



Universidade de Brasília
Instituto de Geociências
Programa de Pós-graduação em Geologia

PETROLOGIA E METALOGÊNESE DO DEPÓSITO DE Au ($\pm$ Cu) RAIMUNDA, SUL DO CRÁTON AMAZÔNICO, PROVÍNCIA MINERAL JURUENA (MT)

ADRIANA ARAUJO CASTRO LOPES

Tese de Doutorado N° 212

Orientadora: Prof^ª. Dr^ª. Márcia Abrahão Moura



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RESUMO

A Província Mineral Juruena, no sul do Cráton Amazônico, é uma província mineral emergente de classe mundial devido aos numerosos depósitos de Au, Cu e metais básicos descobertos e explorados nas últimas décadas. O depósito de Au ($\pm$ Cu) Raimunda, localizado ao leste da província, está inserido no Complexo Novo Mundo, formado por fácies graníticas (monzonito, monzogranito, biotita granodiorito e sienogranito), do tipo I, cálcio-alcálicas, com afinidades com granitos de arco vulcânico, com idades U–Pb variando entre 2,03 e 1,98 Ga, T_{DM} entre 2,52–2,15 Ga e $\epsilon Nd(t)$ entre -2,0 e +1,8. Os dados petrológicos, geoquímicos e isotópicos do Complexo são correlacionáveis aos do Arco Magmático Cuiú-Cuiú, gerado durante a principal fase de arco na região (2,05 – 1,99 Ga). E, juntamente com estudos mais recentes sobre a geotectônica do cráton, permitem sugerir a existência de um arco magmático contínuo do Paleoproterozoico na porção sul do Cráton Amazônico, responsável pela formação das principais rochas graníticas pertencentes às Províncias Minerais Tapajós e Juruena. As fácies monzogranito e sienogranito apresentam mineralização expressiva, e as porções mineralizadas são mais hidrotermalizadas, compreendendo metassomatismo sódico incipiente, microclinização, alteração propilítica, alteração sericitica, silificação, estágio de sulfetação e carbonatação tardia. A mineralização é disseminada e geneticamente associada à alteração sericitica (Au1) e ao estágio de sulfetação (Au2). No estágio inicial de mineralização, o ouro ocorre incluso em pirita, disposto em veios e vênulas de quartzo-muscovita-clorita-pirita-ouro. Os dados de geotermômetro da clorita apontam temperaturas entre 340 e 370 °C para esse sistema. No estágio principal da mineralização, o ouro ocorre incluso em pirita ou preenchendo veios e/ou vênulas de sulfeto maciço ou de quartzo-sulfetos ricos em sulfetos de cobre e bismuto. Dados do geotermômetro da clorita e da microtermometria das inclusões fluidas indicam condições de T entre 325 e 380 °C como o principal intervalo de temperatura da mineralização do depósito. A ocorrência de inclusões aquosas contemporâneas com o fluido aquoso-carbônico de alta temperatura, além dos estágios de homogeneização para o vapor e para o líquido, indicam a imiscibilidade do fluido e/ou mistura de fluidos quentes e mais salinos com fluidos provenientes de fontes meteóricas, mais frias. Fluidos ricos em H_2O -NaCl- $CaCl_2$, contemporâneos à carbonatação tardia, foram aprisionados em temperaturas com intervalo principal de 200–250 °C. Os dados de isótopos estáveis de oxigênio em quartzo mostraram intervalo entre 9 e 12,6‰, e os

dados isotópicos do fluido estão entre 3,11 e 7,86‰. Para o enxofre, o $\delta^{34}\text{S}_{\text{py}}$ mostra intervalos de -1,4 a +0,1 ‰. Em ambos os casos, os fluidos são interpretados como de origem magmática no depósito Raimunda. A coexistência de inclusões fluidas aquosas e aquocarbônicas de alta temperatura revela uma mistura de fluidos quentes de média salinidade com fluidos mais frios e de baixa salinidade. O ouro foi inicialmente transportado como complexos cloretados em fluido de alta temperatura, de salinidade média, ácido e oxidado, provenientes da câmara magmática, e mais tarde como complexos de H_2S^- . A instabilidade físico-química durante a subida do fluido é interpretada como um fator desencadeante da precipitação do minério. Esses resultados oferecem informações valiosas sobre a gênese dos depósitos de Au-pórfiro e as suas implicações para a prospecção mineral no sul do Cráton Amazônico.

Palavras-chave: Cráton Amazônico, Província Mineral Juruena, granito, arco magmático, Paleoproterozoico, geocronologia, petrologia, alteração hidrotermal, inclusões fluidas, isótopos estáveis, ouro pórfiro.

ABSTRACT

The Juruena Mineral Province, located in the southern Amazonian Craton, is an emerging world-class mineral province known for its abundant gold (Au), copper (Cu), and base metal deposits. The Raimunda Au deposit ($\pm$ Cu) sits in the east of the province. It is part of the Novo Mundo Complex, which is composed of I-type calc-alkaline granitic facies (monzonite, monzogranite, biotite granodiorite and syenogranite) related to volcanic arc magmatism yielding U–Pb ages between 2.03 and 1.98 Ga, T_{DM} ages 2.52–2.15 Ga, and $\epsilon Nd(t)$ values of -2.0 and +1.8. The integration of isotopic, geochemical and petrological data of the Complex, which correlate with those of the Cuiú-Cuiú Magmatic Arc formed during peak arc magmatism in the region (2.05 – 1.99 Ga), and recent geotectonic studies of the Craton allow us to suggest the existence of a continuous Paleoproterozoic magmatic arc in the southern sector of the Amazonian Craton. This magmatism was responsible for the genesis of the major granitic rocks within the Tapajós and Juruena Mineral Provinces. The monzogranite and syenogranite facies exhibit significant mineralization and have undergone extensive alteration including incipient sodic metasomatism, microclinization, propylitic alteration, sericitic alteration, silicification, sulfide stage, and late-stage carbonate alteration. Mineralization is sparsely disseminated and genetically associated with sericitic alteration (Au1) and sulfide stage (Au2). At early mineralization, gold occurs as inclusions in pyrite hosted into quartz-muscovite-chlorite-pyrite-gold veins and veinlets. Chlorite geothermometer yields temperatures of 340–370 °C for this system. At peak mineralization, gold occurs either as inclusions in pyrite or fills massive sulfide and Cu-Bi-rich quartz-sulfide veins and veinlets. Chlorite geothermometer and fluid inclusion microthermometric data indicate that the primary temperature range for mineralization in the deposit falls between 325 and 380 °C. The occurrence of inclusions formed from aqueous, high-temperature aqueous-carbonic fluid coupled with stages of homogenization into vapor and liquid phases point out to fluid immiscibility and/or mixing of more saline hot fluids with cooler fluids derived from meteoric sources. H₂O-NaCl-CaCl₂ fluids associated with late-stage carbonate alteration were trapped at temperatures within a range of 200–250 °C. Oxygen stable isotope data of quartz range between 9 and 12.6 ‰, and fluid isotope data range from 3.11 to 7.86‰. Sulfur isotope data $\delta^{34}S_{py}$ range from -1.4 to +0.1‰. In both cases, the fluids are attributed to a magmatic origin at the Raimunda deposit. The coexistence of high-temperature aqueous and

aqueous-carbonic fluid inclusions reveals mixing of hot medium-salinity fluids with cooler low-salinity fluids. Gold was initially transported as chloride complexes in very hot, medium-salinity, acidic, oxidized fluids expelled by the magma chamber, and as H_2S complexes afterwards. The physical-chemical instability caused by the ascent of the fluid likely triggered ore precipitation. These results provide valuable insights into the genesis of porphyry Au deposits and their implications for mineral exploration in the Amazonian Craton.

Keywords: Amazonian Craton, Juruena Mineral Province, granite, magmatic arc, Paleoproterozoic, geochronology, petrology, hydrothermal alteration, fluid inclusions, stable isotope, porphyry gold deposit.

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CAPÍTULO 1

INTRODUÇÃO

1 INTRODUÇÃO

Eventos vulcânicos e plutônicos foram essenciais para a geração da crosta durante o Arqueano e Proterozoico (Cawood *et al.* 2013), compreendendo um dos mais significativos intervalos de geração de crosta, entre 2,1 e 1,7 Ga (Hawkesworth *et al.* 2010).

O centro sul do Cráton Amazônico hospeda múltiplas gerações de granitos com características de arco magmático, em uma região marcada por domínios paleoproterozoicos e expressivo teor de ouro na denominada Província Mineral Juruena (Tassinari e Macambira 1999) (Fig. 1.1). Historicamente, produziu cerca de 300 t de ouro e, em 2017, foi descoberta uma importante jazida de Cu (500 m@0,6% Cu), o que aumentou enormemente os projetos de exploração na região (Crusader 2018; Teixeira *et al.* 2023).

A existência de muitos depósitos minerais paleoproterozoicos de ouro epitermais de alta, intermediária e baixa sulfetação e depósitos do tipo Cu-Mo-Au pórfiro (Fernandes e Juliani 2019) tem atraído atenção mundial para a região. As principais mineralizações da Província Juruena estão espacial e temporalmente relacionadas a granitos paleoproterozoicos predominantemente de assinatura geoquímica cálcio-alcálica (Moura *et al.* 2006; Paes de Barros 2007; Silva & Abram 2008) e com assinaturas de granitos de arco vulcânico.

Mapeamentos geológicos, aumento na geração de dados geocronológicos e isotópicos e estudos mais detalhados sobre o comportamento do fluido mineralizante e da origem do minério permitiram a individualização de novas suítes graníticas e de padrões de mineralizações regionais (Fernandes e Juliani 2019; Guimarães e Klein 2020; Guimarães *et al.* 2021; Tarello Neto e Moura 2021).

No entanto, esses estudos não contemplam toda a região e, ao longo da última década, várias pesquisas têm sido publicadas no sentido de detalhar os depósitos para a melhor compreensão do padrão mineralizante de grande parte da província, destaque para a sua porção leste. Na região de Novo Mundo, porção leste da Província Mineral Juruena, diversos depósitos de ouro disseminados estão hospedados em rochas graníticas hidrotermalizadas, mas também há descrição de corpos de minério filonianos e venulares hospedados em estruturas dúcteis a rúpteis (Paes de Barros 2007; Assis 2015).

A partir do estudo petrológico, geoquímico e isotópico do granito Novo Mundo,

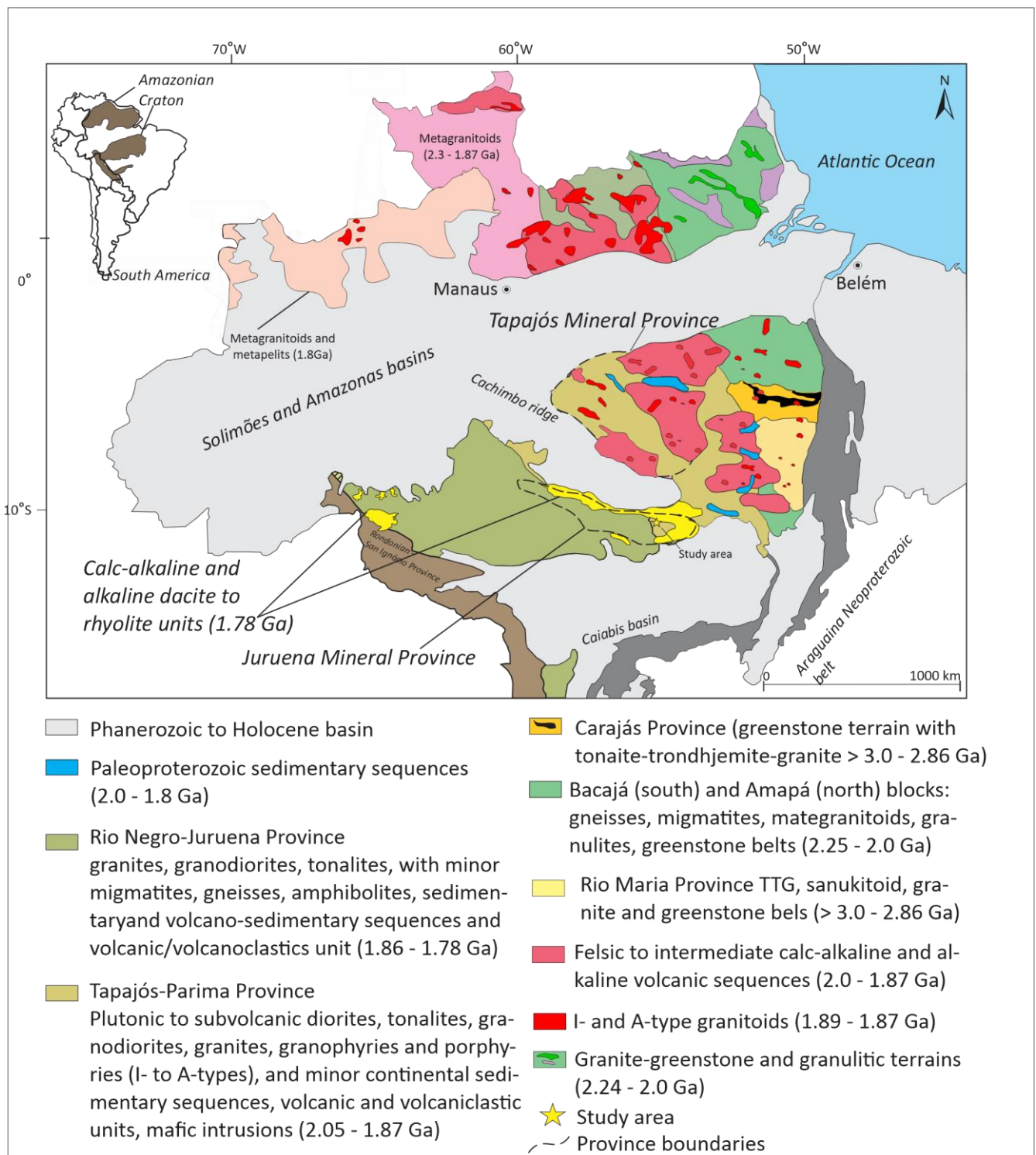


Figura 1. 1. Ambiente geológico e tectônico da Província Mineral Juruena no Cráton Amazônico, baseado em [Juliani et al. \(2021\)](#) e [Fernandes et al. \(2019\)](#). Províncias geocronológicas baseadas em [Tassinari & Macambira \(2004\)](#).

foi possível identificar quatro fácies graníticas: monzonito (2032 ± 6 Ma), monzogranito (2029 ± 4 Ma), biotita granodiorito ($1989 \pm 6,2$ Ma) e sienogranito ($1987 \pm 7,4$ Ma a 1964

± 1 Ma), todas as fácies de composição cálcio-alcálica, para as metaluminosas do tipo I e granitos de arco vulcânico (Lopes *et al.* 2024 *in submission*).

A compreensão da composição, da geoquímica e dos dados isotópicos desses granitos foi de fundamental importância para a elaboração da evolução do Granito Novo Mundo, bem como para a compreensão da abrangência do Arco Cuiú-Cuiú, na porção extremo norte da Província Mineral Juruena, semelhante à Província Tapajós, mostrando indícios consistentes de um arco magmático contínuo paleoproterozoico na porção sul do Cráton Amazônico (Misas 2015; Cruz *et al.* 2016; Fernandes *et al.* 2016; Juliani *et al.* 2021; Lopes *et al.* 2024 *in submission*).

A percolação de fluidos hidrotermais promove alterações mineralógicas e geoquímicas anômalas e sistemáticas nas rochas encaixantes, cuja extensão e magnitude dependem do seu estado de alteração, porosidade, permeabilidade e grau de reatividade química nesses corpos. A interação fluido-rocha gera *pathways* de alteração que podem auxiliar na determinação dos guias de exploração em terrenos maduros (Guillard *et al.* 2020).

O estudo da alteração hidrotermal, da evolução e composição dos fluidos e a compreensão da origem e modelo evolutivo do depósito Raimunda, em Novo Mundo, indicou a origem dos fluidos de porções profundas da crosta, contendo salinidades medianas e alta temperatura, inclusões aquocarbônicas e aquosas, apontando para um sistema que teve origem em profundidade e que misturou, em porções mais rasas com fluido meteórico. O minério foi precipitado em regiões intermediárias a rasas da crosta semelhantes a depósitos do tipo Au pórfiro (Lopes e Moura 2024).

Apesar do grande avanço em relação ao conhecimento sobre os eventos mineralizadores proterozoicos no Cráton Amazônico, nas últimas décadas, poucos estudos tratam, com robustez, dos dados isotópicos e da origem dos fluidos mineralizantes, visando à definição da gênese dos depósitos minerais da região sul do Cráton.

Portanto, este trabalho contém novos dados petrológicos, de inclusões fluidas e de isótopos estáveis inéditos do depósito paleoproterozoico Raimunda, localizado na porção sul do Cráton Amazônico, permitindo a proposição do complexo Novo Mundo, no Arco Magma Cuiú-Cuiú, e reforçando a recente proposta de arco magmático contínuo para porção sul do Cráton (Juliani *et al.* 2021), além de contribuir para a melhor compreensão

do potencial metalogenético da região, com um modelo metalogenético do depósito Raimunda.

1.1. Justificativa e objetivos

O depósito Raimunda difere da maioria dos depósitos conhecidos na porção leste da Província Mineral Juruena por se tratar de um depósito Au ($\pm$ Cu + Te + Bi + Pb), apesar de, até o momento, apenas o ouro ser considerado viável economicamente, com minério disseminado e em níveis de sulfeto maciço. Alguns estudos abordaram a petrologia, a geocronologia e a geologia estrutural do Granito Novo Mundo ([Paes de Barros 2007](#); [Miguel Jr. 2011](#); [Testa 2018](#)), mas a compreensão acerca do sistema mineralizador e do modelo de depósito Raimunda ainda é incipiente, em especial, no âmbito dos parâmetros físico-químicos dos fluidos mineralizantes e da origem do minério.

Ademais, devido à escassez de dados isotópicos robustos para a melhor compreensão da colocação desses corpos e da metalogênese da área, este trabalho é oportuno e se faz necessário para ampliar a compreensão sobre o contexto tectônico e metalogenético local, expandindo-se para correlações regionais e mundiais.

Isso posto, esta tese tem como objetivo principal contribuir para o aprofundamento das discussões e do conhecimento sobre a evolução magmática e tectônica, bem como para a melhor compreensão do sistema metalogenético do ouro que se deu nos granitos cálcio-alcalinos na porção leste da Província Mineral Juruena, sul do Craton Amazônico, desde a fonte do metal, migração do fluido até a formação do depósito mineral Raimunda. Os objetivos específicos são:

- Caracterizar o granito Novo Mundo, por meio de análise de campo, dados petrológicos, geoquímicos e isotópicos;
- Reconhecer o padrão do sistema hidrotermal que afetou as fácies graníticas;
- Determinar os principais parâmetros físico-químicos do sistema mineralizador aurífero;
- Identificar o comportamento dos fluidos que afetaram os corpos graníticos;
- Identificar as fontes dos metais e dos fluidos responsáveis pela alteração hidrotermal e mineralização do depósito;
- Propor um modelo metalogenético para o depósito.

1.2. Métodos de Pesquisa

Para alcançar os objetivos da pesquisa, foram utilizados os seguintes métodos:

1.2.1. Etapa preliminar

Essa etapa compreendeu o levantamento e a revisão bibliográfica dos dados disponíveis na literatura sobre a geologia regional e local, englobando a área de estudo, tanto para a confecção do pré-projeto, nas disciplinas realizadas no Programa de Pós-Graduação em Geologia, do Instituto de Geociências da Universidade de Brasília (IG – UnB), como na preparação para as atividades de campo e as atividades de laboratório.

1.2.2. Trabalhos de campo e de petrografia

Foram realizadas três campanhas de campo na região do município de Novo Mundo e em seu entorno para a melhor compreensão geológica da área de estudo, bem como para a coleta de amostras mais representativas.

Aproximadamente 20 afloramentos foram analisados e dos quais foram coletadas amostras de mão representativas dos corpos graníticos e dos corpos máficos. Descreveram-se 15 furos de sondagem cedidos pela empresa de mineração Nexa Resources e Bemisa S.A., e destes foram coletadas amostras das rochas encaixantes e das rochas hidrotermalizadas e mineralizadas para análises. Quarenta seções delgadas polidas foram produzidas para o estudo de rochas frescas/inalteradas, além de 60 lâminas delgadas polidas, 6 lâminas delgadas bipolidas e 5 seções polidas para o tratamento das rochas não alteradas e hidrotermalizadas/mineralizadas do Complexo Novo Mundo e depósito Raimunda, respectivamente.

As lâminas e seções polidas foram confeccionadas no Laboratório de Laminação do Instituto de Geociências da Universidade de Brasília (LL–UnB); oito lâminas delgadas polidas do minério e seis lâminas delgadas bipolidas foram confeccionada no Laboratório de Laminação da empresa Petrografia BR (Contagem–MG).

Ademais, as descrições petrográficas foram realizadas no Laboratório de Microscopia do Instituto de Geociências da UnB (Mpós-UnB), utilizando microscópio petrográfico Olympus BX-50, bem como no Laboratório de Inclusões Fluidas da UnB (LIF-UnB), utilizando o microscópio petrográfico Olympus BX-50.

1.2.3. Litogeoquímica

33 amostras de rochas foram preparadas e analisadas no Laboratório Global da ALS em Goiânia (GO, Brasil). Os óxidos de elementos maiores (SiO_2 , TiO_2 , Al_2O_3 , Fe_2O_3 , MnO , MgO , CaO , Na_2O , K_2O , P_2O_5 , SrO , BaO) foram analisados por *Induced Coupled Plasma – Atomic Emission Spectrometry* (ICP-AES) e os elementos traço, incluindo as terras raras (Ba, Sr, Rb, Zr, Y, Nb, Ga, Th, Cs, Hf, Ta, U, V, Cr, Sn, As, Cu, Ni, Pb, Sc, Zn, La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu), por *Induced Coupled Plasma – Mass Spectrometry* (ICP-MS), após fusão com metaborato de lítio. Os metais base (Ag, Cd, Co, Cu, Li, Mo, Ni, Pb, Sc, Zn) foram analisados via ICP-AES após digestão ácida. A qualidade e a precisão da análise foram controladas por análises repetidas de brancos (abaixo do limite mínimo de detecção para cada elemento). Os diagramas geoquímicos para o tratamento dos dados foram elaborados utilizando o *software* GCDkit 4.0 (Janousek *et al.* 2006).

1.2.4. Microsonda Eletrônica

As análises de química mineral foram realizadas no Laboratório de Microsonda Eletrônica da UnB (LME-UnB). Análises químicas pontuais foram obtidas usando um microanalisador de sonda eletrônica JEOL® JXA-8230 (EPMA), operando com tensão de aceleração de 20 kV e 40 atm. Para a análise dos cristais de anfibólio, biotita, feldspato potássico, plagioclásio, muscovita e titanita, o sistema dispersivo de comprimento de onda foi configurado para operar com tensão de aceleração de 15 kV, corrente de feixe de 10 nA e tempo de contagem de 10 s. Os padrões de calibração incluíram minerais naturais e sintéticos, quais sejam: albita (Na), forsterita (Mg), ortoclásio (Al, Si, K), topázio (F), vanadinita (Cl), apatita (P), andradita (Ca, Fe), pirofanita (Ti, Mn), eskolaíta (trióxido de cromo – Cr), barita (Ba), óxido de Ni (NiO) e vanadinita (V). Para Ti e V, linhas $k\alpha$ foram empregadas

Para as análises químicas pontuais em cristais de clorita e muscovita, o sistema dispersivo de comprimento de onda (SDO) foi configurado para operar com tensão de aceleração de 15kV, corrente de feixe de 10nA e tempo de contagem de 10s. Os padrões utilizados para a calibração foram os dos minerais naturais e sintéticos, quais sejam: albita (Na), forsterita (Mg), ortoclásio (Al, Si, K), topázio (F), vanadinita (Cl), apatita (P), andradita (Ca, Fe), pirofanita (Ti, Mn), eskolaíta (trióxido de cromo-Cr), barita (Ba), óxido de Ni (NiO) e vanadinita (V). Para Ti e V, linhas $k\alpha$ foram empregadas.

Para os sulfetos e o ouro, as análises químicas foram obtidas utilizando tensão de aceleração de 20 kV, corrente de feixe de 20 nA e tempo de contagem de 10s, exceto para o selênio (25s); o tempo de contagem dos padrões foi metade do tempo de contagem utilizado nos picos. As linhas $K\alpha$ de raios X foram utilizadas para o enxofre, selênio, cobre, ferro, níquel e cobalto; $L\alpha$ para arsênio, prata, antimônio, paládio, cádmio, telúrio e molibdenio; e $M\alpha$ para o chumbo, mercúrio, ouro e platina; para o bismuto, foram utilizados $M\alpha$ e $M\beta$. Os minerais naturais utilizados como padrões em ambas as sondas eletrônica foram a pirita (Fe, S), a calcopirita (Cu), a galena (Pb) e a estibnita (Sb), e os elementos nativos ouro, prata, platina, paládio, bismuto, cobre e antimônio. Os padrões sintéticos foram o arsenieto de gálio (As), o seleneto de zinco (Zn, Se) e o cinábrio (Hg). Foram obtidos os seguintes limites de detecção (em % em peso): 0,018 para o ouro, 0,012 para a prata, 0,008 para o paládio, 0,016 para a platina, 0,032 para o bismuto, 0,025 para o chumbo, 0,016 para o cobre, 0,013 para o ferro, 0,017 para o mercúrio, 0,025 para o arsênio, 0,009 para o antimônio, 0,017 para o selênio, 0,005 para o enxofre, 0,012 para o cádmio, 0,013 para o níquel e o cobalto, 0,017 para o telúrio e 0,030 para o molibdênio.

1.2.5. Geologia isotópica

1.2.5.1. Isótopos radiogênicos

1.2.5.1.1. Método U – Pb em zircão.

Quatro amostras representativas de granito foram selecionadas para análises de zircão U–Pb. Duas amostras foram preparadas no Laboratório de Estudos Geodinâmicos, Geocronologia e Ambientais do Instituto de Geociências da UnB (LEGGA–IG–UnB) e outras duas amostras de granito, no Laboratório de Geoquímica Isotópica da Universidade Federal do Ouro Preto (LGI-UFOP). As amostras foram trituradas mecanicamente e posteriormente peneiradas. A fração menor do que 700 μ m foi peneirada para obter uma separação de minerais pesados, que foi submetida a um separador magnético Franz regulado para 1,5 A.

Os cristais de zircão foram extraídos usando técnicas convencionais de separação gravimétrica e magnética e escolhidos manualmente sob um microscópio binocular para obter frações de tamanho, formato e cor semelhantes. Tanto no Laboratório de Geocronologia da UnB quanto no Laboratório de Geoquímica Isotópica da UFOP, foi

utilizado o equipamento *Thermo Scientific Neptune Plus* ICP-MS para a geocronologia U – Pb.

As imagens de catodoluminescência foram obtidas em Microscópio Eletrônico de Varredura (MEV) JEOL 6510. O padrão aplicado foi o Zircon GJ-1 ([Jackson et al. 2004](#)). A plotagem dos dados U–Pb e os cálculos de idade foram realizados usando ISOPLOT 4.1 ([Ludwig 2003](#)), e os erros para razões isotópicas são apresentados no nível 1σ . ^{238}U , ^{232}Th , ^{206}Pb e ^{208}Pb foram medidos em copos de Faraday, e ^{204}Pb , ^{207}Pb e ^{202}Hg foram coletados em contadores de íons.

1.2.5.1.2. Método Sm – Nd em rocha total

Dados isotópicos de Sm–Nd foram obtidos a partir de nove amostras representativas do Complexo Novo Mundo, utilizando a metodologia de [Gioia e Pimentel \(2000\)](#). As amostras foram processadas no LEGGA–IG–UnB, onde foram pulverizadas em moinho vibratório com cadinho de metal duro e analisadas em espectrômetro de massa de ionização térmica (TIMS) em Multicoletor TRITONTM. Polpas de rocha foram inicialmente misturadas com solução de espigão ^{149}Sm – ^{150}Nd e dissolvidas em cápsulas Savillex, usando técnicas convencionais de troca catiônica com colunas de Teflon contendo resina LN-Spec (HDEHP – ácido fosfórico dietilhexil suportado em pó de PTFE). Uma vez extraídas, as amostras de Sm e Nd foram carregadas no filamento Re do conjunto de filamento duplo. A razão $^{143}\text{Nd}/^{144}\text{Nd}$ foi normalizada usando $^{146}\text{Nd}/^{144}\text{Nd} = 0,7219$, e a constante de decaimento foi $6,54 \times 10^{-12}/\text{y}$. As incertezas para as relações $^{147}\text{Sm}/^{144}\text{Nd}$ e $^{143}\text{Nd}/^{144}\text{Nd}$ são inferiores a 0,2% (2σ) e $\pm 0,003\%$ (2σ), respectivamente. Os resultados obtidos foram tratados no ISOPLOT 4.15 e as idades de residência crustal (idade modelo Nd, T_{DM}) foram calculadas usando o modelo de [De Paolo \(1981\)](#).

1.2.6 Inclusões fluidas

O estudo de inclusões fluidas foi baseado na análise de seis lâminas bipolidas (película com ~120 mm de espessura) de seis diferentes amostras mineralizadas, em paragênese com as principais fases de mineralização do depósito Raimunda. A petrografia das inclusões foi realizada em temperatura ambiente ($\pm 25\text{ }^{\circ}\text{C}$), em microscópio petrográfico do modelo Zeiss Axioplan 2 (Zeiss, Oberkochen, Germany), aplicando os

critérios presentes em Roedder (1984), Shepherd *et al.* (1985) e Van den Kerkhof e Hein (2001).

Os dados microtermométricos foram obtidos a partir do resfriamento e aquecimento dos fragmentos das rochas em platina LINKAM THMSG 600, com intervalo de temperatura entre 200 °C e 600 °C, acoplada ao microscópio petrográfico, com objetivas de aumento de 10x a 100x, presente no LIF-UnB, seguindo os procedimentos de Shepherd *et al.* (1985). A calibração foi realizada usando inclusões fluidas bifásicas e trifásicas sintéticas de H₂O–NaCl e H₂O–CO₂, aplicando taxas de velocidades de 2 a 15°C/min, com acurácia estimada em $\pm 0,5$ °C, para o resfriamento (+25 a -150 °C), e ± 2 °C, para o aquecimento (acima de 350 °C).

1.2.7. Geotermometria da clorita

As condições de temperatura e pressão de formação do depósito Raimunda foram complementadas com o geotermômetro independente da clorita. As temperaturas foram baseadas na proporção catiônica do alumínio tetraédrico da clorita (Al^{IV}), que permite calcular a temperatura de cristalização por meio da equação: $T (^{\circ}C) = 321,98 * (Al^{IV} / 2) - 61,92$ (Cathelineau & Neiva (1985) e Cathelineau (1988)). 18 análises de clorita em lâminas polidas e previamente metalizadas com carbono foram realizadas no LME-UnB, em microsonda do modelo JEOL® JXA-8230 (EPMA), cujas condições operacionais de WDS foram: tensão de aceleração de 15 Kv, corrente de feixe de 10nA e tempo de contagem de 10s.

1.2.8 Espectroscopia Raman

As análises de espectrometria Raman em inclusões fluidas foram desenvolvidas no Laboratório Raman, no Instituto de Física, da Universidade de Brasília, usando Micro Raman Horiba HR, equipado com um controle de detecção CCD, com laser de argônio de 514,5 μm e 130 mW de intensidade, acoplado a um microscópio do modelo Olympus BX41, com objetivas de 100x. As fases voláteis foram analisadas utilizando o tempo de contagem de 30 ou 60 s, repetição de 2x, com variância espectral de 100 a 4000 cm^{-1} e raio laser de 100 a 300 μm .

1.2.9 Isótopos estáveis de oxigênio (O) e enxofre (S)

O estudo de isótopos estáveis foi realizado em minerais de rochas mineralizadas representativas do depósito Raimunda. A seleção e o tratamento das amostras ocorreram no Laboratório de Preparação de Amostras da Universidade de Brasília (LPA-UnB); já os concentrados minerais puros foram obtidos por separação manual em microscópio binocular no LME-UnB.

As análises foram realizadas no laboratório *Queen's Facility for Isotope Research* (*Queen's University*, Ontário, Canadá). Os resultados são apresentados em per mil (‰), usando a notação delta (δ) e os padrões de referência V-SMOW (*Vienna Standard Mean Ocean Water*), para o oxigênio, e V-CDT (*Canyon Diablo Troilite*), para o enxofre.

Os dados de isótopos de enxofre foram obtidos a partir de seis amostras de pirita, contendo de 80 a 95mg cada. Para esta análise, o SO₂ produzido pela combustão de cerca de 0,5 mg de amostra é carregado em cápsulas de estanho reagindo a 1.080 °C. Os gases foram transportados em corrente e reduzidos pelo contato com fios de cobre de alta pureza, e o SO₂ foi separado por cromatografia gasosa. As análises foram realizadas pelo EA-IRMS (*Elemental Analysis and Isotope Ratio Mass Spectrometry*), usando um *Flash Elemental Analyzer* e um espectrômetro de massa *ThermoFisher MAT 253*. As incertezas são estimadas em 0,2‰.

Para os isótopos de oxigênio, foram analisadas seis amostras de minerais de quartzo, entre 20 e 28 mg cada, utilizando o espectrômetro de massa *Thermo Finnigan Delta V Advantage* e *Finnigan Delta XP*. O protocolo analítico aplicou o método de fluoração a laser, no qual o oxigênio é liberado por reação com BrF₅, usando laser de CO₂ como fonte de calor e convertido em CO₂ por reação com grafite a 750 °C, ou é coletado como O₂ em uma peneira molecular e analisado como O₂. As incertezas para $\delta^{18}\text{O}$ foram melhores que 0,1‰ e 0,16 ‰.

1.2.10. Etapa de tratamento e sistematização dos dados

Compreende o tratamento e a integração dos dados obtidos na pesquisa, incluindo:

- a) Elaboração da tese de doutorado;
- b) Publicação de artigos em periódicos especializados e de ampla abrangência.

1.3. Localização e vias de acesso

A área de estudo está localizada na porção extremo-norte do estado de Mato Grosso, a oeste da cidade de Guarantã do Norte, no município de Novo Mundo (Fig. 1.2).

O acesso à área, a partir de Cuiabá, é realizado pela rodovia BR-163 (Cuiabá–Santarém) no sentido N-S, por aproximadamente 700 km, passando pelos municípios de Nova Mutum, Sinop, Peixoto de Azevedo até Guarantã do Norte, onde segue-se pela rodovia MT – 419, por cerca de 65 km no sentido L-W, em que o acesso ao município de Novo Mundo e à área de estudo é realizado por estradas vicinais.

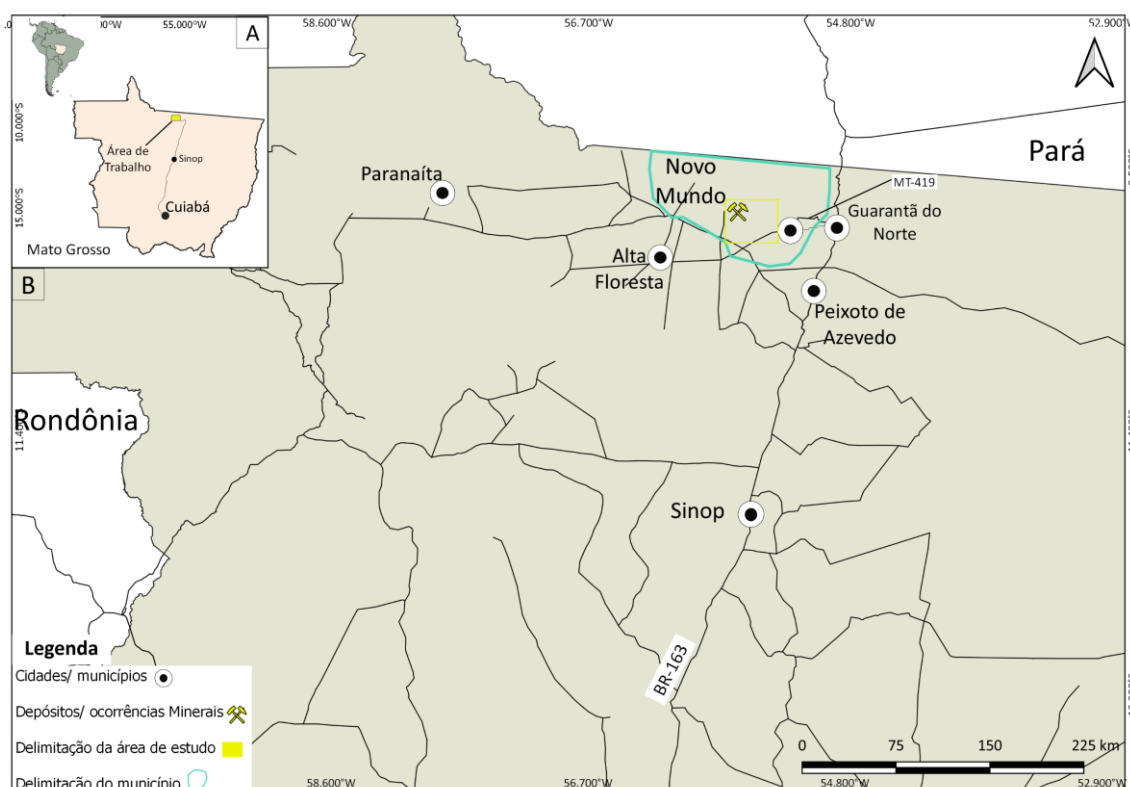


Figura 1. 2. Mapa de localização da área de trabalho, extremo norte do estado de Mato Grosso, Brasil. B) Município de Novo Mundo e a área de trabalho delimitada pela retângulo amarelo na figura, com destaque para as principais cidades e vias de acesso da região.

CAPÍTULO 2

DINÂMICA DE SISTEMAS HIDROTERMAIS GENETICAMENTE ASSOCIADOS A GRANITOS MINERALIZADOS EM OURO

2. DINÂMICA DE SISTEMAS HIDROTERMAIS GENETICAMENTE ASSOCIADOS A GRANITOS MINERALIZADOS EM OURO

2.1 Introdução

Muitos depósitos minerais são formados dentro ou próximos a intrusões magmáticas félsicas e intermediárias, contemplando a maioria dos sistemas hidrotermais relacionados à intrusão ([Pirajno 2009](#)).

Esses depósitos resultam das complexas interações entre magmas com contrastes químicos e físicos. Os fluidos exsolvidos são enriquecidos em H₂O, SO₂, CO₂, H₂S, Cl⁻, Cu, Au, W, Sn e metais básicos, formando concentrações anômalas de minérios. A remoção dos elementos com maior energia de ativação do magma máfico para o félsico (W, Sn, Zr) ocasiona grandes mudanças isotópicas, como a diminuição considerável nos valores de δD , observada em vários depósitos relacionados a magmas félsicos e intermediários ([Hedenquist & Lowenstern 1994](#); [Vigneresse 2007](#); [Kouzmanov & Pokrovsk 2012](#); [Richards 2021](#)).

Variações químicas podem contribuir para a diferenciação e cristalização fracionada no magma, e os processos que levam à formação de depósitos minerais consideram sua origem, os mecanismos de colocação na crosta, os processos de mistura, a distribuição de elementos controlada pela difusão e a partição entre os minerais e a fase fluida, a solubilidade e o estado redox. As variações desses parâmetros são controladas pela temperatura do sistema ([Yang 2012](#); [Heinrich 2024](#)).

Dados estruturais, químicos e isotópicos, além de observações de campo, mostram que os corpos ígneos são formados a partir de uma sucessão de *inputs* magmáticos com assinaturas geoquímicas diferentes.

Mistura de magmas de composição química semelhantes em uma câmara magmática causa poucos efeitos na difusão, no coeficiente de partição e nas condições redox, devido ao baixo contraste de temperatura. No entanto, a mistura com magma máfico induz à variação de até 300 °C em ambos os magmas, e os demais parâmetros são fortemente alterados. Além disso, a adição ou a diluição do enxofre influenciará diretamente o enriquecimento ou não de metais no fluido para a posterior formação de depósitos ([Audétat & Simon 2012](#)).

Este capítulo abordará alguns dos principais parâmetros físico-químicos atuantes

na formação de depósitos minerais a partir do magma (fugacidade de oxigênio (fO_2), condições de acidez e alcalinidade (pH), solubilidade de enxofre e influência da temperatura e pressão), na tentativa de melhor compreender suas influências à formação de depósitos de ouro em magmas graníticos; além de discorrer sobre as principais técnicas para a obtenção desses dados; comparar magmas com estado redox distintos e suas características para a mineralização de ouro.

