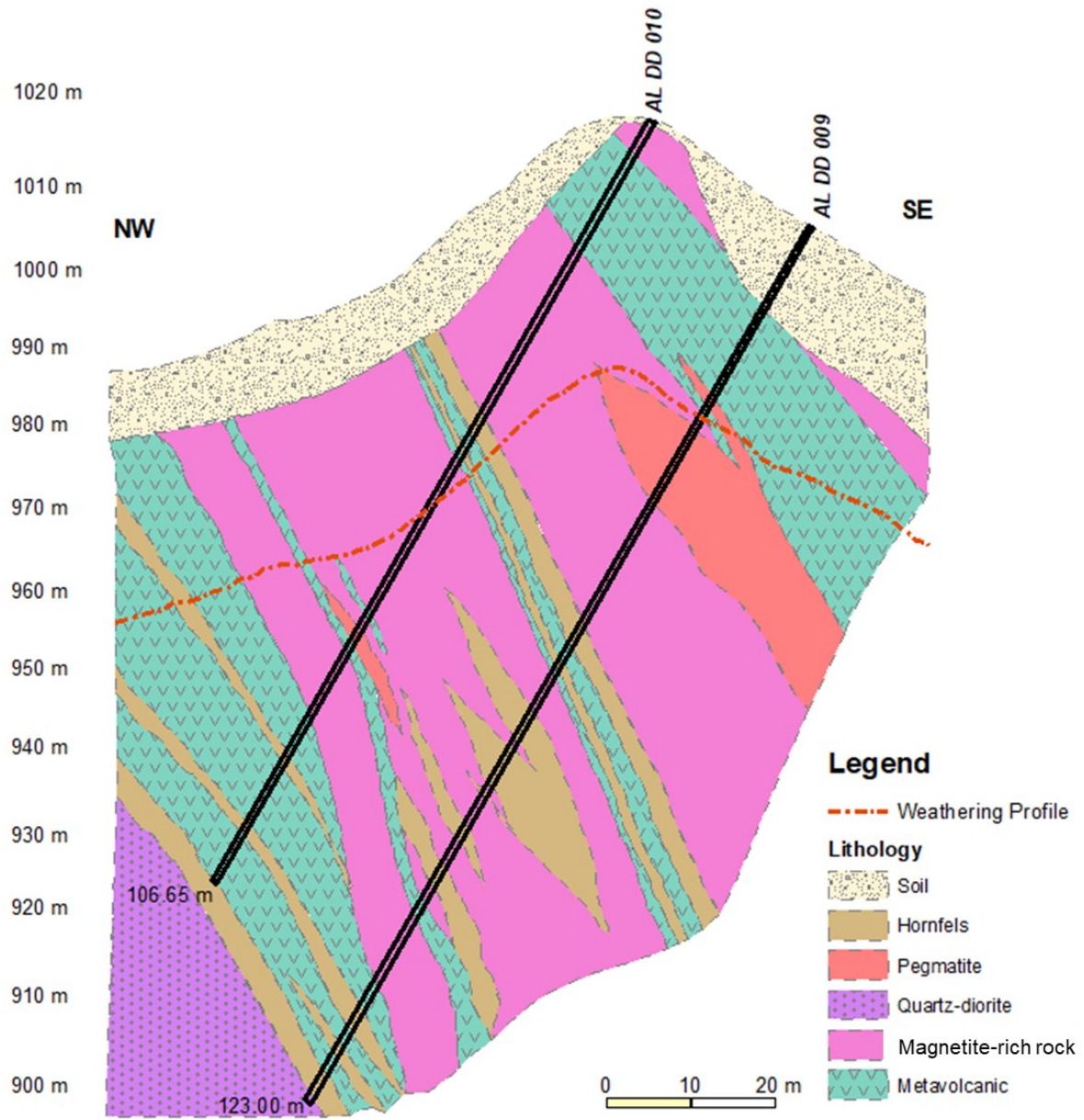


Fig. 3: Geological section in the southern portion of the Água Limpa Fe – (Cu) deposit.

3. Materials and methods

3.1 Petrography

Twenty-two thin and polished samples were collected from 04 drill cores made available from Carijós Mineração, presently implementing a mining project in the area. The samples were selected to show the relation between the intrusive quartz–diorite and Água Limpa pluton in contact with metabasalt.

3.2 Whole rock geochemistry

Geochemical analyses of major, trace, and rare earth elements were conducted at Bureau Veritas Minerals, Vancouver, Canada. The samples were analyzed by XRF for 11 major oxides and ICP-MS for trace and rare earth elements after dissolution with lithium metaborate/tetraborate and digestion with diluted nitric acid, with analytical errors of 5% for most of the major oxides and 10–15% for most of the trace and rare earth elements. Base and precious metal concentrations were determined by digestion in Aqua Regia, followed by ICP-MS analysis. Loss on ignition (LOI) was determined by the weighing difference after ignition at 1.000° C.

Twenty-nine rock chip samples were analyzed: five samples of the Água Limpa pluton (AMS), 11 samples of the metavolcanic rocks (MV), and 13 samples of the magnetite-rich rock (MRR).

3.3 EMPA analysis

Quantitative major elements analysis of magnetite, hornblende, and biotite was carried out at the Universidade Federal de Minas Gerais (UFMG) Microscopy Centre with a Jeol-JXA-8900 RL WD/ED electron microprobe (software EPMA JEOL – version V03.10). The operating conditions were 20 kV and 25 nA. Generally, counting times were 20 seconds on peak and 10 seconds on background for each element. Mg, Al, P, K, Ca, Ti, V, Cr, Mn, Fe, and Ni were analyzed in the magnetite grains. The standards used were MgO, Al₂O₃, Quartzo, A408, Rutilo, V₂O₃, Cr₂O₃, Rodonita,

Magnetite and NiO. A geochemical major elements mapping was performed using the XMapTools software following the instructions of Lanari et al. (2014).

3.4 LAICP-MS analysis

In situ trace element analyses were carried out on the magnetite of two samples: ALDD006 and ALDD009. ALDD006 is in the center of the Fe-(Cu) deposit, and ALDD009 is in the southern portion. The analysis was conducted by laser ablation ICP-MS using a CETAC LSX-213 G2+ laser system coupled with a Quadrupole Analyser ICapQ Thermo at the Geochemistry Lab in the IGC, USP, Brazil. Ablation was performed in a custom-built large-volume cell (Andrade et al., 2014) using high-purity He as the carrier gas, which was subsequently mixed with Ar before introduction into the ICP-MS. The ICP-MS was tuned for maximum sensitivity under robust plasma conditions (Th/U sensitivity ratio ~ 1 ; see Pettke, 2008) and maintained low oxide production rates ($\text{ThO}/\text{Th} < 0.4\%$). For spot analyses, the laser fluence was set to 6 J/cm² at the sample site with a laser repetition rate of 10 Hz and the beam diameter set to 20 μm , respectively. The internal standard isotope used for data reduction was ⁵⁷Fe, based on Fe contents previously measured by WDS at the UFMG Center of Microscopia. NIST SRM 610 was used as a bracketing external standard using the standard reference values of Spandler et al. (2011). We also used BHVO-2G and BCR-2G as secondary standards to evaluate accuracy. Fractionation and mass bias were corrected using standard bracketing techniques, with every fifteen magnetite sample measurements bracketed by BCR-2G measurements. The individual analysis consists of approximately 20 seconds of the signal of the gas background followed by 30 seconds off.

4. Results

4.1 Petrography

Metavolcanic rocks – Metavolcanic rock is characterized by fine to medium grains of hornblende, plagioclase, and quartz, forming a nematoblastic texture. Well-formed euhedral hornblende composes ca. 60 % to 70 % of the minerals. It exhibits a chemical variation from the center to the border of crystals, demonstrated by the dark green in the center to the pallid brownish in the border of crystals. An oxy and hydroxy of iron and biotite substitute the hornblende along the foliation's plan. Plagioclase composes 30% to 40% of the rock and doesn't exhibit twins, making differentiation from the quartz crystals difficult. Both minerals aggregate around the border of the hornblende with polygonal contact. The accessory minerals are apatite, allanite, zircon, and titanite, which occur as inclusion in the hornblende (Fig. 4a).

Hydrothermal alteration is characterized by fine crystals of biotite, epidote, and carbonate substituting the hornblende and forming reaction border products. The epidote also forms as a product of the saussuritization of plagioclase. Carbonate occurs as fine grains in contact with hornblende (Fig. 4b).

Água Limpa pluton - The rocks that compose the Água Limpa pluton have an equigranular medium texture and comprise plagioclase, quartz, and biotite as the main mineral phase. Apatite, titanite, and zircon are the accessory phases (Fig. 4c). The plagioclase exhibits the characteristic twin with local undulose extinction with lobular contact, and it is not well-formed grains. In contact with the metavolcanic rock, there is an intense saussuritization of the plagioclase and the formation of epidote-biotite-sericite-carbonate as its alteration product. It is also possible to see brecciation with epidote in the matrix (Fig. 4d).

Magnetite-rich rock - The Magnetite-rich rock presents a banding of grains of equigranular fine to coarse grain size with a granonematoblastic texture composed of magnetite interlayered with hornblende, epidote and quartz (Fig. 4e). The individual magnetite layers are 1 to 5 mm thick and separated by fine quartz and hornblende layers. Magnetite varies from 100 μm to 10mm, quartz and hornblende exhibit fine grains (e. g. 100 – 500 μm). Hornblende occurs as a granular phase, with a chemical zonation from the center to the border of crystals (Fig. 4f).

Magnetite occurs as (1) granoblastic aggregates of idiomorphic compact crystals with a variable size of 0.1–10 mm (Figs. 5a, b); (2) a porous variety of the

magnetite (1) (Figs. 5a, b); (3) platy-shape musketovite grains associated with pyrite, chalcopyrite, and pyrrhotite (Figs. 5 c, d, e, f).

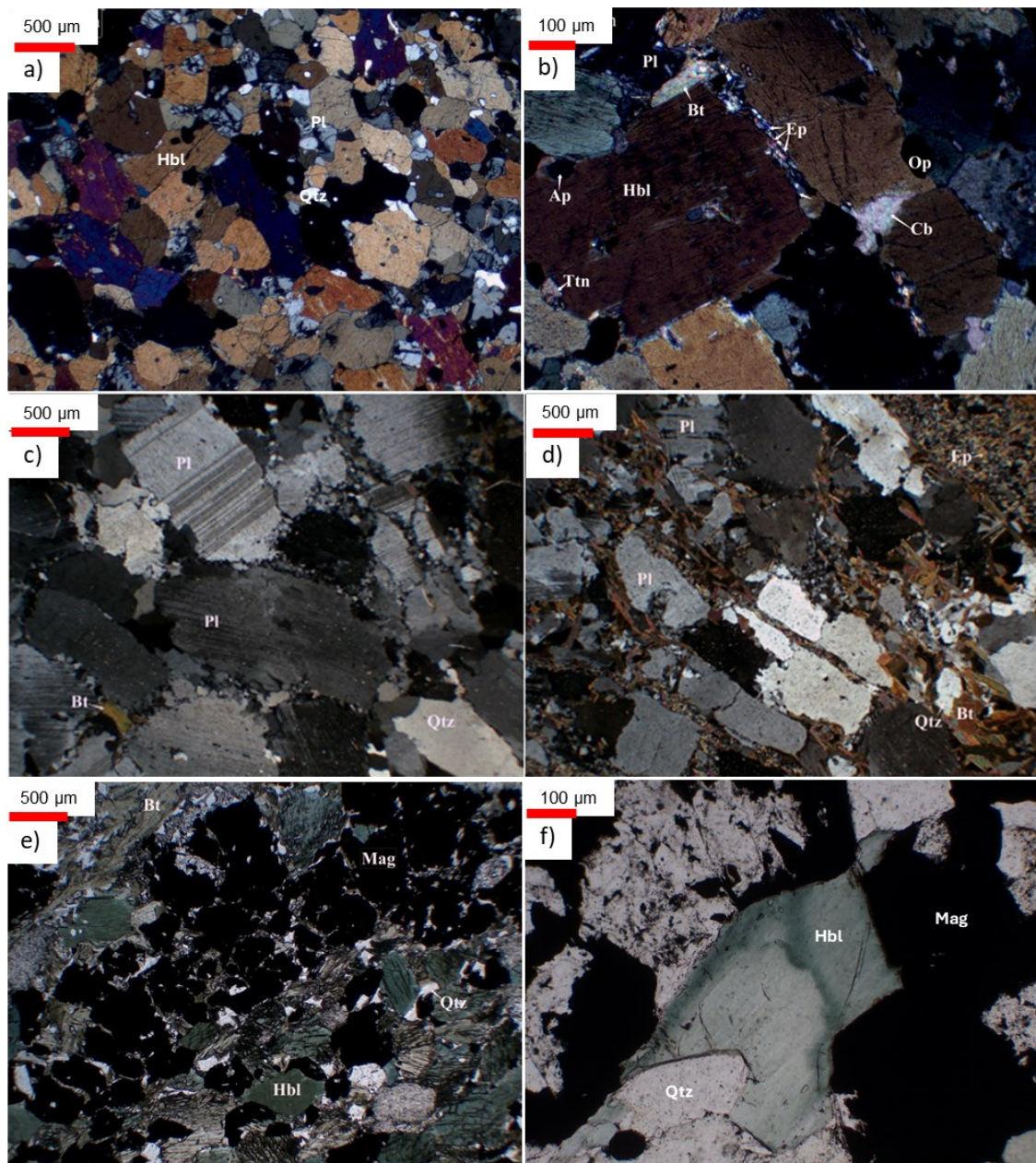


Fig. 4: a) metabasalt composed of hornblende-plagioclase-quartz as the main mineral phase; b) hydrothermal alteration in metavolcanic rock with epidote-carbonate minerals c) quartz-diorite from Água Limpa Pluton; d) metasomatic quartz-diorite brecciation; e) magnetite-hornblende-quartz magnetite-rich rock; f) chemical zonation in hornblende altered with magnetite and quartz in magnetite-rich rock. Mag, Magnetite; Hbl, Hornblende; Qtz, Quartz; Pl, Plagioclase; Ap, Apatite; Bt, Biotite; Chl, Chlorite; Ep, Epidote; Cb, Carbonate; Op, Opaque.

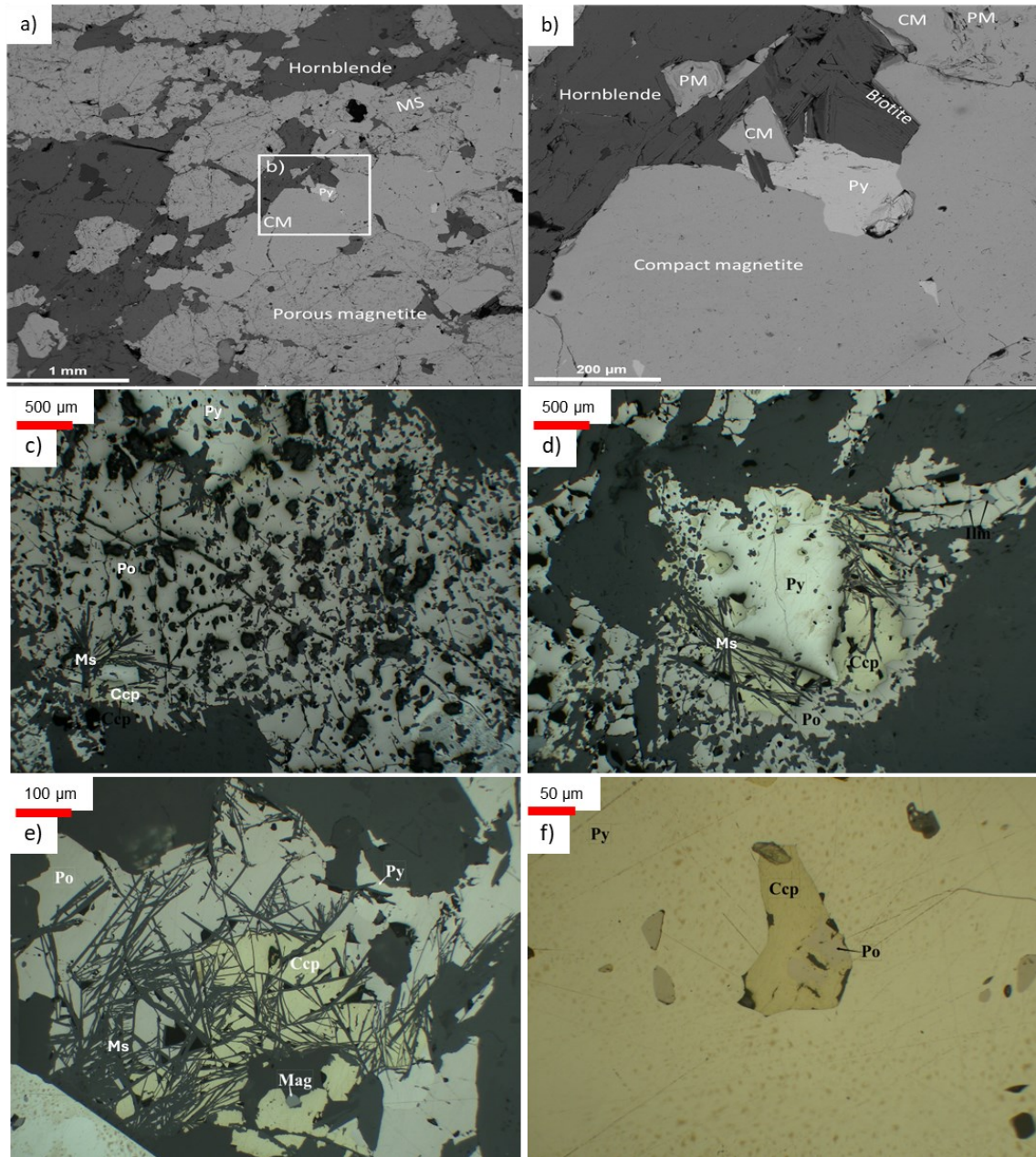


Fig. 5: Electron microscope image showing (a) a magnetite-rich rock composed of three types of magnetite (compact, porous, mushketovite) crystallized with hornblende and pyrite, (b) a zoom showing the contact relationship between the compact and porous magnetite, as well the pyrite and hornblende + biotite minerals; c) Mushketovite, pyrrhotite-pyrite-chalcopyrite; d) pyrite-mushketovite-chalcopyrite-pyrrhotite; e) mushketovite-pyrrhotite-chalcopyrite; f) pyrite-chalcopyrite-pyrrhotite. CM, Compact magnetite; PM, Porous magnetite; Mag, Magnetite; Ms, Mushketovite; Py, Pyrite; Ccp, Chalcopyrite; Po, Pyrrhotite.

The sulfide minerals are associated with the enrichment of magnetite in the magnetite-rich rock (Fig. 7 b, c, d).

4.2 Whole Rock Geochemistry

The whole rock major and trace elements resulting from the rocks that comprise the Água Limpa Fe-(Cu) deposit are demonstrated in Table 2.

Metavolcanic rocks – 11 samples show low SiO₂ content (45.49 – 55.67%), low total alkali (Na₂O + K₂O = 1.02 – 5.47%), and low K₂O (0.26 – 2.23%), characterizing them as basalt to basalt-andesite, with sub-alkaline composition (Fig. 8 a). In the Chondrite Normalized REE diagram (Fig. 9 a), the metavolcanic rocks display a relative LREE enrichment relative to HREE, with (La/Yb)_N and LREE/HREE mass ratios of 2.10 – 87.05 (average value of 20.83) and 1.80 – 30.90 (average value of 8.09), respectively, with positive Eu anomalies (Eu/Eu* = 5.62, average). Primitive-mantle incompatible element patterns (Fig. 9 b) show significant large ion lithophile elements (LILE) enrichment (e.g., Rb, Th, U, and Ba) and high field strength elements depletion (e.g., Nb, K, P, and Ti) (Fig. 9 b). The analysis of Y vs. Sr in the metavolcanic shows a negative trend (Fig. 10).

Água Limpa pluton - The Água Limpa pluton is characterized by an intermediate-to-high silica content (58.53 – 76.63%), medium alkalis (Na₂O + K₂O = 3.32 – 8.30%) and low K₂O (0.96 – 2.11%) (Fig. 8 b). In the chondrite normalized REE diagram (Fig. 9 a), the quartz-diorite displays LREE enrichment relative to HREE, with (La/Yb)_N and LREE/HREE mass ratios of 2.10 – 87.05 (average value of 20.83) and 1.80 – 30.90 (average value of 8.09), respectively, without negative Eu anomalies (Eu/Eu* = 5.62, average). Primitive-mantle incompatible element patterns (Fig. 9 b) show significant large ion lithophile elements (LILE) enrichment (e.g., Rb, Th, U, and Ba) with high field strength elements depletion (e.g., Nb, K, P, and Ti) (Fig. 6 b). A depletion in the negative trend between the Y vs. Sr follows the metavolcanic analysis (Fig. 10).

Magnetite-rich rock – The Magnetite-rich rock is composed of high values of Fe₂O₃ (34.80 – 62.64%). The SiO₂ content varies from high to low (55.46 – 21.69%).

The alkalis are very low (average $\text{Na}_2\text{O} + \text{K}_2\text{O} = 0.18\%$). In the chondrite normalized REE diagram (Fig. 9 a), the magnetite-rich rock displays LREE enrichment relative to HREE, with $(\text{La/Yb})_N$ and LREE/HREE mass ratios of 1.21 – 39.09 (average value of 10.90) and 4.18 – 11.42 (average value of 5.67), respectively. The Eu has both positive and negative values. Different from the metavolcanic rocks and the Água Limpa pluton, which shows the exact characteristics of the Primitive-mantle incompatible element patterns, to the magnetite-rich rock, the Primitive-mantle incompatible elements patterns show high field strength elements depletion (e.g., Rb, Nb, K, P, Zr, and Ti) with enrichment in Ba and Ta (large ion lithophile elements - LILE) (Fig. 9 b). The Y vs. Sr values in the magnetite-rich rock are depleted compared to the metavolcanic and Água Limpa pluton (Fig. 10).