2.2 Parâmetros físico-químicos de sistemas hidrotermais geneticamente associados a granitos mineralizados em Au

2.2.1 Fugacidade de oxigênio

A fugacidade de oxigênio (fO_2) é um dos parâmetros mais importantes na evolução do magma e para a formação de depósitos minerais, pois altera o estado de valência e estabilidade dos elementos (Yardley & Bodnar 2014; Sun *et al.* 2015).

Esse parâmetro corresponde à pressão parcial de oxigênio em determinado ambiente (atmosfera, magma, rocha etc.), geralmente representado como variações em relação a tampões de oxigênio, os quais são uma assembleia de minerais e/ou compostos que restringem a fO_2 em função da temperatura (T) e da pressão (P) (Sun *et al.* 2015).

Os tampões utilizados em condições de altas T e P para avaliar o estado de oxidação de magmas e rochas são: faialita-magnetita-oxigênio (FMQ), Co-CoO, níquel – óxido de níquel (NNO), hematita – magnetita (HM) e MnO – Mn₃O₄ (Sun *et al.* 2015).

A solubilidade de componentes no *melt* magmático (S, composição – FeO, CaO) depende diretamente da fO_2 . Em condições abaixo do tampão FMQ +1, o estado dominante de oxidação do enxofre é 2-, e a fase de saturação pirrotita. Em condições acima de FMQ +1, o dominante é o S⁶⁺, predominando a anidrita e barita. Esse comportamento influencia não somente a solubilidade e eficiência de extração do S por processos de fusão parcial, como também as condições de pH do sistema (Audétat & Simon 2012a; Yang 2012; Sun *et al.* 2013b).

O comportamento geoquímico dos elementos durante a evolução do magma e/ou processos hidrotermais também é afetado pela fO_2 , normalmente plotado em unidades logarítmicas de pressão, sendo NNO e FMQ os principais tampões para representá-los (Yardley & Bodnar 2014; Vigneresse 2007).

Em temperaturas entre 700 °C e 1000 °C, a fO_2 aumenta com a temperatura; granitos tipo – I, com magnetita, estão próximos à linha SO_2 – H_2O do campo da oxidação. Durante a descompressão do magma, a fugacidade alcança valores da ordem de -15 a -11, próximo à curva hematita/magnetita. Por outro lado, granitos tipo–S mostram condições redutoras. Com a diminuição da fO_2 , a razão $Fe^{2+}/(Fe^{2+} + Fe^{3+})$ aumenta com o grau de polimerização, permitindo traçar condições logarítmicas representativas somente da razão $Fe^{2+}/(Fe^{2+} + Fe^{3+})$ (Fig. 2.1) (Vigneresse 2007).

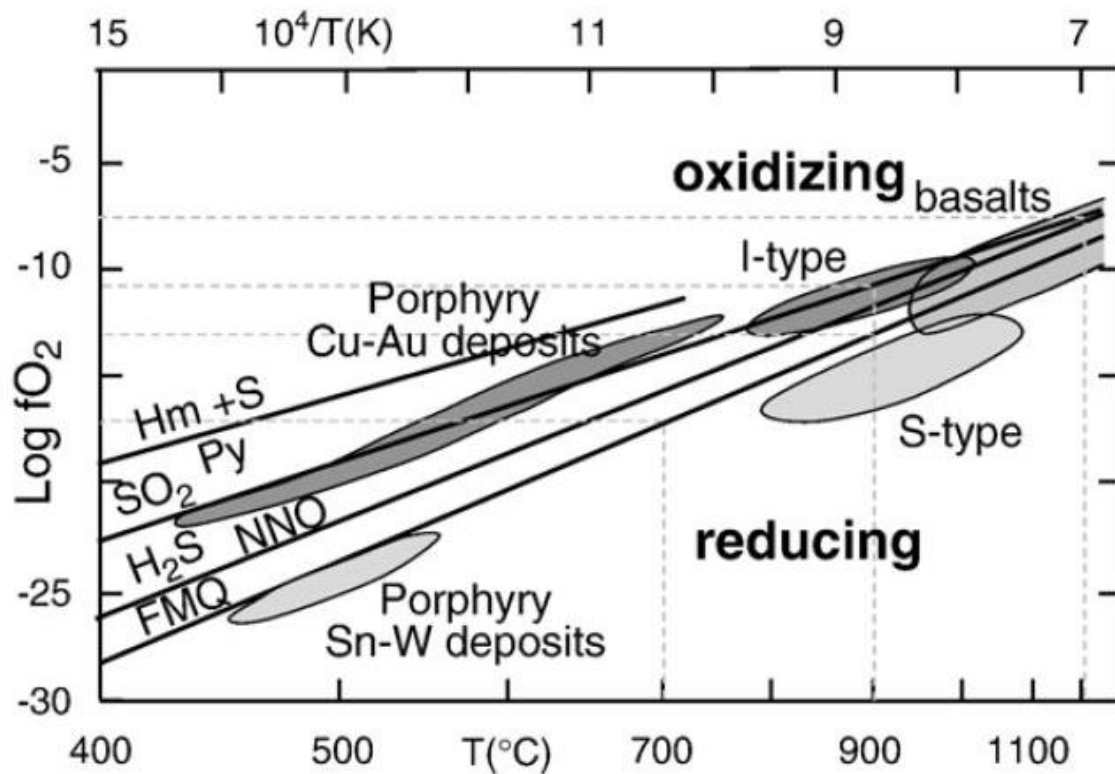


Figura 2. 1. Figura mostrando a correção entre a pressão parcial da fO_2 vs. T (°C) de alguns dos depósitos minerais e tipos de granito em ambiente reduzido e oxidado. Os tampões mais utilizados são o NNO, FMQ e magnetita-hematita (MH) (Vigneresse 2007).

Outro momento importante é o início da cristalização da magnetita, a qual impulsiona a redução do sulfato para sulfeto, o que tende a aumentar a razão $Fe^{2+}/(Fe^{2+} + Fe^{3+})$. Presumivelmente, há flutuações na fO_2 , sendo frequentemente tomadas como o processo controlador para a mineralização Au em diversos depósitos minerais, pois a fase reduzida do S carrega o metal do magma para a fase fluida (Liang *et al.* 2009a; Sun *et al.* 2013a).

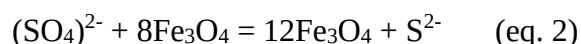
2.2.2 pH

Em sistemas mineralizantes, a redução de sulfato para sulfeto é iniciada pela cristalização de magnetita, acompanhada pela diminuição do pH (liberação de H^+ e diminuição de O_2 no sistema) e pelo aumento do potencial de oxidação do sulfato (Sun *et al.* 2013b; Vigneresse 2019).

Estima-se que a quantidade de H^+ liberada durante a redução do sulfato e a oxidação do Fe^{2+} podem rebaixar o pH para valores abaixo de +2, dependendo da temperatura e da ausência de tampões do pH no sistema, para elevar significativamente a fO_2 (Sun *et al.* 2014a, b), oxidar a magnetita ou Fe^{2+} em fluidos hidrotermais e formar hematita e/ou especularita, conforme as equações abaixo:



Oxidação da magnetita expressa na fórmula a seguir:



A formação da hematita hidrotermal ajudará no rebaixamento do pH, mas a oxidação adicional da magnetita não será modificada. A hematita se cristaliza, uma vez que a redução do sulfato diminui o pH o suficiente para conduzir o potencial de oxidação desta espécie de enxofre até o tampão de fugacidade de oxigênio hematita-magnetita (HM), que corresponde a $\sim \Delta FMQ + 4$.

Assim, o potencial de oxidação do sulfato depende fortemente do pH. A oxidação do ferro ferroso (Fe^{2+}) pelo sulfato, de acordo com a equação (1), leva à diminuição do pH dos fluidos mineralizadores de alguns sistemas graníticos, por exemplo, elevando o potencial de oxidação dos sulfatos. Consequentemente, a fO_2 aumenta durante a redução do sulfato (Fig. 2.2) (Pokrovski & Dubrovinsky 2011; Sun *et al.* 2013b).

2.2.3 Solubilidade do enxofre (S) no fluido

Estudos experimentais mostram que a solubilidade de S em *melts* silicáticos é controlada pela T, P, composição, condições redox e difusão do S no magma (Yang 2012), sendo o principal geossolvente dos elementos calcófilos nesses sistemas, com forte influência durante o processo mineralizador de Au.

A solubilidade do enxofre é inversamente proporcional à pressão, ou seja, à medida que o magma ascende em direção à superfície ($<P$), a solubilidade aumenta (Vigneresse 2007; Yang 2012). O ouro é um elemento calcófilo que está em baixa

abundância na crosta e, assim como cobre (Cu) e molibdênio (Mo), atua como incompatível em magmas subsaturados em sulfeto (Sun *et al.* 2012) e apresenta altos valores de coeficiente de partição entre sulfetos e o *melt* silicático ($D_{Au} = 4500 - 11.200$, Mungall & Brenan 2014).

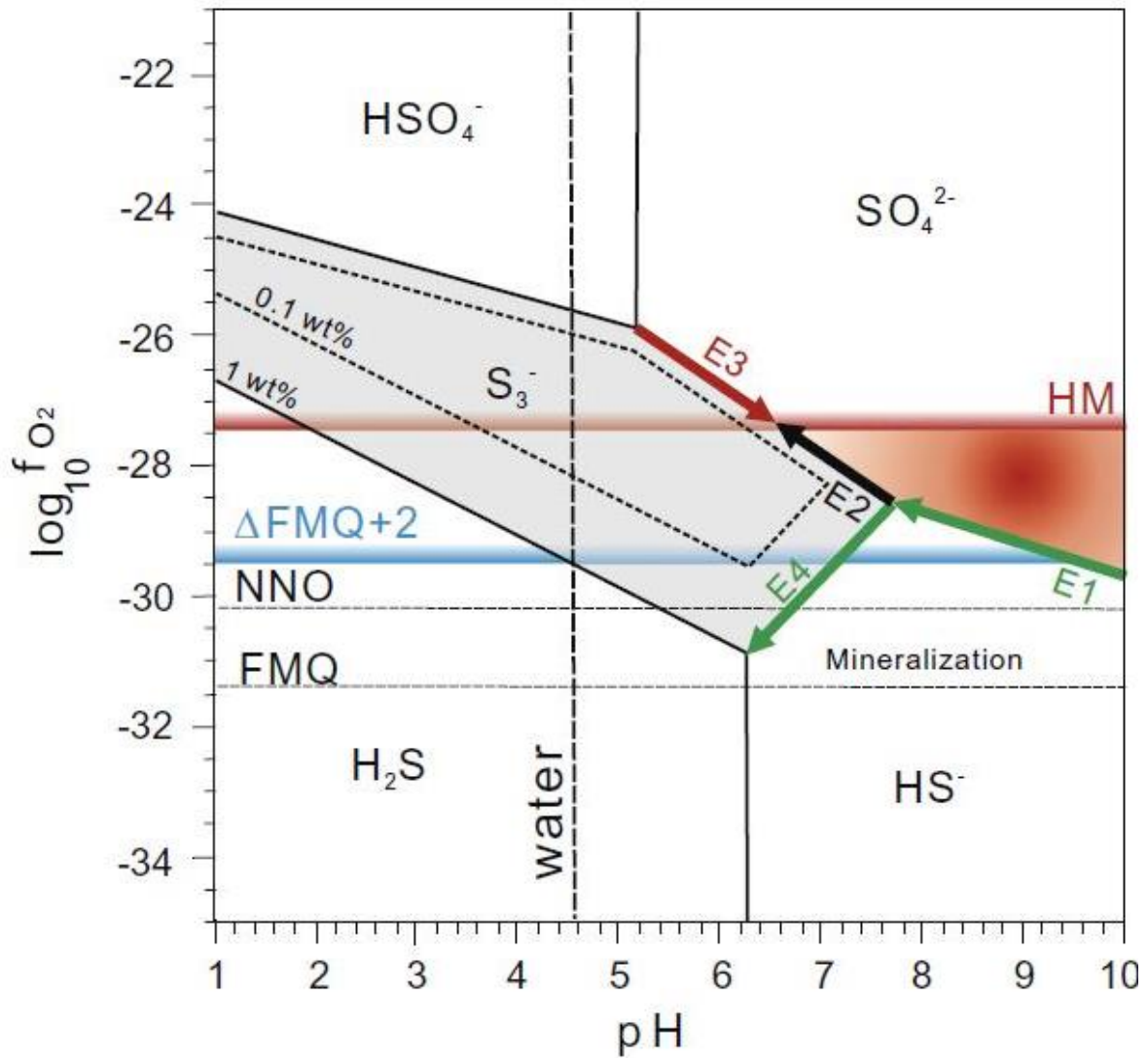


Figura 2. Domínios de estabilidade do íon trissulfeto (S_3^-), sulfato e sulfeto em solução aquosa em função da fugacidade de oxigênio ($\log_{10}f_{O_2}$) e acidez ($pH = -\log_{10}mH^+$, em mol/kg) a 350 °C e 0,5 GPa, ilustrando a redução do sulfato (modificado de Pokrovski & Dubrovinsky (2011) e Sun *et al.* (2013a, b) por Sun *et al.* (2015)). Observa-se a f_{O_2} dos tampões minerais HM, NNO e FMQ e o ponto de neutralidade da água (linha vertical tracejada). O campo laranja entre o HM e o $\Delta FMQ + 2$ é a condição ótima para a mineralização porfírica Cu + Au + Mo (Sun *et al.* 2013b). As linhas E1, E2, E3 e E4 mostram trajetórias para a redução do sulfato (Sun *et al.* 2013b). Domínios de estabilidade do íon S_3^- nas concentrações de 0,1% e 1% de enxofre total dissolvido.

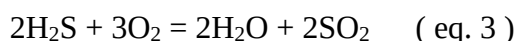
Com isso, a concentração desse elemento é principalmente governada pelo enxofre reduzido, como, por exemplo, sulfetos, complexos hidrossulfetados (Sun *et al.* 2004a) ou complexos polissulfetados (S_2^{2-} , S_3^-) (Pokrovski & Dubrovinsky 2011; Sun *et al.* 2013b).

Durante a evolução do magma, o sulfeto permanece subsaturado; o sulfato (SO_4^{2-}) é a espécie dominante, alcançando valores de solubilidade até 10 vezes maiores do que o da espécie S^{2-} (Jugo 2009; Sun *et al.* 2013b). A extração de S e de elementos calcófilos é mais eficiente em condições oxidantes acima do tampão FMQ (faialita-magnetita-quartzo), por meio do processo de fusão parcial de uma rocha fonte, e o processo de mineralização de magmas graníticos, nessas condições, dá-se pela redução do S, acompanhado pela redução do pH (Yang 2012; Sun *et al.* 2013b).

Todavia, pequenas quantidades de sulfeto residual na rocha fonte são suficientes para reter grandes quantidades dos elementos calcófilos, não permitindo que eles sejam concentrados na fase fluida e, portanto, não sejam exsolvidos do magma, diminuindo, assim, a possibilidade de formar concentrações anômalas de minério (Yang 2012).

Alguns estudos isotópicos consideram os valores de $\delta^{34}S$ como o S total da amostra, porém, estudos outros analisam o $\delta^{34}S$ de sulfetos em paragênese com a mineralização para resultados mais precisos. A maioria dos depósitos magmático-hidrotermais apresenta valores de $\delta^{34}S$ positivos, entre +5‰ e +20‰, dependendo do mineral analisado (Vigneresse 2007).

Autores como Kelley & Cottrell (2012) propõem que a degaseificação de SO_2 também é um fator que deve ser considerado no rebaixamento da fO_2 do magma. Como ilustrado na equação (3), esse processo consome oxigênio, rebaixando a fO_2 do sistema (Sun *et al.* 2015). Entretanto, não ocorre a redução de sulfato para sulfeto no magma; em vez disso, dirige a reação para a direita, a qual leva à oxidação do sulfeto.



Já para a redução do S, a degaseificação de CO_2 pode ser a principal responsável pela redução de sulfato para sulfeto, por meio do processo descrito na equação abaixo:



2.2.4 Temperatura e pressão

Variações nos sistemas graníticos são resultado de vários fatores, entre eles, a rápida oscilação de temperatura (T) e de pressão (P). A variação de parâmetros como

difusão, partição entre mineral/*melt* (K), solubilidade e condições redox é observada considerando a energia de ativação que controla, de forma exponencial, a temperatura dos magmas (Vigneresse 2007; Yang 2012).

No caso de intrusão magmática em câmara magmática contendo composição félsica, semelhante em composição ao que já estava na câmara, a interação resultaria em poucos efeitos sobre a difusão, o coeficiente de partição e o estado redox do magma, devido ao baixo contraste de temperatura resultante da mistura.

No entanto, a intrusão de magma de composição máfica induz à variação de ~ 300 °C em ambos. O comprimento da difusão pode variar em até 2 ordens de magnitude e o coeficiente de partição, em 1 ordem de magnitude. As condições redox correspondem a 2.5 log no magma máfico, mas podem variar até 7 unidades log no magma félsico (Vigneresse 2007).

Em relação ao coeficiente de partição de elementos e polímeros, T e P também são determinantes. O K é uma propriedade física dos elementos em que há a tendência destes se concentrarem na porção magmática (mineral) ou na parte fluida do sistema, a depender, especialmente, do raio iônico e do estado de valência dos elementos analisados.

Entretanto, em fases fluidas e em determinadas composições de magma, e dependendo do stress e do calor envolvidos, há maior facilidade para que um elemento incompatível ($K < 1$) possa se comportar como compatível ($K > 1$) (Vigneresse 2007).

O transporte e deposição do Au, porém, podem ocorrer em ampla margem de temperatura e pressão, levando-se em consideração a composição do magma e do fluido e o ambiente tectônico de formação (Seward 1991).

A pressão é um parâmetro na formação e evolução de um magma e de um depósito mineral que oscila consideravelmente. Não influencia, de forma significativa, a difusão no sistema, porém, a descompressão é relevante quando se trata das interações das fases magmática e fluida, levando a processos classificados como *boiling* (Vigneresse 2007).

Essa pressão tem influência semelhante, de forma diretamente proporcional, na condição de solubilidade da água no *melt* magmático. O enxofre em sua forma reduzida é menos solúvel em altas pressões, implicando o aumento da solubilidade de espécies reduzidas e complexantes de Au à medida que a pressão é reduzida ou que o fluido mineralizante alcança menores níveis de profundidade na crosta (Sun *et al.* 2015).

2.3 Mineralizações de Au relacionada a granitos reduzidos e oxidados

Além de serem distinguidos pela quantidade de magnetita, assembleia mineralógica e enxofre disponível no fluido magmático, os granitos reduzidos e oxidados também diferem em suas composições de elementos metálicos de transição, os quais têm afinidade com a fase volátil do fluido, resultando em associações favoráveis à formação de depósitos minerais (Fig. 2.3) (Takagi & Tsukimura 1997).

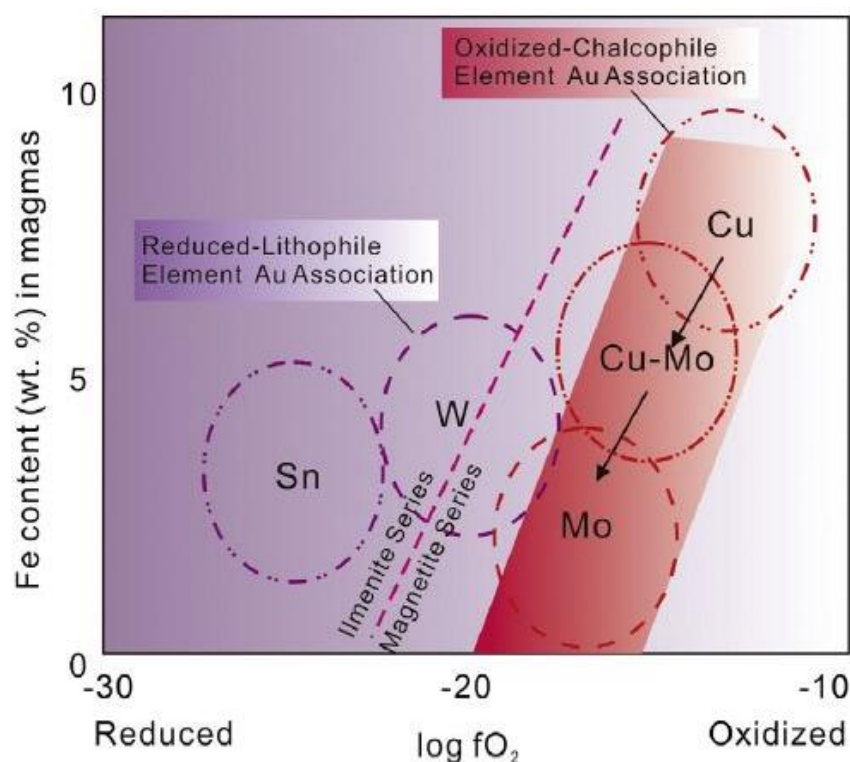


Figura 2. 3. Gráfico esquemático do grau de fracionamento do conteúdo de Fe em magmas vs. fO_2 para magmas cálcio-alcalinos e ricos em Au-Cu-Mo-W. A linha limite do conteúdo entre magmas da série da magnetita e ilmenita corresponde ao exemplo de Ishihara (1997) (Thompson *et al.* 1999).

Magmas oxidados são classicamente observados em zonas de subducção ou em cinturões magmáticos subductados; são relacionados a granitos da série da magnetita (Ishihara 1977), contendo outros minerais como sulfetos, hematita, apatitas com alto SO_3^{2-} e anidrita e outros sulfatos (Santacruz *et al.* 2021).

Ademais, os magmas apresentam alta fugacidade de oxigênio (2 a 3,4; log acima do tampão FMQ), variando entre tampões HM e NNO, a maioria excedendo NNO^{+2} (Richards 2003). O alto oxigênio garante que o sulfato seja a espécie principal de S no

fundido, aumentando significativamente a solubilidade do S no *melt* (Yang 2012; Jugo 2009; Vigneresse 2019).

Esses granitos classificados como do tipo-I (Willian 2017) têm > 0.2% de magnetita e estão próximos à linha $\text{SO}_2\text{--H}_2\text{S}$ no campo de oxidação. Com a diminuição da pressão, a fugacidade pode chegar a valores entre -15 e -11, próximos à curva hematita/magnetita, (H/M) e contém valores superiores a 250 ppm de SO_2 e $f\text{O}_2$ balizado pelo tampão $\text{SO}_2\text{--H}_2\text{S}$ (Vigneresse 2007).

Magmas metaluminosos apresentam mineralização condizente a magmas oxidados, provavelmente devido à alta quantidade de S no fluido. Neste contexto, muitos diagramas de fase relacionando $f\text{O}_2$ e $f\text{S}_2$, ou condições de pH, produzem campos de estabilidade da magnetita e hematita com aumento do valor da $f\text{O}_2$, passando para pirrotita, depois pirita, e estabilidade de Au-Cu, com valores crescentes de $f\text{S}_2$ ou decrescentes de pH (Vigneresse 2007).

No entanto, plutons que são associados a mineralizações de Au em condições reduzidas têm baixa concentração de sulfetos (~ 5 -1 vol.%) e respostas aeromagnéticas baixas (Hart e Goldfarb 2005). Outros critérios para classificar essas rochas são: 1 – falta de hematita primária e de sulfatos; 2 – abundância de pirrotita hipogênica e de CH_4 ; e 3 – os óxidos pertencem à série da ilmenita (Rowins 2000).

Granitos do tipo-S (Willian 2017) têm características redutoras. Em condições de diminuição da fugacidade, a razão $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ aumenta com o grau de polimerização, uma vez que Fe^{2+} e Fe^{3+} formam uma rede de modificações no magma (Ishihara 1997; Vigneresse 2007). Esses granitos, em geral, têm concentrações de SO_2 inferiores a 250 ppm e seus tampões mais comuns são $\text{H}_2\text{O--H}_2$, $\text{CO}_2\text{--CO}$ ou $\text{CH}_4\text{--CO}_2$. Nessas condições de oxirredução, a magnetita não cristaliza, as rochas apresentam altas concentrações de F e Cl, além da afinidade para concentrar Sn, W, Zn e B (Takagi & Tsukimura 1997).

A despeito das distinções físico-químicas e da maior facilidade de formação de depósitos ricos em Au a partir de magmas oxidados, existem diversos exemplos de depósitos de Au relacionados a magmas reduzidos, com $f\text{O}_2$ variando de FMQ (-0,5 a -3) e utilizando tio-complexos (e.g., $\text{H}_2\text{O--CO}_2\text{--CH}_4\text{.N}_2$) como os principais transportadores do minério (Rowins 2000; Yang *et al.* 2008; Smith *et al.* 2012; Li *et al.* 2019; Santacruz *et al.* 2021).

2.3.1 Sistemas mineralizadores pórfiros

As margens convergentes são ambientes geodinâmicos nos quais ocorrem processos de consumo de placa tectônica na litosfera terrestre e alguns constituintes da litosfera continental, e é onde voláteis da hidrosfera são reciclados no manto. Muitos desses componentes, incluindo 11% de voláteis, eventualmente desaparecem na imensidão do manto e podem reaparecer na superfície terrestre como componentes de plumas mantélicas (Richards 2021).

Os depósitos Au pórfiros formam-se comumente em arcos relacionados à subducção, em margem continental e em arcos de ilhas, bem como em ambientes de *backarc*, e uma minoria desses depósitos é encontrada em outros ambientes tectônicos que se formam após cessada a subducção (Sillitoe 2017; Richards 2021).

Este sistema tende a se desenvolver em ambientes sujeitos a um amplo espectro de regimes de *stress* moderadamente extensionais, desde a fase de deslizamentos oblíquos até a fase contracional. Sillitoe (2010), no entanto, chama atenção para a frequente associação espacial dos depósitos do tipo pórfiro com terrenos que estiveram sujeitos a espessamento crustal, soerguimentos e rápida exumação, fatores estes que caracterizam ambientes tectônicos fortemente contracionais.

Takada (1994) propôs que a compressão crustal auxilia o desenvolvimento de grandes câmaras magmáticas na crosta média a superior. Tosdal e Richards (2001) consideram que os momentos de mudança no regime de estresse crustal (de contracional para extensional) são especialmente favoráveis para a geração de depósitos do tipo pórfiro (Cu + Au).

Os depósitos do tipo pórfiro comumente ocorrem como aglomerados de depósitos equidimensionais, que definem faixas lineares paralelas ou transversais a arcos magmáticos, e, menos comumente, são encontrados como ocorrências isoladas (Sillitoe 2017). Formam-se a partir de sistemas magmático-hidrotermais e, geralmente, ocorrem associados a depósitos do tipo *skarn* e epitermais, com idades meso-cenozoicas, embora existam exemplos com idade desde o Arqueano (Sillitoe 2017; Heinrich 2024).

A categoria Cu-Au está relacionada a *stocks* de pórfiro de composição, principalmente, quartzo diorítica a granodiorítica, com uma minoria de depósitos ligada aos seus equivalentes alcalinos (Sillitoe 2017).

O estágio magmático do sistema pórfiro engloba sete etapas, segundo Heinrich (2024): 1– preparação da litosfera e segregação magmática; 2– fracionamento da crosta inferior para a geração de um magma fértil; 3– ascensão para o reservatório de magma

mineralizado; 4 – saturação do fluido e transferência de metais; 5– extração do fluido; 6 –alteração da rocha encaixante e deposição do minério; e 7– soerguimento, erosão e enriquecimento supergênico (Fig. 2.4). O processo magmático de formação dos depósitos do tipo pórfiro começa com a produção de magma no manto e finaliza com a colocação na crosta do corpo rochoso, quando este atinge seu estado *solidus*. Posteriormente, o processo continua com a exsolução de fluidos contendo metais, com o hidrotermalismo e deposição do minério (Audetat e Simon 2012; Heinrich 2024).

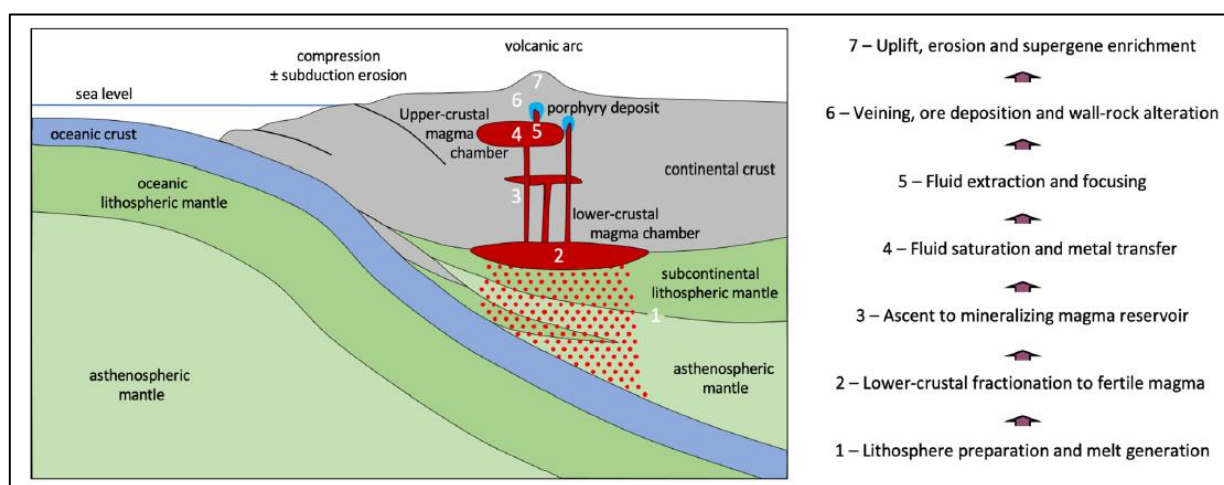


Figura 2. 4. Seção transversal de um arco magmático continental ilustrando a cadeia de processos que segue o percurso da mineralização desde o manto superior até a superfície terrestre (Heinrich 2024).

Os metais, para serem extraídos da cunha mantélica, necessitam de fO_2 2 unidades logarítmicas acima do tampão FMQ, o que acontece mediante fusão parcial do manto da *slab* oceânica ou a partir de um fluido supercrítico (Sun *et al.* 2015).

O enriquecimento em água e a saturação magmática em sulfetos são dois processos decisivos da crosta inferior para a diferenciação dos fundidos primários do manto em magmas formadores de minério (Heinrich 2024). No entanto, além do enriquecimento magmático, a formação de depósitos do tipo pórfiro depende da atuação eficiente e temporalmente colocada de processos de enriquecimento, passando pela profundidade de instalação do pluton na crosta (pressão hidrostática > pressão litostática), para que haja eficiência da exsolução dos fluidos magmáticos, interação com a rocha encaixante e fluidos bacinais e/ou fluidos meteóricos e, conseqüentemente, o devido trapeamento dos fluidos enriquecidos em minério na porção apical do pluton.

A precipitação do minério está associada, geralmente, à mistura de fluidos magmáticos (salmouras + vapores) com água meteórica, acompanhada de arrefecimento de fluidos, diminuição da salinidade e solubilidade retrógrada do quartzo à medida que as rochas transitam de condições litostáticas para hidrostáticas (Heinrich 2024; Sun *et al.* 2024).

2.4 Técnicas analíticas aplicadas ao estudo de depósitos minerais magmático-hidrotermais.

2.4.1 Inclusões fluidas

A formação de depósitos minerais requer a existência de um fluido com capacidade de transportar metais, e as características físico-químicas desses fluidos são variáveis. As inclusões fluidas encontradas em minerais preservam pequenas amostras de fluidos que estavam presentes durante processos geológicos (Bodnar 2003). Depois de formadas, quando trazidas à superfície e fora de suas condições naturais de aprisionamento, o fluido passa por mudanças de fases, formando bolhas de gás, causando imiscibilidade de líquidos, precipitando sólido de saturação ou de aprisionamento (Roedder 1972; Bodnar 2003).

A partição do metal a partir da fase fluida para a vapor, o que ocorre em condições de T e P distintas às ambiente, é descrita como um processo dinâmico controlado por inúmeros parâmetros: temperatura, pressão, composição do *melt*, condições redox e composição química da fase vapor (quantidade de S, salinidade, concentração de HCl no fluido, entre outros). A favorabilidade de extração e transporte dos elementos (Au) para porções superiores do magma depende também de processos físicos, como difusão (cinética) e advecção (transporte) (Vigneresse 2014).

Grande parte das inclusões apresenta, em condições ambiente, mais de uma fase (Fig 2.5). O aprisionamento heterogêneo decorre, em grande parte, da imiscibilidade de fluidos e as bolhas de vapor podem ser formadas em consequência das contrações termais do fluido durante o resfriamento e/ou pelo aprisionamento de fluido efervescente, resultante de uma etapa de *boiling* (alívio de pressão), processo comum em sistemas geológicos com interação de fluidos (Shepherd *et al.* 1985; Bodnar 2003) (Fig. 2.6).

Neste caso, a heterogeneidade se dá ou através de processos de descompressão (e.g., H₂O líquido e vapor de água), ou pela hipersaturação que acusa a precipitação de sólidos acidentais (minerais inclusos), ou pelo aprisionamento de fluidos de composição

distinta (H_2O e CO_2 , H_2O e óleo, metano e H_2O líquido, óleo e metano, entre outros). A presença de CO_2 , por sua vez, pode indicar aumento de pressão ao longo da curva líquido-vapor, sugerindo separação de fases em pressões/profundidades mais altas (Wilkinson 2001).

A microtermometria de inclusões fluidas em minerais em paragênese com a mineralização produz resultados que são muito próximos aos do fluido em profundidade; e, quando aplicada em minerais que não estão diretamente ligados à mineralização, podem ser úteis para a compressão completa do sistema hidrotermal (Shepherd *et al.* 1985; Bodnar 2003).

Essa técnica consiste no aquecimento e resfriamento de inclusões previamente descritas e catalogadas (dados de morfologia, tamanho, cor, estado físico, sistema, entre outros), na intenção de obter dados importantes para a compreensão dos parâmetros físico-químicos que regeram o processo geológico estudado.

Com os dados obtidos, calcula-se a temperatura da primeira fusão ou temperatura eutética (T_e), temperatura de fusão (do gelo ou de outros componentes como hidrohálita, antartícita), temperatura de homogeneização (temperatura em que o fluido foi aprisionado de forma homogênea), salinidade do fluido, densidade, entre outros parâmetros a depender do sistema fluido estudado (Bodnar 2003; Wilkinson 2001; Pestilho e Monteiro 2017); além de possíveis interpretações sobre T, P e X do ambiente de formação do depósito mineral. Diagramas como o de Evolução P-X-T podem corroborar essa interpretação (Fig. 2.7) (Zhao *et al.* 2021; Sun *et al.* 2024).

Técnicas de LA-ICP-MS aplicadas às inclusões fluidas podem expandir o conhecimento sobre a composição dos fluidos e, com isso, melhorar o modelamento do ambiente inicial de formação do depósito mineral. A partir desta técnica é possível obter a composição do fluido de inclusões específicas e, a partir de então, ter mais precisão sobre o sistema fluido que deu origem à mineralização do depósito mineral (Audétat 2020, Sun *et al.* 2024).

2.4.2 Isótopos estáveis (S, O, H, C)

Dados isotópicos são vastamente utilizados em estudos de depósitos minerais devido à eficácia na medição de variações das composições isotópicas dos elementos químicos, por conta de alterações físico-químicas do sistema magmático (Hoelfs 2009). Isótopos são excelentes traçadores de processos naturais e de trocas entre diferentes

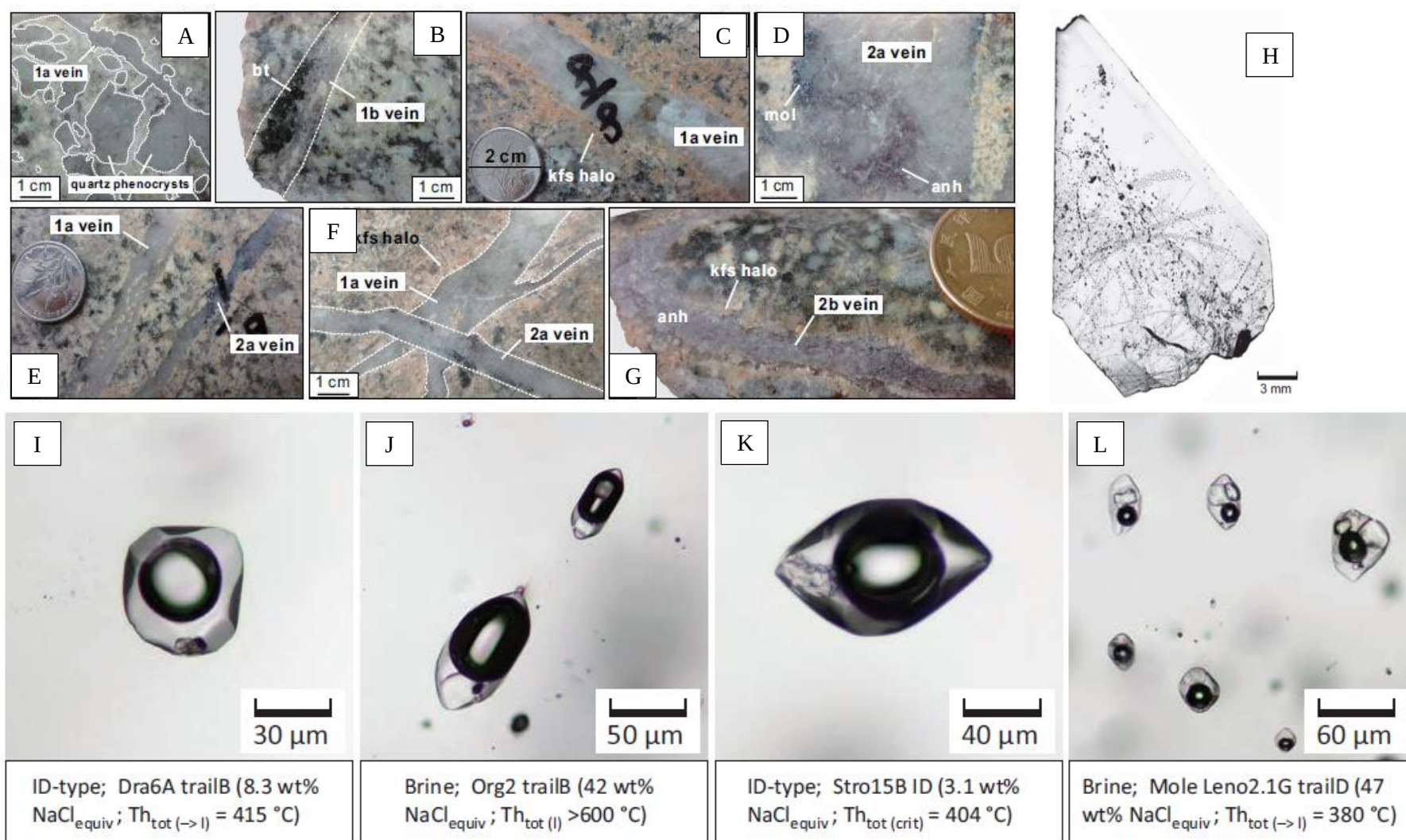


Figura 2. 5.: A – G) Seleção de amostras de rochas para confecção de lâminas bipolizadas para inclusões fluidas; H) Fragmento de quartzo a ser analisado em microtermometria; I – L) Fotomicrografias de porções do fragmento de quartzo mostrando algumas inclusões trifásicas e multifásicas (Audétat 2020; Zhao *et al.* 2021).