	MV	MV	MV	MV	MV	MV	MV	MV	MV	MN	MV		MV	AMS	AMS	AMS	AMS	AMS		AMS	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR	MRR		
													Av ± Stdev					Av ± Stdev					Av ± Stdev														
n	11												5					13																			
SiO ₂ (wt%)	55.67	54.56	54.02	52.32	51.68	46.31	48.23	47.76	45.49	45.93	53.94		50.54	3.67	65.62	76.63	60.56	61.87	58.53		64.64	6.42	55.46	47.24	45.20	47.11	45.97	43.45	43.93	41.48	42.84	42.09	38.35	37.48	21.69	42.48	7.39
TiO ₂	0.83	0.77	0.93	0.96	0.96	1.66	1.75	2.80	1.47	2.82	2.01		1.54	0.72	0.23	0.11	0.26	0.27	0.26		0.23	0.06	0.03	0.14	0.15	0.11	0.21	0.05	<0.01	0.03	<0.01	0.02	0.32	0.08	0.93	0.16	0.25
Al ₂ O ₃	14.80	14.74	14.22	14.10	14.12	15.35	14.15	12.30	12.80	11.87	8.97		13.40	17.4	14.62	13.77	16.82	19.13	20.82		17.03	2.66	0.72	1.34	1.92	0.41	0.96	0.64	0.13	0.67	0.27	0.32	1.42	0.73	3.97	1.04	0.98
Fe ₂ O ₃	9.38	10.04	11.51	12.39	13.02	16.48	17.68	20.11	21.20	22.00	24.78		16.24	5.05	6.16	1.3	7.77	4.28	4.16		4.73	1.27	34.80	48.10	50.20	50.81	52.27	54.22	54.42	56.16	57.04	57.54	57.83	58.94	62.64	53.46	6.61
MnO	0.12	0.12	0.12	0.13	0.13	0.19	0.22	0.25	0.20	0.26	0.46		0.20	0.10	0.06	0.01	0.06	0.03	0.05		0.04	0.02	0.26	0.07	0.07	0.06	0.07	0.06	0.05	0.07	0.04	0.10	0.07	0.08	0.13	0.09	0.05
MgO	6.74	6.38	6.51	6.74	6.96	6.06	5.24	4.32	6.20	4.54	0.81		5.41	1.67	1.16	0.17	1.15	1.36	1.52		1.07	0.47	5.83	1.81	1.51	1.46	0.82	1.46	1.29	1.41	0.05	0.43	1.51	2.05	4.26	1.84	1.49
CaO	5.79	6.48	5.78	6.58	6.49	8.88	7.76	8.52	7.95	8.82	7.00		7.37	1.02	6.92	0.92	6.15	4.92	5.02		4.79	2.07	3.12	1.80	1.90	1.22	0.64	1.20	1.37	1.21	0.05	0.34	1.83	1.93	5.51	1.70	1.34
Na ₂ O	3.94	3.94	3.01	3.06	2.85	2.53	1.05	2.38	1.33	2.12	0.74		2.45	0.12	1.21	4.92	4.48	5.83	7.01		4.69	1.94	0.07	0.14	0.28	0.03	0.04	0.05	0.02	0.05	<0.01	<0.01	<0.01	0.08	0.39	0.10	0.11
K ₂ O	1.24	1.53	2.23	1.96	2.10	0.33	0.26	0.70	1.84	0.71	0.28		1.20	0.74	2.11	1.14	0.96	1.15	1.29		1.33	0.40	0.05	0.07	0.14	<0.01	0.20	<0.01	<0.01	<0.01	<0.01	<0.01	0.06	<0.01	0.57	0.08	0.18
P ₂ O ₅	0.11	0.10	0.11	0.12	0.12	0.12	0.14	0.17	0.24	0.23	0.70		0.20	0.17	0.09	0.02	0.06	0.13	0.12		0.08	0.04	0.04	0.05	0.15	0.04	0.05	0.04	0.11	0.02	0.06	0.03	0.16	0.17	0.51	0.11	0.13
LOI	1.00	1.00	1.20	1.30	1.20	1.80	3.20	0.40	0.90	0.40	0.20		1.15	0.79	1.40	0.00	1.50	0.70	1.00		1.08	0.32	-0.50	-0.80	-1.60	-1.30	-1.30	-1.20	-1.40	-1.10	-0.40	-0.90	-1.70	-1.60	-0.80	-1.12	0.40
Total	99.62	99.66	99.64	99.66	99.63	99.65	99.68	99.71	99.62	99.7	99.89		99.68	0.08	99.58	99.79	99.77	99.67	99.78		99.72	0.08	99.88	99.96	99.92	99.95	99.93	99.97	99.92	100.00	99.95	99.97	99.99	99.94	99.80	99.94	0.07
Rb(ppm)	33.30	39.20	65.40	54.00	55.60	2.80	2.50	5.3	34.80	5.00	3.4		27.39	23.29	44.10	43.70	21.20	27.30	32.00		33.66	9.04	<0.1	0.80	2.20	0.30	4.50	0.30	<0.1	0.10	0.50	<0.1	0.40	<0.1	10.20	1.38	3.14
Ba	455.00	431.00	686.00	590.00	599.00	103.00	198.00	181	780.00	158.00	138		392.64	235.39	1222.00	1376.00	808.00	873.00	675.00		900.80	264.14	32.00	9.00	36.00	28.00	151.00	7.00	6.00	1.00	4.00	5.00	13.00	4.00	409.00	50.36	109.31
Sr	489.30	505.70	474.30	463.70	484.10	152.60	30.70	228.8	157.40	195.40	64.5		295.14	179.65	1679.20	233.00	1043.70	1066.20	990.40		1002.50	459.56	8.40	7.70	23.00	7.90	4.00	19.70	15.90	2.10	2.70	9.00	6.50	156.20	19.16	39.65	
Th	26.90	19.30	3.30	3.40	3.50	0.80	1.00	2.4	5.70	1.40	0.7		6.22	8.25	6.40	20.30	1.30	14.60	4.00		9.32	0.70	0.50	<0.2	5.60	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	2.00	0.70	2.20	0.79	1.83	
Nb	3.80	3.10	4.50	4.70	4.70	5.20	5.80	9.8	9.70	9.80	9.2		6.39	2.54	1.70	5.60	3.00	2.30	2.50		3.02	1.36	0.40	0.80	0.20	0.30	1.60	0.50	0.30	0.50	0.30	0.40	1.70	0.50	6.60	1.15	1.66
Zr	197.00	138.70	79.70	63.70	68.50	100.40	123.60	145	151.60	169.80	146		125.80	0.73	92.80	92.20	74.80	117.90	112.30		98.00	15.49	9.00	5.10	13.40	3.10	5.10	2.50	1.60	1.90	3.20	5.00	14.90	5.20	11.70	5.84	4.32
Hf	4.40	3.50	2.40	2.00	1.80	2.70	3.40	3.9	4.40	4.80	4.1		3.40	0.99	2.40	3.00	1.80	2.80	2.70		2.54	0.42	0.20	0.20	0.40	<0.1	0.20	<0.1	<0.1	<0.1	<0.1	0.40	0.20	0.60	0.16	0.15	
Y	13.60	11.70	15.00	14.30	15.40	35.30	195.90	45	44.30	45.40	55		44.58	50.25	5.00	6.80	5.00	3.80	7.60		5.64	1.37	7.60	4.70	8.30	3.40	2.50	2.40	5.70	1.70	5.80	5.20	6.30	4.70	27.20	6.11	6.25
V	123.00	123.00	128.00	134.00	128.00	610.00	369.00	592	280.00	600.00	<8		280.64	206.19	49.00	15.00	54.00	38.00	33.00		37.80	13.64	<8	63.00	53.00	36.00	48.00	45.00	<8	43.00	<8	71.00	41.00	15.00	29.64	15.14	
La	97.50	68.50	10.70	13.40	10.90	13.00	37.20	17.3	22.70	16.60	11.6		29.12	27.11	21.20	33.90	14.30	34.00	30.70		26.82	7.81	4.00	1.20	25.80	0.40	0.50	0.40	2.50	0.50	6.50	9.60	10.50	23.00	51.50	8.26	14.04
Ce	152.40	110.80	22.90	25.70	23.90	18.20	18.60	38	44.40	39.10	27.5		47.41	41.46	35.80	62.10	24.70	59.80	52.30		46.94	14.44	6.70	3.40	41.30	1.30	1.40	1.00	2.70	1.00	5.30	7.50	17.20	44.40	101.40	13.90	27.09
Pr	13.85	10.48	2.82	2.99	2.90	3.76	7.42	4.76	5.23	5.11	4.61		5.81	3.33	3.82	5.28	5.78	6.39	5.59		4.79	12.46	0.73	0.43	3.89	0.17	0.18	0.11	0.31	0.10	1.38	1.81	1.77	0.56	11.79	1.66	30.7
Nd	43.10	32.10	11.90	11.80	12.30	17.00	35.70	21.6	21.20	22.20	24.6		23.05	9.80	13.60	16.30	11.30	22.50	21.00		16.94	4.26	3.20	2.10	12.40	1.10	1.00	0.40	1.40	0.50	5.70	7.10	6.40	2.40	45.60	6.38	11.66
Sm	5.06	4.20	2.67	2.73	2.68	4.35	8.64	5.96	5.51	5.81	7.5		5.01	1.87	2.03	2.19	1.90	3.37	3.58		2.61	0.71	0.65	0.55	1.93	0.12	0.27	0.08	0.33	0.10	1.15	1.27	1.20	0.64	8.26	1.18	2.09
Eu	1.13	0.99	0.87	0.89	0.93	1.43	3.11	1.96	1.47	1.99	2.2		1.54	0.67	0.60	0.40	0.72	0.71	1.01		0.69	0.20	0.45	0.17	0.31	0.10	0.29	0.03	0.27	0.10	0.43	0.40	0.32	0.16	1.58	0.33	0.38
Gd	3.83	3.27	2.97	2.92	3.11	5.37	16.31	7.32	6.65	7.32	9.45		6.23	3.83	1.47	1.68	1.53	2.04	2.62		1.87	0.42	0.84	0.62	1.82	0.37	0.29	0.23	0.47	0.21	1.16	1.09	1.10	0.69	6.95	1.13	1.71
Tb	0.48	0.43	0.43	0.46	0.49	0.91	2.50	1.28	1.14	1.23	1.58		0.99	0.62	0.18	0.21	0.19	0.23	0.32		0.23	0.05	0.13	0.09	0.25	0.05	0.04	0.03	0.08	0.03	0.18	0.15	0.15	0.10	0.93	0.16	0.23
Dy	2.51	2.28	2.77	2.78	2.77	5.99	16.72	8.06	7.45	7.52	9.83		6.24	4.21	0.98	1.26	0.94	0.96	1.52		1.13	0.23	0.94	0.69	1.29	0.33	0.27	0.24	0.57	0.22	1.29	0.93	0.96	0.60	4.94	0.95	1.19
Ho	0.47	0.42	0.53	0.51	0.54	1.33	4.04	1.75	1.70	1.71	2.09		1.37	1.04	0.19	0.23	0.17	0.14	0.27		0.20	0.05	0.21	0.13	0.29	0.09	0.08	0.07	0.13	0.05	0.24	0.17	0.19	0.15	0.95	0.20	0.22
Er	1.37	1.32	1.50	1.45	1.73	3.98	11.83	5.16	5.10	5.08	6.14		4.06	3.04	0.51	0.90	0.47	0.36	0.75		0.60	0.20	0.72	0.45	1.01	0.30	0.23	0.25	0.49	0.15	0.80	0.54	0.55	0.41	2.63	0.61	0.62
Tm	0.19	0.16	0.20	0.20	0.21	0.57	1.40	0.7	0.73	0.74	0.88		0.54	0.38	0.08	0.12	0.06	0.05	0.11		0.08	0.03	0.10	0.06	0.11	0.03	0.03	0.02	0.07	0.02	0.11	0.08	0.07	0.06	0.36	0.08	0.08
Yb	1.12	1.03	1.28	1.25	1.50	3.65	7.56	4.48	0.73	4.84	5.53		3.00	2.22	0.45	0.94	0.37	0.29	0.66		0.54	0.23	0.71	0.39	0.66	0.33	0.18	0.15	0.51	0.15	0.75	0.49	0.49	0.40	2.18	0.53	0.50
Lu	0.16	0.17	0.19	0.19	0.23	0.56	1.18	0.67	0.63	0.73	0.85		0.51	0.33	0.08	0.16	0.06	0.04	0.09		0.09	0.04	0.11	0.06	0.09	0.05	0.02	0.02	0.08	0.20	0.12	0.08	0.08	0.07	0.33	0.09	0

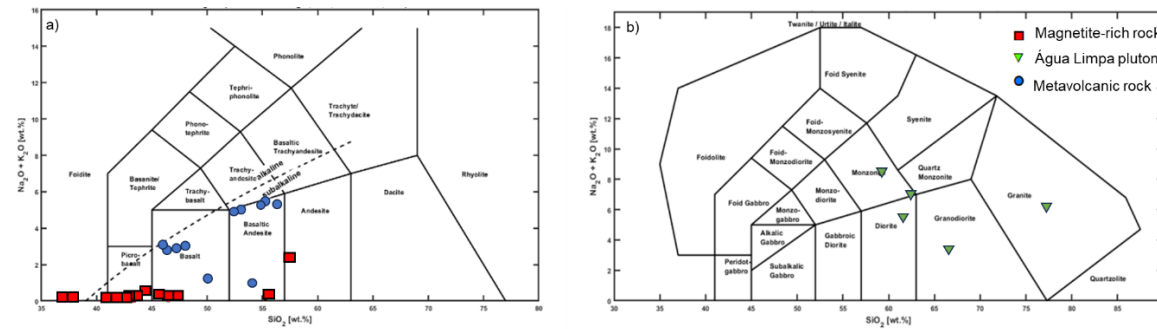


Fig. 6: a) TAS diagram for the metavolcanic rock and themagnetite-rich rock (after Irvine and Baragar, 1971; Le Bas et al., 1986); b) TAS diagram for Alto Maranhão Sanukitoid Suite (after Middlemost, 1994).

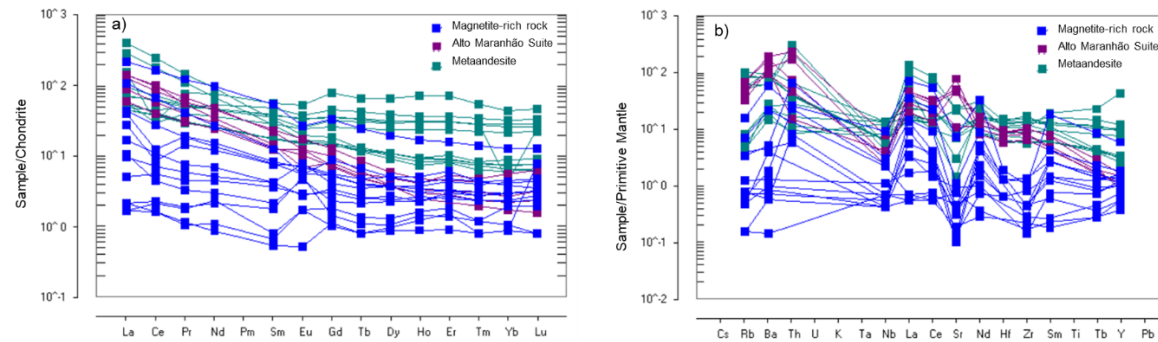


Fig. 7: REE normalized to NMORB, and trace elements normalized to Chondrite modified from Sun and McDonough (1989) from metavolcanic, magnetite-rich, and intrusive quartz–diorite.

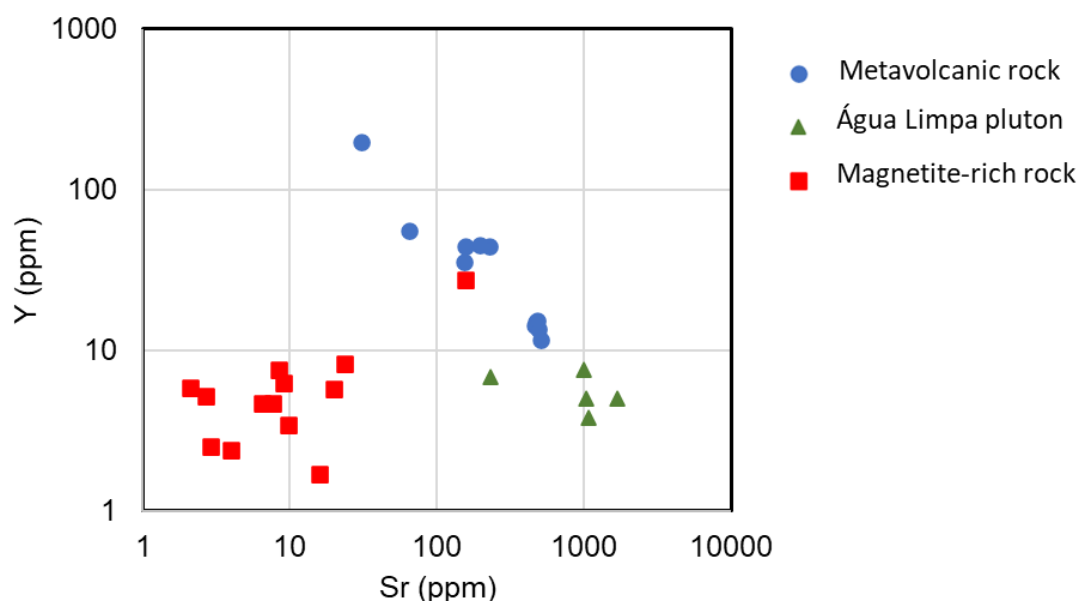


Fig. 8: Y vs. Sr analysis for the rocks that compose the Água Limpa Fe-(Cu) deposit.

4.3 Mineral chemistry

Magnetite forms under various conditions, crystallizing at high temperatures from silicate and sulfide melts or precipitating at lower temperatures from hydrothermal fluids (Dare et al., 2014). With the development of advanced analytical techniques, it is possible to discriminate a broader range of trace elements at lower concentrations that can be used as petrogenetic and provenance indicators in mineral exploration (Dupuis and Beaudoin 2011; Nadoll et al. 2014). Hydrothermal and igneous magnetite can provide vectors to mineralized areas where the primary mineralogical context is lost, deeply covered, or highly altered (Nadoll et al. 2014 and references therein). In this study, we use in situ EMPA and LAICP-MS analysis on magnetite to discuss the origin of the fluids responsible for the mineralization in the Água Limpa Fe-(Cu) skarn deposit.

4.3.1 EMPA analysis

EPMA carried out the major element analysis of magnetite from the Água Limpa deposit, and the results are listed in Table 02. The MgO, P₂O₅, K₂O, V₂O₅, Cr₂O₃, and NiO concentrations are very low, close to the detection limits. Results for Al₂O₃, SiO₂, TiO₂, and MnO are around 0.1 wt%. Consistent with the LAICP-MS results, the MgO, Al₂O₃, V₂O₅, TiO₂, and MnO concentrations are like those of magmatic magnetite (Dupuis and Beaudoin, 2011). The two stages of magnetite present different concentrations of Fe₂O₃: one close to 100% and another of ~60% (Table 2).

Analysis (%)	1	2	3	4	5	6	7	8	9	10	21	22	23	24	25	26	27	28	29	30
MgO	0.04	0.02	0.02	0.00	0.04	0.00	0.04	0.06	0.00	0.00	0.02	0.02	0.00	0.00	0.09	0.00	0.02	0.00	0.04	0.04
Al ₂ O ₃	0.10	0.16	0.12	0.15	0.13	0.09	0.13	0.13	0.08	0.13	0.06	0.02	0.09	0.05	0.06	0.06	0.10	0.13	0.06	0.06
SiO ₂	0.00	0.00	0.09	0.06	0.02	0.16	0.09	0.06	0.09	0.05	0.00	0.02	0.00	0.08	0.05	0.06	0.00	0.00	0.04	0.34
P ₂ O ₅	0.02	0.02	0.00	0.00	0.02	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.04	0.01	0.00	0.01	0.02	0.00	0.00	0.04
K ₂ O	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
CaO	0.00	0.00	0.02	0.00	0.00	0.01	0.04	0.02	0.01	0.01	0.01	0.00	0.01	0.01	0.01	0.05	0.02	0.01	0.00	0.04
TiO ₂	0.13	0.10	0.12	0.11	0.13	0.14	0.11	0.12	0.12	0.12	0.04	0.04	0.03	0.03	0.02	0.05	0.05	0.05	0.05	0.05
V ₂ O ₅	0.03	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.02
Cr ₂ O ₃	0.01	0.01	0.02	0.02	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.04	0.02	0.02	0.00	0.00	0.02	0.00	0.01
MnO	0.07	0.07	0.07	0.04	0.10	0.07	0.05	0.07	0.06	0.12	0.10	0.13	0.08	0.10	0.06	0.06	0.11	0.05	0.08	0.12
Fe ₂ O ₃	101.13	101.41	101.04	101.94	101.48	101.75	100.77	100.17	100.89	100.66	101.01	99.26	100.31	102.36	101.12	101.91	99.94	101.35	100.37	98.75
NiO	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.02	0.05	0.02	0.00
Total	101.54	101.80	101.52	102.32	101.92	102.23	101.23	100.65	101.25	101.10	101.28	99.49	100.60	102.67	101.42	102.19	100.28	101.65	100.68	99.47