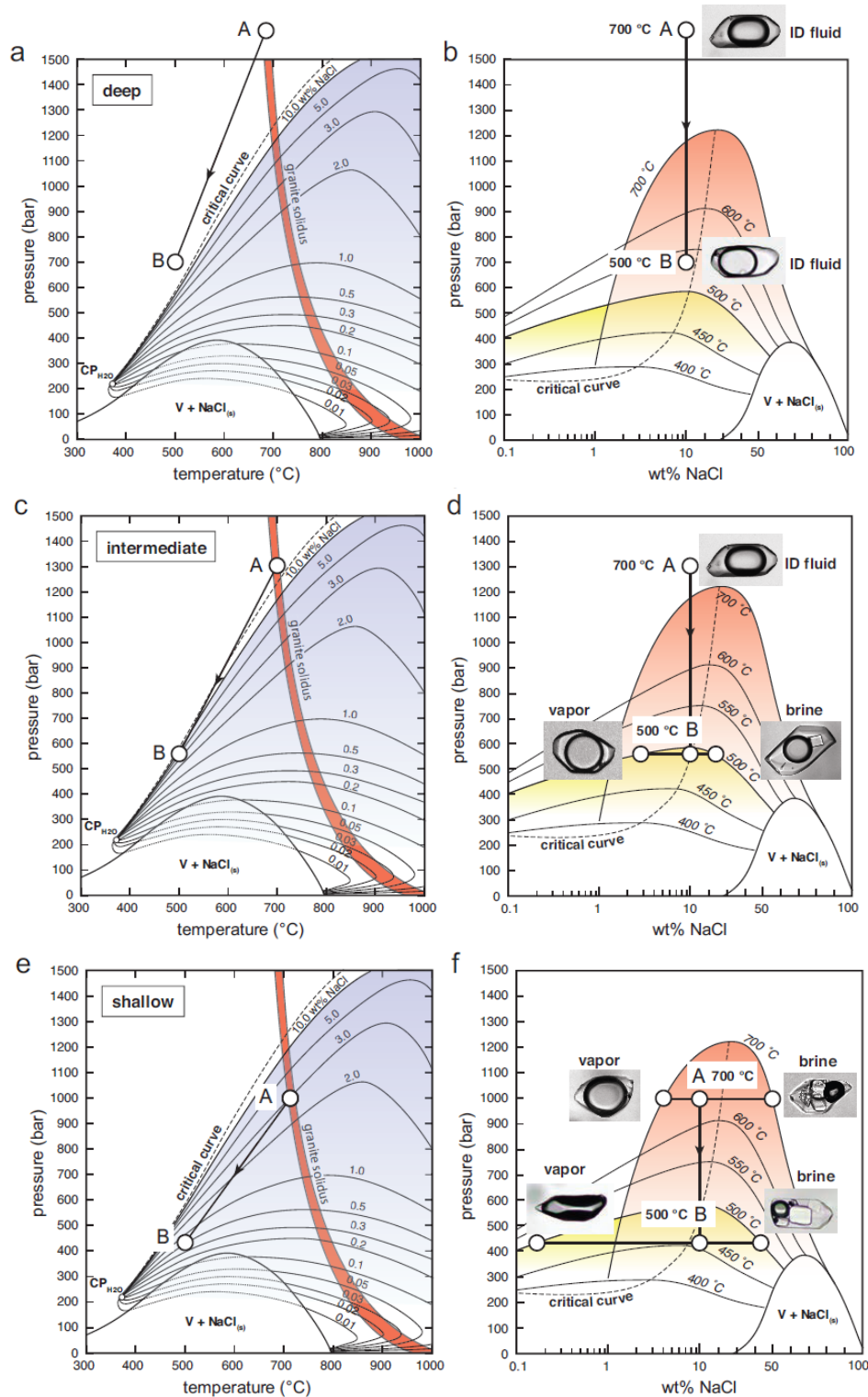


Figura 2. 6. Diagrama P-T-X típico de fluido magmático-hidrotermal: (a, b), intrusão profunda, (c,d) intermediária e (e, f) rasa. As figuras a, c, e correspondem ao campo de imiscibilidade de fluidos H₂O-NaCl projetando várias salinidades para o intervalo de P-T; b, d, f, mostram os campos de imiscibilidade de P-X. Em profundidade, o fluido se comporta como fase-simples em T de 700 e 500 °C. Em intrusões intermediárias, tem fase simples em condições magmáticas, mas bifásica em 500 °C. E em intrusões rasas o fluido é bifásico em ambas as temperaturas (Audétat 2020).

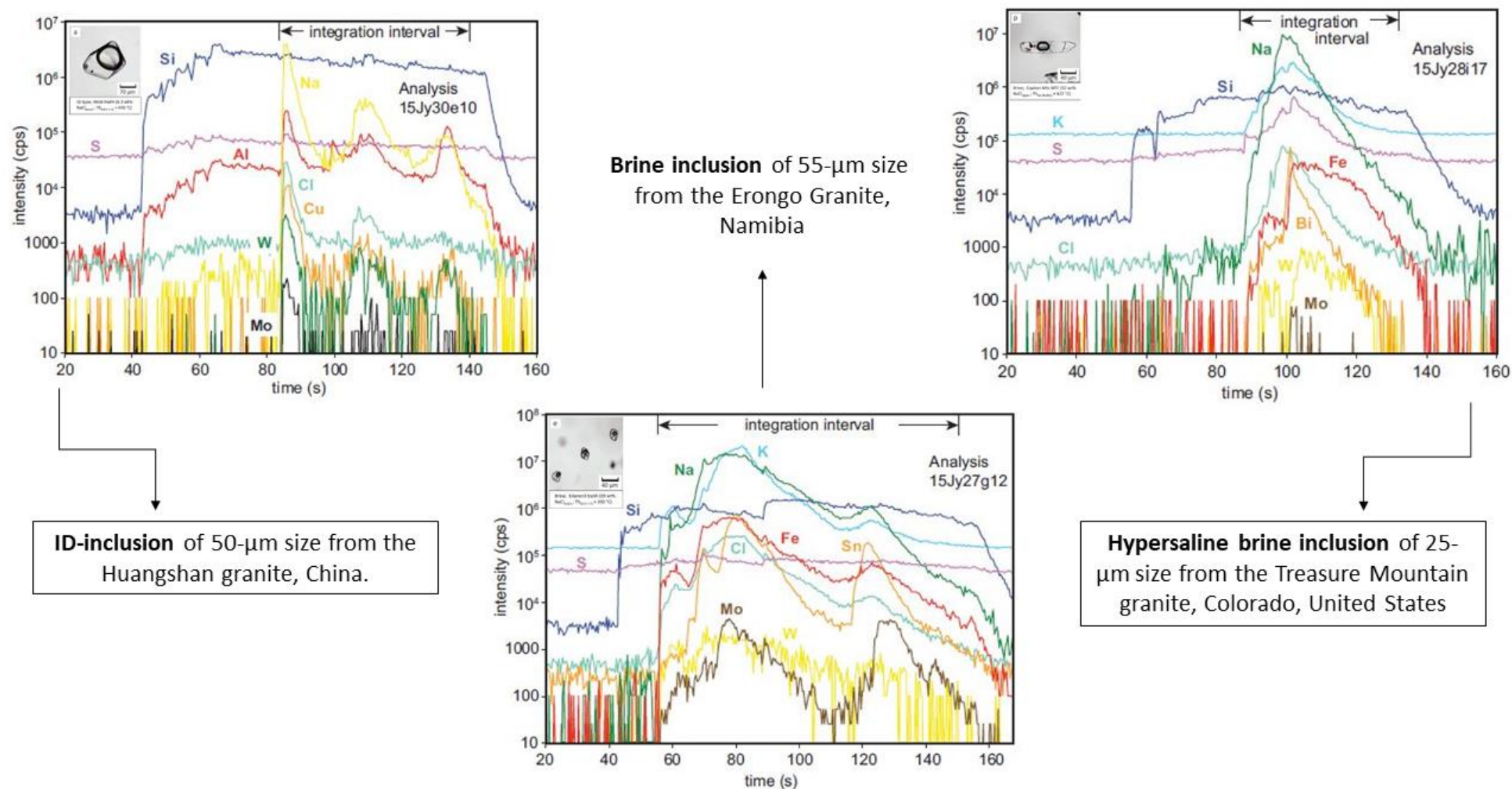


Figura 2. 7. Gráficos construídos a partir de dados de LA-ICP-MS em inclusões fluidas presentes em grãos de quartzo. Com base nos picos gerados durante a análise, é possível identificar todos os componentes presentes no fluido. Em destaque, 3 tipos diferentes de inclusões: A) ricas em vapor, B) salinas e C) hipersalinas.

reservatórios geoquímicos, além de traçadores da fonte do metal ([Allègre 2008](#); [Shanks 2014](#); [Sharp 2017](#)).

Em sistemas magmáticos-hidrotermais, alguns dos mais usados são os de enxofre (S), oxigênio (O) e hidrogênio (D). Os dois principais reservatórios terrestres de S apresentam composição isotópica constante: manto ($\delta^{34}\text{S} \sim 0\text{‰}$) e água do mar ($\delta^{34}\text{S} +20\text{‰}$, como SO_4^{2-}). Depósitos magmáticos tipo pórfiro, por exemplo, apresentam variação isotópica entre -3‰ e $+1\text{‰}$. Aqueles que têm influência da água do mar podem apresentar $\sim \delta^{34}\text{S} +3.5\text{‰}$, indicando influência de ambos os reservatórios ([White 2009](#)).

Esses isótopos podem ser encontrados de 5 formas distintas nos ambientes: S^{+6} (ex. BaSO_4), S^{+4} (ex. SO_2), SO (ex. S), S^{-1} (ex. FeS_2) e S^{-2} (ex. H_2S). Todas as espécies são suscetíveis a fracionamento adicional durante transporte e deposição/precipitação do minério ([White 2009](#)).

A variação da composição de $\delta^{34}\text{S}$, no entanto, sugere oscilação de $f\text{O}_2$ do magma. Valores de $\delta^{34}\text{S}$ constante de sulfetos magmáticos sugerem que $f\text{O}_2$ permaneceu abaixo do limite $\text{SO}_2/\text{H}_2\text{S}$, e não fracionou desde o manto; todavia, o fracionamento sugere aumento na $f\text{O}_2$ do sistema ([Fig. 2.8](#)) ([White 2009](#)). A [tabela 2.1](#), de [Ohomoto e Goldhaber \(1997\)](#), resume valores para o $\delta^{34}\text{S}$ em alguns dos sistemas magmático-hidrotermais mais conhecidos.

O fluido estudado em depósitos minerais pode corresponder a mistura de isótopos de H e O com várias possíveis origens (água meteórica, água oceânica, fluidos bacinais, fluidos metamórficos, fluidos magmáticos e Junevins), distinguindo, de forma clara, o caminho percorrido e possíveis interações/misturas entre moléculas derivadas de diferentes ambientes geológicos ([Taylor 1997](#); [Pirajno 2009](#)).

O H apresenta-se de três possíveis formas isotópicas: ^1H , ^2H , que são estáveis, e ^3H , que é radiogênico e as abundâncias na natureza são, proporcionalmente, 99,98%, 0.0156% e $10^{-18}\%$, respectivamente. O oxigênio apresenta o ^{16}O , ^{17}O , ^{18}O , com abundância de 99,76%, 0.037% e 0.1995%, respectivamente.

Quanto ao sistema magmático, variação de valores de δD pode indicar entrada de magma máfico na câmara magmática em determinado período da evolução do sistema. Quando a entrada ocorre, há forte diminuição de δD , devido à remoção de elementos com alta energia de ativação do magma máfico para o félsico (W, Sn), apontando para o estado de oxirredução do magma resultante ([Vigneresse 2007](#)).

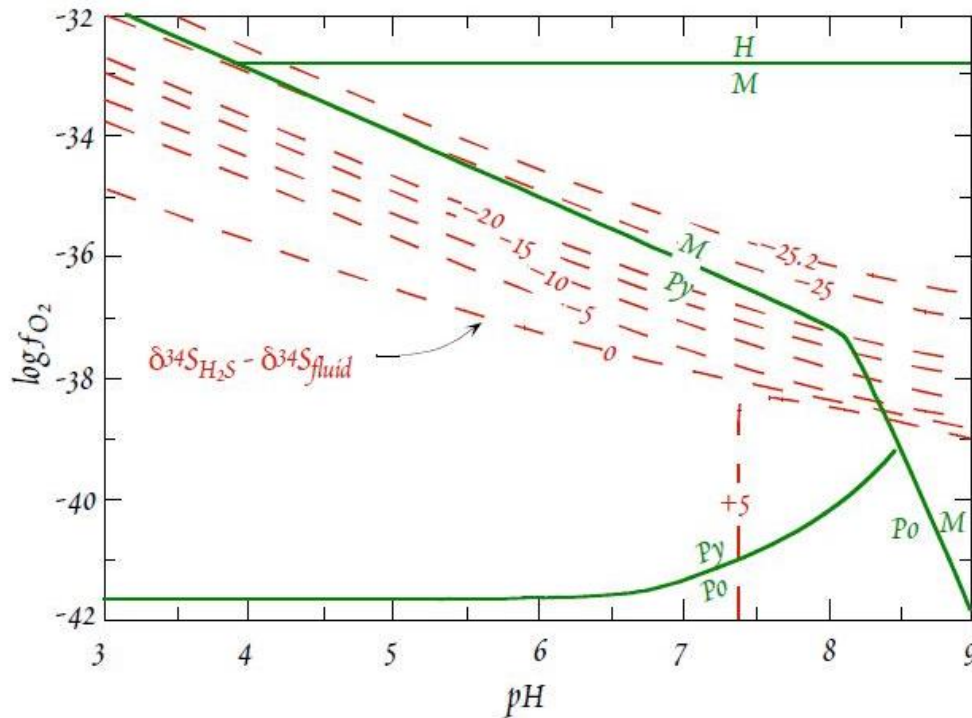


Figura 2. 8. Diferença do $\delta^{34}\text{S}$ entre H_2S e do fluido, em função da T e da razão sulfato/sulfeto em pH neutro. Limites de concentração iguais M: magnetita; H: hematita; Py: pirita; Po: pirrotita; Bn: bornita (Cu_5FeS_4); Cp: calcopirita (White 2009).

O $\delta^{18}\text{O}$ apresenta intervalos característicos para cada fluido/origem/proveniência. Na maioria dos ambientes geológicos há mistura de fluidos, em diferentes proporções. Por tanto, é imprescindível aliar, aos dados de isótopos estáveis, às demais técnicas de investigação (ambiente tectônico, ocorrências minerais ou depósitos minerais nas proximidades, tipos de alterações hidrotermais, temperatura de homogeneização do fluido mineralizantes, entre outros) para entender e interpretar a origem do fluido existente na área de trabalho.

Em sistemas magmático-hidrotermais, o Au é preferencialmente transportado na fase gasosa do fluido mineralizante, a qual é composta, além de outros gases, pelo CO_2 . O *boiling*, resultante da diminuição da pressão local e provável aumento da $f\text{O}_2$, desestabiliza a solução e precipita minerais, como a calcita (CaCO_3) e metais transportados. Neste caso, os valores de $\delta^{13}\text{C}$ da calcita refletirão as condições físico-químicas do momento da deposição do Au, auxiliando na compreensão do sistema mineralizador (Pirajno 2009). Nos fluidos meteóricos, que, em muitos casos, estão misturados a fluidos de origem magmática em geossistemas magmáticos, as variações de

Tabela 2. 1. Características isotópicas do enxofre em grandes depósitos minerais (Ohomoto e Goldhaber 1997).

Tipo de Depósito	Rocha hospedeira	Paragênese do minério	Associação mineral/elemento	Origem do S ⁻²	Intervalo de $\delta^{34}\text{S}$ no mineral	Relação do $\delta^{34}\text{S}$ e o minério
Pórfiros e depósitos tipo skarns associados a granitos tipo-I	Porfiríticas/carbonáticas	Calcopirita, bornita, molibdenita, sheelita, Au	Pirita, anidrita	Fluido magmático que adquiriu S a partir da assimilação de rochas.	$\delta^{34}\text{S} = 0\text{‰}$ a $\pm 5\text{‰}$, Dep. USA e América do Sul; $\delta^{34}\text{S} = +2\text{‰}$ e $+6\text{‰}$ no Japão.	Equilíbrio entre sulfetos. Equilíbrio entre sulfetos e sulfatos.
Pórfiro e skarns, associados a granitos do tipo-S	Porfiríticas/carbonáticas	Cassiterita, estanita, wolframita, sheelita	Pirrotita, bismutinita, arsenopirita, fluorita (sem sulfatos)	Fluidos magmáticos. Maior parte do S provenientes da assimilação de rochas encaixantes locais.	$\delta^{34}\text{S} = +5\text{‰}$ a $+10\text{‰}$ Dep. USA e australianos; $\delta^{34}\text{S} = -10\text{‰}$ a -1‰ , dep. Japoneses.	Equilíbrio entre sulfetos
Sulfetos de Cu-Ni	Intrusões máficas (<i>Sudbury type</i>), komatiitos e lavas máficas (<i>Kambalda type</i>)	Calcopirita, pentlandita	Pirrotita, Pb, Pd, Co, Ir	Origem mantélica; S por assimilação de rochas encaixantes locais	$\delta^{34}\text{S}$ <i>Sudbury types</i> é amplamente variável (-10‰ a $+15\text{‰}$)	
VMS	Tufos vulcânicos; argilitos	Esfalerita, galena, calcopirita	Pirita, barita, anidrita, Au, Ag, pirrotita e magnetita	A partir da lixiviação de sulfetos das rochas encaixantes.	$\delta^{34}\text{S}$ são tipicamente positivos em dep. Arqueanos; VMS paleop. $\delta^{34}\text{S}$ entre $+5\text{‰}$ e $+15\text{‰}$	Equilíbrio e/ou desequilíbrio entre esfalerita e galena, entre esfalerita e calcopirita.

composição são influenciadas pela variação de elevação e latitude, com valores de δD e $\delta^{18}O$ sendo cada vez menores com o aumento da elevação, e vice-versa. Com o aumento de elevação/latitude, menor é a evaporação, que, por sua vez, diminui o fracionamento isotópico. Adicionalmente, a composição desses isótopos é bastante influenciada pela interação fluido/rocha (w/r), em que valores de troca isotópica resultam em $w/r > 1$, com tendência a equilibrar o fluido, enquanto $w/r < 1$ indica equilíbrio com as rochas (Pirajno 2009).

A figura 2.9 representa a relação isotópica de δO e δD em ambientes geotermiais, em condições de mistura de águas/fluidos de *Yellowstone* e outros locais semelhantes mundialmente, relacionando fluidos meteóricos com os de origem magmática. Nesses locais, há aumento de temperatura das rochas devido a atividades magmáticas, estimulando a circulação e infiltração de águas superficiais, além de seu aquecimento e retorno à superfície com maiores pressão e temperatura. Esse processo altera a composição do fluido tanto pela mistura de fluidos de diferentes origens quanto pela interação com as rochas encaixantes.

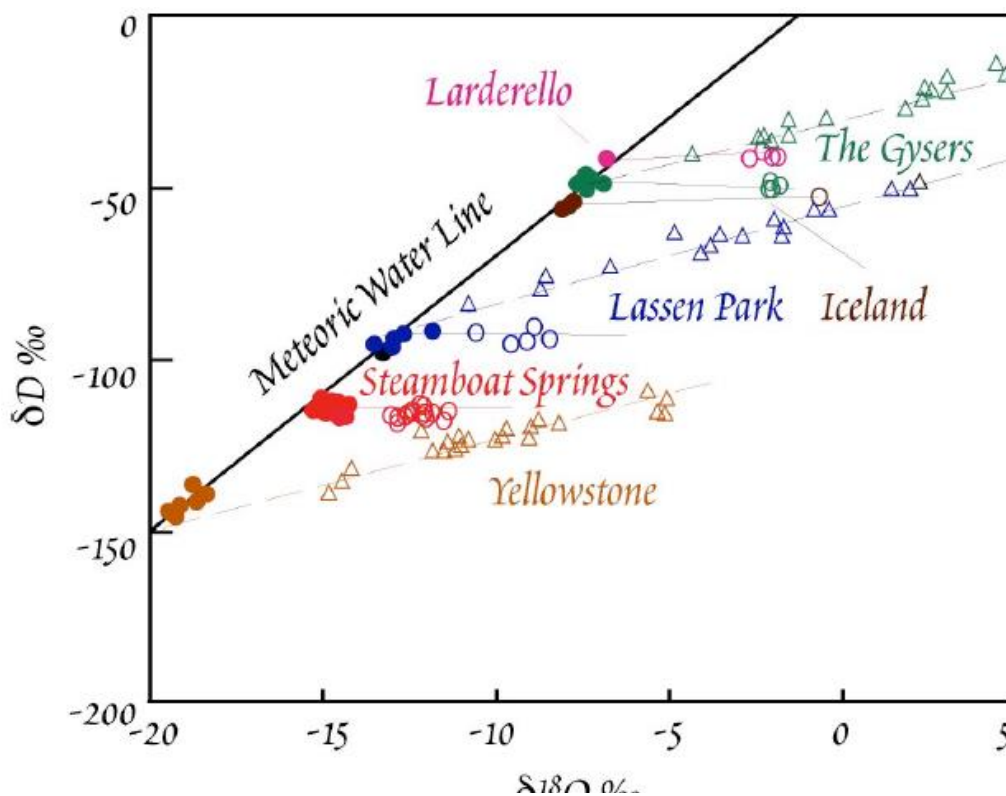


Figura 2. 9. δD e $\delta^{18}O$ em sistemas meteórico-hidrotermais. Círculos fechados mostram a composição da água meteórica nas proximidades de Yellowstone (Colorado, EUA), nas cidades de Steamboat Springs (Colorado), Mt. Lassen (Califórnia), Larderello (Itália) e Geysers (Califórnia). Os círculos abertos mostram a composição isotópica de águas geotermiais cloretadas nesses mesmos locais. Triângulos abertos representam águas ácidas, ricas em enxofre. A linha sólida conecta a água ácida e meteórica. Essa linha da “água meteórica” mostra a correlação entre os valores de δD e $\delta^{18}O$ em água de precipitação (White 2009).

CAPÍTULO 3

CONTEXTO GEOLÓGICO E GEOTECTÔNICO

3.1. Estruturação geológica, divisão tectônica e evolução do Cráton Amazônico.

O Cráton Amazônico está localizado ao norte da América do Sul, com sua maior porção dentro do território nacional, contudo, também se estende pelo leste da Bolívia, pelo sudeste da Colômbia, pela Guiana e Guiana Francesa, leste do Paraguai e sul da Venezuela ([Dardenne & Schobbenhaus 2000](#)).

Há décadas, o Cráton Amazônico tem sido compreendido como originado a partir de uma série de arcos acrescionários amalgamados, desde o Arqueado até o Paleo- e Mesoproterozoico ([Santos *et al.* 2000](#); [Tassinari & Macambira 2004](#)).

Os dois principais modelos de compartimentação do Cráton Amazônico são os modelos propostos por [Tassinari e Macambira \(1999\)](#) e [Santos *et al.* \(2000\)](#). No primeiro modelo, o cráton está compartimentado em seis províncias geocronológicas: províncias Amazônia Central (> 2,3 Ga), Maroni – Itacaiúnas (2,2-1,95 Ga), Ventuari – Tapajós (1,95-1,80 Ga), Rio Negro – Juruena (1,8-1,55 Ga), Rondônia – San Ignácio (1,55-1,3 Ga) e Sunsás (1,3-1,0 Ga).

Neste cenário, as províncias Ventuari – Tapajós e Rio Negro – Juruena e parte das províncias Maroni – Itacaiúnas e Rondônia – San Ignácio teriam evoluído a partir de sucessivos episódios de acreção continental com rochas derivadas do manto superior, através de uma sucessão de arcos magmáticos. Por outro lado, as províncias Sunsás e partes das províncias Rondônia – San Ignácio e Maroni – Itacaiúnas estiveram associadas, ao longo de sua história geológica, a eventos de colisão continental ([Tassinari & Macambira 1999](#); [Tassinari & Macambira 2004](#)).

De acordo com a proposta de [Santos \(2003\)](#) e [Santos *et al.* \(2006\)](#), o Cráton está subdividido em sete províncias tectônicas, as quais incluem: província Carajás (3,0-2,5 Ga), Amazônia Central (supostamente arqueana), Transamazonas (2,26-2,01 Ga), Tapajós–Parima (2,03-1,88 Ga), Rio Negro–Juruena (1,82-1,52 Ga), Rondônia–Juruena (1,82-1,54 Ga) e Sunsás–k’Mudku (1,45-1,10 Ga).

Alternativamente, [Juliani *et al.* \(2013; 2021\)](#) têm proposto um modelo de estruturação para a parte sul do Cráton Amazônico em dois arcos magmáticos paleoproterozóicos, de direção E-W, denominados de Arcos Tapajônicos I e II, sendo um mais antigo (2,13-1,95 Ga), a sul, superposto por outro mais jovem (1,89-1,87), a norte. Na contramão dos modelos anteriores, [Juliani *et al.* \(2013\)](#) ainda sugerem que esses Arcos

Tapajônicos tenham sido produzidos por subducção, de sul para norte, de placa oceânica sob uma placa arquena a paleoproterozoica.

Esses estudos mais recentes têm propostos, com base no conhecimento de novos dados isotópicos e geofísicos, que o Cráton é resultado da formação de arcos continentais (Juliani *et al.* 2013; Galé & Juliani 2016; Carneiro *et al.* 2019; Juliani *et al.* 2021; Tarelow Neto & Moura 2021). Carneiro *et al.* (2019), baseando-se em um robusto processamento magnético, propuseram uma nova compartimentação tectônica para o Cráton Amazônico, no sentido dos lineamentos E–W, intrinsecamente relacionados ao embasamento.

De acordo com essa interpretação, a formação e a evolução das unidades paleoproterozoicas ocorreram durante uma orogenia oceano-contidente de longa duração (aproximadamente 200 M.yr), com subducção E–W e mergulho para norte, a qual originou dois arcos magmáticos continentais – Arco Magmático Tapajônico I e Arco Magmático Tapajônico II – entre 2.1 e 1.87 Ga (Juliani *et al.* 2021).

Além disso, a progressiva migração da litosfera oceânica subductada para posição horizontal (*flat-subduction*), à medida que adentrou sob a litosfera continental, há cerca de 1,88 Ga, formou magmas cálcio-alcálico de alto K e shoshonítico distantes da fossa oceânica e desencadeou um processo de fusão crustal, que gerou magmas félsicos tipo A contemporâneos. Durante esta evolução, houve intenso soerguimento, erosão e rifteamento. A formação de bacias *pull-apart* também contribuiu para a preservação de unidades vulcânicas. A ocorrência de associações cálcio-alcálicas mais jovens na região de São Félix do Xingu (ca. 1,88 Ga) resultou do afundamento da crosta oceânica após um estágio de subducção de placa plana. O movimento do cinturão móvel em direção a arcos magmáticos mais antigos gerados na porção sul do Cráton Amazônico é confirmado pelas estruturas orientadas E–W (Fernandes e Juliani 2019; Juliani *et al.* 2021).

Ademais, os dados geoquímicos e isotópicos das rochas plutônicas, subvulcânicas e vulcânicas são mais característicos de formação em arcos magmáticos continentais (Pinho *et al.* 2003; Juliani *et al.* 2013, 2014, Carneiro *et al.* 2014; Galé *et al.* 2015; Tokashiki 2015; Bettencourt *et al.* 2016; Scandolara *et al.* 2017).

A porção oriental da Província Juruena é interpretada como um bloco cratônico por Trevisan *et al.* (2021), que é considerado um remanescente da parte oriental do Domínio Tapajós dentro da Província Ventuari–Tapajós. Ademais, é reconhecido como

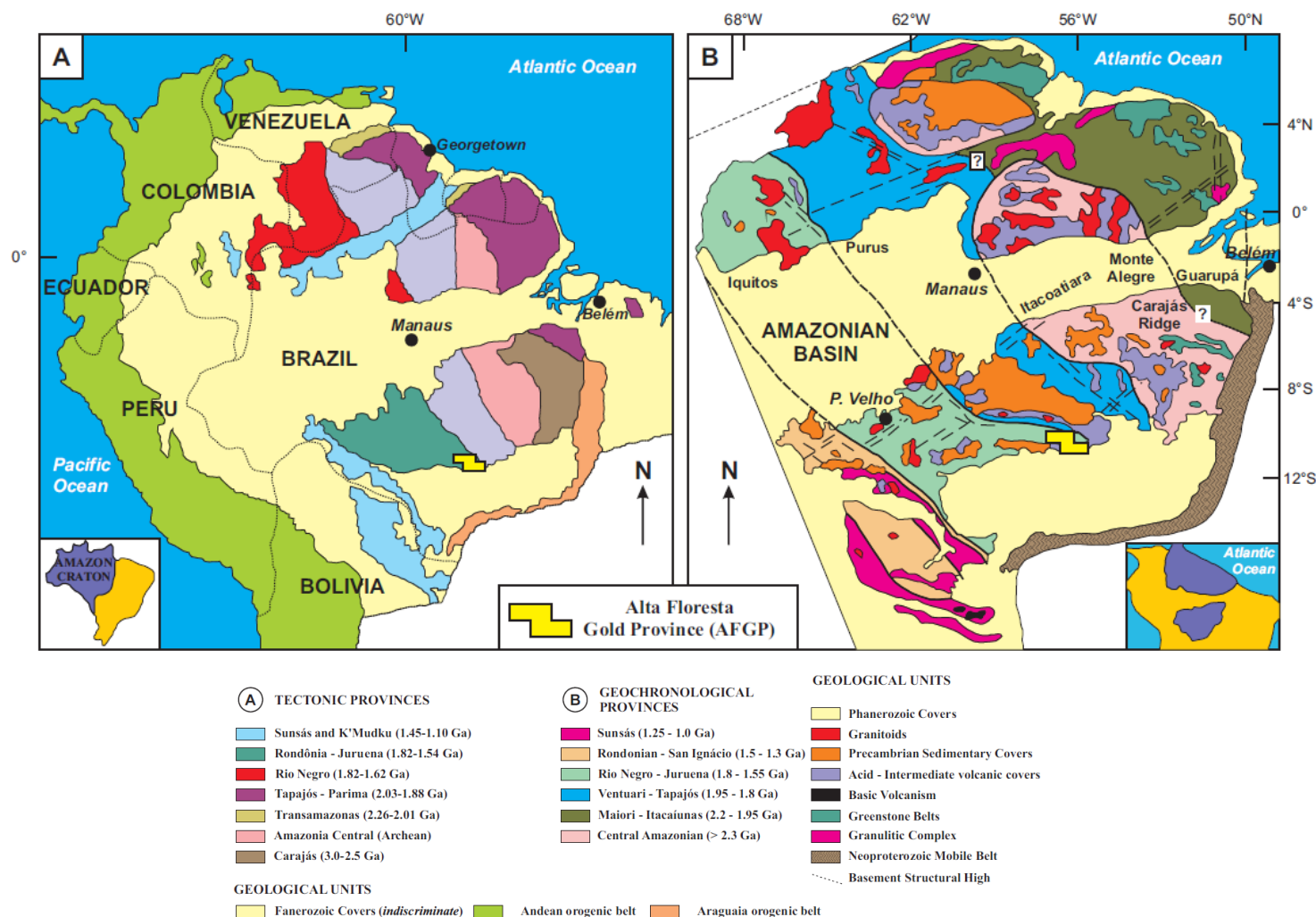


Figura 3. 1. Mapas de localização das Província geocronológicas e tectônicas do Cráton Amazônico, com destaque para a Província Mineral Juruena (Alta Floresta) (Assis *et al.* 2017). A) Modelo de Santos *et al.* (2006); B) Modelo de Tassinari e Macambira (1999).

contendo vestígios do magmatismo do cinturão ígneo da Amazônia Ocidental (1,85-1,74 Ga) na Província Rio Negro–Juruená.

As idades obtidas a partir de zircões herdados (1.960-1.900 Ma) estabeleceram semelhanças entre as idades da Suíte Creporizão (1.960 Ma), Suíte Intrusiva Tropas (1900 Ma) e Suíte Intrusiva Parauari (1.880 Ma), todas interpretadas como parte do Domínio Tapajós, Província Tapajós. Essas semelhanças sugerem que a variação de idade pode ser resultado da reformulação de antigos terrenos de arco magmático que foram sucessivamente aglutinados ao protocraton (Juliani *et al.* 2014; Juliani *et al.* 2021).

As rochas meta- a peraluminosas, oxidadas, cálcio-alcalinas, plutônicas e vulcânicas, encontradas na porção leste da Província Juruená, possuem médio a alto potássio, pertencem à série magnetita (granitos tipo I), porém também ocorrem rochas vulcânicas, subvulcânicas e granitos alcalinos (Assis 2015; Alves *et al.* 2019), e há semelhanças significativas nos aspectos texturais e no comportamento geoquímico aos dos granitos descobertos no sudoeste da cidade de Jacareacanga, província do Tapajós (Lopes e Moura 2019).

3.1.1. Província Tapajós – Parima

A Província tectônica Tapajós–Parima (2,1-1,87 Ga) de Santos *et al.* (2000; 2006) é constituída por quatro domínios principais: Peixoto de Azevedo, Uaimiri, Parima e Tapajós. O embasamento da província é truncado por gnaisses de composição granodiorítica a tonalítica, representantes de um primeiro arco magmático insular cálcio-alcalino primitivo denominado Complexo Cuiú-Cuiú (2,03-2,0 Ga) (Pessoa *et al.* 1977; Santos *et al.* 2000). O arco magmático Creporizão (1,98-1,95 Ga) é formado por ochas ígneas mais evoluídas (monzogranitos e andesitos) com contrinuição crustal. Segundo Santos (2003), esses dois primeiros arcos constituem a Orogênese Mundurucus.

Ainda conforme o autor, seguiu-se um hiato tectônico de aproximados 40 a 50 Ma, antes do desenvolvimento de um terceiro arco insular primitivo denominado de Tropas (1,9 e 1,89 Ga), e predominantemente constituído por tonalitos, andesitos e basaltos. O arco Parauari (1.879-1.883 Ga) é formado por monzogranitos e vulcânicas félsicas, com sienogranitos a tonalitos subordinados, truncando as unidades supracitadas (Santos 2003).

De modo geral, essas sequências de arcos magmáticos estão recobertas por

sedimentos fluviais do Grupo Buiçu (1.879 ± 12 Ma/ zircão detrítico) e por *sills* e diques máficos da Suíte Crepori (1.780 ± 9 Ma) (Santos 2003).

3.1.2. Província Rondônia-Juruena

A Província Rondônia-Juruena ($1.82 - 1.54$ Ga), regionalmente disposta na direção WNW-ESSE a E-W, pode ser subdividida em dois domínios principais: domínios Roosevelt-Juruena ($1.84-1.79$ Ga) e Jamari ($1.76 - 1.65$ Ga). Grande parte das unidades pertencentes ao domínio Roosevelt-Juruena ocorrem no segmento leste da Província Juruena.

O embasamento do Domínio Jamari, constituído por ortognaisses cálcio-alcálicos de baixo K (associação quartzo-diorito, tonalito, metabasalto, meta-andesito) e paragnaisses (Santos *et al.* 2000) originados em fase de arco magmático juvenil tipo Andino (1.67 e 1.63 Ga) mostra significativa contribuição crustal, possivelmente da orogênese Tapajós-Parima (Santos 2003). Essas unidades são cobertas por rochas sedimentares clásticas da Formação Mutum-Paraná e do Grupo Beneficente e são interpretadas como remanescentes de bacias intracratônicas ou de *rift* continental (Santos *et al.* 2000; Santos 2003).

Unidades pós-orogênicas do final da orogenia Juruena (1.606 a 1.532 Ma) estão incluídos na Suíte Serra da Providência, representada por charnokitos e granitos rapakivi tipo-A gnaissificados e milonitizados no decorrer da deformação da Orogênese Candeias do Ciclo Sunsás e representantes de diversos episódios magmáticos intrusivos (Santos 2003).

O Domínio Roosevelt-Juruena é formado por quatro associações distintas: embasamento, granitos cálcio-alcálicos, sequências vulcânicas e vulcanossedimentares e assembleias pós-orogênicas com granitos anorogênicos e sedimentares clásticas. No conjunto, esses domínios teriam evoluído a partir de sucessivos arcos magmáticos desenvolvidos por subducção de crosta oceânica sob a crosta Tapajós-Parima (Santos *et al.* 2000; Santos 2003).

O embasamento é formado pelos complexos Monte Verde e Bacaeri-Mogno (2.243 ± 130 a 1.850 Ma), a norte e mais antigo, além do Grupo Jauru ($1.78-1.74$ Ga), mais jovem e situado ao sul da província. É intrudido por diversas suítes graníticas orogênicas e cálcio-alcálicas de idade 1.84 a 1.77 Ga, a exemplo das suítes Juruena (1.823 ± 35 Ma a 1.817 ± 57 Ma; U-Pb em zircão), Paranaíta (1.793 ± 6 Ma a 1.816 ± 57 Ma; U-

Pb em zircão) e Sienito Cristalino (1.806 ± 3 Ma; U-Pb SHRIMP em zircão), produtos do primeiro arco magmático com participação crustal, posterior à formação do arco Bacaeri, juvenil e mais antigo (Santos 2003).

O segundo arco continental é formado por um conjunto de granitos mais jovens, representado pelo Complexo Monte Verde (dominantemente juvenil), Suíte São Pedro (1.784 ± 17 Ma), Tonalito Vitória (1.785 ± 8 Ma), Suíte São Romão (1.770 ± 9 Ma), Granito Nhandu (1.889 ± 17 Ma a 1.848 ± 17 Ma), o granito tipo-S de Apiacás (~ 1.78 - 1.77 Ga) e os granitos pós-colisionais com características anorogênicas vinculados à Suíte Teles Pires (1.78 - 1.75 Ga). A maior quantidade de vulcânicas félsicas associada a esse arco continental (Suíte Colíder; 1.786 ± 17 a 1.773 ± 5 Ma; Silva & Abram 2008) possivelmente reflita a sua maior preservação em relação ao primeiro arco.

Cobrindo todas as unidades, estão as vulcânicas da Suíte Colíder (1.786 ± 17 a 1.773 ± 5 Ma), as vulcanossedimentares do Grupo Roosevelt (1.740 ± 8 Ma) e sedimentares clásticas e carbonáticas com camadas subordinadas de piroclásticas, pertencentes ao Grupo Beneficente (2.64 a 1.69 ± 73 Ga; Santos *et al.* 2000).

3.1.3. Contexto geológico e geotectônico da Província Mineral Juruena.

A Província Mineral Juruena insere-se no centro-sul do Cráton Amazônico, entre as Províncias Proterozóicas Ventuari–Tapajós (1.95 – 1.80 Ga) e Rio Negro–Juruena (1.80 – 1.55 Ga) (Tassinari & Macambira 1999, 2004), ou nas Províncias Tapajós–Parima (2.03 – 1.88 Ga) e Rondônia–Juruena (1.82 – 1.54 Ga) (Santos *et al.* 2000).

A Província Mineral Juruena é composta por granitos paleoproterozoicos oxidados, cálcio-alcalinos, de médio a alto K, para a meta- a luminosos, do tipo–I, além de sequências vulcânicas e vulcanossedimentares, com granitos do tipo–A, tardios e subordinados e rochas vulcânicas (Santos *et al.* 2000). As unidades são agrupadas em: 1) granitos deformados do embasamento (2.81 – 1.99 Ga); 2) granitos félsicos do tipo–I, unidades vulcânicas e vulcanossedimentares (1.97 – 1.87 Ga); 3) granitos pós-orogênicos do tipo–I altamente evoluído e do tipo–A, subvulcânicas a vulcânicas e vulcanossedimentares (1.78 – 1.77 Ga); e 4), além de sequências sedimentares clásticas (ca. 1.37 Ga).

As mineralizações na Província Mineral Juruena são predominantemente auríferas e grande parte ocorre alinhada ao longo de um cinturão de direção NW–SE, denominado Peru–Trairão (Miguel Jr. 2011; Assis 2015).

Com base na alteração hidrotermal, na paragênese, na geoquímica do minério e na forma de ocorrência, os depósitos minerais foram agrupados em quatro grupos distintos: (i) depósitos de Au ($\pm$ Cu) disseminados (*e.g.*, depósitos Luizão, Pé Quente, Serrinha, X1 e Raimunda); (ii) depósitos filoneanos de Au ($\pm$ Cu) (*e.g.*, Edu, Peteca, Paraíba e Basílio), ambos predominantemente representados por pirita com concentrações variáveis de calcopirita e hematita; (iii) depósitos disseminados de Cu + Mo ($\pm$ Au) (*e.g.*, Ana e Jaca), com pirita, molibdenita e concentrações variáveis de hematita e calcopirita; (iv) depósitos de Au + Zn + Pb ($\pm$ Cu) (*e.g.*, depósito do Francisco, Bigode e Luiz) representados por pirita, esfalerita e galena, além de calcopirita e digenita subordinadas (Assis 2015; Galé *et al.* 2018; Lopes e Moura 2024).

Corpos graníticos ricos em magnetita, de arco magmático continental, situados em nível crustal catazonal-mesozonal a epizonal, intrudem as unidades do embasamento, a exemplo do Complexo Novo Mundo, com idade U–Pb em zircão de 2,03–1,98 Ga (Lopes *et al.* 2024 *in submission*). Unidades vulcânicas dacíticas a riolíticas cálcio-alcalinas associadas a arcos também estão presentes. Tais unidades compreendem o monzonito, monzodiorito e quartzo monzonito Pé Quente (ca. 1,97 Ga) (Paes de Barros 2007; Assis 2011; Miguel Jr. 2011; Rios 2019); o sienogranito e o monzogranito Flor da Mata e Aragão (ca. 1,93 Ga) (Miguel Jr. 2011); e o granodiorito e o tonalito *adakite-like* Guarantã (1,90–1,89 Ga) (Rios 2019; Assis 2015; Alves *et al.* 2019). A unidade metamáfica-metaultamáfica composta por flogopita xisto (1,93 Ga) está restrita à jazida de ouro venular de Serrinha do Guarantã. Com exceção do granito Flor da Mata, todos hospedam mineralização hidrotermal de ouro.

3.1.3.1. Novo Mundo

O denominado Granito Novo Mundo, classificado por Paes de Barros (2007), ocorre na porção centro sul do Cráton Amazônico, norte do estado de Mato Grosso. Foi distinguido por Paes de Barros *et al.* (2004) e teve seus limites definidos por Moura (2004) por meio de dados gamaespectrométricos.

Corresponde a um plúton alongado na direção regional NW-SE, no qual é possível reconhecer a recristalização e o estiramento dos cristais de quartzo ($L_x=N15W/10^\circ$), possivelmente decorrente do alojamento mediante controle estrutural tardio ao desenvolvimento das zonas de cisalhamento que delimitam suas bordas NE e SW (Assis 2015).

[Souza et al. \(2005\)](#) classificam este corpo granítico como pertencente à Suíte Intrusiva Nhandu. Segundo [Paes de Barros \(2007\)](#), esse granito exhibe variação faciológica de sienogranito a monzogranito, além granodiorito, quartzo monzonito e monzonito subordinados, posteriormente cortados por múltiplos diques de diabásio e corresponde a um corpo intrusivo em rochas do embasamento e o individualizou em duas porções, Centro-Sul e Norte. A porção Centro-Sul é constituída principalmente por sienogranitos (1.964 ± 1 Ma - Pb/Pb), com monzogranito (1.970 ± 3 Ma – Pb/Pb), quartzo monzonito e monzonito (1.956 ± 12 Ma – U-Pb) subordinados. O sienogranito exhibe idade $T_{DM} = 2,76$ Ga e $\epsilon Nd_{(1.964)} = -7,62$; enquanto o monzogranito apresenta idade $T_{DM} = 2,55$ Ga e $\epsilon Nd_{(1.964)} = -4,48$.

Dados recentes obtidos por [Lopes e Moura \(2024 in submission\)](#) mostram intervalo de idade obtidos a partir de U–Pb em zircão entre 2,03 a 1,99 Ga para os granitos de Novo Mundo, com idade modelo $T_{DM} = 2.52 - 2.14$ Ga e $\epsilon Nd = -2.0$ a $+1.81$. Os dados isotópicos apontam a participação de crosta continental e para a presença de fonte mantélica arqueana na geração desse magma.

Apresenta ainda enxame de diques basálticos e andesíticos, por vezes balizando corpos de minérios. Na porção norte ([Paes de Barros 2007](#)), predominam monzogranito, com granodioritos e sienogranitos subordinados, com diques de gabros e dioritos.

Baseando-se em critérios estabelecidos por [Chappell & White \(1974, 2001\)](#), [Paes de Barros \(2007\)](#) acentua semelhanças entre o granito Novo Mundo e os granitos Tipo–I pela sua natureza cálcio-alcalina à subalcalina, associação com intrusivas básicas e vulcânicas andesítica, e titanita como fase acessória, bem como com os granitos Tipo–S, pela ausência de restitos e de hornblenda e biotita, teores altos de K_2O (5,6-7,1 %), médios de Rb (197-222 ppm) e baixos de Sr (53-103 ppm) ([Testa 2018](#)).

CAPÍTULO 4

GEOLOGY, PETROLOGY AND GEOCHRONOLOGY OF 2.03 TO 1.96 GA MAGMATIC ARC GRANITES FROM JURUENA MINERAL PROVINCE, BRAZIL: IMPLICATIONS FOR THE AMAZONIAN CRATON TECTONIC EVOLUTION.

[ARTIGO 1 – SUBMETIDO AO PERIÓDICO JOURNAL OF BRAZILIAN JOURNAL
OF GEOLOGY](#)

Highlights

- Long-lived ocean-continent subduction in the south Amazonian Craton during the Paleoproterozoic.

- Juruena and Tapajos Mineral Provinces formed by granites with similar compositions and ages.
- Novo Mundo complex formed by 2.03–1.96 Ga I-type calc–alkaline granites and diabase dikes.
- Different degrees of melting and crustal contamination in the formation of the studied granites.

Abstract

New petrological, geochemical, U–Pb and Sm–Nd isotopic data allow us to propose a new petrogenetic model for the granite production in the southernmost portion of Amazonian Craton, Brazil, Novo Mundo complex, extreme north of the Juruena Mineral Province and the south of Tapajós-Parima Province. The rock facies include monzonite, monzogranite, biotite granodiorite and syenogranite. Andesitic dikes occur by cutting the granitic body. Biotite from the granites is classified as formed in a volcanic arc environment. The I-type, meta to peraluminous, calc–alkaline to alkali–calcic granites have petrographic, lithogeochemical, mineral chemistry and isotopic characteristics of volcanic arcs granites, with different degrees of crustal contribution. Magmatic zircon crystals revealed U–Pb LA–ICP–MS ages of 2032 ± 6 Ma for monzonite, 2029 ± 4 Ma for monzogranite, 1989 ± 6.2 Ma for biotite granodiorite and 1987 ± 7.4 Ma to 1964 ± 1 Ma for syenogranite. The granites exhibit $T_{DM\ Nd}$ model ages of 2.52–2.15 Ga and $\epsilon Nd(t)$ from -2.0 to +1.81, indicating that the parental magma of the granites originated from the mantle or melting of the juvenile Orosian crust. The weak positive $\epsilon Nd(t)$ anomalies suggest an upwelling asthenospheric mantle that supplied heat to remelt the lower continental crust and the isotopic data highlights different levels of crustal reworking during arc evolution. These ages can be correlated with the data evolution of the Paleoproterozoic Cuiú–Cuiú Magmatic Arc emplaced during the main phase of the arc subduction evolution in the region (2.05–1.99 Ga). The monzonite and monzogranite are associated with the initial stages of arc evolution. The granodiorite shows the contribution of juvenile sources correlating with mafic magma inflow events, while the syenogranite represents the final stages of the arc formation, with crustal contribution. The obtained data, together with recent studies, allow suggest the existence of a continuous magmatic arc in the Paleoproterozoic at the south part of the Amazonian craton, which was responsible for the formation of the Tapajós and Juruena provinces main granitic rocks.

They are also crucial to expanding our knowledge for the magma generation and granite-related mineral deposits and in south Amazonian Craton.

Keywords: Amazonian Craton, Granite, Magmatic arc, Paleoproterozoic, Geochronology, Petrology.

1. Introduction

Plutono-volcanic activities were essential in generating the Earth's crust in the Archean and Proterozoic (Cawood *et al.* 2013) and one of the most significant continental crust generation intervals occurred between 2.1 and 1.7 Ga (Hawkesworth *et al.* 2010).

In the Amazonian Craton, at least two episodes of large-scale volcano-plutonism mark the generation of Paleoproterozoic crust in the Central Amazonian and Tapajós-Parima Provinces (Santos *et al.* 2008), first: according to Bryan and Ferrari (2013) are called Orocaima SLIP (Silicic Large Igneous Province, ~200.000 km²; 2.0 Ga) (Barbosa *et al.* 2021); and Uatumã SLIP (~1.500.000 km²; 1.88 Ga) (Roverato *et al.* 2019).

In the southern portion of the Amazonian Craton, two of the country's main metallogenic provinces occur, the Tapajós Mineral Province and the Juruena Mineral Province. This last one is one of the Brazil's gold-producing regions, located in the extreme south of the Amazonian Craton, historically, has produced around 300 t/au and, in 2017, an important Cu deposit was discovered (500m @ 0.6% cu), which increased significantly the exploration projects in the region (Crusader 2018).

The Juruena Mineral Province is placed along a set of NW–SE trending tectonic zones. The region is made up of calc–alkaline granites (2.05–2.03 Ga) intruded by younger calc–alkaline granites with geochemical signatures of volcanic arc granites (Moura *et al.* 2006, Paes de Barros 2007, Juliani *et al.* 2014) with ages between 2.03–1.98 and 1.93–1.89 Ga.

Large-scale magmatism during the Orosirian period resulted in the emplacement of many granite bodies and several plutono–volcanic episodes between 2020 and 1860 Ma (Vasquez *et al.* 2017, Cassini *et al.* 2020), occupying more than one million square kilometers in the central and south portions of the Amazonian Craton. A significant part of these felsic calc–alkaline rocks (plutonic rocks and volcanic associated) occur in the Tapajós-Parima geochronological province.

This work characterized the new facies variation of Novo Mundo complex in Juruena Mineral Province based on textural aspects, geochemical data and new isotopic results. The obtained data, including new *in situ* zircon geochronology and Nd isotopes, allow

suggest mingling-mixing of granitic magma, resulting in the Novo Mundo facies and a new geological model and petrogenetic evolution for the Cuiú–Cuiú Magmatic Arc are proposed.

2. Regional Geology

The Amazonian Craton has constituted a tectonically stable unit since the Grenvillian Orogeny (1.2 Ga) event, and it is one of the largest cratonic regions in the world, covering approximately 430,000 km² (Tassinari & Macambira 1999). In the last decade, a new craton formation model has emerged based on multidisciplinary data, showing that the Amazonian Craton is related to the formation of continental arcs (Fig. 1A) (Juliani *et al.* 2013, Galé and Juliani 2016, Carneiro *et al.* 2019, Fernandes and Juliani 2019; Juliani *et al.* 2021).

According to this interpretation, the formation and evolution of Paleoproterozoic units occurred through a long-term ocean–continent orogeny (approximately 220 M.yr), with an E–W subduction and a northward dip, which would have originated two continental magmatic arcs (Tapajônico Magmatic Arcs I and II) between 2.1 and 1.87 Ga (Juliani *et al.* 2021). During this evolution, there was intense uplift, erosion and rifting. Pull-apart basin formation also contributed to preserving volcanic units.

The inherited zircon ages (1960–1900 Ma) established similarities between the ages of the Creporizão Suite (1960 Ma), Tropas Intrusive Suite (1900 Ma), and Parauari Intrusive Suite (1880 Ma), all of which are interpreted as part of the Tapajós Domain, Tapajós Province. These similarities suggest that the age variation can be the result of reworking of ancient terranes of the magmatic arc that were successively agglutinated to the protocraton (Juliani *et al.* 2014, Juliani *et al.* 2021).

The eastern portion of the Juruena Province is interpreted as a cratonic block, which is considered a remnant of the eastern part of the Tapajós Domain within the Ventuari-Tapajós Province (Trevisan *et al.* 2021, Fig. 1B). Additionally, it is recognized as containing vestiges from the magmatism of the Western Amazon igneous belt (1.85–1.74 Ga) within the Rio Negro-Juruena Province.

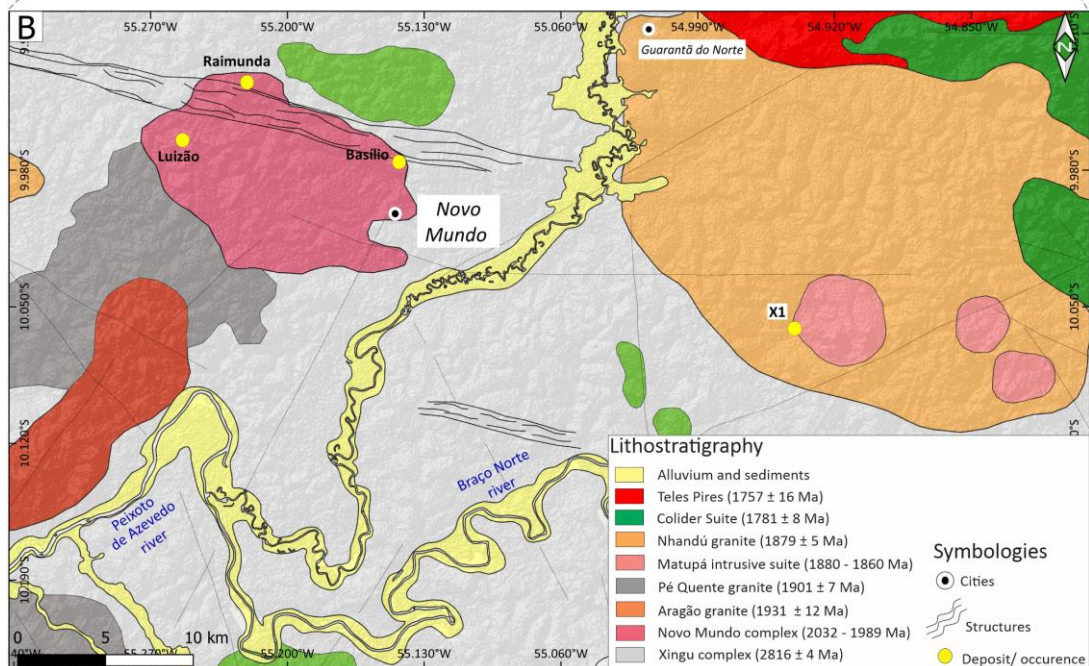
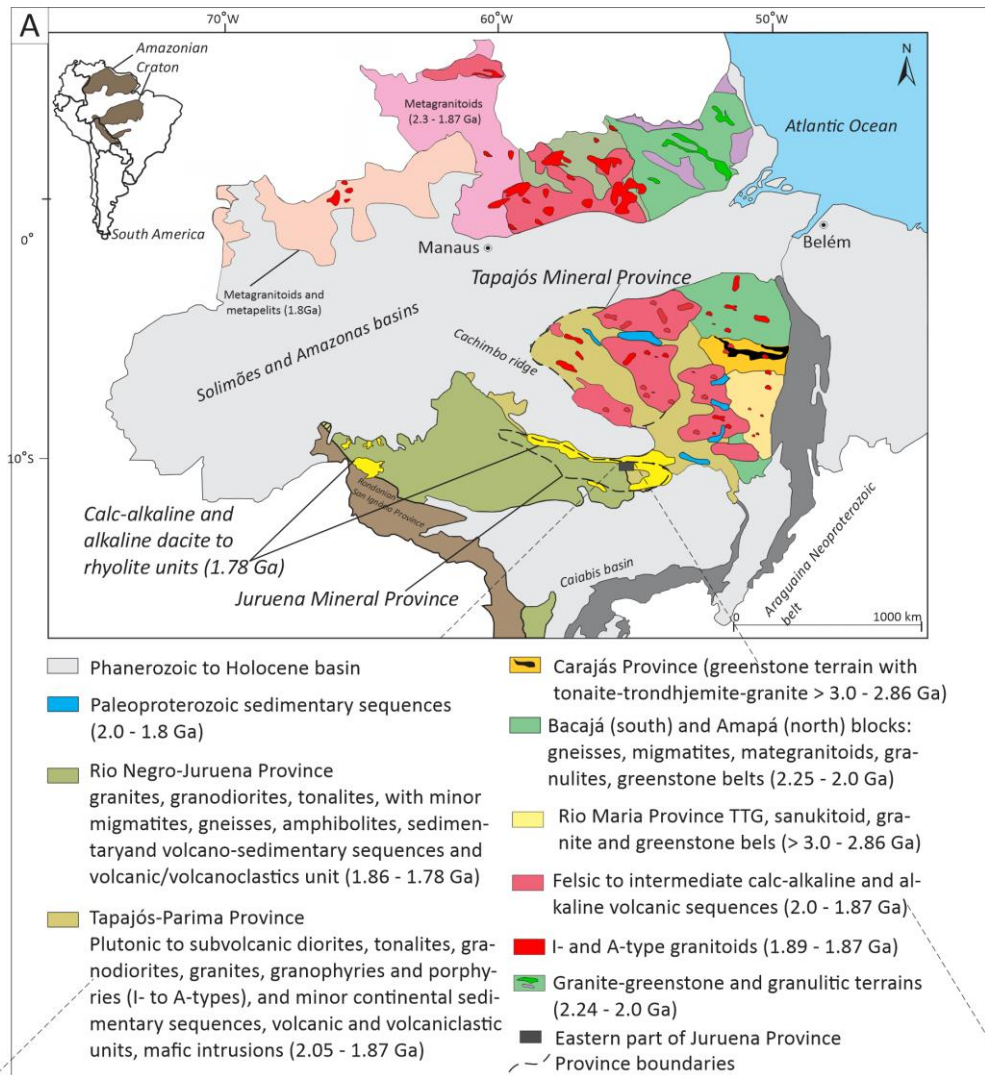


Fig. 1. A) Geotectonic map of the Amazonian Craton and the major tectonic provinces, from [Juliani *et al.* \(2021\)](#), showing the study area (area B); B) Geological map of the eastern part of Juruena Province (modified by [Paes de Barros \(2007\)](#) and [Miguel Jr. \(2011\)](#)).

The meta-peraluminous, oxidized, calc-alkaline, plutonic, and volcanic rocks found in the eastern section of the Juruena Mineral Province have a medium- to high K, of the magnetite series (I-type granites), volcanic, subvolcanic rocks, and alkaline granites ([Alves *et al.* 2019](#)), and there are similarities in the textural aspects and geochemical behavior of granites discovered in the southwest of Jacareacanga City, Tapajós Province ([Lopes and Moura 2019](#)).

3. Analytical Methods

3.1. Fieldwork and Petrography

Three field campaigns occurred in the Novo Mundo area and fifteen drill holes were selected for this study. Forty thin and polished sections were produced at the University of Brasília (UnB). Petrographic analyses were performed at the petrographic microscopy laboratories of UnB, using an Olympus BX-50 petrographic microscope. Point counting was carried out on 25 thin sections.

The modal classification of granitic rock was based on modal parameters obtained by counting 1500 points for medium- to fine-grained rocks and 1800 points for coarse-grained types. The conventional Michel-Lèvy method was used to determine the plagioclase zone composition and, when possible, by cleavage 010-twinning Law of Albite or twinning Law of Albite–Carlsbad methods ([Deer *et al.* 2013](#)).

3.2. Lithogeochemistry

Twenty rock samples were prepared and analyzed in the ALS Global Laboratory (GO, Brazil). Major elements were analyzed by ICP-AES (Induced Coupled Plasma Atomic Emission Spectrometry), and trace elements, including the Rare Earths, by ICP-MS (Induced Coupled Plasma Mass Spectrometry), after fusion with lithium metaborate. The base metal was performed via ICP–AES after acid digestion. Geochemical discriminant diagrams were elaborated using the GCDkit 4.0 software ([Janousek *et al.* 2006](#)).

3.3. Electronic Microprobe

Mineral chemistry analysis was conducted at the Electronic Microprobe Laboratory at UnB. Analyses were obtained by amphibole, biotite, k-feldspar, plagioclase, muscovite, and titanite crystals using a JEOL® JXA– 8230 electronic probe microanalyzer (EPMA) operating with an acceleration voltage of 20 kV and 40 atm. The wavelength dispersive system was set to operate with an accelerating voltage of 15 kV, a beam current of 10 nA, and a counting time of 10 s.

3.4. Isotopic geology

Thermo Scientific Neptune Plus field ICP–MS was used for the U–Pb geochronology at the Laboratory of Geochronology of UnB (two samples) and the Laboratory of Isotope Geochemistry of the Federal University of Ouro Preto, Brazil (UFOP) (two samples). The samples were mechanically crushed and subsequently sieved. The fraction smaller than 700 μm was planned to obtain a heavy mineral separation and submitted to a Franz magnetic separator set at 1.5 A.

Zircon grains were then manually selected with a binocular loupe, mounted on epoxy resin, and imaged by cathodoluminescence. The applied standard was Zircon GJ–1 (Jackson *et al.* 2004). Plotting of U–Pb data and age calculations were performed using ISOPLOT 4.1 (Ludwig 2003), and errors for isotopic ratios are presented at 1σ level. ^{238}U , ^{232}Th , ^{206}Pb , and ^{208}Pb were measured in Faraday cups, and ^{204}Pb , ^{207}Pb , and ^{202}Hg were collected in ion counters. Cathodoluminescence (CL) images of zircon crystals were obtained using scanning electron microscope JEOL 6510, in the Microanalysis Laboratory (LabMev) at UnB and UFOP.

Sm–Nd from nine samples were processed at the Geochronology Laboratory at the University of Brasília (UnB) utilizing the Gioia and Pimentel (2000) methodology. Samples were pulverized in a vibration mill with a metal carbide crucible and analyzed in a Thermal Ionization MS (TIMS) in a TRITON™ Multicollector. Rock pulps were mixed with ^{149}Sm – ^{150}Nd spike solution and dissolved in Savillex capsules were carried out using conventional cation exchange techniques with Teflon columns containing LN-Spec resin (HDEHP–diethylhexil phosphoric acid supported on PTFE powder). Once extracted, Sm and Nd samples were loaded on the Re filament of double filament assembly. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratio was normalized using $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, and the decay constant was $6.54 \times 10^{-12}/\text{y}$. The uncertainties for the $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are less than 0.2% (2σ) and $\pm 0.003\%$ (2σ), respectively. The results were treated

in ISOPLOT 4.15, and crustal residence ages (Nd model ages, T_{DM}) were calculated using [DePaolo \(1981\)](#).

4. Results

4.1. *Novo Mundo geology*

The Novo Mundo granitic complex is located in the central–southern portion of the Amazonian Craton in the state of Mato Grosso, eastern of Juruena Mineral Province. Dozens of mineral deposits, mainly gold, can be found along a NW–SE lineament. The basement rocks are intruded by granite, covering approximately 60 km², that follows a W–NW trend, coinciding with the direction of the main regional lineaments that affect the oldest terrains in that region ([Paes de Barros 2007](#)).

The petrographic study included analyses of hand rock samples, collected from outcrops, and samples from drill holes distributed throughout the body. The Novo Mundo granitic facies vary from monzonite, monzogranite, biotite granodiorite and syenogranite. They are phaneritic, inequigranular, fine–medium-grained rocks, greenish gray (monzogranite), reddish gray (syenogranite), and grayish (biotite granodiorite and monzonite) (Fig. 2). Figures 2A, 2B and 2C show outcrops of monzogranite, syenogranite and mafic rock in the region. Figures 2D to 2H highlight drill hole samples, where monzogranite, syenogranite, granodiorite, monzonite and andesite occur respectively.

The contact between the granitic facies is locally gradational/transitional but is mainly observed as subangular to rounded microgranular enclaves as fragments of the other facies, especially monzonite enclaves included in syenogranite. The mafic rocks occur as discordant contact, cutting the granitic facies abruptly. Occur throughout the region and the boreholes crop up more frequently below the average depth of 60–50 m (Fig. 2C). In drill holes, the mafic dikes (basalt, basaltic–andesite to andesite composition) are fine-grained and showing veinlets with sericitic alteration (Fig. 2H).

4.2. *Petrography and mineral chemistry*

Based on their textural aspects and compositional data, four major petrographic granite varieties were recognized, which occur spatially associated: monzogranite, syenogranite, biotite granodiorite, monzonite. The mineral chemistry is summarized in supplementary data file 1–5. These four granitic rocks consist of quartz (25–30 %), plagioclase (20–

25 %), orthoclase (20–25 %), biotite (8–10 %), hornblende (5–8%), muscovite (2–5 %) and magnetite (1–3 %). Accessory minerals are monazite, zircon, apatite, allanite, and titanite.



Fig. 2. Field and petrographic features of the Raimunda deposit showing macroscopic characteristics of the Novo Mundo rocks. A) Novo Mundo monzogranite block. B) syenogranite block showing k-feldspar phenocrysts. C) Mafic blocks emplaced in monzogranite and syenogranite terrain. D) Monzogranite sample: inequigranular phaneritic texture, locally containing plagioclase phenocrysts. E) Syenogranite drill hole samples: inequigranular phaneritic texture. F) Granodiorite: inequigranular texture. G) Monzonite: inequigranular texture. H) Mafic dike: fine-grained, locally containing plagioclase porphyries.

The monzogranite is dominant in the area and is characterized by a medium-grained size and a gray coloration (Fig. 2A, D). The quartz (0.3–1.4 mm) is subhedral to anhedral, and locally exhibits undulose extinction. The albite (An_{10-20}) (0.5–3 mm) is subhedral, and displays a prismatic crystal habit, except when in alteration masses. The potassic feldspar (0.3–3 mm) is subhedral to anhedral.

Biotite (0.1–2 mm) occurs as subhedral, brownish lamellae, and its pleochroism varies from light-to-dark brown (Fig. 3A, B). It is classified based on the cation contents of the normalized structural formula at 22 oxygen atoms and all iron expressed as Fe^{2+} , and shows TiO_2 (1.32–2.68 %), Al_2O_3 (14.61–16.99 %), and K_2O (8.08–10.10 %) values.

According to the $\text{FeO}-(10*\text{TiO}_2)\text{-MgO}$ diagram they are primary magmatic biotites (Fig. 4A; Suppl. Data Table 1), are in peraluminous field (Fig. 4B) and indicates annite classification (Fig. 4C).

The muscovite (0.5–1.0 mm) is euhedral lamellae (Fig. 3B). In addition to the textural aspect, the mineral chemistry of muscovite indicates its primary (TiO_2 0.79–11.462 %; K_2O 9.424–10.54 %; Al_2O_3 29.76–32.041 %) and secondary (TiO_2 0–0.691 %; K_2O 9.16–10.15 %; Al_2O_3 28.94–32.22 %) origins (Fig. 4C; Suppl. Data Table 2). Titanite (0.1–0.5 mm) is reddish brown and occurs in lenticular or diamond-shaped forms as well as granular habits.

The syenogranite is reddish and/or pinkish, and portions of milky quartz spread throughout its body (Fig. 2B, E). Quartz (0.3–1.5 mm) is subhedral to anhedral, with undulose extinction and jagged to lobular boundaries. The plagioclase (An_{10-20}) (0.5–4 mm), occurs as subhedral lamellae with a prismatic crystal habit. Orthoclase (0.3–4 mm), subhedral to anhedral.

Few preserved biotite lamellae were found in the syenogranite samples (0.1–1.5 mm), and it's occurs as subhedral, brownish lamellae (Fig. 3C), most of which are altered to chlorite, epidote, and muscovite. The mineral chemistry data (TiO_2 1.70–1.96 %; Al_2O_3 13.01–14.56 %; K_2O 9.14–9.68 %) indicates primary magmatic to re-equilibrated biotites (Fig. 4A), fall into the field of calc-alkaline suite bioites (Fig. 4B) and is classified as annite (Fig. 4C).

The muscovite (0.4–1.0 mm) is a euhedral lamellae with sizes ranging from. The textural relationships and chemical analysis (TiO_2 0–0.69 %; K_2O 9.436–9.71 %; Al_2O_3 30.39–32.22 %) indicate the secondary origin.

The biotite granodiorite is medium-grained, and gray–white (Fig. 2F). Quartz (0.2–3 mm) occurs similar to a subhedral to euhedral habit, and is recrystallized with plagioclase, potassic feldspar, and alteration minerals (chlorite + sericite + muscovite + epidote) locally.

Plagioclase crystals (An_{40-30} tending to An_{25}) (0.5–3 mm) are well-preserved, locally sericitized, and fractured. The orthoclase (0.2–3 mm) displays Carlsbad twinning, and, locally, albite–pericline. It is fractured but shows minor alteration and locally occurs as a mineral aggregate with plagioclase.

Biotite (0.2–1 mm) is brown, its pleochroism ranges from light-to-dark brown (Fig. 3D). It occurs in paragenesis with quartz, plagioclase, and potassic feldspar and,

frequently, inside the masses of opaque minerals (magnetite) alongside amphibole. The textural relationships and chemical analyze (TiO_2 0.9–1.87 %; Al_2O_3 4.22–16.18 %; K_2O 9.11–10.10 %) indicate magmatic origin, with some samples at the limit or in the re-equilibrated field (Fig. 4A) and according to the $\text{FeO}^*\text{--MgO--Al}_2\text{O}_3$ biotite discrimination diagram, the samples are in the calc–alkaline field and are classified as annite (Fig. 4B, C).

Titanite (0.2–1 mm), subhedral, with a brownish-red color, and occurs along veinlets and in paragenesis with biotite and hornblende. It was classified as primary due to its textural aspects and chemical data (TiO_2 27.90–32.70 %; Al_2O_3 1.80–3.25 wt.%; CaO 24.16–26.22 %) (Suppl. Data Table 3).

Amphibole (0.5–2 mm) is anhedral to subhedral, with dark green to light green pleochroism. Locally, basal section is observed (Fig. 4D); and, also, amphibole aggregates containing biotite, magnetite, and, more rarely, pyrite. The values from Al_2O_3 (9–10.88 %), FeO (16.25–18.33 %), MgO (9.68–11.17 %), and CaO (10.56–11.04 %), joint with the textural aspects, indicate primary origin and an mg-rich hornblende classification (Suppl. Data Table 4; Fig. 4D).

The monzonite is a medium-grained, mesocratic, and gray–white (Fig. 2G). Plagioclase crystals (An_{40-30} tending to An_{25}) (3–0.5 mm) are subhedral and are well-preserved, but locally sericitized and fractured. K-feldspar (2–0.2 mm) displays Carlsbad twinning and, locally, albite–pericline (Fig. 3E). They are fractured but show little alteration and locally occur as mineral aggregates with plagioclase. Quartz (1–0.2 mm) occurs as a subhedral to euhedral habit. In general, it contains sutured contact. Locally, it recrystallizes with plagioclase and potassium feldspar. Epidote crystals (2–1 %) (0.5–0.1 mm) are subhedral to anhedral and are secondary and associated with biotite and amphibole.

Biotite lamellae (0.2–1 mm), is brown, with pleochroism from light-to-dark brown. They occur in paragenesis with hornblende and titanite, including magnetite and pyrite. The textural and chemistry data (TiO_2 1.38–2.63 %; Al_2O_3 14.40–16.45 %; K_2O 9.63–10.10%) shows primary origin (Fig. 4A, field A), are calc–alkaline (Fig. 4B) and classified as annite (Fig. 4C).

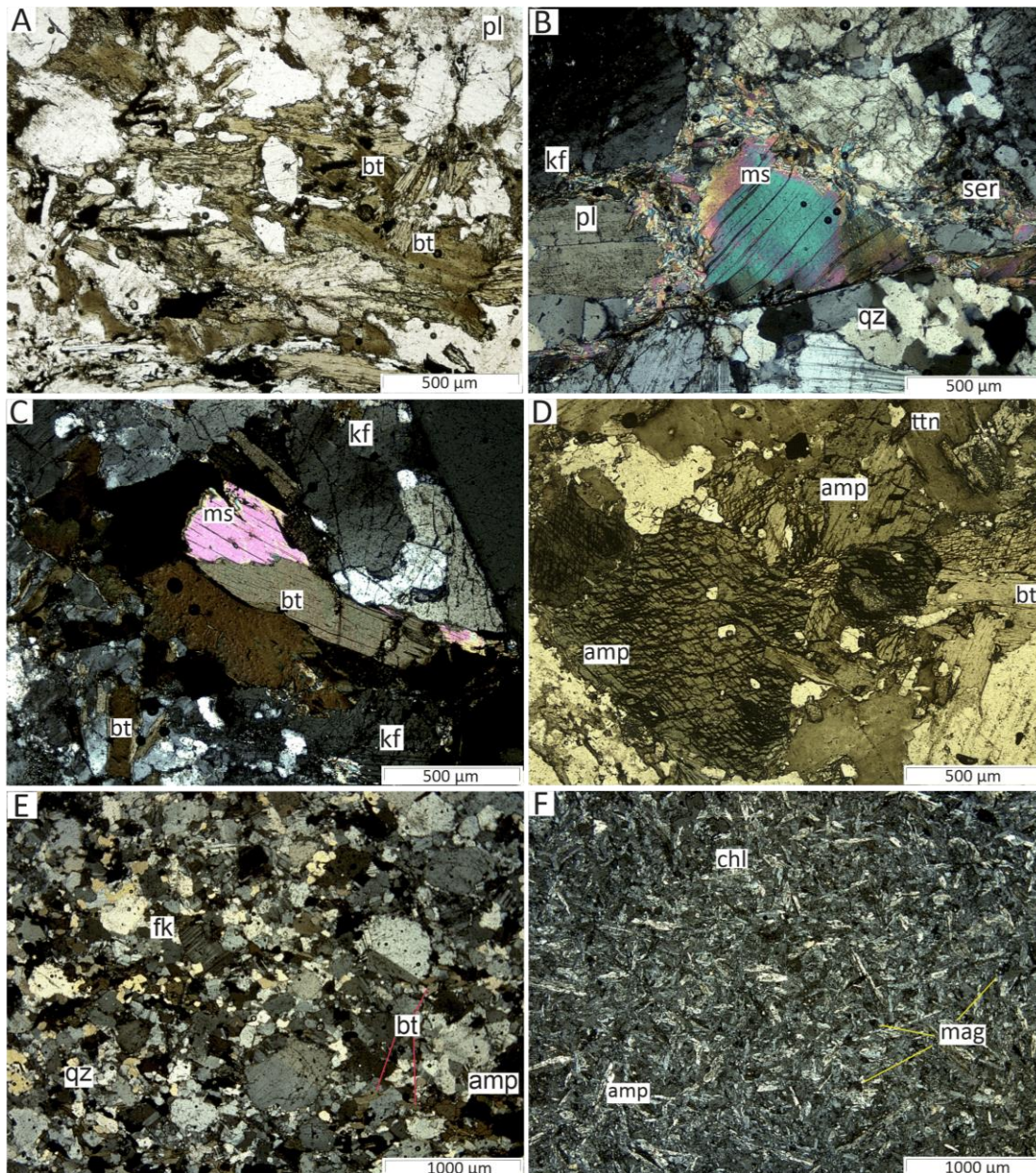


Fig. 3. Microscopic characteristics of the Novo Mundo Complex rocks. A) Monzogranite, highlighting the biotite lamellae; B) Primary muscovite lamellae associated with quartz, plagioclase in monzogranite; C) Biotite lamellae contacting muscovite, K-feldspar, plagioclase and quartz in syenogranite samples; D) Amphibole associated with biotite and titanite in granodiorite; E) Texture of monzonite: biotite, hornblende, plagioclase, and quartz association; F) Texture of the diabase dikes containing plagioclase laths and interstitial amphibole.

Titanite (0.2–1 mm), occur along the veinlets and in paragenesis with biotite and hornblende. The textural aspects and chemical data (TiO_2 31.59–35.22 %; Al_2O_3 1.31–1.91 wt.%; CaO 25.01–25.97 %) classified as primary origin. The amphibole (0.2–1.5 mm) is anhedral to subhedral, with dark green to light green pleochroism (Fig. 4G, H). The mineral chemistry data (Al_2O_3 9.08–10.24 %; and FeO 18.54–19.78 %; MgO 9.13–

9.73 wt.%) shows it is classified as Mg-rich hornblende (Fig. 4D).

The mafic dikes are isotropic, fine-grained and, locally, contain plagioclase phenocrysts that are dark green and strongly magnetic. Comprise plagioclase (40–50 %; An_{20-50}), calcium amphibole (pargasite) (15–35 %), pyroxene (10–15 %), biotite already altered to chlorite (8–10 %), magnetite (10–15 %), titanite (~1 %), and chlorite (15–20 %) (Fig. 2C, H).

Plagioclase (0.5–4 mm) occurs in a prismatic, elongated as phenocrystals immersed in the rock matrix. It is, in general, well-preserved but locally sericitized. Magnetite (0.1–1 mm) is subhedral, locally fractured, and disseminated throughout the entire rock. Amphibole (0.2–3 mm) is dark green with a moderate relief. Locally, it is altered to chlorite and carbonate, which occur as agglomerates with other minerals (biotite, pyroxene, magnetite, and, locally, pyrite) within alteration masses (Fig. 3F). Biotite (0.1–1.5 mm) is brownish and has pleochroism varying from light-to-dark brown. It is altered to chlorite and muscovite.

Pyroxene (0.2–3 mm) is subhedral and brownish to reddish brown, with pleochroism between light and dark brown. It is sometimes completely altered to biotite. Magnetite (0.1–1 mm) is subhedral, locally fractured, and disseminated throughout the entire rock. Titanite (0.20–0.80 mm) occurs as anhedral to subhedral crystals, dispersed throughout the rock or associated with the amphibole. Its mean Al_2O_3 and Fe_2O_3 values are 1.22 wt.% and 1.33 wt.%, respectively. For the amphibole, FeO (6–13.21%), MgO (13.85–15.81 %) and CaO (9.64–11.78 %), respectively, and it is classified as pargasite (Suppl. Data Table 5, Fig. 4D).

4.3 Geochemistry

4.3.1 Granites

The whole rock analysis results of representative studied rocks are founded Table 1. The granitic samples used for geochemical study have L.O.I. values mainly less than 1.5 %. The monzogranite and the syenogranite have SiO_2 values of 73.10–75.31 % and mean Al_2O_3 value is 12.85 %. The biotite granodiorite has an average SiO_2 of 67.1 % and Al_2O_3 is 15.48 %; monzonite has an average SiO_2 61.3 wt.%, and Al_2O_3 is 17.32 %. The average total REE content is 187.40 and 158.76 ppm for monzogranite and syenogranite, respectively; 279.70 ppm for granodiorite, and 228.10 ppm for monzonite.

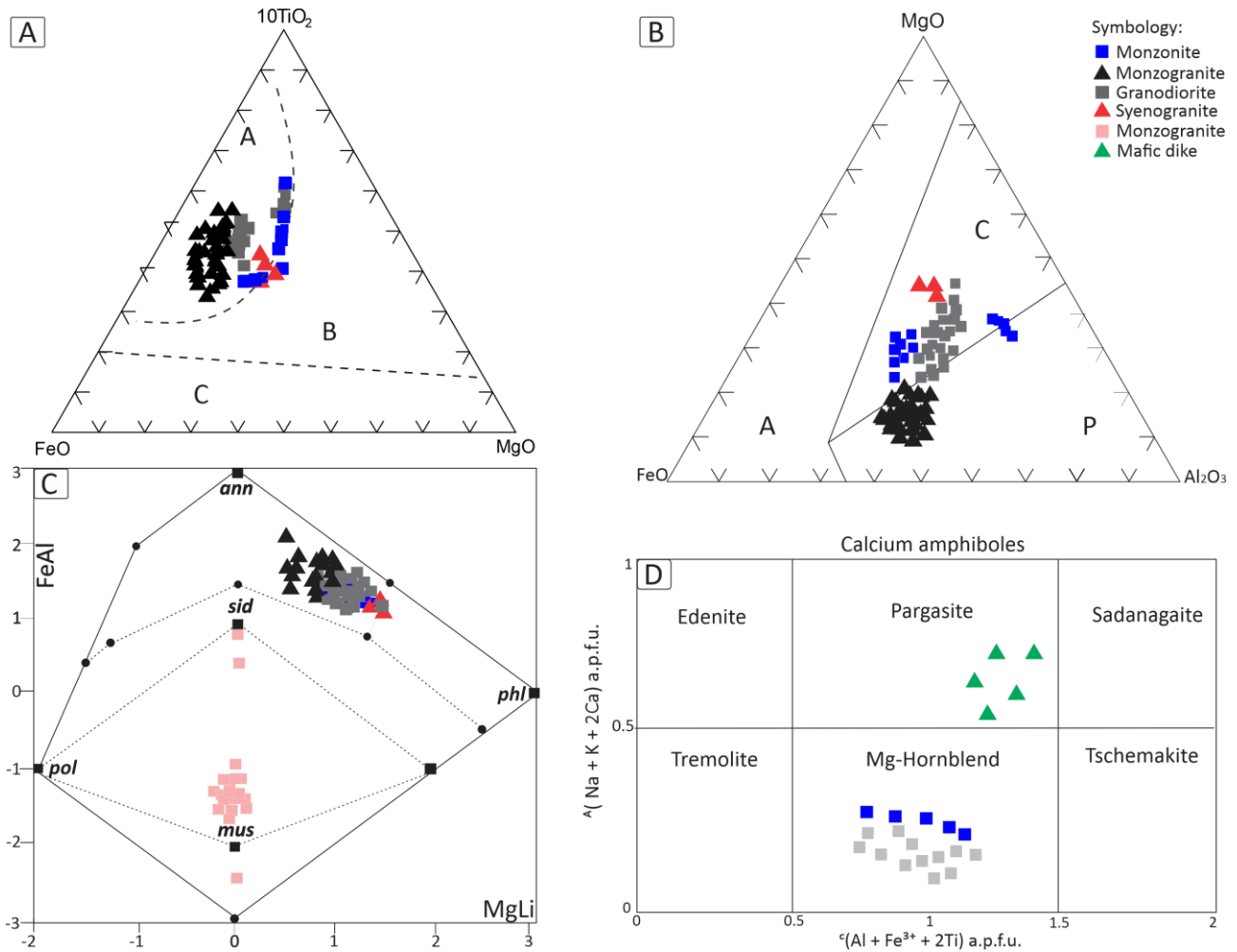


Fig. 4. Biotite, muscovite, and amphibole composition of the Novo Mundo granite. A) FeO – $(10 \times \text{TiO}_2)$ – MgO by [Nacht et al. \(2005\)](#), (A) domain of the primary magmatic biotites, (B) re-equilibrated biotites and (C) neoformed biotites; B) The FeO^* – MgO – Al_2O_3 biotite discrimination diagram ([Abdel-Rahman 1994](#)) for alkaline (A), peraluminous (P), and calc-alkaline (C) suites; C) MgLi vs. FeAl mica classification diagram by [Tischendorf et al. \(2007\)](#), placing the three mica species in a solid solution. The samples fall mainly in the annite field, for biotites. The primary muscovite from monzogranite samples plot in the muscovite field. Abbreviations: Ann: annite; Phl: phlogopite; mus: muscovite; pol: polyolithionite; Sid: siderophyllite; D) [Hawthorne et al. \(2012\)](#) classification diagram for calcium amphiboles.