Analysis (%)	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60
MgO	0.04	0.07	0.00	0.04	0.06	0.01	0.07	0.02	0.01	0.00	0.01	0.00	0.00	0.01	0.04	0.03	0.00	0.00	0.05	0.03
Al ₂ O ₃	0.04	0.11	0.00	0.04	0.02	0.09	0.08	0.04	0.08	0.03	0.00	0.01	0.00	0.02	0.04	0.03	0.00	0.08	0.10	0.05
SiO ₂	0.18	0.33	0.25	0.05	0.06	0.13	0.18	0.17	0.08	0.13	0.05	0.12	0.08	0.04	0.11	0.16	0.04	0.44	0.21	0.14
P ₂ O ₅	0.00	0.00	0.01	0.00	0.01	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.04	0.01	0.06
K ₂ O	0.00	0.10	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.02	0.05	0.09	0.14	0.07	0.07	0.12	0.08	0.44	0.02	0.02
CaO	0.00	0.02	0.00	0.03	0.00	0.01	0.00	0.02	0.00	0.01	0.00	0.01	0.00	0.01	0.00	0.01	0.01	0.03	0.03	0.02
TiO ₂	0.01	0.06	0.04	0.03	0.05	0.04	0.02	0.05	0.04	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.05	0.06
V ₂ O ₅	0.01	0.03	0.01	0.01	0.00	0.04	0.02	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.01
Cr ₂ O ₃	0.00	0.00	0.01	0.00	0.00	0.02	0.02	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.03	0.00	0.01	0.01	0.02	0.02
MnO	0.11	0.11	0.07	0.05	0.15	0.10	0.08	0.08	0.09	0.09	0.03	0.01	0.03	0.00	0.00	0.02	0.02	0.03	0.09	0.08
Fe ₂ O ₃	101.80	101.21	101.01	101.32	101.83	101.49	100.44	101.40	101.70	101.09	64.33	64.06	63.51	63.92	63.74	64.08	63.89	62.71	99.82	101.60
NiO	0.00	0.03	0.01	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.01	0.02	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.01
Total	102.19	102.06	101.42	101.56	102.17	101.95	100.91	101.80	102.00	101.39	64.48	64.32	63.77	64.12	64.04	64.44	64.08	63.78	100.42	102.11

Table 2: EMPA for the magnetite analyzed.

The chemistry mapping of the magnetite-rich rock was conducted in the XMapTools software (Fig. 9), which provides a detailed analysis of the EMPA results. The concentrations of each element analyzed in the magnetite, hornblende, and biotite crystals are visible. It is possible to see the alteration in Mag 2 exhibiting a core enrichment in iron while the rim is depleted. The platy-shaped magnetite displays fractures with iron depletion. The hornblende is depleted in Fe and Al, while it's enriched in Ca, with medium concentrations of Si and Mg. The biotite is enriched in Al, K, and Mg, with a medium concentration of Si, and depleted in Fe, Ca, and Mn.

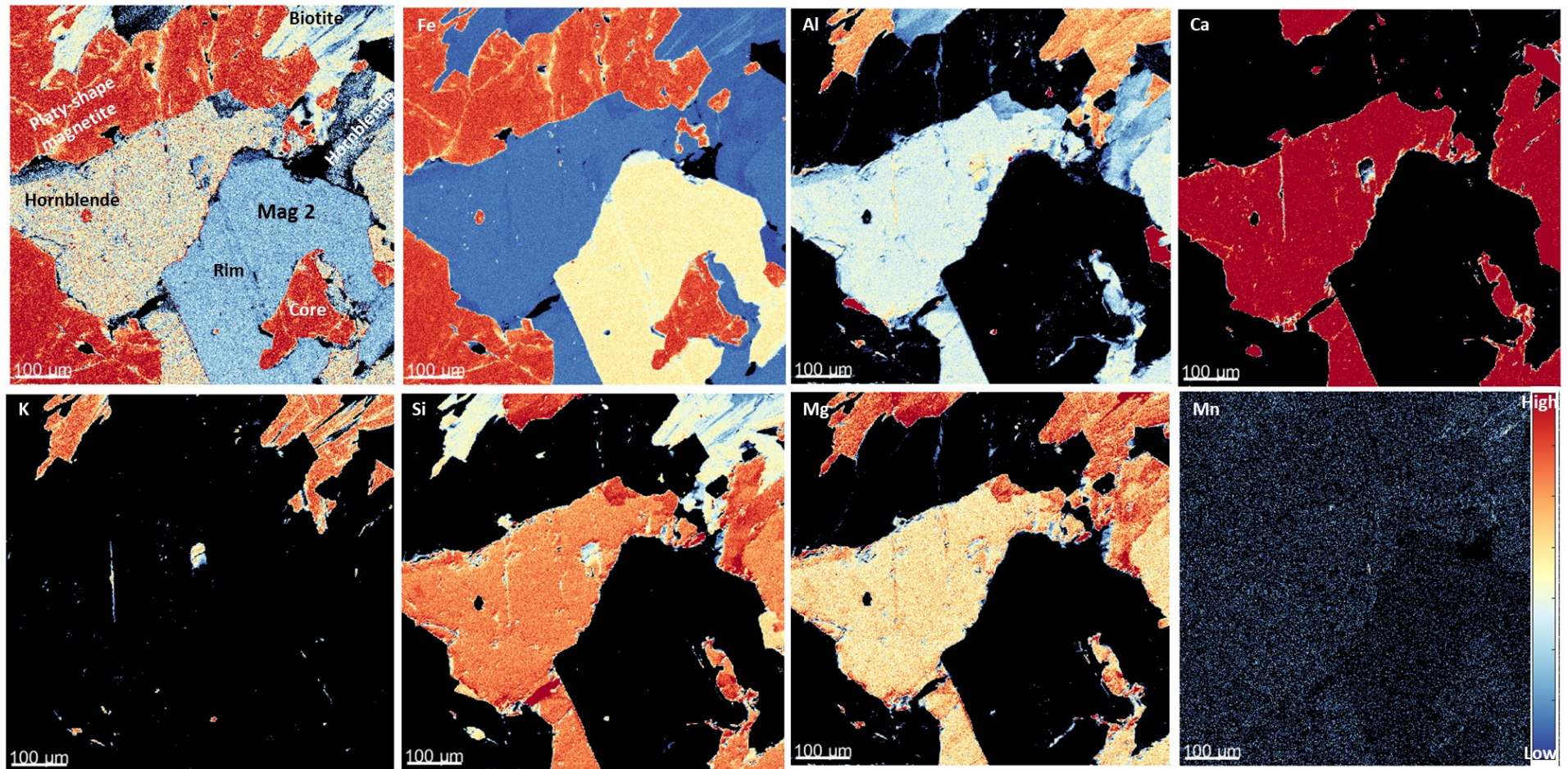


Fig. 9: EMPA mapping of the magnetite-rich rock, showing magnetite, hornblende, and biotite alteration. The elements selected to analyze were Fe, Al, Ca, K, Si, Mg, and Mn.

4.3.2 LA-ICP- MS analysis

Sixty spots were measured in each sample analyzed. Significantly higher amounts of Si, Ca, Mg, Al, Ti, Co, and Zn were found in ALDD006 than in ALDD009, and the opposite trend for V and Mn are reported in Table 4. Chromium and V, however, are consistently low for both magmatic and hydrothermal origin. Ni and Sn are below the detection limit. The average concentrations of Ti (247.54 ppm), V (62.93 ppm), Al (620.89 ppm), Mn (617.51 ppm), and Ga (~20 ppm) are lower than the average concentrations of hydrothermal magnetite (Nadoll et al., 2014) in the ALDD009. In the ALDD006, the average concentrations of Ti (1566.44 ppm), V (24.22 ppm), Al (945.79 ppm), Mn (484.26 ppm), and Ga (~20 ppm) are like magmatic magnetite (Dupuis and Beaudoin, 2011). Yttrium, Zr, and Mo show low-value concentrations (Fig. 12).

Table 3 presents the average, maximum, and minimum concentrations of selected elements in parts per million (ppm) of granular magnetite from samples ALDD006 and ALDD009 of the Água Limpa Fe-(Cu) deposit. These data provide a comprehensive overview of the elemental composition of the magnetite samples.

Elements		Si	Ca	Mg	Al	Sc	Ti	V	Cr	Mn	Co	Ni	Zn	Ga	Sn	Y	Zr	Mo
Number of spots analyzed																		
ALDD006 (60)	Average	2998.98	625.67	245.97	945.79	2.89	1566.44	24.22	3.77	484.26	9.27	<d.l.	132.45	19.11	<d.l.	0.30	0.44	0.77
	Maximum	17334.39	3008.94	1041.00	2279.57	5.34	6283.24	37.25	6.46	918.24	10.63	<d.l.	233.69	22.14	<d.l.	0.34	0.68	1.19
	Minimum	986.81	252.80	98.29	503.26	1.51	638.84	9.58	2.00	244.57	5.28	<d.l.	75.53	11.22	<d.l.	0.22	0.14	0.32
ALDD009 (60)	Average	1047.59	281.81	119.50	620.89	2.71	247.54	62.93	2.54	617.51	0.64	<d.l.	88.13	18.78	<d.l.	0.49	0.24	1.10
	Maximum	1883.85	579.36	352.76	1282.70	4.83	400.29	88.23	18.56	679.53	1.79	<d.l.	130.75	20.65	<d.l.	1.90	0.51	2.10
	Minimum	283.76	91.92	55.52	222.49	0.81	139.92	18.45	0.60	369.00	0.27	<d.l.	50.50	10.34	<d.l.	0.13	0.15	0.40

d.l.: detection limits

Table 3: LA-ICP-MS analysis in magnetite from samples ALDD006 and ALDD009 of the Água Limpa Fe-(Cu) deposit.