Table 1. Lithogeochemistry data for Novo Mundo plutonic rocks.

Sample ID																				
Rock (%)	Monzogranite				Syenogranite									Biotite granodiorite				Monzonite		
	F06-02	F06-05	LNM-25	LNM-51	F06-01	F06-09	F70-06	F73-09	F75-10	LNM-46	DNFD-03	DNFD-15	DNFD-19	F64-10	DNFD 01-B	DNFD 01-C	DNFD 01-D	F06-10	F17-07	DNFD 01-A
SiO ₂	75.7	74.2	71.8	69.5	74.1	73.1	75.8	77.1	74.5	75.9	74.8	77.0	76.5	61.7	67.9	69.0	69.8	62.8	63.7	57.5
TiO ₂	0.1	0.1	0.4	0.5	0.1	0.4	0.1	0.1	0.1	0.3	0.3	0.2	0.2	0.7	0.5	0.6	0.5	0.7	0.7	0.7
Al ₂ O ₃	12.3	12.4	14.6	14.0	12.5	13.8	12.9	12.7	13.1	12.1	12.2	13.0	13.2	16.4	16.1	15.0	14.5	17.4	16.6	18.0
Fe ₂ O ₃	1.0	1.2	3.2	3.4	1.1	2.4	1.2	1.2	1.5	1.4	1.5	0.8	0.7	6.1	3.8	4.7	3.6	5.3	5.8	6.0
MnO	0.0	0.0	0.1	0.1	0.0	0.1	0.0	0.0	0.0	0.1	0.0	0.0	0.0	0.2	0.1	0.1	0.1	0.1	0.1	0.2
MgO	0.5	0.5	0.6	0.7	0.4	0.7	0.3	0.4	0.4	0.4	0.4	0.5	0.9	3.0	1.4	1.1	1.1	1.5	1.6	2.2
CaO	0.9	0.8	1.9	2.0	0.8	1.2	0.7	0.6	0.4	0.8	1.0	0.4	0.2	5.0	3.2	2.6	2.3	3.6	4.2	4.7
Na ₂ O	2.8	1.3	3.0	2.8	3.2	3.1	3.2	3.0	3.1	2.9	2.8	3.3	1.7	3.9	4.1	2.7	2.8	4.9	4.1	4.7
K ₂ O	5.2	6.3	4.6	4.1	5.1	4.9	5.4	5.3	5.4	4.7	4.9	5.4	7.2	1.7	2.6	4.2	4.7	2.6	3.6	2.4
P ₂ O ₅	0.0	0.0	0.1	0.2	0.0	0.1	0.1	0.0	0.1	0.1	0.1	0.0	0.0	0.2	0.2	0.1	0.1	0.3	0.4	0.3
SrO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	<0.01	0.1	0.0	0.0	0.0	0.0	0.1	0.1
BaO	0.0	0.0	0.1	0.1	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	0.1	0.0	0.1	0.1	0.1	0.0	0.1	0.1
LOI	1.6	1.7	0.7	1.1	1.3	1.3	1.0	1.0	1.0	1.0	1.8	0.9	1.1	1.4	1.1	1.2	1.6	1.1	0.7	1.7
Total	100.0	98.6	101.1	98.3	98.6	101.0	100.8	101.5	99.5	99.6	99.6	101.6	101.7	100.3	100.9	101.4	101.0	100.1	101.7	98.4
ppm																				
Ba	238.0	387.0	767.0	850.0	276.0	928.0	644.0	315.0	257.0	366.0	374.0	409.0	487.0	346.0	566.0	1010.0	815.0	327.0	1385.0	488.0
Sr	45.6	38.1	146.5	173.5	58.2	277.0	92.6	62.5	51.5	140.0	85.2	56.7	35.0	519.0	323.0	229.0	268.0	244.0	566.0	498.0
Rb	149.5	208.0	214.0	170.5	191.5	157.5	200.0	166.5	181.0	191.0	194.0	221.0	253.0	105.5	105.0	181.0	196.5	118.5	115.0	94.1
Zr	117.0	118.0	189.0	176.0	117.0	276.0	69.0	108.0	72.0	169.0	180.0	156.0	159.0	115.0	161.0	261.0	231.0	280.0	228.0	317.0
Y	4.3	6.5	22.0	22.5	6.3	4.4	9.9	5.2	10.3	28.1	27.3	17.4	14.8	52.9	12.2	17.4	15.4	15.5	29.4	50.7
Nb	4.1	2.3	12.6	14.0	5.0	5.3	4.2	5.1	2.4	19.3	16.1	14.2	14.5	19.2	8.3	12.3	9.7	9.3	68.5	15.3
Ga	14.5	18.4	26.8	24.7	17.8	14.2	15.1	16.3	20.4	18.9	18.6	19.3	19.8	23.1	24.4	25.9	23.6	24.5	22.0	29.6
Th	30.2	45.7	18.4	19.7	17.0	30.2	12.4	23.0	9.9	24.6	22.3	24.4	24.0	4.8	1.7	36.6	66.0	2.0	13.6	5.0
Cs	1.3	1.8	4.1	1.6	1.3	1.3	0.7	0.7	0.9	0.5	0.7	1.6	2.2	5.0	2.2	2.4	1.4	4.2	1.7	3.1
Hf	4.7	4.4	5.5	5.2	4.7	8.5	2.6	4.0	3.8	5.4	6.3	4.7	4.7	4.3	4.6	7.0	7.1	7.2	6.4	8.8
Ta	0.3	0.2	1.3	1.4	0.5	0.5	0.4	0.4	0.4	2.5	2.3	1.9	1.8	1.5	1.1	1.0	0.9	0.5	3.1	1.4
U	3.5	4.4	2.7	1.3	3.6	1.8	1.9	11.0	3.7	3.1	2.5	1.7	2.6	2.5	0.9	3.0	2.8	1.2	3.6	2.0
V	15.0	19.0	27.0	33.0	22.0	33.0	22.0	14.0	12.0	14.0	15.0	8.0	7.0	137.0	46.0	49.0	54.0	76.0	108.0	101.0
C	0.2	0.2	0.0	0.0	0.2	0.1	0.1	0.1	0.1	0.1	0.2	0.1	0.0	0.1	0.0	0.0	0.1	0.1	0.0	0.1
Sn	0.0	0.0	4.0	3.0	1.0	2.0	2.0	1.0	2.0	3.0	3.0	2.0	2.0	6.0	2.0	4.0	3.0	3.0	5.0	5.0
As	0.2	1.8	0.3	0.3	0.2	0.3	0.1	0.3	1.0	0.3	0.6	0.3	0.8	1.0	0.5	0.3	0.6	0.2	0.5	0.7

Cu	1.0	4.0	15.0	36.0	2.0	2.0	2.0	2.0	6.0	1.0	6.0	2.0	3.0	1.0	22.0	18.0	27.0	13.0	10.0	17.0
Ni	2.0	3.0	5.0	8.0	1.0	6.0	1.0	1.0	10.0	1.0	1.0	1.0	2.0	22.0	4.0	12.0	3.0	3.0	4.0	6.0
Pb	18.0	24.0	29.0	26.0	36.0	18.0	26.0	31.0	31.0	25.0	14.0	19.0	16.0	6.0	17.0	38.0	37.0	15.0	19.0	20.0
Sc	1.0	1.0	8.0	7.0	2.0	2.0	1.0	2.0	3.0	2.0	2.0	3.0	3.0	24.0	11.0	10.0	6.0	4.0	12.0	18.0
Zn	10.0	17.0	65.0	57.0	18.0	31.0	15.0	15.0	12.0	27.0	10.0	16.0	16.0	94.0	60.0	74.0	54.0	71.0	74.0	82.0
La	48.0	26.0	44.9	52.4	21.9	74.8	36.9	29.5	8.5	48.8	46.2	50.3	42.3	23.8	29.1	84.5	107.5	21.0	81.9	30.5
Ce	91.8	57.4	88.9	105.5	43.6	139.0	68.1	56.4	16.1	97.5	90.3	89.3	80.2	59.8	53.5	167.5	212.0	43.4	165.5	75.1
Pr	8.7	6.2	10.0	11.4	4.6	13.2	6.9	5.7	1.8	10.3	9.8	8.5	7.9	8.3	6.0	17.9	22.9	5.1	17.8	10.5
Nd	31.2	21.8	35.0	40.6	16.6	40.4	23.8	19.8	7.6	35.2	33.1	25.9	24.6	37.1	21.8	64.0	78.6	20.6	66.9	45.1
Sm	4.3	3.7	6.9	7.9	3.8	4.4	4.6	4.1	2.8	6.3	5.6	3.8	3.7	10.5	3.8	10.8	12.9	4.2	11.8	10.8
Eu	0.4	0.4	1.1	1.1	0.4	0.9	0.8	0.4	0.6	0.7	0.7	0.5	0.7	1.6	1.0	1.4	1.3	1.0	2.6	1.3
Gd	2.4	2.6	5.6	6.3	3.2	1.8	3.7	3.0	2.6	5.0	4.4	2.7	2.9	9.6	3.0	7.7	9.1	4.0	9.0	9.7
Tb	0.2	0.3	0.9	0.8	0.4	0.3	0.6	0.4	0.6	0.8	0.7	0.4	0.4	1.6	0.5	1.0	1.0	0.5	1.2	1.5
Dy	1.1	1.2	4.4	4.7	1.7	1.2	2.8	1.6	2.7	4.7	4.5	2.4	2.3	10.0	2.3	4.3	4.5	3.2	6.5	8.9
Ho	0.2	0.2	0.8	0.8	0.3	0.3	0.5	0.3	0.6	1.0	1.0	0.5	0.5	2.0	0.5	0.7	0.7	0.6	1.1	1.8
Er	0.3	0.7	2.1	2.1	0.9	0.9	1.4	0.8	2.2	2.9	2.7	1.5	1.6	6.8	1.2	1.6	1.3	1.6	3.3	4.9
Tm	0.1	0.1	0.3	0.4	0.1	0.2	0.2	0.1	0.4	0.5	0.4	0.3	0.3	1.1	0.2	0.2	0.2	0.2	0.4	0.8
Yb	0.4	0.5	1.8	2.1	0.6	0.9	1.0	0.6	2.1	2.9	2.8	2.0	1.9	6.4	1.1	0.9	0.9	1.3	2.8	4.6
Lu	0.0	0.0	0.3	0.3	0.1	0.2	0.2	0.1	0.4	0.5	0.4	0.4	0.3	0.9	0.2	0.2	0.1	0.2	0.4	0.8
Lu _N	0.9	1.2	7.7	9.4	3.5	5.0	5.3	2.9	11.2	12.9	10.9	9.4	13.8	27.4	4.4	4.4	4.1	22.7	5.3	12.1
Eu/Eu*	0.4	0.4	0.6	0.5	0.3	1.0	0.6	0.3	0.7	0.4	0.5	0.6	0.4	0.5	0.9	0.5	0.4	0.4	0.8	0.8
La _N /Yb _N	88.9	36.9	16.3	16.6	23.2	54.2	24.4	32.8	2.8	10.9	16.5	14.8	11.2	2.5	18.3	61.2	84.3	4.4	10.9	19.4
La _N /Sm _N	6.8	4.4	4.0	4.1	3.6	10.6	4.9	4.4	1.9	5.1	8.1	7.0	4.8	1.4	4.7	4.8	5.2	1.7	3.1	4.3
Ce _N /Yb _N	64.9	31.1	12.3	12.8	17.6	38.4	17.2	23.9	2.0	8.1	11.2	10.7	8.5	2.4	12.8	46.3	63.4	4.2	8.6	14.9
Ce _N /Sm _N	4.0	3.0	3.0	3.0	2.0	7.0	3.0	3.0	1.0	3.0	5.0	5.0	3.0	1.0	3.0	3.0	3.0	1.0	2.0	3.0
Eu _N /Yb _N	99.0	67.0	3.0	13.0	70.0	50.0	46.0	21.0	35.0	78.0	50.0	6.0	65.0	34.0	29.0	64.0	87.0	63.0	44.0	29.0

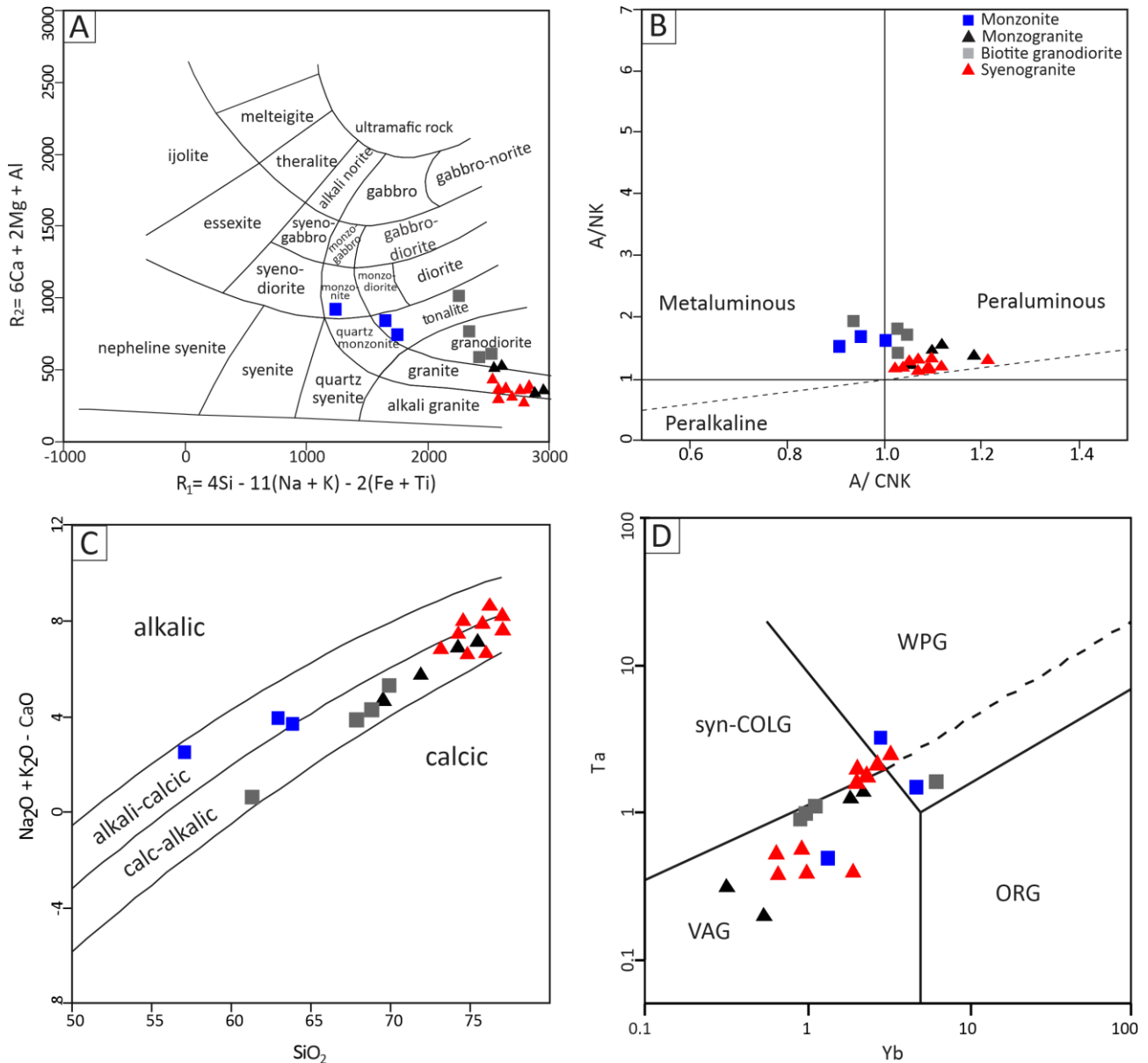


Fig. 5. Diagram classification and geotectonic approach of Novo Mundo plutonic rocks. A) R_1 – R_2 classification diagram by De La Roche *et al.* (1980) that places the samples between the monzonite–tonalite–granodiorite and granite fields. B) Aluminum Saturation Index (Maniar and Piccoli 1989). C) $Na_2O + K_2O - CaO$ vs. SiO_2 classification diagram by Frost & Frost (2008): granite samples are placed mainly in the calc-alkaline rocks, whereas monzonite is placed in alkali-calcium. D) Ta vs. Yb geotectonic classification diagram by Pearce *et al.* (1984). The Novo Mundo samples are predominantly in the volcanic arc field (VAG). Abbreviations: (Oceanic Ridge Granites (ORG); Syn- and Post-collisional Granites (Syn-COLG); Within Plate Granites (WPG); Volcanic Arc Granites (VAG).

In the R_1 – R_2 diagram (De La Roche *et al.* 1980; Fig. 5A), syenogranites and monzogranites are classified as granite and alkali granite, respectively.

Three samples of the granodiorite are situated in the field of granodiorite in Fig. 5A, and one sample is in the tonalite field. The monzonite samples fall in the monzonite/tonalite fields.

According to the [Maniar and Piccoli \(1989\)](#) diagram (Fig. 5B), the Novo Mundo granites are metaluminous to peraluminous, with $(\text{Al}_2\text{O}_3/\text{CaO} + \text{NaO}_2 + \text{K}_2\text{O})_{\text{molar}}$ varying from 0.7 to 1.23. The syeno- and the monzogranite have a predominant peraluminous character in the alumina saturation index diagram, whereas the granodiorite and monzonite are displayed between the peraluminous to metaluminous fields.

The granites are mainly classified as calc-alkaline in the diagram by [Frost & Frost \(2008\)](#) and the monzonite samples lie in the alkali-calcic field (Fig. 5C). In the Rb vs. Y + Nb diagram (Fig. 5D), the samples are predominantly in the volcanic arc field, whereas the monzonite samples are in the within-plate field.

Figure 6 shows the correlation between major and trace elements vs. SiO_2 values from the Novo Mundo granites. Figures 6A–B show the inverse correlation between MgO and TiO_2 vs. SiO_2 for all granitic rocks. Despite some dispersion of samples in the trace element diagram, important distinctions between the rocks are evident in the Figs. 6C–6J.

According to the REE pattern, the distribution is more homogeneous for the granodiorite and monzonite facies. The normalized Eu (Eu/Eu^*) anomalies were calculated following the method of Nakamura (1974). The negative Eu anomalies for monzogranite and syenogranite are 0.46 and 0.54, respectively; for granodiorite is 0.55; and for monzonite, 0.66.

The total REE values are 236.39–121.10 ppm in the monzogranite, 217.74 – 49.00 ppm for the syenogranite, 452.76–123.97 ppm for the granodiorite and 371.19–106.88 ppm for the monzonite (Fig. 7A–D). The diagram reveals a moderate to high pattern of fractionation of light REEs (LREE) in relation to heavy REEs (HREE), indicating the following ratios for La/Yb_N and $(\text{Gd}/\text{Yb})_N$, respectively: 39.66 and 1.04 for monzogranite, 21.18 and 0.99 for syenogranite, 41.58 and 4.65 for granodiorite, and 11.55 and 2.22 for monzonite.

4.3.2 Diabase dikes

Lithogeochemical data for Novo Mundo mafic dikes are shown in supplementary data Table 6. The diabase dikes have SiO_2 value ranges of 46.1–57.7 %, 4.9–12.55 % MgO, 6.7–15.15 % FeO_t , 5.4–7.99 % CaO, 0.5–2.73 % TiO_2 , 2.10–3.50 % Na_2O and 0.81–2.46 % K_2O . Ni is 85–288 ppm and #mg ($\text{MgO}/(\text{MgO} + \text{FeO}^*)$) is 0.313–0.538.

They are calc-alkaline, classified as andesite and basalt, according to the TAS diagram

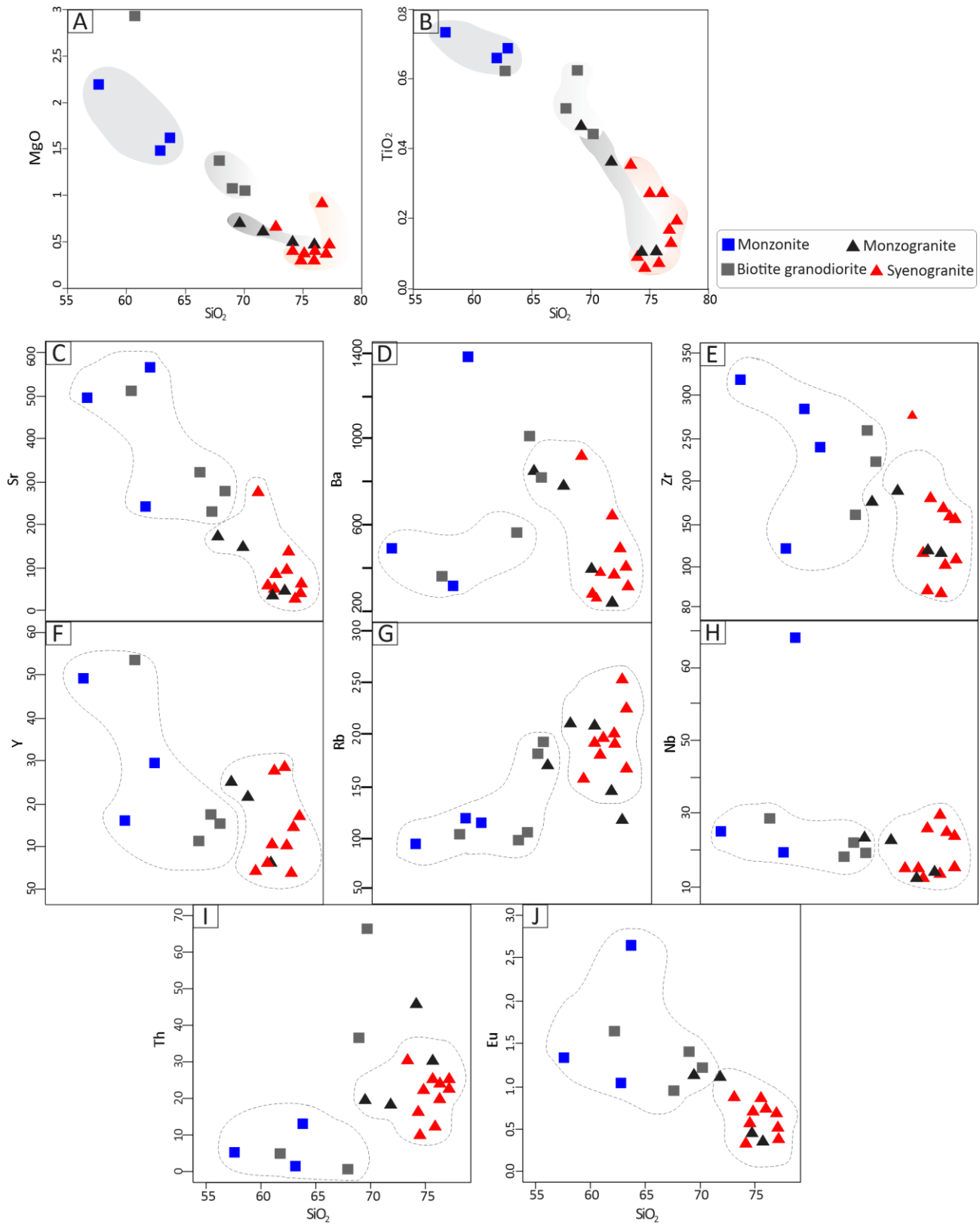


Fig. 6. Harker-type variation diagram of major and trace elements vs. SiO_2 for the Novo Mundo granites.

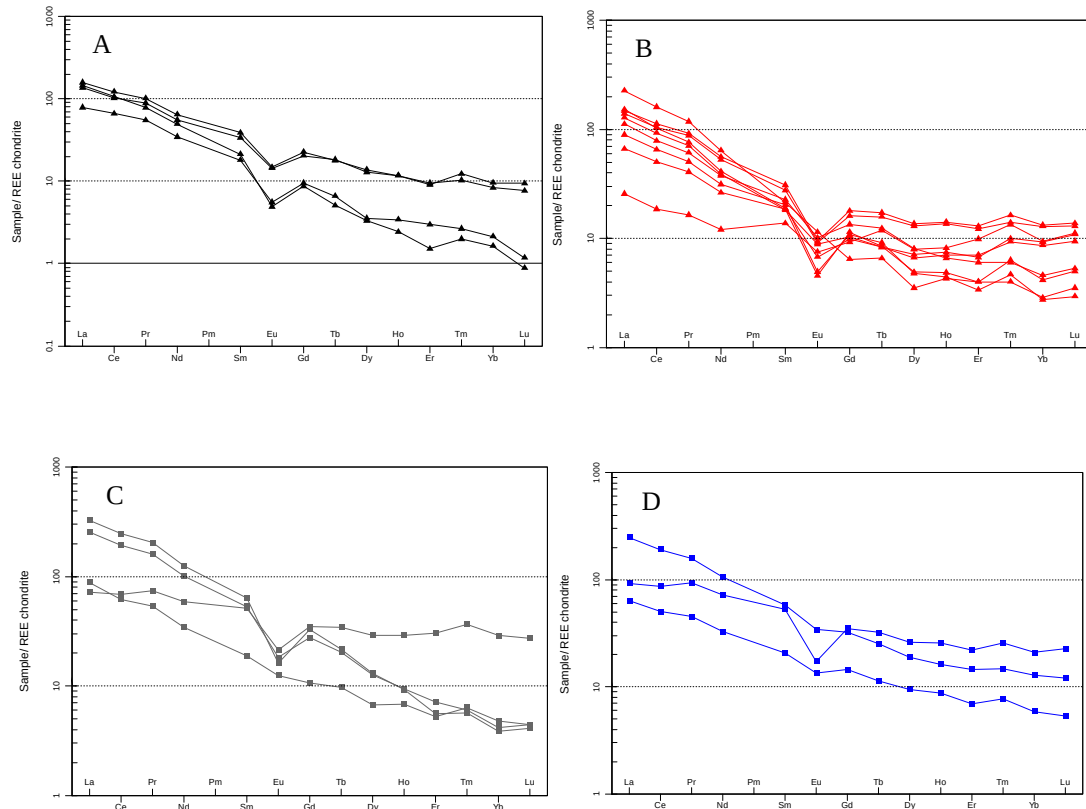


Fig. 7. Distribution patterns of normalized rare earth elements (Nakamura 1974) for the four granitic facies from the Novo Mundo Complex. Symbology as in Fig. 6.

(Cox *et al.* 1979) and to Pearce (1995) trace elements classification diagram (Fig. 8A). The distribution of Large-ion lithophile elements (LILEs) from Novo Mundo diabase dikes samples is displayed in Fig. 8D. All samples are depleted in Ta, Nb, Hf, and Ti. In the range between Gd and Yb there is a gradual and subtle depletion in all samples. Elements such as Th, La, Sm, and Gd are enriched in all samples.

4.4. Geochronology and geochemistry of radiogenic isotopes

4.4.1. U–Pb

The U–Pb isotopic data from the four selected Novo Mundo granitic rocks are summarized in supplementary data Table S7 (Supplementary data).

For the monzonite (F06-10), eight zircon prismatic crystals from sample F06-10 were analyzed. They are light-to-dark brown short prisms are, with a mean size smaller than 0.1 mm, moderately fractured, metamict, with moderate luminescence in backscattered electron (BSE) images. The U–Pb concordia diagram reveals an upper intercept age of

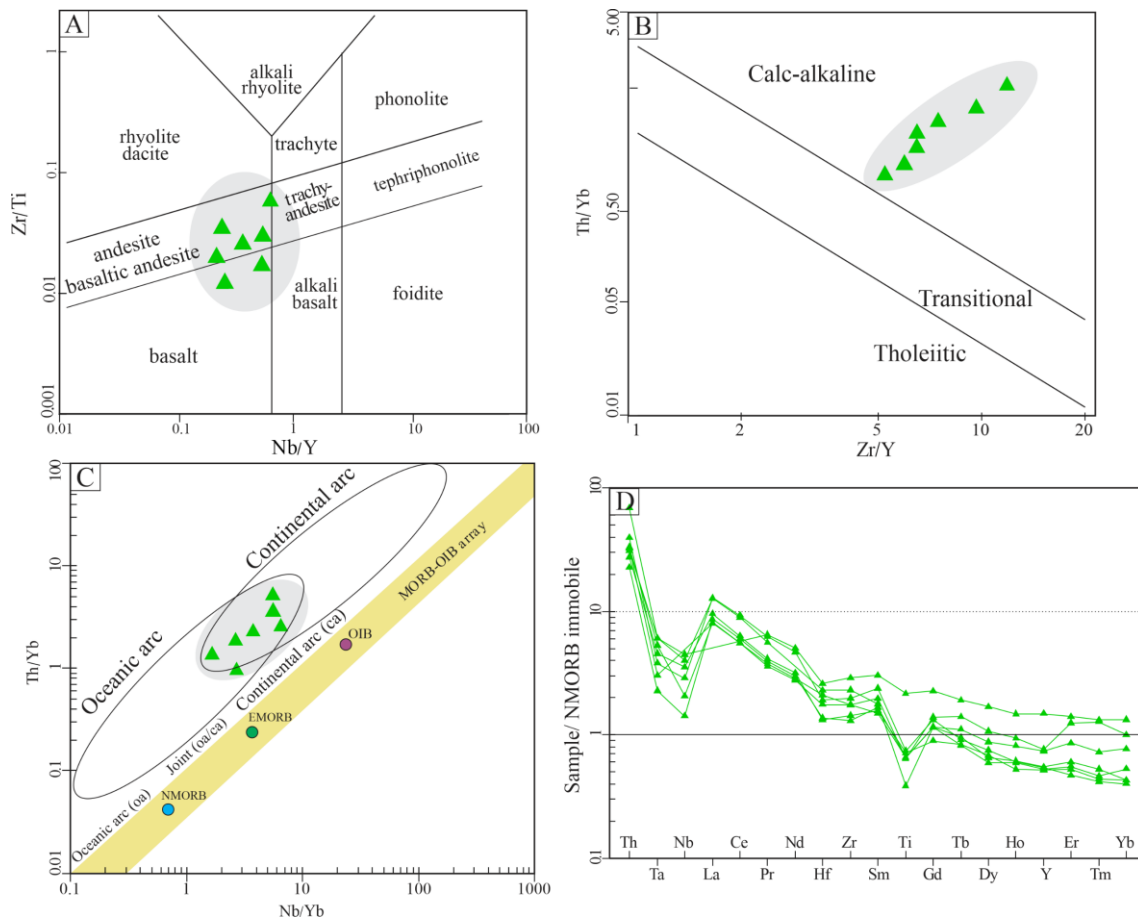


Fig. 8. Geochemical diagrams for the Novo Mundo Complex dikes. A) Nb/Y vs. Zr/Ti classification diagram, modified from the original (Pearce 1995): samples show basalt and basaltic–andesite and andesite composition; B) Tectonic environment discrimination diagram: Nb/Yb vs. Th/Yb from Pearce (2014). The samples are in the continental arc sector (CA); C) Diagram Zr/Y vs. Th/Yb by Ross & Bédard (2009): the samples are placed in a calc–alkaline basalt environment; D) Distribution diagram of trace elements normalized to N-MORB (Sun & McDonough 1989), in Pearce (2014).

2032.4 ± 6.6 Ma (MSWD = 0.01, Fig. 9A), which is interpreted as the crystallization age.

U–Pb monzogranite (LNM-25) data were obtained from twenty-five zircon prismatic crystals. These crystals dominantly had bipyramidal prismatic habit and were sometimes rounded, with a mean size smaller than 0.1 mm. They are brownish and show a moderate degree of metamictization, albeit preserving their igneous characteristics; they are without intergrowth features and are slightly zoned. The U–Pb results yield an upper intercept age of 2029.9 ± 4.1 Ma (MSWD = 0.5, Fig. 9B), which is interpreted as the crystallization age.

However, four zircon crystals display older ages than the average value calculated from the main population (2151 ± 6.0 Ma to 2070 ± 5.4 Ma). These data are probably

related to inherited zircon from older events in the region.

The biotite granodiorite (F64-10) had eleven prismatic zircon crystals analyzed, with a mean size smaller than 0.1 mm. They are moderately fractured, metamict and exhibit regular oscillatory zoning with moderate luminescence in BSE images. The U–Pb concordia diagram reveals an upper age of 1989.8 ± 6.7 Ma (MSWD = 1.3, Fig. 9C), interpreted as the crystallization age.

Twenty-nine prismatic zircon crystals from the syenogranite (LNM-46) were analyzed. They are bipyramidal, also with subhedral and anhedral habits, and a mean size smaller than 0.1 mm. The crystals are dark brown and translucent to subtranslucent, with moderate degree of metamictization and moderate luminescence in BSE images. The U–Pb upper intercept age of 1987 ± 7.4 Ma (MSWD = 1.5, Fig. 9D), interpreted as the crystallization age.

4.4.2. Isotopic geochemistry of Sm–Nd

Table 2 shows the Sm–Nd isotopic data for the studied granitic rocks. The monzonite shows a $T_{DM} = 2.14\text{--}2.52$ Ga and $\epsilon Nd(t) = +1.81$ and -2.0 , respectively. The two monzogranite samples have a $T_{DM} = 2.27\text{--}2.28$ Ga, and the isotopic signature is slightly positive, between $\epsilon Nd(t) = +0.16$ to $+0.22$. The T_{DM} age of the biotite granodiorite is $2.19\text{--}2.48$, and its $\epsilon Nd(t)$ is slightly different at $+0.44$ and -1.39 , respectively. The three syenogranite have a T_{DM} between 2.24 and 2.27 Ga and a negative $\epsilon Nd(t)$ between -1.12 and -0.03 .

5. Discussion

5.1. Petrogenetic model for the Novo Mundo complex.

The new geological, petrological and geochronological data presented in this paper reveal the existence of four types of granitic rocks that crystallized from 2.03 to 1.99 Ga in the northern Juruena Mineral Province, south of the Amazonian Craton, composing the Novo Mundo granitic complex. The obtained geochemical and Sm–Nd data are consistent with different episodes of granite formation and mixing of crustal and mantle sources. The T_{DM} ages are interpreted as an average guide for the separation of the bulk mineral content from a parent reservoir, probably being the result from mixing materials generated from sources of different ages and isotopic compositions (DePaolo 1981, Motta *et al.* 2022).

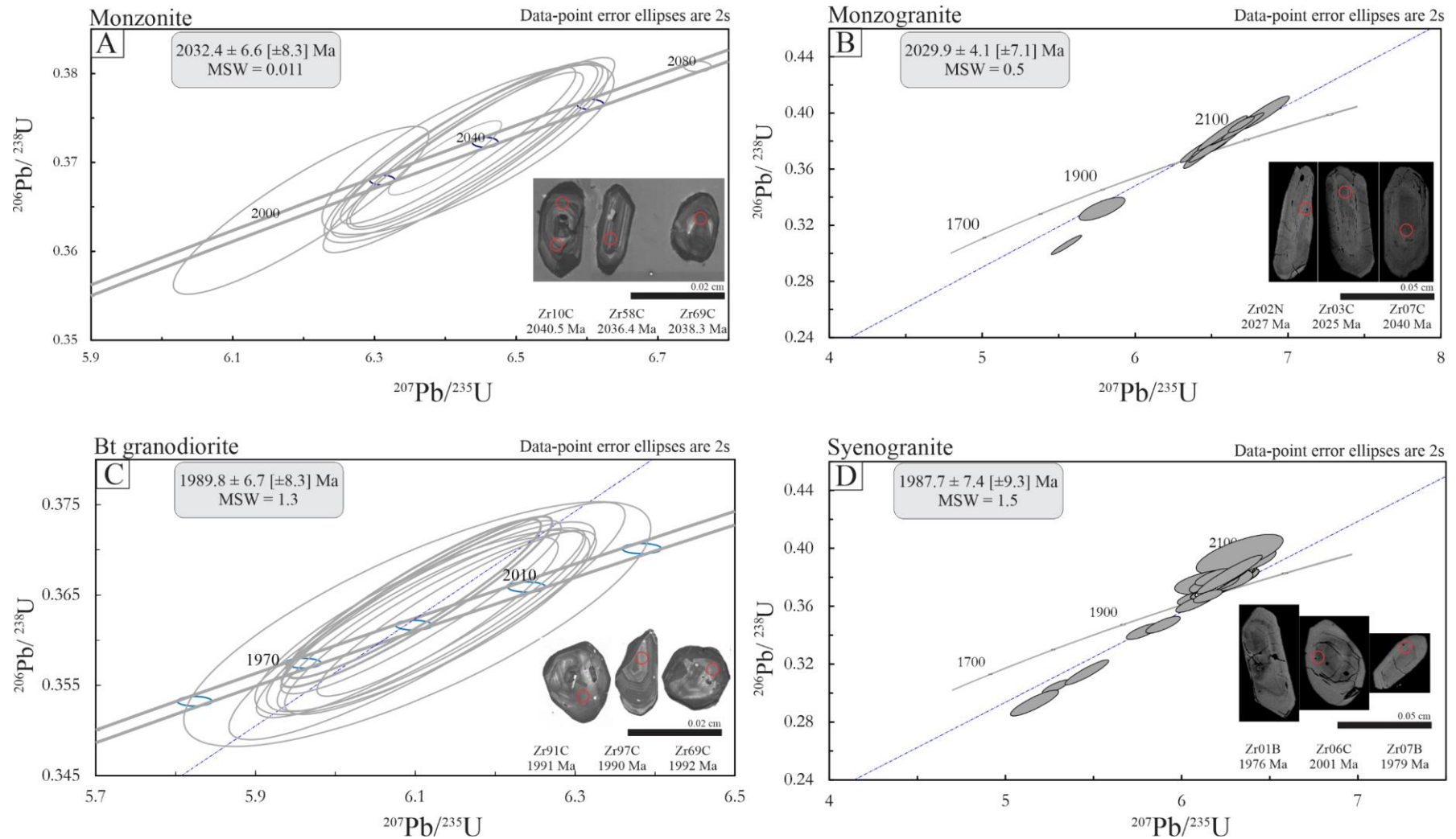


Fig. 9. –Pb Concordia diagrams, with backscattered electron (BSE) images of zircon crystals inside, from Novo Mundo upper intercept granites ages. The red circles shown in the images represent the U–Pb analysis sites.

Table 1. Sm-Nd isotopic data of the Novo Mundo granites.

Samples	Petrology	Age (Ma)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$ $\pm 2\text{SE}$	$\text{Nd}_{(0)}$	$\epsilon\text{Nd}_{(t)}$	T_{DM} (Ga)
F06-10	Monzonite	2032	4.453	21.881	0.123001	0.51155+03	21.22	-2.00	2.52
DNFD-01A	Monzonite	2032	9.672	46.753	0.1251 3834	0.511791+03	-16.52	1.81	2.14
LNM 51	Monzogranite	2029	5.745	31.523	0.110241744	0.511492 +03	-22.36	0.22	2.27
LNM 25	Monzogranite	2029	6.059	33.427	0.109644572	0.511481+03	-22.58	0.16	2.28
DNFD-01C	Granodiorite	1989	8.824	58.073	0.091912551	0.511285+03	-26.39	0.44	2.19
F64-A10	Granodiorite	1989	7.977	36.443	0.132302	0.511724+03	-17.83	-1.39	2.48
LNM 05	Syenogranite	1987	0.898	6.212	0.0874	0.511152+06	-28.99	-0.36	2.27
LNM 46	Syenogranite	1987	5.892	36.155	0.08744369	0.511341+04	-25.30	-1.12	2.24
DNFD-03	Syenogranite	1987	5.643	34.959	0.097641	0.511328+04	-25.55	-0.03	2.24

The monzonite (2.032 Ga) is primarily alkali–calcic, metaluminous, dominant, to peraluminous, and has petrographic and geochemical characteristics of volcanic arc granites. Its positive $\epsilon\text{Nd}_{(t)}$ value (+1.81) indicates mantle contribution in its formation or a short time of crustal residence for the protolith. Its T_{DM} value (2.16 Ga) and the geochemical data (Y, Ba, Sr) indicate possible mixing of mantellic and crustal sources.

The monzogranite (2.029 Ga) is meta– to peraluminous also with characteristics of volcanic arc granite. It contains primary muscovite and in [Abdel-Rahman \(1994\)](#) diagram (Fig. 5B), the rock has biotite composition of peraluminous granites. The $\epsilon\text{Nd}_{(t)}$ data (0.16–0.22) is coherent with mixing of crustal and mantle contribution.

The youngest granites (biotite granodiorite – 1.989 Ga; and syenogranite – 1.987 Ga) are meta– to peraluminous, with calc–alkaline compositions, also with petrographic and geochemical characteristics of volcanic arc granites. The negative $\epsilon\text{Nd}_{(t)}$ values for the syenogranite (-1.12 to -0.03) indicate predominance of crustal contribution to its formation.

The granitic magma origin in magmatic arc settings results from the interaction of crustal and mantle components in subduction zones ([Castro 2014](#)). Positive Rb and Th slight enrichment in syenogranite and monzogranite can suggest a characteristic pattern of magmas of a calc–alkaline series-type (Fig. 6I, G).

The Zr values (Fig. 6E) return scattered or display more complex linear trends, indicating fractional crystallization, where the composition evolves during fractionation. The negative anomaly of Nb and Ti, shown in multielement plot, can indicate magma

production by melting of subduction-related lithospheric mantle (Gavryushkina *et al.* 2018).

The REE spectrum (Fig. 7) shows enrichment in LREEs relative to HREEs and slightly negative Eu anomalies, similar to that of high-fractionated high-K calc–alkaline magmas (Rollinson & Pease 2021). The negative Eu anomaly indicates that feldspar underwent fractionation during the evolution of granitic magmatism; moreover, it signals that the fluid phase would have fractionated this mineral (Fig. 7). For rocks whose SiO₂ content is above 65 wt.%, the negative anomalies for Ce and Sm can be used as a discriminating index between different granite types, indicating possible fractional crystallization and potential crustal contamination through magma (Pearce & Peate 1995).

The REE behavior, whose increase with the increasing of SiO₂ is prominent, and the alignment of the trace element contents observed in the Harker diagram, suggest that the magma that generated the rocks of the Novo Mundo complex was formed from sources related to melting or fractional crystallization under simultaneous high pressures and degrees of magmatic differentiation and/or source enrichment in residual garnet (Laurent *et al.* 2014, Loucks 2014).

Following the definition by Pearce *et al.* (2014), Fig. 8B displays the ratio of the incompatible elements Nb/Yb vs. Th/Y for the diabase dike. The Pearce (2008) diagram is a proxy for classifying basalts generated in subduction environments and those related to subduction environments (mid-ocean ridge basalt–ocean island basalt array). Within this model, the Novo Mundo mafic dikes are positioned in the continental arc and ancient volcanic rocks enriched in LREEs and LILEs and depleted in Nb–Ta–Ti usually have a calc–alkaline affinity and are commonly interpreted as arc-derived, a behavior analogous to that of the Novo Mundo andesite dikes (Fig. 8D) (Ross & Bédard 2009).

5.2. Geotectonic evolution and regional implications

Several studies involving U–Pb zircon ages (LA–ICP–MS, TIMS, SHRIMP) and Sm–Nd T_{DM} model ages have been carried out in Juruena Province (Paes de Barros 2007, Miguel jr. 2011, Dezula *et al.* 2018, Alves *et al.* 2020, Tarello Neto & Moura 2021).

Rocks from the Novo Mundo were correlated to the Cuiú–Cuiú complex (U–Pb from 2.03 ± 7 to 1.95 ± 7 Ga) of the Tapajós Province by Testa (2018), based on their petrological similarities and geochronological data, which is relatively similar to the data obtained by Paes de Barros (2007) (1970 ± 3 Ma and 1964 ± 1 Ma, for syenogranite and

monzogranite, respectively). The obtained data indicates that the magmatic event that gave origin to the Novo Mundo granites occurred between 2032.4 ± 6.6 and 1964 ± 1 Ma, initially with monzonite crystallization.

The T_{DM} ages of the analyzed granites are close, indicating a common reservoir. However, the monzonite and the biotite granodiorite have more prominent positive ϵ_{Nd} values ($1.81 - 0.40$) than monzogranite (slightly positive ϵ_{Nd} to slightly negative: -0.20 to 0.16). Such data indicate different degrees of crustal contamination in the initial stage of the evolution of the Cuiú–Cuiú arc in Juruena Mineral Province. These crystallization ages correlate with I-type calc-alkaline granites of Nhandu intrusive suite (Silva & Abraham 2008), Aragão granites (Dezula *et al.* 2018), and the Pé Quente intrusive suite (Assis 2015) in Juruena Mineral Province.

Moreira (2019) obtained U–Pb LA–ICP–MS data for the Paraíba tonalite rocks, in Peixoto de Azevedo Domain, of 2.03 ± 26 Ga, with inherited zircons with ages between 2.7 and 3 Ga, and $\epsilon_{Nd(t)}$ data of -1.38 . Near that area, Moura and Botelho (2002) obtained U–Pb ages between 1872 ± 12 and 1869 ± 10 Ma, Sm–Nd (T_{DM}) model ages from 2.15 to 2.47 Ga and $\epsilon_{Nd(t)}$ values between -0.98 and -4.3 for the The Matupá intrusive Suite, interpreted as arc-related I-type calc-alkaline granites.

Rizzotto *et al.* (2019) proposed that the Peixoto de Azevedo Domain is an extension of the southern portion of the Tapajós Mineral Province, such as a cratonic block, based on age, rock, and magmatism types. According to Tarelow Neto and Moura (2022), the Matupá Intrusive Suite may be linked to the Cuiú–Cuiú Magmatic Arc, and the Teles Pires Intrusive Suite, to the Juruena Magmatic Arc. The authors also suggest an alternative model, in which both suites originated and evolved from a single magmatic arc.

In the Tapajós domain, Santos *et al.* (2004) described the Mundurucus Orogeny ($2.04 - 1.95$ Ga) as three successive magmatic arcs: Cuiú–Cuiú island arc (oceanic arc; $2040 - 2000$ Ma), Cumaru ($2000 - 1980$ Ma), and Creporizão continental arcs ($1980 - 1957$ Ma). According to those authors, late granites (Creporizão suite) and correlated volcanic rocks (Vila Riozinho formation), with ages between 1997 and 1957 Ma, mark the post-collision phase. New data from Guimarães and Klein (2020) show U–Pb between 2033 and 1996 Ma and $\epsilon_{Nd(t)}$ between -2.1 and -3.2 for the Vila Riozinho rocks.

Alves *et al.* (2020) reviewed the main rocks, U–Pb ages, and Sm–Nd isotopic data of the Cuiú–Cuiú Magmatic Arc region. They obtained U–Pb LA–ICP–MS ages ranging

from 2.05 – 1.99 Ga and $\epsilon\text{Nd}_{(t)}$ -3.24. According to [Silva \(2017\)](#), the granodiorites studied in the Cuiú–Cuiú Complex have U–Pb age 2.03 ± 17 Ga, with $\epsilon\text{Nd}_{(t)}$ between -1.9 and -2.42. The Serra Granite (2020–1968 Ma), in Tapajós Province, is alkali–calcic and ferroan, with moderate to strongly negative $\epsilon\text{Nd}_{(t)}$ values between -0.8 and +0.1, and T_{DM} model ages of 2.27–3.15 Ga. These rocks have been interpreted as a differentiated variety of the Creporizão Suite (2002–1956 Ma) ([Guimarães and Klein 2020](#)).

The $\epsilon\text{Nd}_{(t)}$ values from the Novo Mundo granites are slightly negative to slightly positive (+1.81 to -2.0). Such isotopic signature is likely due to the mixing process between melts from the asthenospheric (highly positive ϵNd_t) and lithospheric mantles (moderately negative ϵNd_t) resulting in a mixed material ([Valério *et al.* 2021](#)).

The juvenile melt compatible with Nd– T_{DM} ages of the Novo Mundo granites is possibly represented by 2.52–2.15 Ga from the Cuiú–Cuiú Complex. The interaction of mantle and crustal sources in this accretionary environment generated hybrid magmas and subsequent magmatic differentiation processes supported by rocks with ϵNd values ranging from slightly negative to positive ([Duarte *et al.* 2019](#)).

In a volcanic arc environment, it is common to find igneous rocks that derive from different sources or are the result of mixing or assimilation processes. Magma mixing may be connected to this plutonic rock variation observed in Novo Mundo. Felsic magmas derived from the partial melting of various crustal rocks might have physical and chemical properties in common. This process has been suggested to play a crucial role in granite petrogenesis from subduction zones to collisional orogens ([Gao *et al.* 2016](#)).

The proposed evolution of the Cuiú–Cuiú magmatic arc and its impact on the Novo Mundo region are shown in Fig. 10. During stage 1 (2.03–1.99 Ga), the subduction process produced mafic magmas, which rise through the lithosphere and stall in the subcontinental–lithospheric mantle (SCLM) or mafic lower crust near the Moho. These magmas accumulate in the lower-crustal melting–assimilation–storage–homogenization (MASH) zones, where produce felsic magmas. Then migrate to higher crustal levels and might release magmatic fluid and cause ore formation ([Zhou and Li 2000](#)).

This evolutionary petrogenetic model could explain the I-type monzonite–tonalite–monzogranite generation from the Cuiú–Cuiú magmatic arc, which resulted from sequential magmatic pulses that produced rocks predominantly of felsic composition at several points along the eastern portion of the Juruena Mineral Province.

In stage 2, the transition zones separating mafic from felsic magmas contain hybrid magmas, presenting a strong mixing character (Vigneresse 2007), forming granodiorite rocks (1.98 – 1.96 Ga), forming granodiorite rocks (1.98–1.96 Ga). The syenogranite emplacement marked the final stage of the formation process of the Cuiú–Cuiú magmatic arc, justifying its more crustal influence.

The obtained data are coherent to Juliani *et al.* (2021) model, according to which, in the Tapajós and Juruena mineral provinces, a long-lived (almost 220 M.yr) ocean-continent orogeny event produced ca 2.0 to 1.77 Ga continental magmatic arcs, with an approximately E–W-oriented subduction, dipping to north.

6. Conclusions

The Novo Mundo complex (2.032–1.987 Ga) is composed of a series of granitic rocks represented by monzonite, monzogranite, biotite granodiorite and syenogranite, that are interpreted as representing the orogenic phase of the Cuiú–Cuiú Magmatic Arc (2050–1950 Ma).

The monzonite is an alkali–calcic peraluminous granite with hornblende and biotite as the primary mafic minerals. We can interpret the isotopic data (T_{DM} age of 2.14 – 2.52 Ga and $\epsilon Nd(t) = -2.0$ to $+1.81$) as resulting from an initial stage plutonic event connected to the Cuiú–Cuiú magmatic arc.

The monzogranite has primary muscovite, which indicates a peraluminous character, and it is classified as generated in a volcanic arc with crustal contribution.

The biotite granodiorite has a magnesian tendency and a meta–peraluminous affinity, typical of I-type granitic bodies, with calc–alkaline to alkali–calcic compositions from volcanic arc. $T_{DM} = 2.20 - 2.48$ Ga and $\epsilon Nd = 0.44$ to -1.39 , indicating the participation of a juvenile source or mixing of felsic magmas of different compositions or linked to the entry of mafic magma into the system.

The syenogranite has a calc–alkaline signature and volcanic arc environment. The T_{DM} signature of 2.29–2.23 Ga and $\epsilon Nd = -0.03$ to -1.12 indicate a crustal contribution. Hybrid magmas could be products of the interaction of magmas generated from mantle and crustal sources in accretionary environment, supported by rocks with ϵNd values from slightly negative to positive.

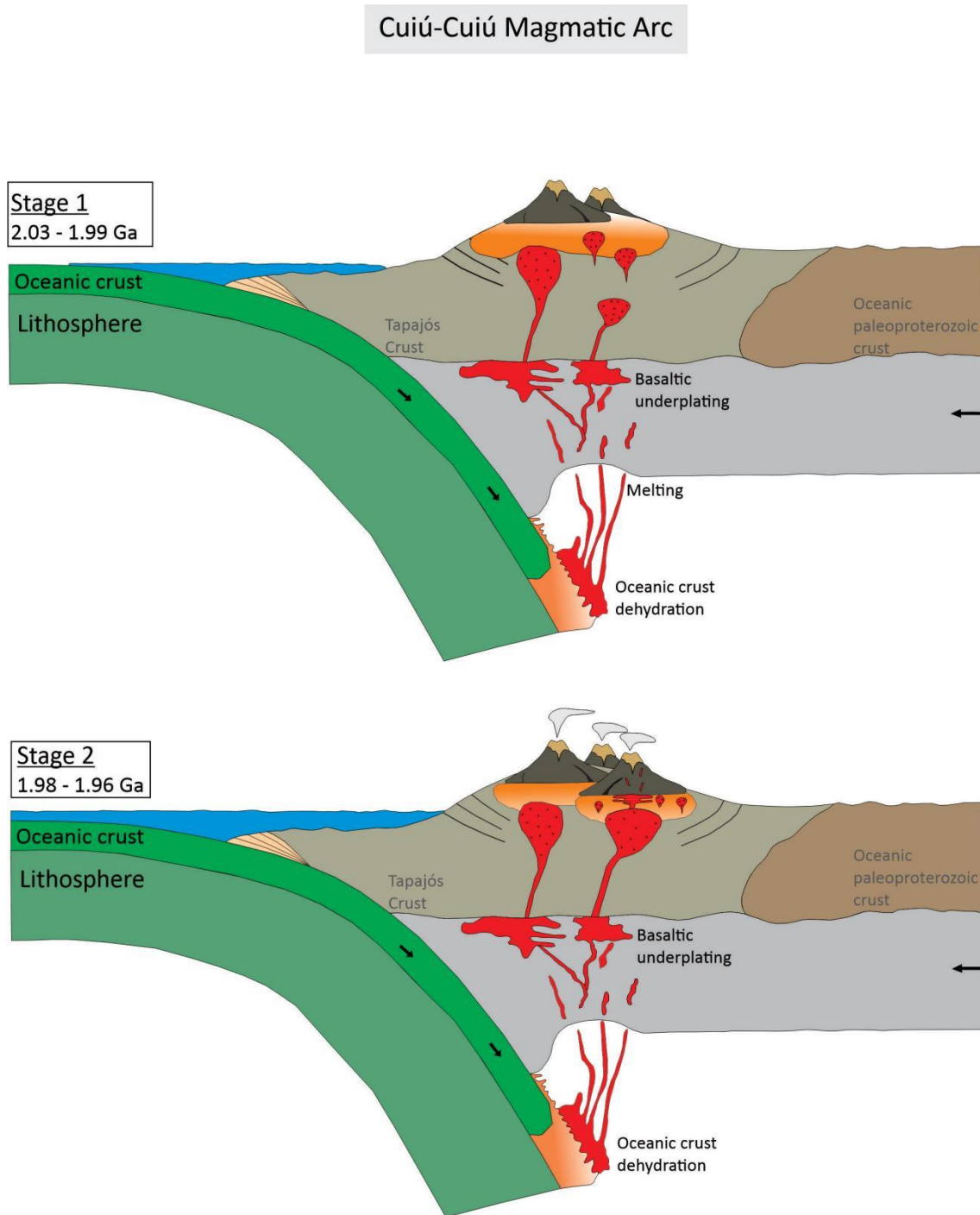


Fig. 10. A schematic representation of granite generation in the Novo Mundo region during the evolution of the Cuiú-Cuiú magmatic arc (modified from [Tarelou Neto and Moura 2022](#)).

In a collisional tectonic setting, the generating source would be a lithospheric mantle from the fractional crystallization under different pressure conditions (thick continental crust vs. thinner crust) and the magma mixing between magmatic chambers in a subduction environment, with crustal contribution of varied proportions during the Cuiú-

Cuiú magmatic arc emplacement period.

The obtained data, together with recent studies, allow suggest the existence of a continuous magmatic arc in the Paleoproterozoic at the south part of the Amazonian craton, which was responsible for the formation of the Tapajós and Juruena mineral provinces main granitic rocks.

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CAPÍTULO 5
THE PALEOPROTEROZOIC RAIMUNDA PORPHYRY-TYPE
GOLD DEPOSIT, JURUENA MINERAL PROVINCE,
AMAZONIAN CRATON (BRAZIL): CONSTRAINTS BASED
ON PETROLOGICAL, FLUID INCLUSION AND STABLE
ISOTOPE DATA

[ARTIGO 2 – PUBLICADO NO PERIÓDICO MINERALS.](#)

Article

The Paleoproterozoic Raimunda Porphyry-Type Gold Deposit, Juruena Mineral Province, Amazonian Craton (Brazil): Constraints Based on Petrological, Fluid Inclusion and Stable Isotope Data

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Abstract: The Juruena Mineral Province is an emerging world-class mineral province in the southern Amazonian Craton, due to numerous Au-Cu and base metal deposits, as the Raimunda deposit, related to Novo Mundo 2.03–1.98 Ga I-type calc-alkaline granites. Its hydrothermal alteration zones comprise Na-metasomatism, microclinization, propylitic and sericitic alteration, silicification, sulfide stage and late carbonate alteration. The disseminated mineralization, associated with the sulfide stage, the main mineralization stage, is represented by gold inclusions and fracture-filling grains in pyrite, chalcopyrite and Cu-Bi sulfides. Chlorite geothermometer and fluid inclusion data indicate temperature conditions of 325–380°C for the mineralization. The coexistence of high-temperature aqueous and aqueous-carbonic fluid inclusions, based on microthermometric study of fluid inclusions, reveals a mixing of medium-saline hot fluids with cooler and low-saline fluid. The $\delta^{18}\text{O}_{\text{fluid}}$ (3.11 – 7.86‰) and $\delta^{34}\text{S}_{\text{py}}$ data (–1.4 – 0.1‰) are coherent to magmatic origin for the mineralizing fluid. Gold was initially transported as chlorine complexes in a hot, high-salinity, acidic and oxidized fluid from the magma chamber, and later as H_2S complexes. The chemical-physical instability during fluid ascent is interpreted as triggering factors for ore precipitation. The results offer valuable insights into the genesis of porphyry–Au deposits and their implications for prospecting in the Amazonian Craton.

Keywords: Paleoproterozoic; Amazonian Craton; fluid inclusions; stable isotope; porphyry–Au deposit.

1. Introduction

The Amazonian Craton (Brazil) has undergone a series of calc-alkaline granitogenic events over its geological history, some of which led to gold and other metals mineralization [1]. The Tapajós and Juruena Mineral provinces are the major metallogenic provinces in Brazil, located in the southern Amazonian Craton. The Juruena Province, especially, has been an important gold producer in the country. Its main mineralization is predominantly spatially and temporally related to Paleoproterozoic calc-alkaline granites formed in volcanic-arc settings [2–5].

In the Juruena Mineral Province, mainly in the surroundings of the cities Nova Santa Helena, Peixoto de Azevedo, Matupá, Garantã do Norte and Novo Mundo, primary gold

mineralization associated mainly with granite systems and, subordinately, with volcanic sequences occur either disseminated or hosted into veins and stockwork-veins [5,6].

The emplacement of these bodies into continental magmatic arcs has been essential for forming disseminated, vein-type, and $\text{Au} \pm \text{Cu} \pm$ base metals stockwork-type magmatic-hydrothermal systems. The systematic and integrative study of these orebodies points out to a magmatic-hydrothermal model comparable to that of porphyry gold deposits [7], low-, intermediate-, and high-sulfidation epithermal systems [8], and structurally controlled [9] placed at variable crustal levels.

The Novo Mundo granite complex, which hosts the Raimunda deposit, the main object of this study, is in the homonymous municipality in the mid-northern region of the state of Mato Grosso, east of the Juruena Mineral Province. This granite body hosts several mineral deposits, including the Raimunda deposit, located in its northwestern region.

The percolation of hydrothermal fluids promotes anomalous and systematic mineralogical and geochemical changes in the host rocks, the extent and magnitude of which depends on its alteration state, porosity, permeability, and degree of chemical reactivity of the host rocks. The fluid-rock interaction generates alteration footprints that may assist in determining the exploratory guides in mature terrains [10].

Despite recent advances in knowledge about Proterozoic mineralizing events in the Amazonian Craton, prospecting complex Paleoproterozoic gold-copper deposits in Juruena Mineral province is a challenge, especially due to few studies that have investigated the origin of mineralizing fluids using stable isotopic data to constrain the genesis of mineral deposits in the southern region of the Amazonian Craton.

This paper integrates new petrological (i.e., drill core samples, whole-rock assay results), ore mineral chemistry (e.g., chlorite, muscovite, sulfides and gold), fluid inclusion and stable isotope data (O and S) from a preserved Paleoproterozoic deposit, located in the southern Amazonian Craton, and proposes a metallogenetic model for the Raimunda deposit and other analogous deposits in Brazil and worldwide.

2. Geological Settings

The Amazonian Craton has been a tectonically stable area since the Grenvillian Orogeny (1.2 Ga). It has been understood for decades as the result of successive amalgamation of accretionary arcs from the Archean to the Paleo- and Mesoproterozoic [11,12]. However, newer alternative concepts and models have been proposed based on new isotopic and geophysical data (Figure 1).

Based on the progressively younger igneous bodies striking NE, at 1.88 Ga, the central region of the Amazonian Craton underwent a period of flat-slab subduction, which caused the migration of arc-related units towards the Central Amazonian block [13]. The sequence related to the Paleoproterozoic Tapajós Province arc was developed on an Archean basement and the well-defined E-W-striking lineament, which extends from the Tapajós Province to the Carajás Province, represents the ancient cratonic basement [14]. The lack of fold belts and/or metamorphic terranes, as well as the orogenic and post-orogenic tectonics, led Juliani et al. [15] to suggest weakly extensional to slightly compressive settings for the region.

Magnetotelluric profiles carried out in the eastern region of the province and found a high-resistivity anomaly approximately 20 km deep, which is ascribed to the Peixoto de Azevedo Domain, surrounded by conductive anomalies related to the Statherian magmatism and mafic enclaves [16]. Furthermore, this domain presents a controversial tectonic evolution linked to long-lived shear deformation comparable to that of the Tapajós Province [17].

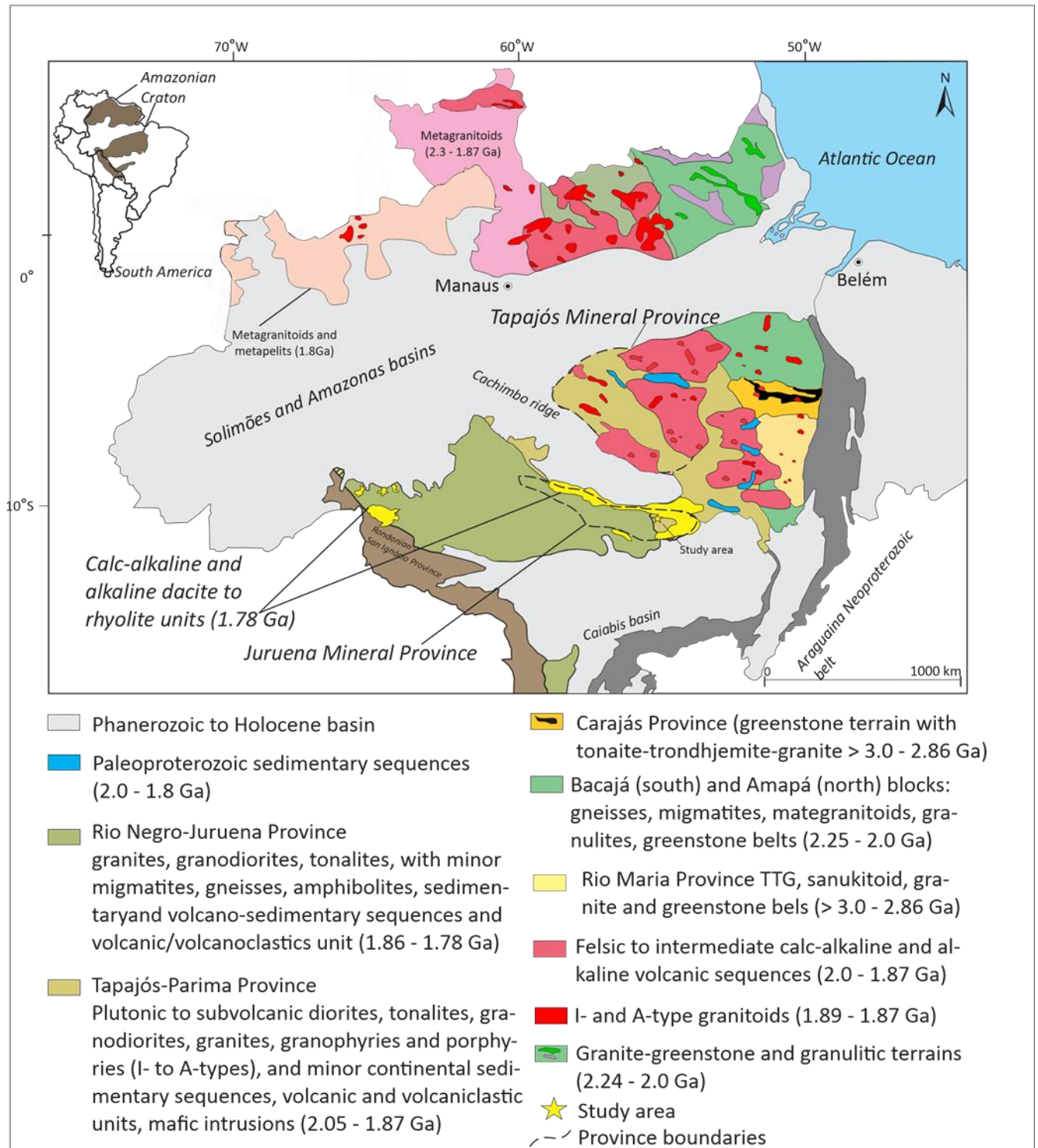


Figure 1. Geological and tectonic environment of the Juruena Mineral Province in the Amazonian Craton, based on [5,21]. Geochronological provinces based on [11].

In the southern segment of the Amazonian Craton, there are two extensive Paleoproterozoic domains: an older one, named Peixoto de Azevedo (2.05 to 1.98 Ga), and a younger one, named Juruena (1.85 – 1.75 Ga). The Peixoto de Azevedo Domain comprises the Cuiú-Cuiú Magmatic Arc and post-magmatic arc rocks, mainly composed of Paleo- to Mesoproterozoic volcano-plutonic sequences formed in arc settings [6,18,19].

The eastern segment of the Juruena Province is mainly composed of oxidized, medium- to high-K calc-alkaline, volcanic-plutonic, and meta- to peraluminous rocks.

These rocks mostly belong to the magnetite series (I-type granites), and are volcanic, subvolcanic, and alkaline granites [20].

2.1. Geology of the Deposit

The exploration area of Novo Mundo Paleoproterozoic Complex exhibits surface coverage that contains dozens of mineral deposits which host Paleoproterozoic Au mineralization [22], Au-Cu-Mo [9] and Cu [23] along a NW-SE-striking lineament.

The Novo Mundo complex intrudes Cuiú-Cuiú Archean rocks (granitic- to tonalitic gneisses and migmatites) and shows U-Pb data between 2.03 and 1.98 Ga [24]. It has approximately 60-kilometre-diameter and is composed mainly by granites, along W-NW, which is compatible with the preferred orientation of the major regional lineaments mapped from the oldest terranes in that region [3], although there is no evidence of structural controls [19].

The Raimunda deposit is associated to monzonite, biotite granodiorite, monzogranite and syenogranite facies. The last two are the main host rocks of the deposit. The granites are cut by dykes of basaltic-andesite to andesite compositions.

The overview of the deposit is shown in figures 2A, and 2B, with a detailed view of the open-pit constructed by artisanal miners in the late 1990s and early 2000s for ore extraction. Figures 2C–E show rare blocks of outcrops classified as monzogranite, syenogranite and mafic bodies, respectively, all found around the deposit. The contact between the granitic facies is rarely observed. In drill holes, subangular to rounded microgranular enclaves are locally registered, especially monzonite enclaves into syenogranite.



Figure 2. Raimunda deposit outcrops, Novo Mundo, Juruena Mineral Province, Brazil. (A,B) Photograph of old ore extraction pit area, highlighting the location of some of the drill holes; (C) Syenogranite block showing potassic feldspar phenocrysts; (D) Monzogranite block found around the area; (E) Mafic rocks blocks.

Although the study area is in a region with local NW-SE-striking ductile-brittle shear zones (Novo Mundo Shear Zone – NMSZ) [25], which host several gold-bearing mineral deposits in the eastern region of the Juruena Mineral Province [6] and display a high density of microfractures, the rocks of the Raimunda deposit have not undergone significant deformation. The strongly fractured and crushed appearance of the minerals results from the action of the hydrothermal system over those rocks. Centimeter-to-millimeter-thick veins and veinlets are mostly non-oriented and contain alteration minerals (e.g., quartz, chlorite, muscovite, epidote, calcite, pyrite and rutile).

2.1.1. Monzonite Facies

The monzonite is dark gray, medium-grained, inequigranular, and exhibits a phaneritic texture (Figure 3A). It contains andesine crystals (An_{40-30} tending to An_{25}) (20–25 %), orthoclase (18–20 %), quartz (18–20 %), biotite (3–5 %), and hornblende (2–5 %). Accessory minerals are magnetite (2 %), titanite (1%), apatite (<0.1 %), and zircon (<0.1%), as well as hydrothermal minerals, such as epidote (1–2 %), chlorite, muscovite, and carbonate (~5 %) (Figure 4A).

2.1.2. Monzogranite Facies

The monzogranite is the more representative facies of the deposit. It has greenish-gray color and is composed of quartz (25–30 %), albite (15–20 %), orthoclase (20–25 %) and biotite (3–5 %); accessory minerals (magnetite <1%), and hydrothermal minerals (muscovite, chlorite, carbonate, epidote and rutile, 12–15 %) (Figure 3B). Microtextures show widespread veins and veinlets through the rock in association with potassic alteration (Figure 4B–D). Locally, open-space filling textures (“comb texture”) are also observed, especially proximal to or in sulfide-bearing strongly hydrothermalized zones. Quartz megacrysts (0.5–3 mm) are restricted to strongly silica-altered zones (Figure 4B).

2.1.3. Syenogranite Facies

The syenogranite is reddish to pinkish and, in places, displays porphyritic texture with quartz megacrystals/phenocrysts (Figure 3C). It is composed of quartz (30–35 %), orthoclase (20–25 %), albite (15–20 %), and biotite (1–2 %). Accessory magnetite (<1 %), apatite (<0.1 %) and zircon (<0.1 %) feature as inclusions in quartz, plagioclase, and potassium feldspar. Muscovite, carbonate and epidote commonly replace feldspar and plagioclase (12–15 %) (Figure 4F, G). These samples also contain quartz megacrystals (0.5–3.5 mm) in strongly silicified zones, which are similar to those of monzogranites.

2.1.4. Granodiorite Facies

The granodiorite is greyish, medium-grained, composed of quartz (25–30 %), andesine (An_{40-30} tending to An_{25}) (25–30 %), orthoclase (15–20 %), biotite (2–5 %), and hornblende (1–3 %) (Figure 3D). The accessory minerals are magnetite (1–3 %), titanite (1 %), apatite (<0.1 %) and zircon (<0.1%); and the secondary minerals are muscovite, carbonate, epidote, and rutile (10–12 %) (Figure 4E).

3. Materials and Methods

3.1. Fieldwork, Sampling, and Petrography

Three fieldwork campaigns were carried out for mapping and sampling. 80 samples were collected for petrographic analysis for identifying the main gangue and ore mineral associations of hydrothermal alteration and ore-bearing zones, as well as their patterns

and styles [26,27]. This enables a better understanding of their evolution throughout time in the system.

The polished thin sections and polished sections were made at the thin-section laboratory of the Geosciences Institute of the University of Brasília (LABLAM-UnB). The doubly polished thin sections were made at the thin-section laboratory of the Petrografia-BR company (Contagem-MG). The petrographic descriptions were carried out at the Microscopy Laboratory of the UnB Geosciences Institute using an Olympus BX-50 petrographic microscope (Olympus Coporation, Tokyo, Japan), as well as at the UnB Fluid Inclusions Laboratory (LIF-UnB) using an Olympus BX-50 petrographic microscope.



Figure 3. Host rocks from the Raimunda deposit. (A) Monzonite, showing microclinization and veinlets composed of Cal, Chl, Qz and Ser with no preferred direction pattern; (B) Weakly microclinized monzogranite, cut by veinlets without preferential orientation of Qz + Chl + Ms + opaques; (C) Microclinized syenogranite, cut by venules of Cb, Qz and Ms; (D) Microclinized granodiorite, cut by venules of Qz + Chl + Ms. Abbreviations: Cal: calcite; Chl: chlorite; Ms: muscovite; Qz: quartz.

3.2. Electron Microprobe

The mineral chemistry analyses were determined using a JEOL® JXA-8230 Electron Probe Micro-analyzer (EPMA) (JOEL Ltda., Tokyo, Japan), at the Institute of Geosciences, University of Brasília. For silicates, the chemical analyses were obtained using an

acceleration voltage of 15 kV, beam current of 10 nA and a nominal beam diameter of 1 μm in the spot mode. The counting time for each element during the analysis was 10 s on the peak and 5 s on the background. Interference correction was implemented for Ti and V. Matrix effects were corrected using an in-house ZAF program. The natural and synthetic calibration standards used were albite (Na), forsterite (Mg), orthoclase (Al, Si, K), topaz (F), vanadinite (Cl, V), apatite (P), andradite (Ca, Fe), pyrophanite (Ti, Mn), barite (Ba), Ni oxide (Ni), and vanadinite (V).

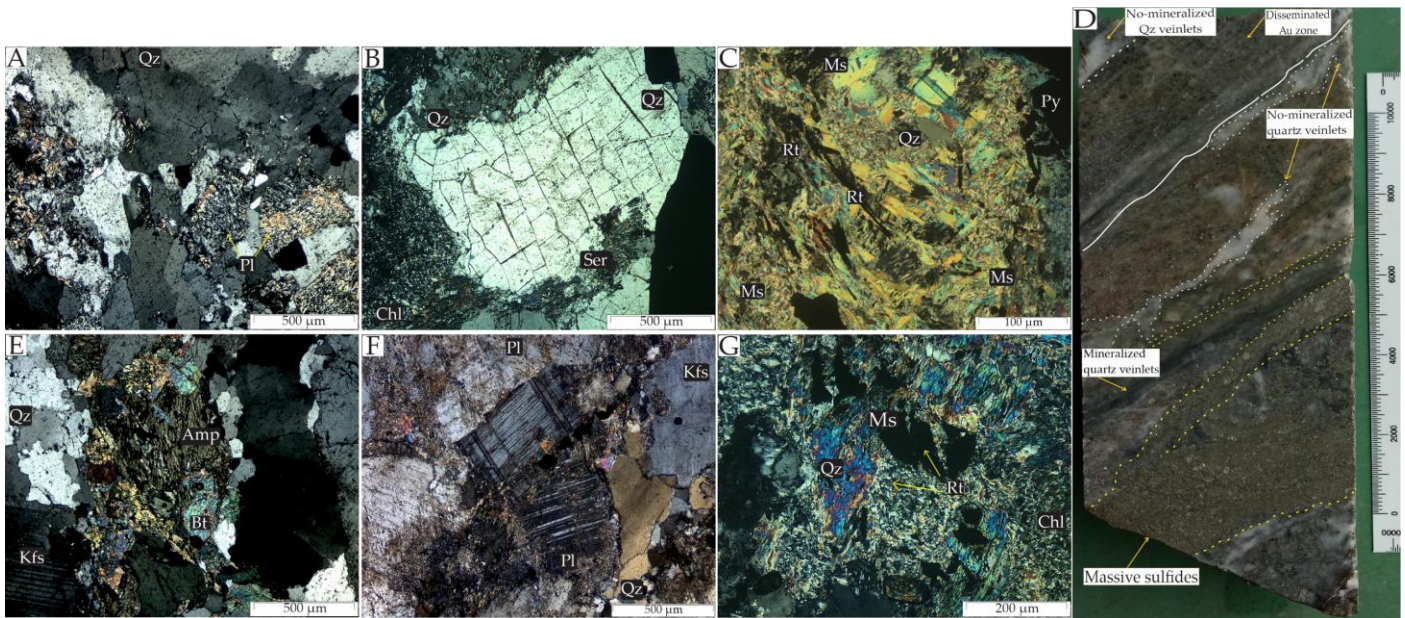


Figure 4. Hydrothermal alteration features in drill core samples from Raimunda gold deposit. (A) Monzonite micrographic texture, with portions of agglomerated and recrystallized quartz, along with Pl and Kfs, in addition to Ms, Chl and opaque minerals; (B) Megacrystal of Qz strongly fractured in monzogranite rock, occurring in clusters of hydrothermal minerals containing Chl, Ser and opaque minerals in zones of strong silicification; (C) Monzogranite alteration cluster containing Ms, Chl, Rt and mafic minerals; (D) Monzogranite showing pervasive silicification, along with either monomineralic with only quartz or containing other phases; massive sulfide levels are observed by cutting the sample; (E) Granodiorite showing recrystallized Qz, Pl and Kfs together with mafic minerals such as Amp, Bt, Chl, Ep and mass of Ms; (F) Cluster of recrystallized minerals in alteration mass containing Pl, Qz, Kfs, in addition to Ms, Chl in syenogranite sample; (G) Mass of hydrothermal minerals in syenogranite containing Ms, Qz, Rt lamellae and opaque minerals. Abbreviations: Amp: amphibole; Bt: biotite; Chl: chlorite; Kfs: potassium feldspar; Ms: muscovite; Pl: plagioclase; Py: pyrite; Qz: quartz; Rt: rutile; Ser: sericite/fine muscovite.

For the sulfides and gold were used an accelerating voltage of 20 kV, a beam current of 20 nA and a counting time of 10 s, except for selenium (25 s); the counting time for backgrounds was always half of the used-on peaks values. The X-ray $K\alpha$ -lines were used for sulfur, selenium, copper, iron, nickel and cobalt; $L\alpha$ for arsenic, silver, antimony, palladium, cadmium, tellurium and molybdenum; and $M\alpha$ for lead, mercury, gold and platinum; $M\alpha$ and $M\beta$ were used for bismuth. The natural minerals applied as standards in both electron probes were pyrite (Fe, S), chalcopyrite (Cu), galena (Pb), and stibnite (Sb), and the native elements gold, silver, platinum, palladium, bismuth, copper and antimony. The synthetic standards were gallium arsenide (As), zinc selenide (Zn, Se) and cinnabar (Hg). The following detection limits were obtained (in wt.%): 0.018 for gold, 0.012 for silver, 0.008 for palladium, 0.016 for platinum, 0.032 for bismuth, 0.025 for lead, 0.016 for copper, 0.013 for iron, 0.017 for mercury, 0.025 for arsenic, 0.009 for antimony, 0.017

for selenium, 0.005 for sulfur, 0.012 for cadmium, 0.013 for nickel and cobalt, 0.017 for tellurium and 0.030 for molybdenum.

3.3. Chlorite Geothermometry

The studies of formation temperature and pressure conditions at the Raimunda deposit were complemented using a chlorite-independent geothermometer. The structural formulas of chlorite and muscovite were calculated based on 28 and 11 oxygen atoms, respectively. The temperatures are based on the cationic proportion of tetrahedral aluminum (Al^{IV}) in chlorites, which allows the crystallization temperature to be calculated using the following equation: $T (^{\circ}C) = 321.98 * (Al^{IV}/2) - 61.92$ [28,29]. 18 analyses of chlorite were carried out on carbon-coated polished thin sections at the LME-UnB using a JEOL® JXA-8230 EPMA (JOEL Ltda., Tokyo, Japan). The WDS operated under the conditions of an acceleration voltage of 15 Kv, a beam current of 10nA, and a count time of 10s.