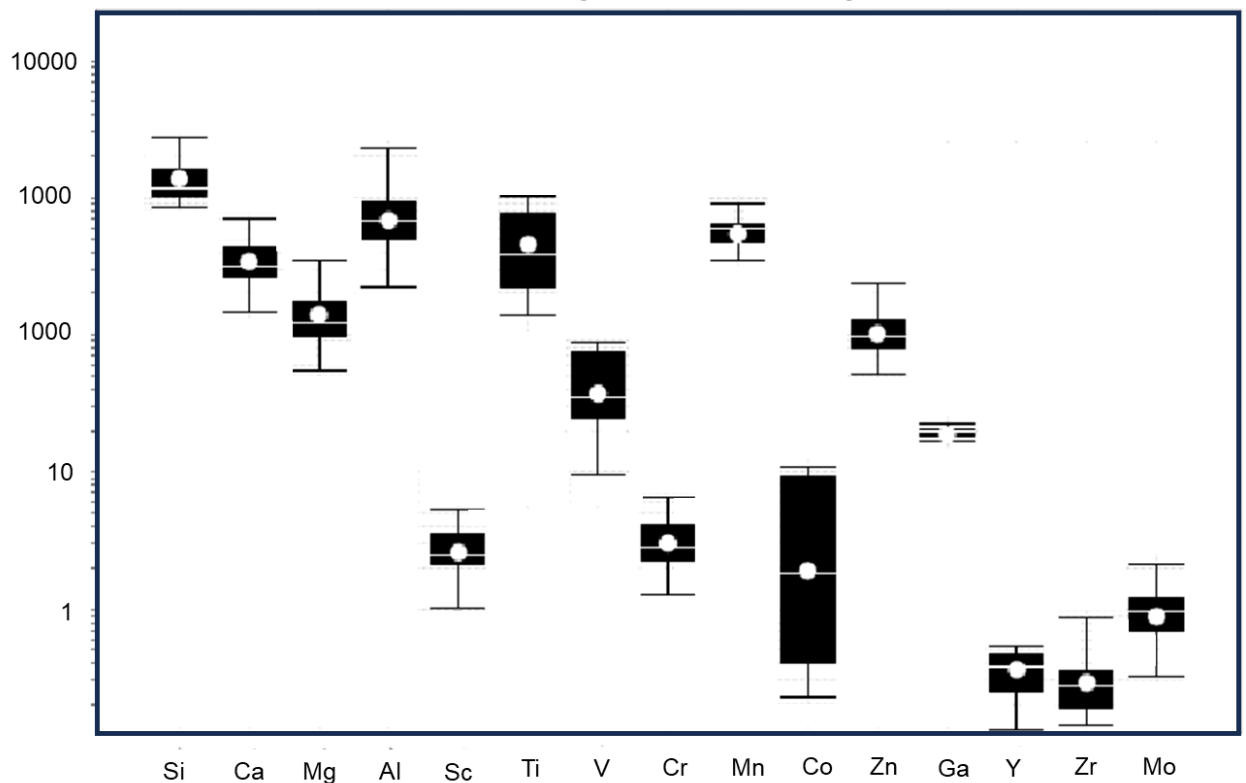


Fig. 10: A box and whisker plot visually represent the magnetite grain's LAICP-MS data. The boxes outline the 25th to 75th percentiles, visually representing the data's distribution. The whiskers extend to the minimum and maximum values, indicating the range of trace element contents. The short white line in the box represents the median value, while the filled white circle represents the average value. It's important to note that trace element contents below the detection limit were removed from the plot to ensure clarity.

5. Discussion

5.1 Geology of the Água Limpa Fe-(Cu) Deposit

Analyzing the contact relation between the host rock of the mineralization and the intrusive quartz-diorite, we can observe that the Água Limpa Fe-(Cu) deposit can be related to a metasomatic formation process. The mineralization is in contact between a metavolcanic mafic rock and an intrusive quartz-diorite. The metabasalt is characterized by fine to medium grain size of hornblende and plagioclase with a

pervasive calcic hydrothermal alteration overprinted, forming an epidote-carbonate rim over the hornblende. The intrusive Água Limpa Pluton is related to the Alto Maranhão Sanukitoid Suite. The Água Limpa Pluton has the magma composition of high-Mg andesites in modern subduction zones; interpretations imply that this suite originated from partial melting from a mantle wedge, which was metasomatized to different degrees by slab-derived melts (Seixas et al., 2013). Some portions of the intrusive pluton are brecciated with epidote-carbonate minerals, altering the plagioclase's paragenesis. A point that attests to the hydrothermal influence of the crystallization of the Água Limpa Pluton affecting the metavolcanic in the metasomatic contact is the Y-Sr ratio. The fresh rocks display high Y and low Sr. At the same time, the hydrothermally altered epidote-rich quartz-diorite presents low Y and high Sr concentration values (Fig. 8). The barium enrichment in the epidote-rich quartz-diorite is another evidence of the hydrothermal process that evolved in the intrusive plutonic rocks. This enrichment in Ba and Sr could be justified by the location of the collected samples in the contact between the quartz-diorite and metavolcanic rocks that host the magnetite-rich rock of the Água Limpa Fe-(Cu) deposit. The enrichment in LREE can be related to the introduction of sediments in the material-subducted enriched mantle, according to Fonseca et al. (2018).

5.2. Chemistry of magnetite

The major and trace elements analyzed in the magnetite by EMPA and LAICP-MS provide critical input to the definition of the magmatic-hydrothermal characteristics of the deposit. Magnetite from Fe-skarn ores contain lower Ti, V, Co, Ni, and Ga (Dare et al., 2014; Nadoll et al., 2014; Hu et al., 2015). The chemical evidence of magmatic-hydrothermal characteristic of the magnetite studied is observed by analyzing the EMPA results in the correlation between Ti (ppm) vs. Ni/Cr (ppm), as proposed by Dare et al. (2014), demonstrated in Fig. 11a. The relation of Mg+Mn (ppm) vs. (Si+Al)/(Mg+Mn) (ppm) further contributes to this interpretation, reflecting a decrease in fluid/rock ratios that, according to Hu et al. (2014), characterize the Fe-Cu deposits analyzed in their work (Fig. 11b).

Following the proposition of Dupuis and Beaudoin (2011) two clusters of magnetite are identified in the Água Limpa Fe-(Cu) deposit in Ca + Al + Mn (wt.%) vs. Ti + V (wt.%) diagram (Fig. 11c) in the EMPA analysis: one forms a trend between BIF-Skarn, and the other stays in the IOCG field (Fig. 11c). The magnetite core shows a skarn characteristic, while the rim shows an IOCG chemistry characteristic (Fig. 11d). Analyzing Ni/(Cr+Mn) (wt.%) vs. Ti+V (wt.%) diagram analysis in EMPA, the chemistry of the magnetite indicates a skarn deposit (Fig. 11d).

The LAICP-MS analysis in the magnetite samples delivered a possible magmatic-hydrothermal signature of the deposit. The analyses from the ALDD006, which are located in the center of the deposit, show a preserved magmatic signature to the magnetite, imprinted by the higher average concentrations of Al and Ti (945.79 ppm and 1566.74 ppm, respectively) when compared with the ALDD009 (620.89 ppm and 247.54 ppm, respectively). On the other hand, ALDD009, positioned in the southern portion of the deposit presents low concentrations average of Ti (247.54 ppm) and V (62.93 ppm) in combination with relatively high Mn (617.51 ppm) concentrations average, attesting, according to Nadoll et al. (2015), the characteristic features of skarn deposits. The Co average concentration in the ALDD006 also shows the magmatic signature features compared to the lower concentration average of ALDD009 with hydrothermal signature, 9.27 ppm and 0.64 ppm, respectively.

Analyzing the Mg + Mn (ppm) vs. Ga + Sn (ppm) diagram proposed by Xie et al. (2022) based on the LAICP-MS measures in magnetite from ALDD006 and ALDD009, it is possible to distinguish different styles of skarns deposits associated with the Fe, Cu, and Au mineralization (Fig. 11e). This variety of mineralization types can be observed analyzing the Al + Mn (ppm) vs. Ti + V (ppm) diagram proposed by Nadoll et al. (2015) that shows a Skarn + Porphyry mineralization to the main concentration of the analysis, with two points in the Porphyry chemistry signature to the magnetite analyzed in the center of the deposit (Fig. 11f) related to the high concentration of Ti.

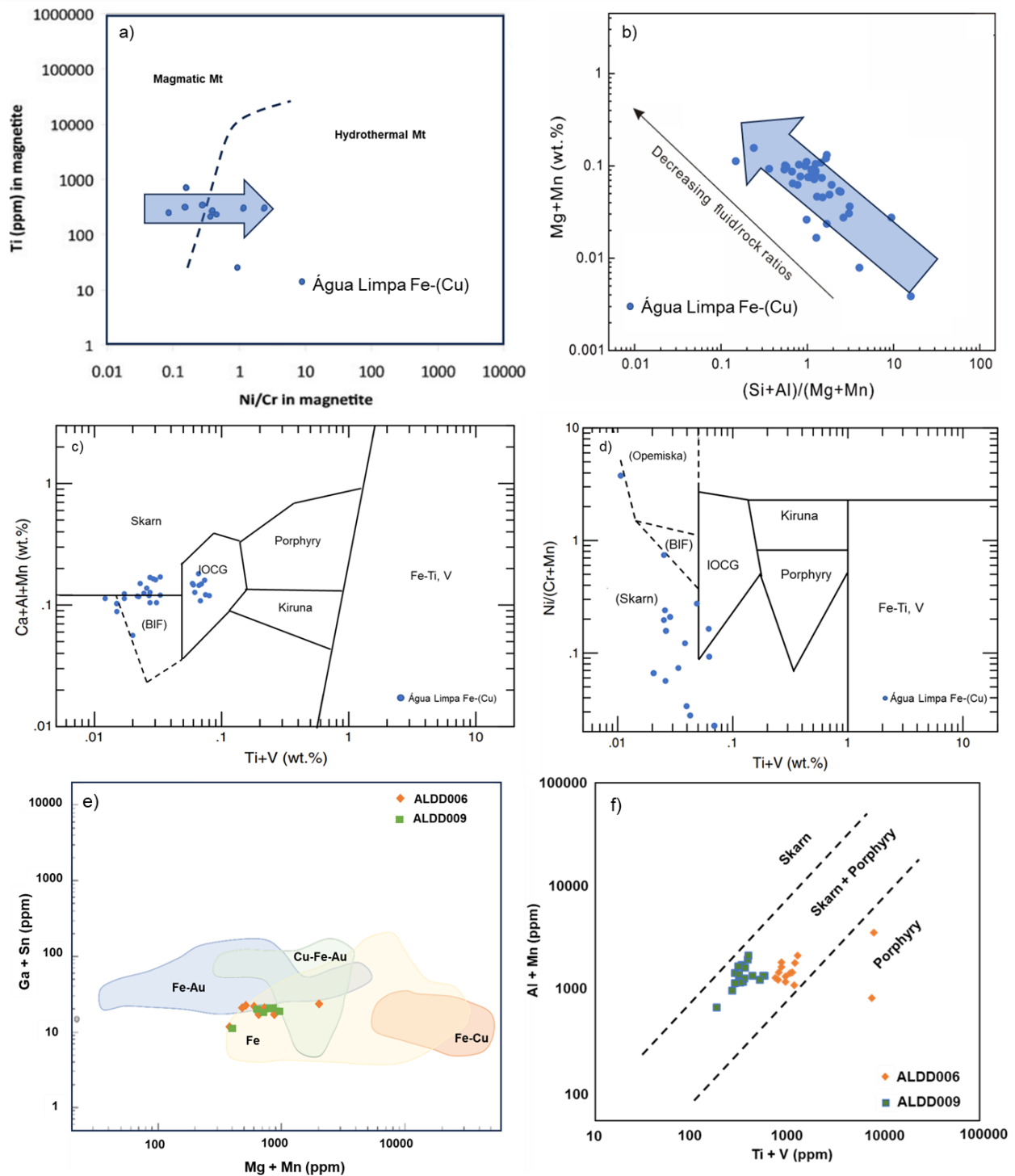


Fig. 11: a) Magmatic-hydrothermal diagram characteristics for magnetite analyzed in EMPA, modified from Dare et al. (2014). b) Relationship of the decreasing fluid/rock ratios to the magnetite analyzed in the EMPA and the relation with the Fe-Cu deposits described by Hu et al. (2014); c) d) Diagram showing the composition of skarn to magnetite analyzed in the EMPA, modified from Dupuis and Beaudoin (2011); e) Ga + Sn (ppm) vs. Mg + Mn (ppm) diagram for magnetite from different styles of Fe-

Cu-Au skarn analyzed in LAICP-MS to the Água Limpa Fe-(Cu) deposit, modified from Xie et al. (2022); f) Ti + V (ppm) vs. Al + Mn (ppm) diagram for magmatic-hydrothermal magnetite from Água Limpa Fe-(Cu) deposit, analyzed in LAICP-MS, modified from Nadoll et al. (2015)

5.3 Genetic interpretations

Based on the petrography descriptions of the samples collected along the Água Limpa Fe-(Cu) deposit, we propose a mineral paragenesis of the host rocks and the mineralization presented in Table 4. Clinopyroxene, plagioclase, and quartz comprise the primary magmatic mineralogy of the host rock. This mineral assemblage is changed to magnetite (Mag 1), quartz, hornblende, ilmenite $\pm$ titanite, epidote, $\pm$ clinozoisite in the first phase of the alteration phase (Fig. 12 a). The second phase of alteration is characterized by the dissolution and reprecipitation of magnetite (1) forming magnetite (Mag 2) + allanite $\pm$ carbonate $\pm$ apatite and micas (biotite $\pm$ chlorite $\pm$ sericite) (Fig. 12 b). In the later alteration, a sulfide phase is characterized by pyrite, pyrrhotite, and chalcopyrite crystallized contemporarily with mushketovite (Mag 3) $\pm$ carbonate $\pm$ apatite.

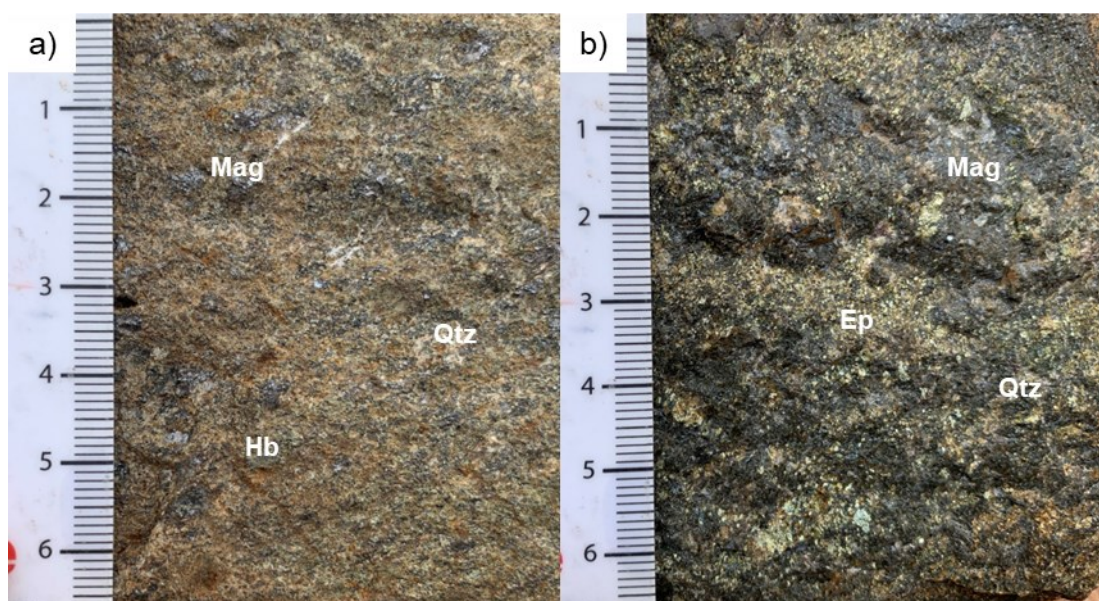


Fig. 12: a) First phase of magnetite associated with hornblende and quartz; b) Second phase associated with epidote and quartz. Mag – Magnetite; Qtz – Quartz; Hb – Hornblende; Ep – Epidote.

Minerals	Magmatic	Hydrothermal		
		First Phase	Second Phase	Third Phase
Clinopyroxene	_____			
Plagioclase	_____			
Quartz	_____			
Hornblende		_____		
Magnetite		_____	_____	_____
Ilmenite		-----		
Titanite		-----		
Epidote		_____	_____	
Clinozoisite		_____		
Allanite			-----	
Carbonate			-----	-----
Apatite			-----	-----
Biotite			_____	
Chlorite			-----	
Sericite			-----	
Mushketovite				_____
Pyrite				_____
Chalcopyrite				_____
Pyrrhotite				_____

Table 4: A mineral assemblage of the Água Limpa Fe-(Cu) deposit. Line widths indicate the relative mineral abundances. Dotted lines indicate the minor and trace phases.

Despite the banding of granular magnetite (Mag 1) and quartz, the association with hornblende and biotite shows that the mineralization can be temporarily related to the formation of metavolcanic mafic rock. According to Tornos et al. (2024), the initially high fO_2 in the mineral systems dominated by magnetite-(apatite) rocks associated with mafic magmas promotes the formation of low-Ti (< 1 wt%) magnetite as observed in the magnetite studied. The porosity observed in the Mag 2 (Fig. 5a, b) can be associated with the dissolution and re-precipitation process (DRP) originating from high-temperature fluid and the initially high fO_2 activity (Putnis 2009; Putnis and John 2010; Dare et al., 2015; Hu et al., 2015; Heidarian et al., 2016; Hu et al., 2023). The DRP can result in the formation of a new mineral or the same mineral with a different composition (Putnis, 2002, 2009, 2015; Harlov et al., 2005, 2011; Ruiz-Agudo et al., 2014). Magnetite compositions can be modified significantly by the DRP process (Hu et al. 2014, 2015; Hu et al., 2023). Major elements such as Si, Ca, Al, Mg, and Mn behaved as mobile elements during the dissolution and re-precipitation, so the contents of these elements in the secondary magnetite are different from those in the primary magnetite (Hu et al., 2023). The hydrothermally altered magnetite Mag 2 is

replaced to various extents by a secondary platy-shaped variety (Mag 3) along fractures and/or grain boundaries that have appreciable porosity (Fig. 5c, d, e). The platy-shaped magnetite is generally in sharp contact with the Mag 2 and contains numerous solid inclusions, such as sulfide and silicate minerals. In the later stage of hydrothermal alteration, smaller temperatures, less fO_2 activity associated with high fS activity and reduced fluids work to crystallize a mushketovite associated with pyrite-chalcopyrite-pyrrhotite that occurs as disseminated and in small veins. Mushketovite is a pseudomorphic replacement of primary hematite by magnetite under reducing conditions (Mücke and Cabral, 2005). The crystallization of the mushketovite in the same paragenesis of sulfides marks the later ore-forming environment, and the iron-rich hydrothermal fluids transformed from high temperature and oxidized to low temperature and reduced features, with less oxygen fugacity associated with high sulfur fugacity active (Hong et al., 2021). Pires (1977) associates the pyrite-pyrrhotite-chalcopyrite as a later hydrothermal phase in the Morro da Mina Mn deposit, that is located in the same trend of metavolcanic-sedimentary rocks of Água Limpa Fe-(Cu) Deposit. Fig. 12 shows the three stages of the magnetite formation in the Água Limpa Fe-(Cu) deposit.

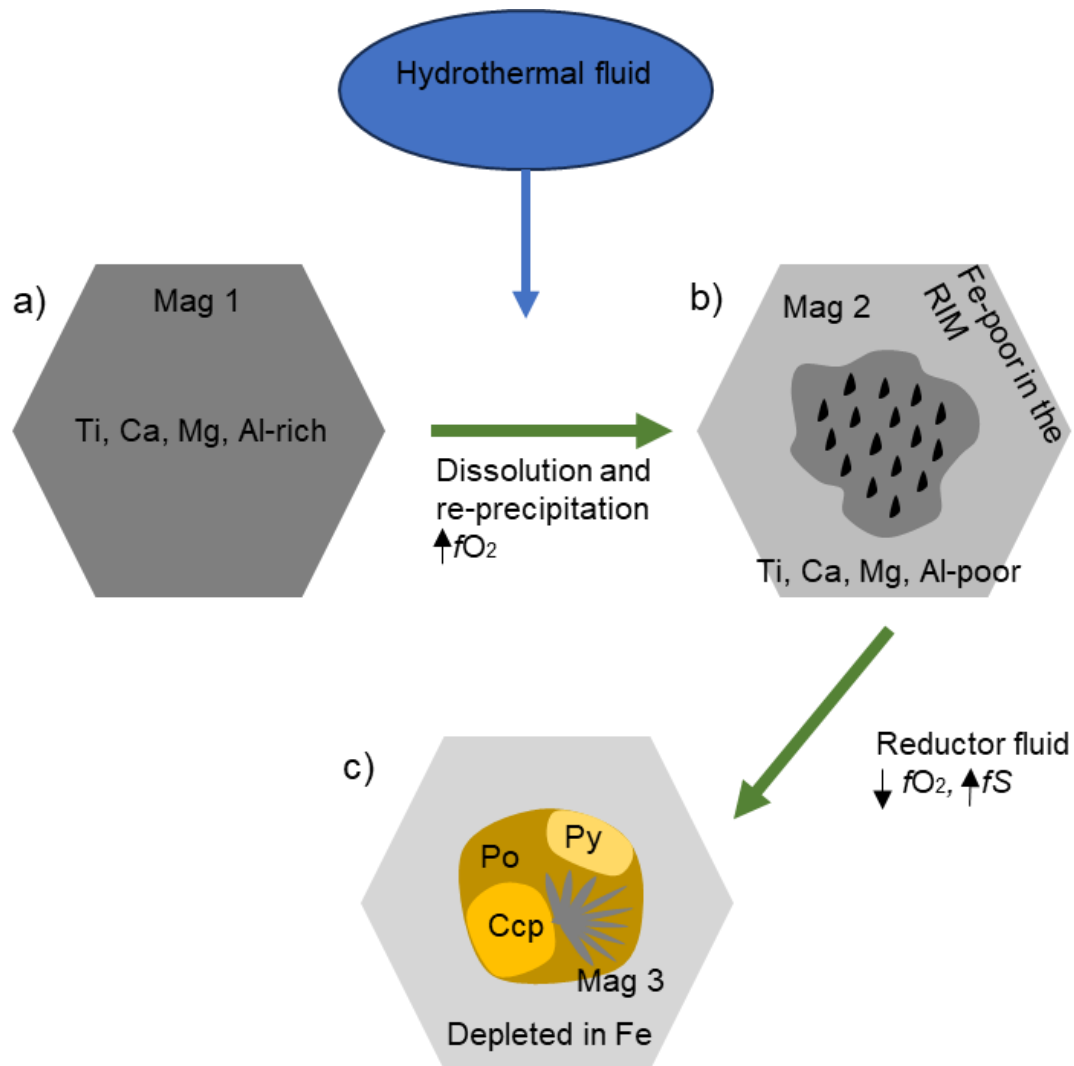


Fig. 13: Schematic illustration showing three stages of magnetite in the Água Limpa Fe-(Cu) deposit: a) Mag 1 is a magmatic hypidiomorphic with high content of Ti, Ca, Mg, Al; b) Mag 2, which is depleted in Ti, Ca, Mg, Al, forming a porous anhedral crystal with Fe-poor RIM; c) Mag 3 (mushketovite), related with the crystallization of pyrrhotite, chalcopyrite, and pyrite. Po, Pyrrhotite; Ccp, Chalcopyrite; Py, Pyrite.