3.4. Fluid Inclusions

The study of fluid inclusions was performed on doubly polished thin sections from six different mineralized samples. Petrography of fluid inclusions was carried out at room temperature ($\pm 25^{\circ}C$) under a Zeiss Axioplan 2 petrographic microscope (Zeiss, Oberkochen, Germany) according to the criteria of [30–32].

Microthermometric measurements were obtained with a LINKAM THMSG600 heating-freezing system mounted on an Olympus BX-50 petrographic microscope, at the UnB Fluid Inclusions Laboratory (LIF-UnB). Rock fragments were observed at variable temperatures between $-195^{\circ}C$ and $600^{\circ}C$ under 10x to 100x objective lens following the procedures of [31].

The calibration was carried out using synthetic H_2O – $NaCl$ two-phase and H_2O – CO_2 three-phase fluid inclusions at speed rates of 2 to $15^{\circ}C/min$, with an estimated accuracy of $\pm 0.5^{\circ}C$ for freezing (± 25 at $-150^{\circ}C$), and $\pm 2^{\circ}C$ for heating (above $350^{\circ}C$).

3.5. Stable Isotopes

The study of stable isotopes was performed on minerals of representative mineralized samples from the Raimunda deposit. Sample selection and sample preparation took place at the sample preparation laboratory at University of Brasília (LPA-UnB). The pure mineral concentrates were handpicked under a binocular microscope at LME-UnB.

Analyzes were performed at the Queen's Facility for Isotope Research (QFIR) laboratory at Queen's University, Ontario, Canada. Results are reported as delta (δ) values per mil (‰) relative to V-SMOW (Vienna Standard Mean Ocean Water) for oxygen, and V-CDT (Canyon Diablo Troilite) for sulfur.

Oxygen isotope data were obtained for nine samples, between 20 and 28 mg each, using the Thermo Finnigan Delta V Advantage and Finnigan Delta XP mass spectrometers (Thermo Fisher Scientific, Waltham, MA, United States). The analytical protocol applies the laser fluorination method to release oxygen from minerals through the reaction with BrF_5 by using a CO_2 laser as a heat source. It is then either converted into CO_2 through the reaction with graphite at $750^{\circ}C$, or it is collected as O_2 in a molecular sieve and analyzed as O_2 . The analytical uncertainties for $\delta^{18}O$ were lower than 0.1 and 0.16 ‰.

For sulfur, six samples of pyrite were prepared, each weighing between 80 and 95mg. For this analysis, the SO_2 produced through combustion of about 0.5 mg of sample is loaded into tin capsules reacting at $1,080^{\circ}C$. The gases were transported through current and reduced by contact with high-purity copper wires. The SO_2 was eventually measured using gas chromatography. Analyzes were performed using a Flash Elemental Analyzer (EA) coupled with a ThermoFisher MAT 253 Isotope Ratio Mass Spectrometer (IRMS) (Thermo Fisher Scientific, Waltham, MA, United States). Analytical uncertainties are estimated at 0.2 ‰.

4. Results

4.1. Hydrothermal Alteration and Mineralization

The Au ($\pm$ Cu) Raimunda deposit sits in the NW of the Novo Mundo Complex, being hosted by variably hydrothermally altered granite facies (i.e., monzonite, biotite granodiorite, monzogranite and syenogranite).

The types of hydrothermal alteration identified at the Raimunda deposit, and their respective mineral assemblages, are summarized in Table 1. The hydrothermal system and the associated mineralization are mainly observed through pervasive zones, and, subordinately, along veins and veinlets and the more affected rocks are the monzo- to syenogranite facies.

Based on overprinting relationships, mineralogy and alteration assemblages, seven alteration zones are recognized: (1) Early sodic metasomatism (albite); (2) Microclinization (orthoclase/microcline-quartz1-hematite); (3) Early or intermediate propylitic alteration (epidote-chlorite1-carbonate-rutile-titanite-pyrite1); (4) Intermediate to proximal sericitic alteration (muscovite-chlorite2-quartz2-pyrite2-gold1); (5) Silicification (quartz3); (6) Main mineralization phase: sulfide stage (gold2-pyrite3-chalcopryrite-bismuthinite and other Bi- and Cu sulfides); (7) Late-stage carbonate alteration and chloritization (carbonate-chlorite3-quartz4).

The alterations occur throughout the ore-bearing host granite at the deposit. At the field scale, and even when analyzing the different drill cores, it is not possible to macroscopically delineate hydrothermal alteration zones.

The early Na-metasomatism alteration modified the essential mineralogy of the granites. The oligoclase is completely or partially replaced by albite and the magmatic biotite is locally preserved.

Subsequently, microclinization represents another outward alteration halo surrounding the previous one, and it is commonly observed at the periphery of the alteration profile. Microclinization commonly occurs within all granite facies; although more intensely in the syenogranite, through the alteration of K-feldspar and the formation of microcline rims. The orthoclase is microclinized and the feldspar and biotite crystals are partially to completely altered to fine muscovite. Hydrothermal quartz grains (quartz1) are generally milky.

The propylitic alteration can be noticed in all granitic facies of Raimunda deposit, but is more perceptible in non-mineralized samples, including monzonite and granodiorite facies. It is pervasive, affecting especially feldspar crystals, which are partially or completely replaced by epidote-chlorite1-calcite. It is common to occur veins and/or veinlets with epidote-chlorite1-carbonate crossing the samples and rutile, titanite and fine-grained pyrite1 in agglomerates mass.

Sericitic alteration is marked by its pale green color throughout all granite facies, where they occur pervasively or selectively altered, and becomes more intense towards the monzogranite facies. The alteration mineral assemblage is mainly composed of muscovite, chlorite2 (Chl2), pyrite2 (Py2), quartz2 (Qz2) and chalcopryrite. The main feature is the texture obliterated by the replacement of feldspars by fine muscovite along with disseminated fine-grained pyrite and quartz, locally. In this case, pyrite (Py2) occurs disseminated either into quartz veins and veinlets or associated with quartz2 and muscovite. This phase was identified as the first stage of mineralization (i.e., muscovite, pyrite2, quartz2 and gold1). Euhedral to subhedral pyrite is classified as pyrite2, which is poorly fractured and displays a low degree of fracture filling, occurring disseminated in host rocks and, locally, into veins or veinlets that cut through the rock (Figure 5A). Gold (Au1) occurs either as inclusions or as fracture-fills in pyrite2.

Silicification occurs pervasively or as veins and veinlets (Figure 5C). The veins/veinlets consist of microcrystalline quartz (massive and mosaic texture – Qz3), associated with sulfides and other minerals, such as muscovite and carbonate.

Sulfide stage is both disseminated and forming massive sulfide zones that crosscut the rocks. At the deposit, that corresponds to the main mineralizing stage with a high

content of sulfides associated (i.e., pyrite³, chalcopyrite, chalcocite, bismuthinite, hodrushite, krupkaite, wittchenite) and gold². The pyrite³ occurs in massive sulfides levels, strongly fractured and with a high degree of fracture filling (Figure 4E, G). Gold (Au²) occurs as fracture-fills in pyrite³ and as seldom inclusions in quartz³.

The samples highlight pressure relief features, vuggy silica in quartz veins/ veinlets and levels of massive sulfide (Figure 5E, F). Late hydrothermal alteration is represented by carbonate alteration. It is less expressive than the last two stages and restricted to the final stages of hydrothermal alteration. In general, potassium feldspar and albite are affected by carbonate alteration. Carbonate-quartz⁴-chlorite³ veins and veinlets crosscut the rock samples (Figure 5E).

4.2. Ore and Hydrothermal Mineral Chemistry

Tables S1–S7 display EPMA analyses of pyrite, chalcopyrite, gold, copper-rich minerals, bismuth-rich minerals, chlorite and muscovite and can be found in the Supplementary Material. The selection considered the variable styles of mineralization (veins/veinlets and disseminated) for each mineral as well as their ore mineral paragenesis.

4.2.1. Sulfides and Gold

The analysis of pyrite focused on types 2 (Py2) and 3 (Py3), which are associated with the mineralization stages at the deposit (Table S1, Figure 6). Py3 (Figure 6A) contains slightly higher contents of Au (Py3: 0–0.09 %; Py2: 0.02–0.07 %), Te (Py3: 0–0.24 %; Py2: 0–0.12 %), and Mo (Py3: 0.01–0.17 %; Py2: 0–0.08 %). Furthermore, there are higher contents of Pb (Py3: 0.1–0.34 %; Py2: 0.19–0.23 %), as well as higher contents of Cu (Py3: 0–0.82 %; Py2: 0–0.04 %).

The significant amount of these elements in Py3 compared to Py2 reveals the genetic differences between those types based on chemical evidence, which had been previously corroborated by textural features. This points out variable timings and fluctuating element availability throughout ore precipitation.

The chalcopyrite occurs in fracture-fills and inclusions in Py2 and, more intensely, in Py3 (Figure 6A, C, D, F). Chemically, there are no significant differences between them, with average Au contents of 0–0.06 %, Ag contents of 0–0.16 %, Te contents of 0–0.23 %, Mo contents of 0–0.12 %, Pb contents of 0.04–0.30 % and Bi contents of 0–0.23 %.

Chalcocite (Cu₂S) occurs together with chalcopyrite and other copper sulfides, such as larosite ((Cu, Ag)₂₁(Pb, Bi)₂ S₁₃), filling fractures in pyrite³. Bismuthinite is the most abundant bismuth mineral and, likewise, occurs as fracture-fills. Other Cu- and Bi-rich sulfides identified as part of the mineral paragenesis at the deposit are hodrushite (Cu₈Bi₁₂S₂₂), krupkaite (PbCuBi₃S₆), and wittchenite (Cu₃BiS₃) (Figure 6D).

All sulfides are in equilibrium with Py3 making up the main mineralization stage at the Raimunda deposit. Gold occurs either as inclusions (Au¹) or fracture-fills (Au²) (Figure 6B – F). In both cases, gold inclusions and fracture-fills in pyrite showed no significant compositional variation between them (Table S3, Supplementary Material). Fracture-filling gold grains (Au²) display average Au values of 57.03–66.77 %, and Ag values of 28.45–28.74 %, while gold inclusions yield Au contents of 65.91–66.80 % and Ag contents of 27.90–28.70 %. Therefore, their chemistry reveals relative constant silver values and variable gold values, which are related to Cu, Te, and Mo input.

Table 1. Hydrothermal alteration stages and paragenetic evolution table of the Raimunda Au ($\pm$ Cu) deposit, Juruena Mineral Province, Brazil.

Event Alteration Stages Mineral	Pre-Mineralization		Mineralization Stages				Post-Mineralization
	Na-Metassomatism	Microclinization	Propylitic Alteration	Sericitic Alteration (Ms-Qz2-Py2-Au1)	Silicification (Qz3)	Sulfide Stage (Py3- (Au2-Ccp-Cc-Bin)	Carbonate Alteration (Cal-Chl3-Qz4)
Quartz 1		██████████	██████████				
Quartz 2				██████████			
Quartz 3					██████████		
Quartz 4							██████████
K- Feldspar		██████████					
Albite	██████████						
Magnetite		██████████	██████████	-----			
Titanite			██████████				
Rutile			██████████	██████████			
Chlorite		-----	██████████	██████████			██████████
Muscovite		-----	-----	██████████		██████████	
Calcite		-----	██████████				██████████
Epidote			██████████				
Pyrite 1		██████████					
Pyrite 2				██████████			
Pyrite 3						██████████	
Chalcopyrite				██████████		██████████	
Chalcocite						██████████	
Larosite						██████████	
Bismuthinite						██████████	
Hodrushite/krupkaite /wittchenite						██████████	
Gold				██████████		██████████	
	Traces -----	Minor -----	Abundant ██████████				
Temporal evolution	→						

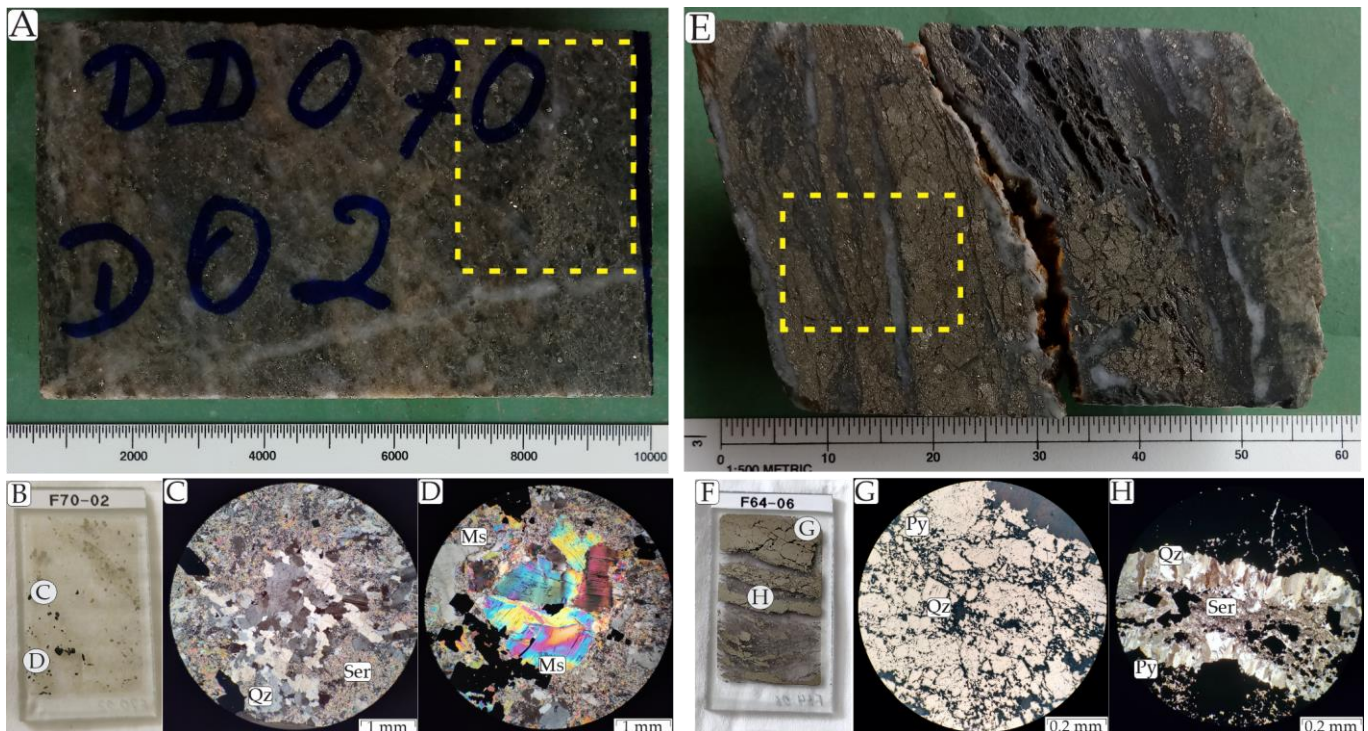


Figure 5. Hydrothermal alteration features and mineralized samples from the Raimunda Au ($\pm$ Cu) deposit. (A) Drill hole sample with pervasive silicification, generalized sulfidation, chloritization and sericitic alteration; (B) Polished thin of the yellow delimited field in figure A; (C) photomicrograph showing strong silicification with recrystallized Qz surrounded by intense sericitic alteration; (D) Ms lamellae with recrystallized Qz, opaques and sericitic alteration; (E) Drill hole sample highlighting pressure relief features, vuggy silica in quartz veins/ veinlets and levels of massive sulfide interspersed with monomineralic Qz veinlets; (F) Polished thin of the yellow delimited field in figure E; (G) Massive sulfide levels showing Py crystals strongly fractured; (H) Sulfide veinlets, with Qz and Ms. Abbreviations: Ms: muscovite; Py: pyrite; Qz: quartz; Ser: sericite/fine muscovite.

4.2.2. Chlorite Chemistry and Geothermometry

The chemical composition of chlorite contributes to the study of hydrothermal alteration processes to delineate the former physical-chemical conditions (T, P, fO_2 , pH, and host rock composition) of the geological environment at the time of its formation, which makes it an important indicator of the rock formation temperature [28,33].

Table S6 (Supplementary Material) presents 18 microprobe analyses of chlorite (i) after biotite (Chl1), (ii) in equilibrium with muscovite and epidote (sericitic alteration) (chl2) and (iii) in association with the late stage carbonatization (Chl3). The compositional variation in chlorite might be related to the partial replacement of igneous biotite by chlorite, which had been depicted from textural features, and is used for a classification scheme.

Magnesium-rich chlorite yields Fe/ (Fe + Mg) ratios varying between 0.33 and 0.56, and Si/Al ratio varying between 1.02 and 1.39. Si atoms range between 5.24 and 5.98 a.p.f.u., and Al^{IV} ranges from 2.02 to 2.76 a.p.f.u. Iron-rich chlorite exhibits Fe/ (Fe + Mg) ratios of 0.67-0.84, and less varying Si/Al ratios of 1.03-1.09. Si atoms range between 6.67 and 0.84 a.p.f.u., and Al^{IV} values fall within the range of 2.49-2.70 a.p.f.u. Based on the classification of [34] these compositions plot at the transition between clinocllore-chamosite (Figure 7A, B).

Empirical geothermometers were proposed by several authors [28,29,34,36]. Each one applies correction criteria based on the characteristics of the given chlorite, as follows: (i) tetrahedral Al (Al^{IV}) contents [28,29]; (ii) corrected Al^{IV} contents based on the

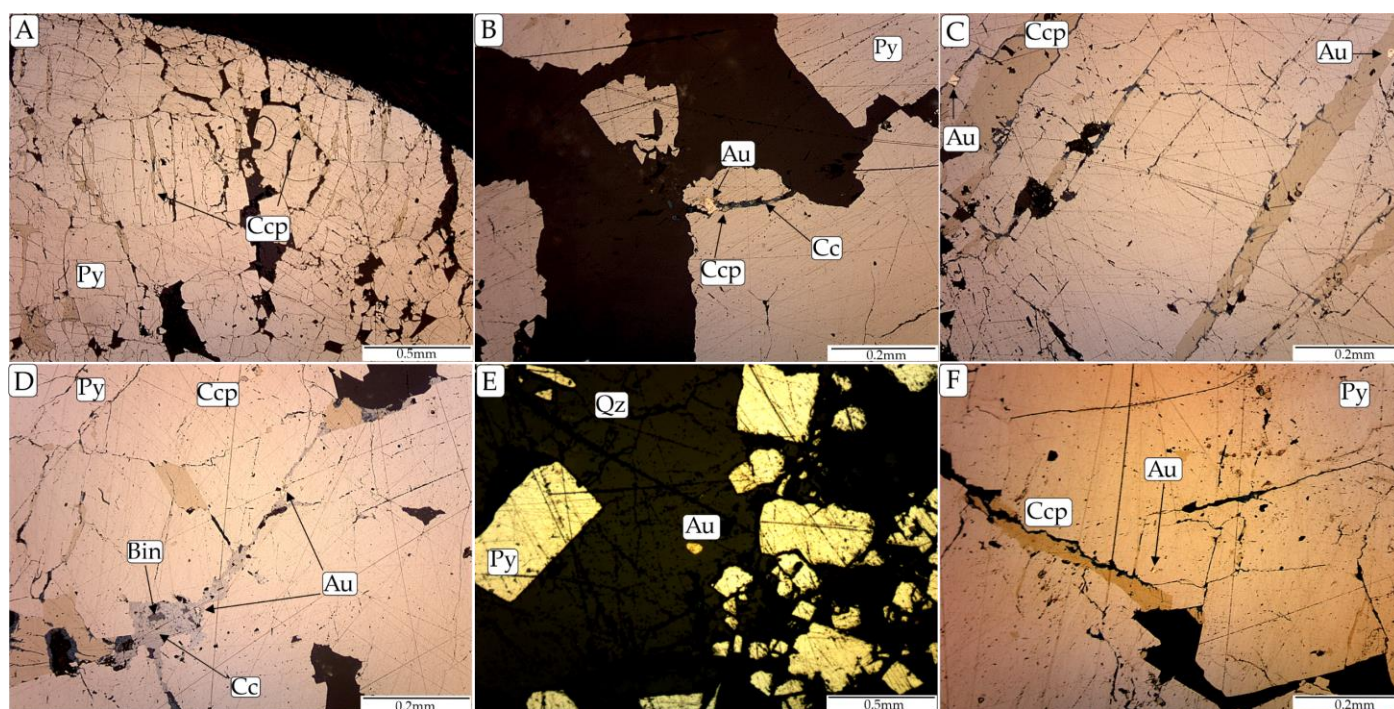


Figure 6. Ore mineral association in the Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil. The images were obtained in reflected light and with parallel polarizers. (A) Massive sulfide texture, with anhedral and strongly fractured Py filled with Ccp, Cc, Bin, au and other copper sulfides; (B, C and D) Strongly fractured pyrite filled by Au-Ccp-Cc-Bin and Cu sulfides association; (E) Au Included in Qz-Py association; (F) Au included in Py crystals strongly fractured and filled with other sulfides. Abbreviations: Au: gold; Bin: bismuthinite; Ccp: chalcopyrite; Cc: chalcocite; Qz: quartz; Py: pyrite.

relationship between Al^{IV} and $Fe/(Fe + Mg)$ ratios [35–38]. Empirical geothermometers used in this study are summarized in Table S6 (Supplementary Material). The peak temperature was obtained through the equation of Cathelineau [29]. The values calculated according to Cathelineau and Nieva [28] are similar to those calculated according to Cathelineau [29]. The lowest temperatures were obtained using the equation of Zang and Fyfe [36]. Based on the features and compositions of the different types of chlorites here identified, the formation temperatures estimated through Cathelineau [29] were selected.

The chlorite geothermometry data are coherent with the petrographic data: the hydrothermal alteration had its onset with propylitic alteration, marked by the replacement of biotite by chlorite (Chl1) along with the formation of other Fe-rich minerals (e.g., rutile) at temperatures between 355 and 382.4 °C. As alteration progressed, chl2, which is associated with sericitic alteration, began to nucleate, at temperatures between 338.9 and 372.8 °C, under the conditions where the first mineralization stage took place. During late hydrothermal alteration, Chl3 was predominant in association with late quartz and carbonate veinlets, at temperatures ranging between 263.3 and 297.87 °C.

4.2.3. Muscovite

Electron microprobe data and the chemical formula of muscovite from the Raimunda deposit are available in Table S7 (Supplementary Material). The muscovite samples chosen for analysis are those associated with the first (1) mineralization stage (sericitic alteration) due to their greater occurrence. 16 spot analyses reveal a relative homogeneous composition with small variations in Al_2O_3 (29.8–31.4 %), MgO (0.6–1.18 %), FeO (1.54–5 %), and Na_2O (0.13–0.4 %) [39] (Figure 7C).

4.3. Fluid Inclusions Data

4.3.1. Petrography and Classification

The fluid inclusion study was carried out on quartz grains from the mineralization stage (Qz2 and Qz3) and late alteration stage (Qz4), using doubly polished thin sections from the Raimunda deposit. Primary and secondary fluid inclusions were described according to the fluid inclusion assemblage (FIA) criteria proposed by [40], which is based on the classification criteria of [31].

Based on distribution, number of phases present at room temperature, patterns of occurrence, phase transitions throughout heating and freezing runs and Raman spectrometry data, nine subtypes of fluid inclusions were recognized. The main characteristics are summarized in Table 2.

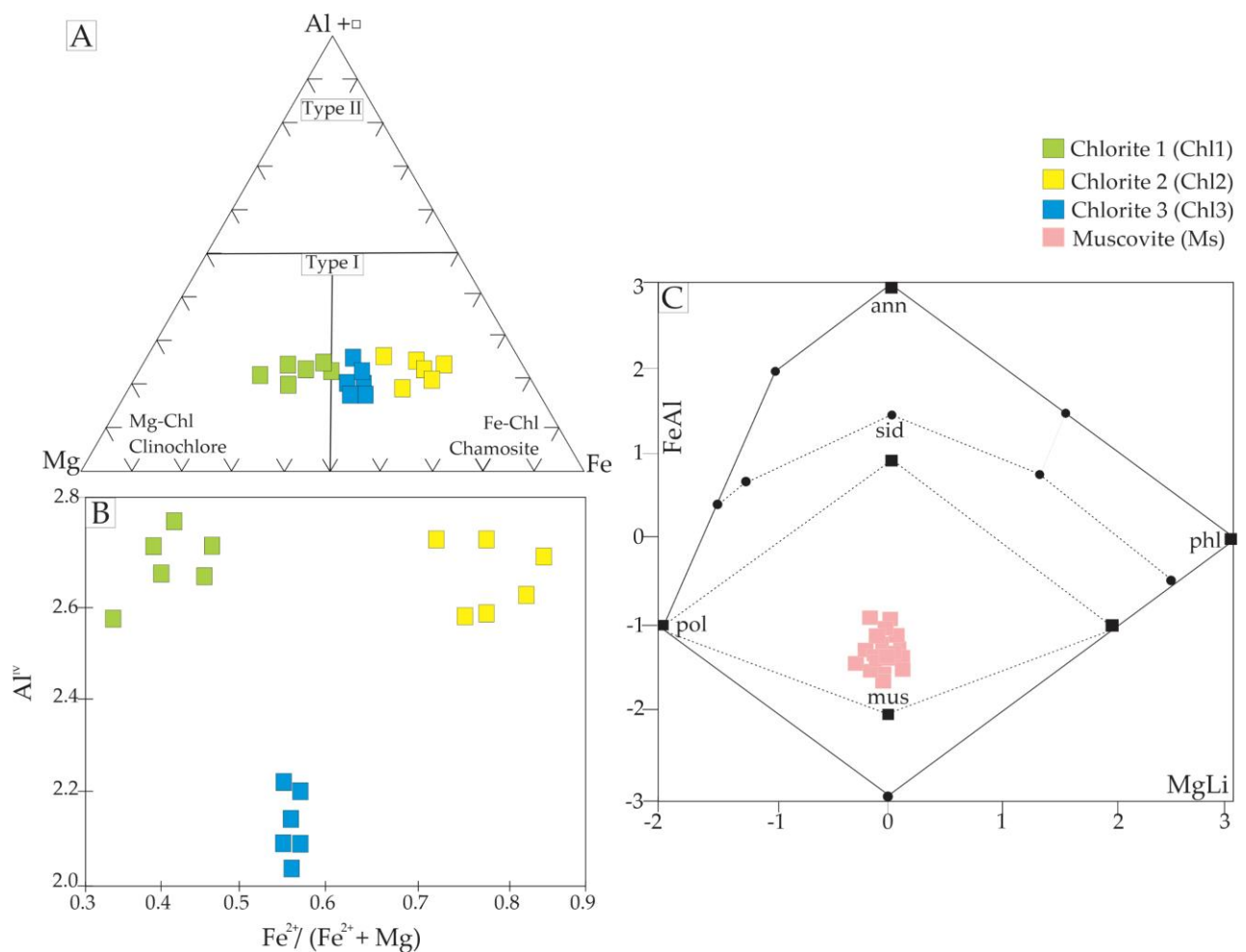


Figure 7. Classification diagrams of chlorite and muscovite. (A) Ternary diagram Mg, Al + □ (□ corresponds to the octahedral vacancy) and Fe, allocating the Fe-rich (chamosite) and Mg-rich (clinochlore) types; (B) Diagram showing the covariance between the Fe/(Fe + Mg) ratio and Al^{IV} for the chlorites of the deposit; (C) FeAl vs. MgLi classification plot of muscovite by [35], placing the three mica species in solid solution. The samples fall into the muscovite field. Abbreviations: Ann: annite; Chl: chlorite; Ms: muscovite; Phl: phlogopite; Pol: polyolithionite; Sid: siderophyllite.

Type 1 (t1): two-phase (liquid + vapor) aqueous liquid-rich primary (t1-a) and secondary (t1-b) inclusions. Type t1-a inclusions show clear evidence of primary fluid, and occur in Qz3, which is associated with sulfidation, as clusters or isolated inclusions with liquid fill ration (F) of 0.70-0.8. They are usually colorless, with a light-colored bubble

surrounded by a dark rim. They are rounded, oval and, more rarely, irregularly shaped, varying between 8-20 μm (Figure 8A, C). Type t1-b inclusions encompasses secondary inclusions in Qz2 associated with sericitic alteration. They mostly occur along trails and subordinately as clusters. They are colorless and display an F of 0.7-0.8. The light-colored bubble moves at room temperature in some samples. They range between 3 and 12 μm , and are rounded, sub-rounded, elongated and, more rarely, rectangular.

Type 2 (t2): primary, single-phase carbonic vapor-rich, two-phase ($\text{L}_{\text{H}_2\text{O}} + \text{V}_{\text{CO}_2}$) and three-phase ($\text{L}_{\text{H}_2\text{O}} + \text{L}_{\text{CO}_2} + \text{V}_{\text{CO}_2}$) aqueous-carbonic fluid inclusions, that occur in Qz3. Single-phase inclusions are gray, with F of 100% CO_2 , varying between 5 and 15 μm . Two-phase and three-phase inclusions display bubbles with a dark rim, varying between 5 and 20 μm . They are sub-rounded, elongated and, more locally, irregularly-shaped. Their F is 30-50 vol% (Figure 8A-C). They occur as single inclusions or in clusters being oval to irregularly shaped. Three-phase inclusions usually display CO_2 bubbles only during reheating, before the homogenization temperature for CO_2 is reached. They are considered part of the same fluid inclusion assemblage (FIA).

Type 3 encompasses three-phase primary (t3-a) and three-phase secondary (t3-b) inclusions ($\text{L}_{\text{H}_2\text{O}} + \text{V}_{\text{H}_2\text{O}} + \text{S}_1$) containing daughter minerals and multiphase-solid phases ($\text{L}_{\text{H}_2\text{O}} + \text{V}_{\text{H}_2\text{O}} + \text{S}_1 + \text{S}_2$). The t3-a inclusions correspond to primary in Qz4, associated with late-carbonatization. It varies between 10 and 20 μm , and is elongated, and sub-rounded; colorless to greenish. They contain cubic solid phases (daughter mineral) that are colorless, isotropic and interpreted as halite based on petrographic and microthermometric data as well as in the lack of Raman signal. Overall, F for both subtypes is 15-20 vol% (vapor), 15-20 vol% (solid), and 60-70 vol% (liquid).

The t3-b fluid inclusion is secondary in Qz3 that occurs as trails and/or aligned associated with sulfidation. They vary between 8 and 22 μm and are square-shaped, elongated and sub-rounded; greenish. The cubic solid phases (daughter mineral) are colorless and, locally, also occur a trapped solid (opaque mineral) in the fluid inclusion (Figure 8D). The F for both subtypes is 15-20 vol% (vapor), 15-20 vol% (solid), and 60-70 vol% (liquid) (Figure 8D, E).

Type 4 (t4): liquid-rich two-phase aqueous inclusions. Primary fluid inclusions (t4-a) in Qz3 are associated with sulfidation. Secondary inclusions (t4-b) predominate in all samples studied and are associated with sericitic alteration (Qz2). In both types, the liquid phase is dominant, with F of 0.8-0.85 and with a vapor bubble content of 15-20 vol%. They are colorless, 8 to 20 μm , and display rounded, sub-rounded, elongated, and irregular shapes. They are generally found as trails/aligned along micro-fractures in the crystal (t3-b), clusters and isolated (t3-a) (Figure 8E).

Type 5 (t5): secondary, liquid-rich two-phase aqueous (L+V) inclusions (Figure 8F). The degree of fill (F) is 0.75-0.8. They are irregular to sub-rounded, vary between 15 and 50 μm , and occur only in Qz2 which is associated with sericitic alteration.

Type 6 (t6): single-phase liquid-rich and/or vapor-rich inclusions. They are colorless to greenish, rounded/oval to irregular, and locally occur along trails varying between 4 and 20 μm . Vapor-rich inclusions are dark, irregular, and vary between 2 and 15 μm . They occur in all samples studied.

4.3.2. Microthermometric Data

Table 2 summarizes fluid inclusions microthermometric data of the Raimunda deposit. Microthermometry was carried out on all types of inclusions in quartz grains previously selected according to petrographic criteria. A total amount of 182 valid measurements were obtained (Table S8, Supplementary Material). The salinity and density of types 1, 4 and 5 aqueous fluid inclusions were calculated based on the equation of Bodnar [40], using the final ice melting temperature (T_{mice}) to estimate salinity in the H_2O -NaCl system. For the fluid inclusions interpreted to be of the system H_2O -NaCl- CaCl_2 , salinities were calculated using the Steele-Macinnis et al. equation [41]. The salinities in H_2O -NaCl- CO_2 inclusions (type 2) were estimated using the final CO_2 -

clathrate melting temperatures [42], and the densities were estimated using the equations of Parry 1986 [43].

Type t1-a shows temperature of first visible melting (T_e) ranging between -24.5 and -21.0 °C and encompasses inclusions with the final ice melting temperatures (T_{mice}) between -6.4 and -0.8 °C (Figure 9A). The total homogenization temperatures (T_{hi}), into the liquid phase, took place between 233.5 and 471.2 °C (Figure 9C), with average values of 325 – 380 °C. The salinities data are between 1.40 and 9.72 wt.% NaCl eq. and the density values between 0.41 and 0.89 g/cm³. In type t1-b, T_{mice} took place between -6.4 and -0.5 °C (Figure 9B), with an average from -4.2 to -0.5 °C and the T_{hi} varies from 200.5 to 297 °C (Figure 9D). The salinity varies from 0.88 to 6.74 wt.% NaCl eq. and densities of 0.71-0.92 g/cm³.

Type 2 has an average CO₂ melting temperature (T_{mCO_2}) values ranging between -56.8 and -56.6 °C for two-phase and three-phase aqueous-carbonic inclusions (Figure 10A). For carbonic inclusions, T_{mCO_2} is -57.3 to -57 °C. These values vary to either more positive or more negative values compared to the pure H₂O-CO₂ system (-56.6 °C). This points out to the presence of other volatile compounds apart from CO₂ (e.g., CH₄ and N₂). The clathrate melting temperatures (T_{mCl_a}) are between 6.2 and 8.5 °C and the CO₂ homogenization temperatures (T_{hCO_2}) vary between 26.8 and 30.1 °C for aqueous-carbonic inclusions, and between 19 and 26 °C for single-phase carbonic inclusions. T_{hi} vary from 210 to 390 °C, but the main range for three-phase (L_{H2O} + L_{CO2} + V_{CO2}) inclusions take place between 280 and 340 °C; and, for two-phase (L_{H2O} + V_{CO2}) inclusions, the main temperature range is 300 – 364.7 °C. Salinities are between 3.00 and 5.94 wt.% NaCl eq. and the densities range from 0.87 to 0.88 g/cm³.

Type t3-a primary inclusions showed T_{mhh} values ranging between -10.7 and -3.8 °C. These inclusions homogenized into the liquid and vapor phase between 290.5 and 367.5 °C (T_{hi}). The salinities are between 38.88 and 50.25 wt.% NaCl + CaCl₂ eq. with densities varying from 0.78 to 0.82 g/cm³.

Type t3-b inclusions show the first observed melting temperature range between -55 to -51.2 °C, interpreted as a eutectic temperature (T_e) and suggesting the composition of H₂O-NaCl-CaCl₂ system [41]. The final melting of ice is between -43.2 and -37.8 °C, and the hydrohalite dissolution (T_{mhh}) occurs between -10.7 and -1.1 °C, with main interval between -9 to -5 °C. During heating, the vapor bubble homogenizes first (132 to 202 °C) and, as temperature increases, daughter minerals homogenize (T_{hi}) in the range of 192 – 278.6 °C, with an average between 210 and 250 °C. The salinities show range values of 33.02–38.21 wt.% NaCl + CaCl₂ eq. and density values range from 0.87 to 0.97 g/cm³. The H₂O–NaCl–CaCl₂ ternary diagram plot the composition for type t3–b inclusions at the Raimunda deposit based on Steele-MacInnis et al. [41] (Figure 10B).

Type 4-a primary inclusions display T_{mice} values between -5.4 and -1.0 °C (Figure 9A) and T_{hi} took place in the range between 200 and 288 °C (Figure 9C). The with corresponding salinity of 1.56 – 8.41 wt.% NaCl eq. and the density was estimated between 0.74 and 0.92 g/cm³. Type t4-b secondary inclusions yielded T_{mice} values in the range of -4.7 and -0.1 °C (Figure 9B). This type homogenized into the liquid phase between 101 and 195 °C (Figure 9D). The salinity is between 0.53 and 6.59 wt.% NaCl eq. with densities between 0.74 and 0.92 g/cm³.

In type 5, the T_{mice} fall between -4.2 and -0.8 °C and total homogenization temperatures (T_{hi}) into the liquid phase range between 98 and 190 °C. The salinity is between 1.39 – 6.74 wt.% NaCl eq. and densities between 0.92 and 0.97 g/cm³ (Figure 9B, D). This type homogenized both into the liquid and vapor phase and the latter showed some inclusions crepitating while other inclusions did not return to their original state before heating

4.4. Stable Isotopes Data

Stable isotope data are summarized in Table 3. The oxygen data were obtained for Qz2, which is associated with sericitic alteration (Ms-Qz2-Py2-Au1), and for Qz3, which

is related to the sulfide stage (Py3-Au2-Ccp-Cc-Bin and Cu sulfides); sulfur data were obtained for Py2 (sericitic alteration) and Py3 (sulfide stage), both associated with the mineralization stages at the Raimunda deposit (Table 1).

The $\delta^{18}\text{O}$ values of quartz vary between 9.0 and 12.6‰. The isotopic values of the fluid ($\delta^{18}\text{O}_{\text{H}_2\text{O}}$) were calculated using the total homogenization temperature range (T_{H}) of primary t1 fluid inclusions (325 – 380 °C) as the isotopic equilibrium temperature for the main mineralized phase, sulfide stage; and the chlorite geothermometric temperature (340 – 370 °C) for the sericitic alteration stage.

These fluid inclusions are associated with the main mineralization stage at the Raimunda deposit. For the quartz-water system, the oxygen isotopic fractionation factor and the isotopic balance equation were used [44]; for oxygen, the following formula was applied to determine the temperature range between 250 and 500 °C: $1000 \ln \alpha_{\text{Qz-Water}} = 3.34 \cdot (10^6 T^{-2}) - 3.31$.

$\delta^{34}\text{S}_{\text{py}}$ values for sulfides slightly varied between -1.4 and + 0.1 ‰. Pyrite associated with sericitic alteration (Py2) yielded a $\delta^{34}\text{S}$ value of -0.5‰; pyrite from massive sulfide zones (Py3) provided $\delta^{34}\text{S}$ values from -1.4 to +0.1‰. The isotopic fractionation factor between pyrite – H_2S were retrieved from [45].

5. Discussion

5.1. Tectonic Settings of Host Rocks, Hydrothermal Zonation and Mineralization

The Au ($\pm\text{Cu}$) Raimunda hydrothermal deposit is hosted in the monzogranite (2.029 ± 4 Ma) and syenogranite (1987 ± 7.4 Ma to 1964 ± 1 Ma) facies, which are part of the Cuiú-Cuiú granite complex (2.03 and 1.98 Ga), eastern region of the Jurueña Mineral Province [24]. These rocks are related to a plutonic-volcanic event in volcanic arc settings [46].

The zones more intensely affected by hydrothermal alteration are indicative of ore precipitation at intermediate- to shallower levels. This is corroborated by pressure-release structures, vuggy quartz, as well as breccias and brittle-like structures typical of shallower depths (Figure 5E).

Gold-rich mineral deposits in Novo Mundo were studied by [19] and [22]. These authors found disseminated gold hosted in intensely hydrothermally-altered syenogranite and monzogranite crosscut by andesitic to basaltic dikes. Those authors observed that mineralization was associated with intense phyllic alteration overprinting strong pervasive potassic alteration [22]. These other deposits share strong similarities with the Raimunda deposit, which might be indicative of similar mineralization events in the region. Other neighboring mineral deposits are hosted in 1.99 – 1.97 Ga granites [22].

In Novo Mundo, the mineralogical composition, the magmatic and hydrothermal textural features, the mineral parageneses, the cutting and overprinting relationships among host rocks and alteration and mineralization zones, microthermometric data, as well as chlorite and muscovite chemistry, allow us to propose the following paragenetic sequences for the Raimunda deposit presented in Table 1.