6. Conclusions

The Água Limpa Fe-(Cu) deposit is positioned in contact with metabasalt and intrusive quartz-diorite sanukitoid suite related to the end period of the arc-related formation of the Mineiro Belt during Rhyacian Minas-Bahia orogeny.

Three types of magnetite were identified in the Água Limpa Fe-(Cu) deposit. Petrography and trace-element characteristics indicate that: (1) Mag 1 is magmatic in origin, with Ti, Ca, Mg, and Al-rich content; (2) Mag 2 is magmatic-hydrothermal in origin and formed as a product of dissolution and re-precipitation of iron-rich fluid with high oxygen fugacity activity, crystallizing porous magnetite with an iron-depleted RIM, and also depleted in Ti, Ca, Mg, Al concerning Mag 1; (3) Mag 3 is hydrothermal magnetite, crystallized as platy-shape form (mushketovite), in smaller temperatures, less fO_2 activity associated with high fS activity, and reduced fluids associated with pyrite-chalcopyrite-pyrrhotite that occurs as disseminated and in small veins cross-cutting the magnetite-rich rock.

Integrating the geology, mineral assemblage, and chemistry of magnetite confer the magmatic-hydrothermal characteristics of the Água Limpa Fe-(Cu) deposit.

7. Author statement

Eduardo Oliveira: Conceptualization, Methodology, Formal analysis, Investigation, and Writing – Original Draft. Carlos Rosière: Conceptualization, Methodology, Validation, Investigation, Writing, Review, and Editing, Supervision, and Funding Acquisition.

8. Declaration of competing interests

The authors declare that they have no competing financial interests or personal relationships that could have influenced the work reported in this paper.

9. Data availability

Data will be made available on request.

10. Acknowledgments

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5 CONCLUSÕES

5.1 - Evolução gequímica, petrológica e litoestratigráfica

Os trabalhos de geologia de campo, geoquímica de rocha total, geocronologia, e geologia estrutural que deram a base para o desenvolvimento dessa tese de doutorado se concentraram em uma sequência vulcanossedimentar localizada na porção norte do Cinturão Mineiro, em contato com o embasamento arqueano que compõem o Bloco Minas, na porção sul do cráton São Francisco. Essa unidade geológica, definida por Ebert (1963, 1965) como Formação Lafaiete foi aqui definida como Sequência Lafaiete, em função de novos dados petrológicos e geocronológicos, que revelaram um período de vulcanismo e sedimentação na porção norte do Cinturão Mineiro durante a orogênese riaciana Minas-Bahia e fazem parte da evolução final das unidades que compõem o Cinturão Mineiro:

1. As idades U-Pb obtidas nesse trabalho, a partir da datação da sequência vulcanossedimentar de Lafaiete e da suíte intrusiva correlacionada à Suíte Alto Maranhão, inseridas na porção norte do Cinturão Mineiro, apresentam uma história de 100 milhões de anos de evolução geodinâmica na porção sul do cráton São Francisco, entre vulcanismo e sedimentação: A idade de 2170.3 ± 8.2 Ma obtida no basalto definido como base da Sequência Lafaiete representa o início do vulcanismo após um período de quiescência global, denominado “magmatic lull”, associado à formação de uma bacia forearc. Em seguida há uma segunda injeção de um magma ultramáfico, responsável pela formação de um magma da série picrito-boninito em 2149.8 ± 5.9 Ma, formado a partir de um magma de composição dunitica a peridotítica, contaminado na fonte em função do início do processo de subdução responsável pela formação da bacia forearc.
2. O boninito (2150 Ma) forma-se nos estágios iniciais da subdução, que foi responsável pela formação da Suíte Alto Maranhão, cristalizada entre 2140 a 2130 Ma.
3. Os zircões do boninito apresentam uma idade metamórfica em 2100 Ma, relacionado à idade da colisão do Arco Alto Maranhão e o protocraton.

4. As rochas vulcânicas da Sequência Lafaiete se formaram em um ambiente de forearc que foi amalgamado com o embasamento arqueano-proterozóico do bloco Minas em 2100 Ma. Após a colisão desses dois terrenos há o relaxamento do orógeno, com subsequente erosão e deposição de sedimentos com uma primeira idade de máxima deposição em 2099 ± 9 Ma, composta essencialmente por quartzito, que exhibe duas fontes de material sedimentar: uma proveniente do embasamento arqueano do Cráton do São Francisco e outra formada pela erosão das rochas que compõem o orógeno no Cinturão Mineiro. No topo da sequência sedimentar, existe uma sedimentação mais proximal associada à gonditos, folhelhos negros, grauvacas, datada com período de máxima deposição em 2079 ± 6.3 Ma (gondito) e 2064.5 ± 6.2 Ma (grauvaca). O pico metamórfico é marcado no boninito em 2098.9 ± 6.2 Ma e representa o fechamento do orógeno. com idade de máxima deposição em 2099 Ma, e com sedimentos relacionados aos granitos da suíte Alto Maranhão tendo idades de máxima deposição em 2075 e 2065 Ma, respectivamente.

5. A sequência vulcânica é crono-correlata à deposição do Grupo Sabará no Supergrupo Minas. O vulcanismo incia-se com uma zona de subducção sob o protocraton gerando uma bacia forearc e gerando os sedimentos que foram depositados no grupo Sabará, que recebe aporte de sedimentos das rochas formadas no sistema orogênico ao sul do protocraton.

6. O enriquecimento em elementos incompatíveis na série picrito-boninito, além de ETRL, ϵHf negativos e idades modelo arqueanas revela a influência da contaminação por uma crosta continental arqueana durante a formação do magma responsável pela série picrito-boninito.

7. As idades magmáticas e de máxima deposição para as rochas que compõem a Sequência Lafaiete sugerem que essas rochas se formaram em um ambiente de arco que foi amalgamado contra o embasamento arqueano-proterozóico do Bloco Minas sendo crono-correlatas à deposição dos grupos Sabará e Itacolomi, dentro do Supergrupo Minas.

5.2 - Compartimentação tectônica-estrutural da porção sul do cráton São Francisco, no contato entre o Cinturão Mineiro, Complexo Mantiqueira e Bloco Minas

Em relação ao modelo de evolução estrutural da porção sul do cráton São Francisco, entre o embasamento arqueano-proterozóico do Bloco Minas e as rochas provenientes de arcos magmáticos, vulcânicos e sequências intraoceânicas que compõem o Cinturão Mineiro e o Complexo Mantiqueira, foram definidos 06 domínios estruturais que representam o arcabouço tectono-estrutural dessa porção crustal. As estruturas mapeadas na porção norte do Cinturão Mineiro revelam um sistema acrescionário-colisional durante o fechamento do orógeno Minas-Bahia, delimitado por três zonas de cisalhamento transcorrentes que definem a geometria da porção sul do cráton São Francisco.

Os limites norte-noroeste do Cinturão Mineiro e norte-nordeste dessa unidade tectônica são definidos por duas estruturas de direção NE-SW e NW-SE, definidas pelas zonas de cisalhamento Jeceaba-Bom Sucesso e Congonhas-Itaverava, que mostram uma movimentação de deslocamento lateral direito, sendo reativadas no sentido inverso sinistral.

Na porção leste do Cinturão Mineiro, a zona de cisalhamento Lenheiros de direção de deslocamento lateral dextral se desenvolveu ao longo do contato entre o Cinturão Mineiro e os Complexos Mantiqueira-Juiz de Fora, durante o fechamento da orogenia Minas-Bahia. A zona de cisalhamento Lenheiros exibe uma lineação mineral com plunge entre 20° a 60° para sul, com lineações de estiramento com plunge em 50° para NE, indicando um carácter oblíquo de deslocamento.

A zona de cisalhamento Jeceaba-Bom Sucesso define a borda norte do Cinturão Mineiro e está em contato com o embasamento arqueano do protocontinente São Francisco. Todo esse processo de transcorrência ocorreu entre 2130 a 2060 Ma durante o fechamento do orógeno Minas-Bahia, existente na porção sul do cráton São Francisco.

Um componente de movimentação reversa sinistral também é observada nas Zonas de Cisalhamento Congonhas-Itaverava e Lenheiros. A foliação nesses domínios tem direção NE-SW com mergulho para SE, relacionando-se à colagem

Neoproterozóica dos crátons São Francisco e Congo, entre 620-520 Ma (Pedrosa-Soares et al., 2001; Noce et al., 2007; Kuribara et al., 2019).

5.3 - Metalogênese do depósito magmático-hidrotermal de Fe-(Cu) Água Limpa

O depósito de Água Limpa é um depósito mineral magmático-hidrotermal hospedado em rochas metavulcânicas da Formação Lafaiete, em contato com a suíte intrusiva Alto Maranhão, aqui definida pelo Plúton Água Limpa, que tem sua gênese associada ao Orógeno Minas-Bahia. Os estudos petrográficos e de química mineral (microsonda e LA-ICP-MS) em magnetita revelam três fases de cristalização de magnetita: (1) uma magmática compacta, com formato granular euédrico enriquecida em Si, Ca, Mg, Al e Ti; (2) uma magnetita porosa, formada à partir de dissolução e reprecipitação da magnetita (Mag 1) através da passagem de fluídos hidrotermais de alta temperatura e alta fugacidade de oxigênio, recristalizando uma magnetita depletada nos elementos Si, Ca, Mg, Al e Ti e formando uma borda também depletada em ferro; (3) uma magnetita esqueletóide/especular denominada de mushketovita, cristalizada nos processos hidrotermais tardios, com baixa temperatura e baixa fugacidade de oxigênio e alta fugacidade de enxofre, formando um fluído redutor, associada associada à uma paragênese de pirita-calcopirita-pirrotita.

O processo de alteração hidrotermal foi caracterizado analisando os dados de elementos maiores e traços por microsonda e LA-ICP-MS em magnetita, hornblenda e biotita, no qual mostram alteração química do centro para as bordas desses minerais. Também foram usados dados petrográficos para caracterização das alterações hidrotermais que atuaram no depósito de Água Limpa, onde os minerais primários compostos por plagioclásio e clinopiroxênio foram substituídos por hornblenda, epidoto, clinozoisita, carbonato e allanita como fases de alteração hidrotermal.

Integrando os dados de geologia de campo, relações de contato, análises petrográficas e geoquímicas de elementos maiores e traços, define-se o depósito Água Limpa, como um depósito magmático-hidrotermal de Fe-(Cu).

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