The evolution of the mineralizing hydrothermal system goes through the development of sericitic alteration (muscovite + pyrite2 + quartz2 + gold1), a lower-temperature stage compared to the previous ones (Na-metassomatism, microclinization and propylitic alteration). Sericitic alteration represents the first mineralization stage, with temperature estimate between 340 and 370 °C based on chlorite chemistry (Table S6, Supplementary Material).

This stage is followed by the main ore precipitation stage, with average temperature estimates between 325 and 380 °C, based on the homogenization temperature of type 1-a fluid inclusions.

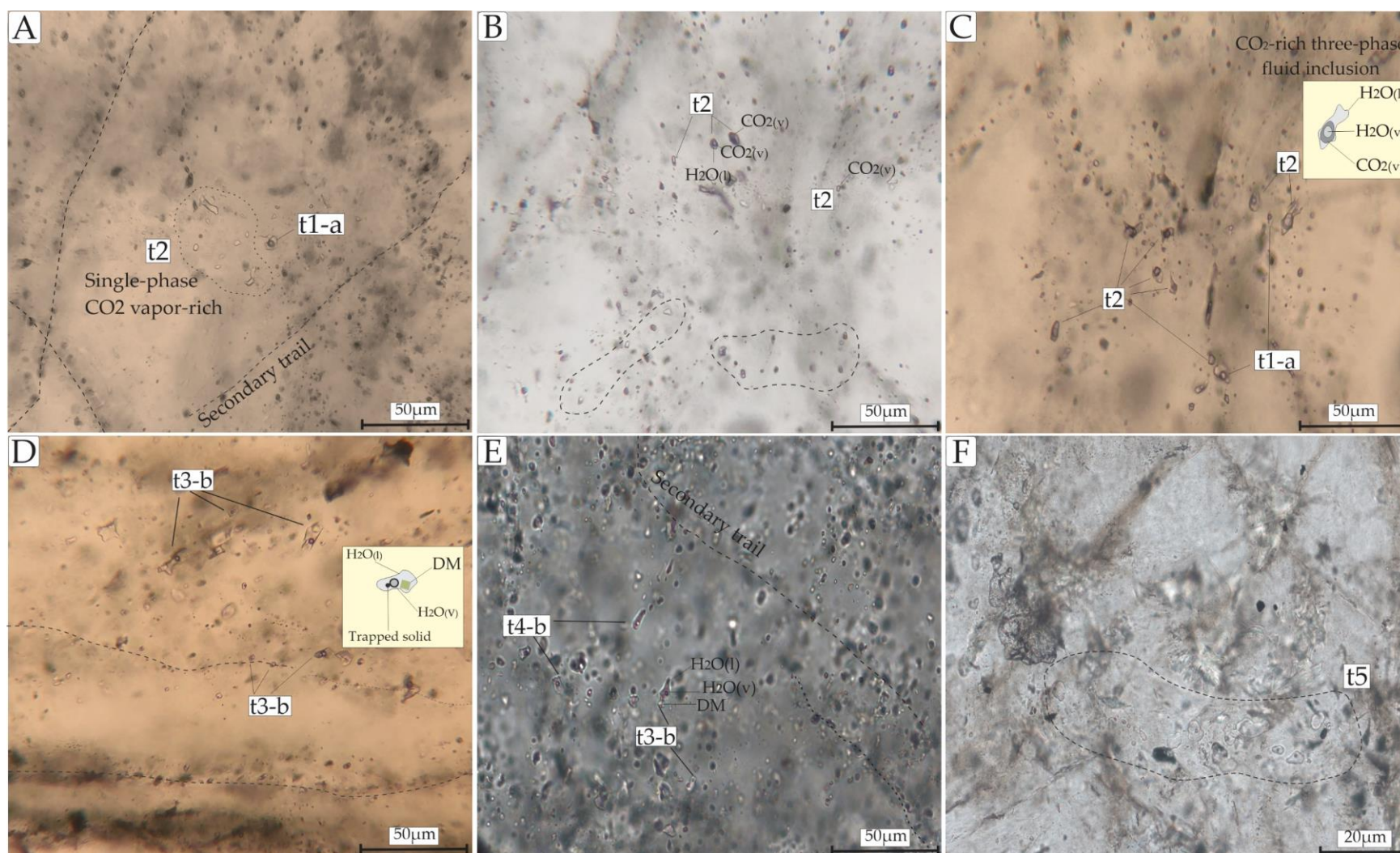


Figure 8. Fluid inclusions mapped in samples from the Raimunda deposit, Juruea Mineral Province, Brazil. (A) t1-a inclusions, primary liquid-rich aqueous and t2, primary, single-phase carbonic vapor-rich inclusions, occurring in clusters or isolated; (B) t2 inclusions, single-phase carbonic and two-phase aqueous-carbonic; (C) t1-a fluid inclusions occurring together with t2 fluid inclusions, two-phase and three-phase aqueous-carbonic; (D) tracks of secondary inclusions of type t3-b,

three-phase (liquid, vapor and halite crystal– DM daughter mineral); (E) 4-b fluid inclusions, two-phase aqueous, and t3-b, three-phase aqueous inclusions; (F) t5 occurring in parallel tracks/lines and inclusions.

Table 2. Petrographic and microthermometric data of fluid inclusions from the Raimunda deposit, Juruena Mineral Province, Brazil. Abbreviations: L – liquid; V – vapor; S – daughter mineral (saturation solid); Ts – trapped solid. T_e : eutectic temperature; T_{mhh} – hydrohalite melting temperature; T_{mice} – ice melting temperature; T_{mCO_2} – CO₂ melting temperature; T_{mcla} – clathrate melting temperature; T_{hCO_2} – CO₂ homogenization temperature; T_{ht} – total homogenization temperature.

Fluid Inclusion Type	Host Rock	Hydrothermal Alteration Stage	Mineral Association	Petrographic Characteristics	Microthermometric Data	Model System
Type 1-a (L + V); $n = 28$	Monzogranite; syenogranite	Sulfide stage	Py3 + Ccp + Qz3 + Au2	Primary; random; colorless; ellipsoidal, oval to irregularly shaped; 8–20 μm ; F: 0.70–0.8	$T_e = -24$ to -21 °C; $T_{mice} = -6.4$ to -0.8 °C; $T_{ht} = 233.5$ to 471.2 °C; salinity = 1.396–9.72 wt.% NaCl eq.; density; 0.409–0.895 g/cm ³	H ₂ O–NaCl
Type 1-b (L + V); $n = 25$	Monzogranite; syenogranite	Sericitic alteration	Ms + Qz2 + Au1	Secondary; planar grouping and trails; oblate; 3–12 μm ; F: 0.7–0.8	$T_{mice} = -4.2$ to -0.5 °C; $T_{ht} = 200.5$ to 297 °C; salinity = 0.879–6.737 wt.% NaCl eq.; density = 0.715–0.918 g/cm ³	H ₂ O–NaCl
Type 2 (V; L + V; L + L + V); $n = 40$	Monzogranite	Sulfide stage	Py3 + Ccp + Qz3 + Au2	Primary; random distribution or in clusters; grayish; dark bubble; elongated and oval; 5–20 μm ; F: 30–50	$T_{mCO_2} = -57.3$ to -56 °C; $T_{mice} = -3.8$ to -2.9 °C; $T_{mcla} = 6.2$ – 8.5 °C; $T_{hCO_2} = 19$ – 30.1 °C; $T_{ht} = 210$ to 364.7 °C; salinity = 3.00–5.94 wt.% NaCl eq.; density = 0.87–0.88 g/cm ³	H ₂ O–NaCl– CO ₂
Type 3-a (L + V; L + V + S); $n = 3$	Monzogranite	Carbonatization	Cal + Chl3 + Qz4	Primary; random distribution; square-shaped, elongated, and sub-rounded; colorless or greenish; 8–22 μm ; F: 60–70; with one cubic crystal (halite daughter mineral)	$T_{mhh} = -10.7$ to -3.8 °C; $T_{ht} = 290.5$ to 367.5 °C; salinity = 38.88–50.25 wt.% NaCl + CaCl ₂ eq.; density = 0.78–0.82 g/cm ³	H ₂ O–NaCl– CaCl ₂
Type 3-b (L + V; L + V + S + Ts); $n = 19$	Monzogranite	Sericitic alteration	Py + Ccp + Qz3	Secondary; planar grouping and trails; oblate square-shaped, elongated, and sub-rounded; 3–12 μm ; F: 60–80; with colorless cubic crystals (halite daughter mineral) and trapped solid locally	$T_e = -55$ to -51.2 °C; $T_{mice} = -43.2$ to -37.8 °C; $T_{mhh} = -10.6$ to -1.1 °C; $T_{ht} = 100.6$ to 251.5 °C; salinity = 33.02–38.21 wt.% NaCl + CaCl ₂ eq.; density = 0.87–0.97 g/cm ³	H ₂ O–NaCl– CaCl ₂
Type 4-a (L + V); $n = 19$	Monzogranite; syenogranite	Silicification; Sulfide stage	Py3 + Ccp + Qz3 + Au2	Primary; clusters and isolated; colorless; 8–20 μm ; F: 0.8–0.85	$T_{mice} = -5.4$ to -1.0 °C; $T_{ht} = 199.2$ to 288 °C; salinity = 1.57–8.41 wt.% NaCl eq.; density = 0.74–0.92 g/cm ³	H ₂ O–NaCl
Type 4-b (L + V); $n = 41$	Monzogranite; syenogranite	Sericitic alteration	Ms + Qz2 + Au1	Secondary; trails/aligned along micro-fractures in the quartz; 8–20 μm ; F: 0.8–0.85	$T_{mice} = -4.7$ to -0.1 °C; $T_{ht} = 101$ to 195 °C; salinity = 0.53–6.59 wt.% NaCl eq.; density = 0.87–0.97 g/cm ³	H ₂ O–NaCl
Type 5 (L + V); $n = 8$	Monzogranite; syenogranite	Sericitic alteration	Ms + Qz2 + Au1	Secondary; irregular to sub-rounded shaped; colorless; 15 and 50 μm ; F: 0.75–0.8	$T_{mice} = -4.2$ to -0.8 °C; $T_{ht} = 98$ to 190 °C; salinity = 1.39–6.74 wt.% NaCl eq.; density = 0.92–0.97 g/cm ³	H ₂ O–NaCl
Type 6 (L; V)	Monzogranite; syenogranite	All assemblages		Colorless to greenish; rounded/oval to irregular; 4 and 20 μm ; it occurs in all samples studied		H ₂ O–NaCl

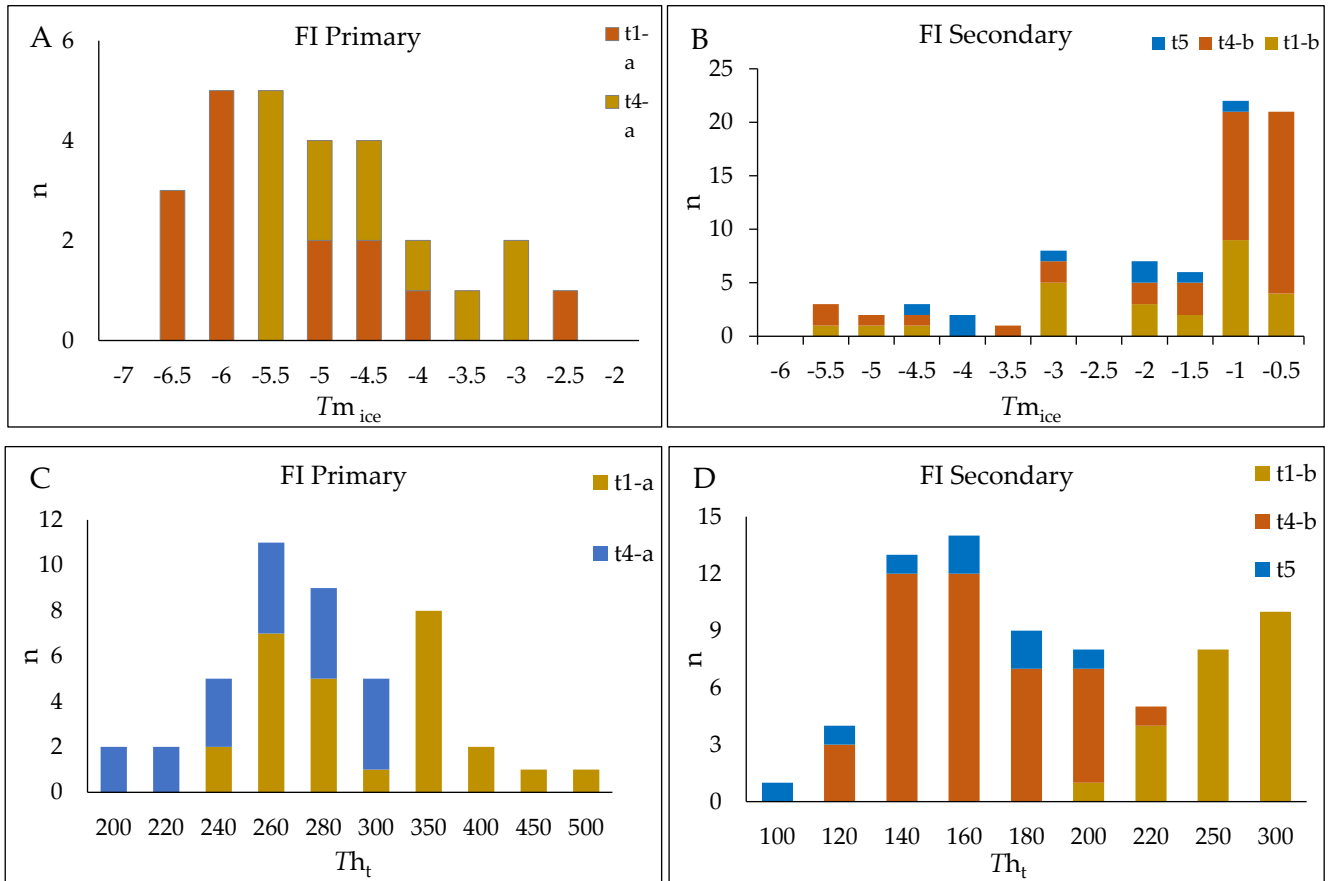


Figure 9. Microthermometric histograms of H₂O–NaCl aqueous fluid inclusions from the Raimunda deposit, Novo Mundo, Jurueña Mineral Province. (A, B) Variation of T_m_{ice} vs. frequency of primary (t1-a and t4-a) and secondary (t1-b, t4-b and t5) inclusions; (C and D) Th_t of primary (t1-a and t4-a) and secondary (t1-b, t4-b and t5) inclusions.

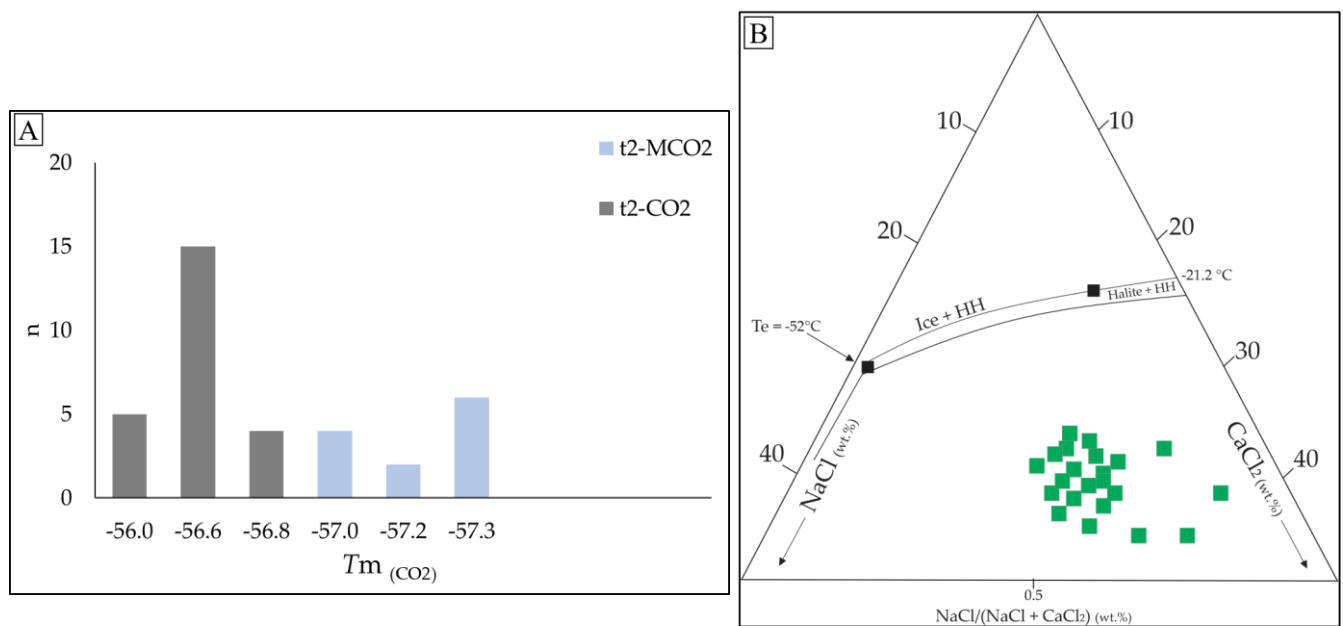


Figure 10. A) Fluid inclusions microthermometric T_mCO₂ histogram vs. frequency; (B) Ternary diagram showing the composition of t3 fluid inclusions (green square boxes) in terms of mass percent equivalent in NaCl and CaCl₂ [41].

Table 3. Isotopic composition of hydrothermal minerals and fluids from the Raimunda deposits, Novo Mundo granite, Juruena Mineral Province, Brazil.

Sample	Mineral	Hydrothermal alteration stage	$\delta^{34}\text{S}_{\text{Py}}$ (‰)	$\delta^{18}\text{O}_{\text{Qz}}$ (‰)	$\delta^{18}\text{O}_{\text{H}_2\text{O}}$ (‰) [44]	Estimated isotopic equilibrium temperature (°C) from fluid inclusion data
F75-05	Qz2	Sericitic alteration		9	3.11 – 3.99	340 – 370
F70-02				9.6	3.72 – 4.59	
F73-02				10.3	4.72 – 5.29	
F70-03	Py2		-0.5			
	Qz2	Sulfide stage		11.5	5.62 – 6.50	325 – 380
F73-03	Py3		0.1			
	Qz3			12.6	6.22 – 7.86	
F48-A15	Py3		-1			
	Qz3	Sulfide stage		12.2	5.82 – 7.46	325 – 380
F70-01	Py3		-0.1	11.1	4.73 – 6.36	
F78-05	Py3		-0.7			
	Qz3			11.6	5.23 – 6.85	
F73-04	Py3		-1.4			
	Qz3			10.7	4.32 – 5.95	

The following post-mineralization stage is marked by cooling and decompression of hydrothermal fluids [47] and is represented by chloritization-carbonate alteration, with average temperatures of 262 – 280 °C, based on chlorite geothermometer (chl3).

5.2. Origin, Hydrothermal System, and Evolution of Fluid Composition

Isotopic and fluid inclusion data indicate an evolution from hot, medium-salinity fluid (> 450 °C and ≤10 wt.% NaCl eq.) to cooler, medium- to low-salinity fluids (< 300 °C and ≤5 wt.% NaCl eq.).

The petrographic characteristics and fluid inclusion microthermometric data reveal the existence of two fluids: a low- to medium salinity H₂O–NaCl–CO₂ aqueous-carbonic fluid (3.0–5.94 wt.% NaCl eq.) and a low-temperature high-salinity H₂O–NaCl–CaCl₂ aqueous late fluid (33.02–50.25 wt.% NaCl + CaCl₂ eq.), and that make up the fluid system of the Au (± Cu) Raimunda deposit.

The CO₂ content in aqueous-carbonic and carbonic fluid inclusions might be related to hydrolysis during hydrothermal alteration of the host rock (propylitic alteration), which consumed water from the fluid, enriching it in CO₂. The high homogenization temperatures observed at these inclusions (up to 472 °C) corroborate this interpretation.

Aqueous fluids are divided into two chemical systems, based on the relationship *Te* versus salinity: 1) H₂O–NaCl fluid in Qz2 and Qz3, with *Te* between -24.5 °C and -21.4 °C, and salinity between 0.87 and 9.72 wt.% NaCl eq.; 2) H₂O–NaCl–CaCl₂ fluid in Qz3 and Qz4, with the first melting temperature observed, interpreted as a *Te*, between -55 and -51.2°C, and salinity between 33.02–50.25 wt.% NaCl + CaCl₂ eq. (Table S8, supplementary Material).

The saturated fluid displays homogenization temperatures T_h (290.5 – 367.5 °C, for t3-a; 200 – 250 °C, for t3-b main interval) with complete dissolution of the daughter mineral, thereby indicating homogeneous entrapment. And, despite of the occurrence of two primary fluids (H_2O-CO_2 and $H_2O-NaCl$) deriving from a single fluid and the variable homogenization into the vapor and liquid phases (t3 and t5), these inclusion types do not belong to the same assembly (FIA), and the homogenization occur in a medium- to large range of temperature, which is compatible with a homogeneous entrapment.

The occurrence of brines in aqueous and aqueous-carbonic primary fluid inclusions can yield for the fluid mixing hypotheses, led to brine dilution and an increase in salinity for aqueous-carbonic fluid therefore explaining the presence of the late-carbonate alteration at the deposit. Fluid mixing may have taken place at the time of formation of Qz3, with ongoing brine activity through the formation of Qz4 at a lower temperature, when the aqueous-carbonic fluid began to reduce its activity and precipitate calcite, which is corroborated by late hydrothermal alteration (Cal + Chl3).

Thus, the evolution of the medium-salinity and very hot system (> 470 °C) is linked to the stages of hydrothermal alteration prior to mineralization (sodic metasomatism, microclinization, and propylitic alteration); and the medium- to low salinity (< 10 wt.% NaCl eq.) cool system or progressively cooler (< 380 °C) is associated with the main to late alteration stages sericitic alteration and contemporaneous Sulfide stage associated with silica veins and carbonate alteration. Furthermore, the transition from microclinization into late carbonate alteration can suggest a progressive increase in pH values [48–50], which likely lasted throughout ore precipitation.

The calculated isotopic composition of the hydrothermal fluid responsible for both mineralization stages shows $\delta^{18}O_{H_2O}$ values for stage 2 in the range between +4.32 and +7.86‰ (sulfide stage), while those calculated for stage 1 vary between +3.11 and +6.49‰ (sericitic alteration) (Table 3), which is coherent to isotopic compositions of fluids associated with felsic magmas [51], and are compatible with the magmatic compositional isotopic range expected for world-class porphyry-type magmatic-hydrothermal deposits, with a homogeneous isotopic fluid ($\delta^{18}O_{H_2O}$, 6–10‰) [52–55]. The results are similar to paragenetic and fluid inclusion data, and point to mixing between magmatic and meteoric fluids, that have values within the $\delta^{18}O$ range of –20 to –10‰ according to Craig [56] but remained predominantly with characteristics of magmatic fluid [57].

Figure 11 compares $\delta^{18}O_{H_2O}$ values, based on temperatures provided by chlorite chemistry, which is associated with sericitic alteration and the first mineralization stage, and $\delta^{18}O_{H_2O}$ values based on homogenization temperatures from fluid inclusions associated with the main mineralization stage at the Raimunda deposit. The two groups show an isotopic range similar to those obtained for magmatic-hydrothermal deposits with mantle-derived fluid.

The coexistence of t1-a and t2 primary inclusions at the deposit, with similar temperatures, can indicate simultaneous fluid entrapment. Phase changes among the different types of inclusions may suggest that fluid immiscibility preceded their entrapment; homogenization temperatures decrease from type t1-a to t4-a on the salinity vs. homogenization temperatures diagram (Figure 11B), which is likely due to progressive cooling and/or mixing with relatively lower-temperature fluids, such as meteoric water [58].

$\delta^{34}S$ data directly reflects its source given that under proper redox and pH conditions for precipitation, much or all the sulfur available in the system may have been converted into H_2S at the first mineralization stage [59]. In turn, once

the second mineralization stage is composed of sulfides only, the isotopic composition of sulfur in pyrite matches that of the sulfur source [59].

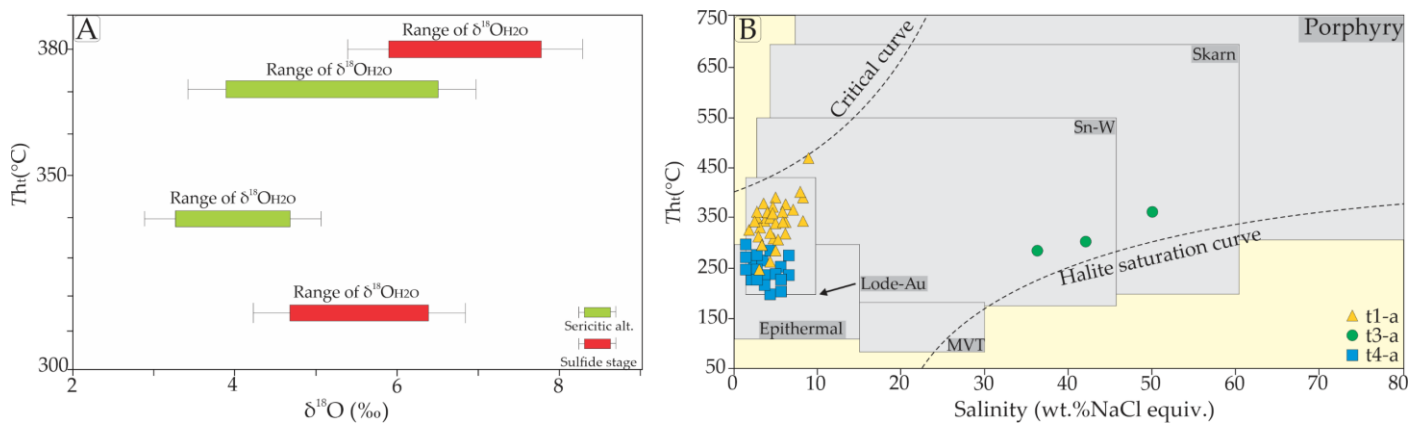


Figure 11. A) Diagram relating the $\delta^{18}O$ values of the hydrothermal fluid vs. the homogenization temperature range of fluid inclusions related to the main stage of mineralization, based on [54]. The diagram shows green bars, representing the fluid obtained from quartz grains in paragenesis with sericitic alteration, and red bars, for quartz grains in paragenesis with sulfide stage (Py3-Au2-Ccp-Cc-Bin), considering the minimum and maximum temperature range of the main stage of mineralization of the deposit; B) Diagram of salinity vs. homogenization temperature of fluid inclusions corresponding to the main stage of mineralization, in relation to the main magmatic-hydrothermal deposits. The Raimunda deposit analyses plotted in the porphyry field. The salinity-temperature curves and fields are based on [58].

Figure 12 correlates $\delta^{34}S$ values of the Raimunda deposit (-1.4 to $+0.1$ ‰) with the isotopic ranges of several deposits in the Tapajós and Juruena Provinces, as well as with world-class deposits and deposits and geological environments available in the literature and are compatible with those of other world-class magmatic-hydrothermal deposits [60–62].

Reference values of $\delta^{34}S$ from [63], combined with petrological and mineral textural evidence at the Raimunda deposit, are indicative of mineralization of magmatic origin, which may be related to the fluids that hydrothermally altered the Novo Mundo granite. This is corroborated by the similarity between $\delta^{34}S$ values and hydrothermal patterns at the Raimunda deposit and those reported for other world-class magmatic-hydrothermal deposits and/or granite-related mineral deposits [64–68].

The occurrence of primary gold in the Juruena Mineral Province demonstrates a close spatial relationship with (i) oxidized I-type calc-alkaline granites, (ii) volcanic rocks, (iii) volcano-sedimentary sequences linked to continental magmatic arcs [4]. And the common occurrence of native Au as inclusions in pyrite and chalcopyrite are indicative of early Au oversaturation in the fluid which we relate to sulfide precipitation and phase separation, destabilizing the $AuHS^0$ or $Au(HS)^2^-$ complex, leading to the accumulation of Au particles [67].

At the Raimunda deposit, ore is composed of native gold and has a mantle-like polymetallic signature ($Au \pm Ag \pm Cu \pm Bi \pm Te \pm Pb$) associated with pyrite, chalcopyrite, chalcocite, larosite, bismuthinite and other bismuth sulfides emplaced at a zone of intense sulfidation.

This copper-rich sulfide ore assemblage suggests that the Raimunda deposit formed under fO_2 conditions above the PPM buffer, pyrrhotite + pyrite + magnetite, which is close to $mSO_2/mH_2S = 1$ [19]. It is compatible with moderate to high fO_2

conditions where fluids oscillate from low to high temperatures; the latter being represented by moderate-salinity aqueous-carbonic fluids (up to ≤ 370 °C and up to ≤ 10 wt% NaCl eq.). This scenario suggests that ore precipitation is linked and/or similar to porphyry magmatic-hydrothermal systems [71].

Hence, our hypothesis is that during the final stage of granite crystallization (~ 2.03 – 1.98 Ga), high-salinity and higher-temperature single-phase fluids, with an intermediate density, were exsolved from a magma at high depth. These fluids went through immiscibility and separated into H_2O -NaCl and CO_2 -rich fluids. During their ascent, fluids decompressed and cooled down [56], while ions (e.g., Au^+ , Ag^+ , Te^{2-} and S^{2-}) precipitated along favorable structures, under proper P–T conditions, giving rise to the first mineralization stage.

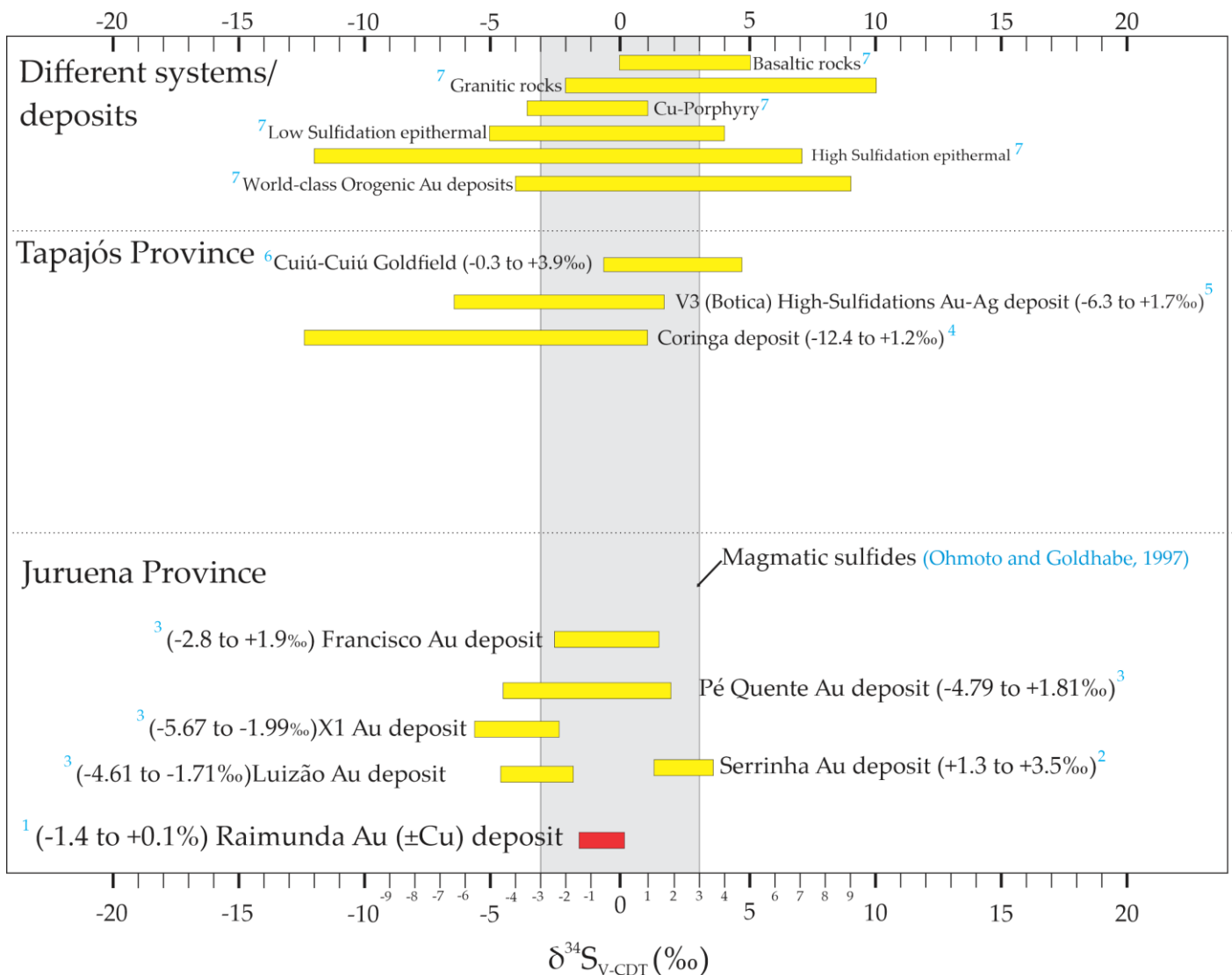


Figure 12. Sulfur isotope values ($\delta^{34}\text{S}_{\text{CDT}}$) of different geological mineral deposits, isotopic data from the Tapajós and Juruena Province, Amazonian Craton. References: 1 – this study; 2 – [2]; 3 – [20]; 4 – [8]; 5 – [68]; 6 – [71]; 7- Mineral deposits from Shanks [51]. The magmatic sulfides range (-3 to +3‰) corresponds to [62].

Other ore-forming elements (i.e., Au, Cu^+ , Ag^+ , Fe^{2+} , S^{2-} and Te^{2+}) were exsolved from the magmatic fluid and precipitated along microfractures and stockwork

sulfide veins. This last stage gave rise to the main stage of mineralization at the deposit, which is coherent with the ore mineral paragenesis.

The ore precipitation was triggered by the cooling of magmatic fluid expelled from magmatic chambers at shallower depths [67]. This model outlines that the input of meteoric water would only affect peripheral and/or post-mineralization hydrothermal alteration, such as propylitic alteration [57, 67], without playing a significant role in the main mineralization stage, and the homogeneous $\delta^{18}\text{O}_{\text{fluid}}$ values (3.11 – 7.86 ‰) corroborates this scenario. And the relative depth gold precipitation and, consequently, deposit formation can be helpful to preserve the Raimunda deposit until these days (Figure 13).

All these features are compatible with those of porphyry magmatic-hydrothermal deposits, commonly found in Phanerozoic volcano-plutonic settings [e.g., 72]. Although they lack some classic features of Phanerozoic analogous surrounding the Raimunda deposit, which may be due to variable erosion levels, it is likely associated with mineralizing events reported earlier for the Tapajós-Parima Paleoproterozoic Province [8].

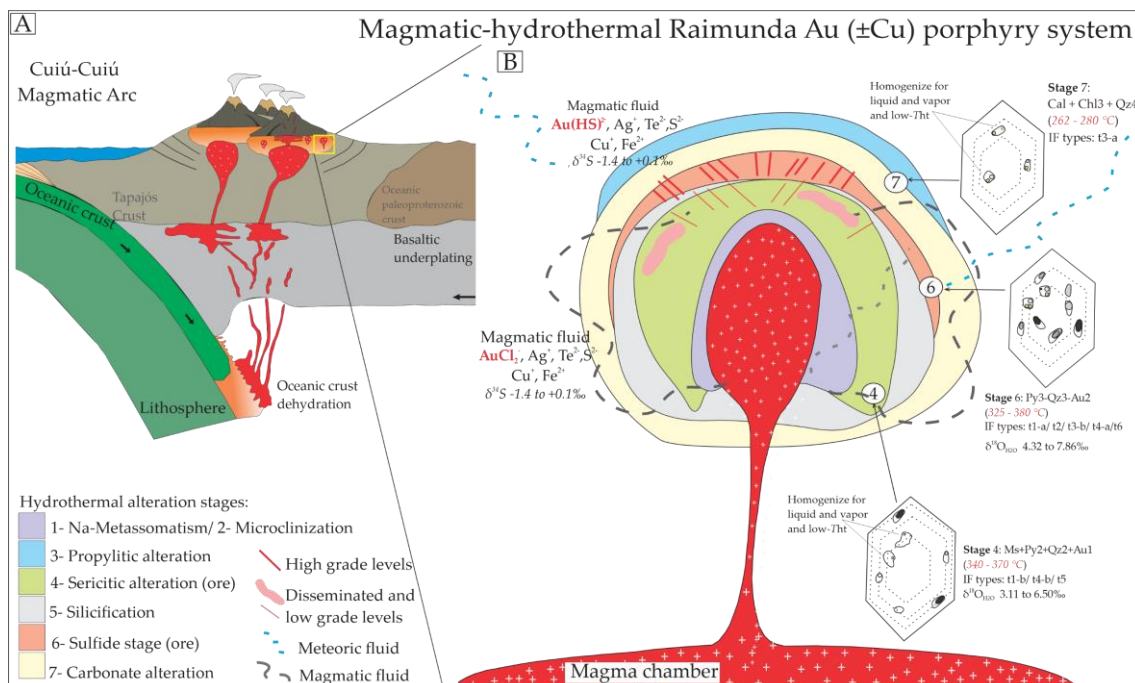


Figure 13. Graphic mineralization model showing the magmatic and hydrothermal periods of the Raimunda Au (±Cu) deposit, Jurueña Mineral Province, Amazonian Craton. (A) The schematic representation of granite generation to Novo Mundo granites [24] during the evolution of the Paleoproterozoic Cuiú-Cuiú magmatic arc (2.03 – 1.96 Ga); (B) The hydrothermal and fluid evolution, indicating the main fluid stages, temperatures and isotope composition of Raimunda deposit. The zonation is only illustrative, as there is overlap of stages in Raimunda system.

6. Conclusions

The current study allows us to draw the following conclusions about the Au (±Cu) Raimunda deposit:

The gold mineralization is hosted in arc I-type calc-alkaline, peraluminous to metaluminous granitic rocks, subdivided into four granite facies (monzonite, monzogranite, biotite granodiorite and syenogranite), with U – Pb zircon ages of

2.03-1.98 Ga, named Novo Mundo Complex, eastern region of the Juruena Mineral Province, Amazonian Craton.

The hydrothermal system resembles that of mineral deposits previously studied in the Novo Mundo Complex and in the eastern region of the province, which reveals an important regional association.

Hydrothermal alteration data and mineral and fluid chemistry data allow us to establish the following relationships: (a) the initial stages of alteration indicate temperatures $> 470\text{ }^{\circ}\text{C}$; (b) progressive evolution of the hydrothermal system led to a temperature decrease that culminated into the first mineralization stage associated with sericitic alteration (340 and $370\text{ }^{\circ}\text{C}$); (c) it was followed by the main stage of ore precipitation, sulfide stage (Py3-Ccp-Cc-Bin-Au2), with average temperatures of $325 - 380\text{ }^{\circ}\text{C}$, as demonstrated by the homogenization temperature of type 1-a primary fluid inclusions; (d) the post-mineralization stage, marked by carbonate alteration and chloritization (chl3), yielded average temperatures from 262 to $280\text{ }^{\circ}\text{C}$ for the system.

Au ($\pm$ Cu) mineralization occurs as Ag-rich ($25 - 28\%$, electrum) gold inclusions and fracture-fills in pyrite and the main mineralization stage mineralization at the deposit is composed of the following sulfide assemblage: pyrite, chalcopyrite, chalcocite, larosite, bismuthinite, hoderushite, krupkaite and wittchenite.

The fluid inclusion study identified nine different subtypes of inclusion, varying from a $\text{H}_2\text{O}-\text{NaCl}$, $\text{H}_2\text{O}-\text{NaCl}-\text{CO}_2$ and $\text{H}_2\text{O}-\text{NaCl}-\text{CaCl}_2$ model systems. And based on the relationship T_e versus salinity, were possible to identified two chemical systems in the Raimunda deposit: 1) $\text{H}_2\text{O}-\text{NaCl}$ fluid in Qz2 and Qz3, with T_e between -24.5 and $-21\text{ }^{\circ}\text{C}$ and salinity between 0.87 and $9.72\text{ wt.}\%$ NaCl eq.; 2) $\text{H}_2\text{O}-\text{NaCl}-\text{CaCl}_2$ fluid in Qz3 and Qz4, with T_e between -55.3 and $-51.2\text{ }^{\circ}\text{C}$, and salinity of $33.02-50.25\text{ wt.}\%$ NaCl + CaCl_2 eq.

The Raimunda deposit evolved from essentially high- to medium temperature ($\leq 472\text{ }^{\circ}\text{C}$) and medium salinity ($\leq 10\text{ wt.}\%$ NaCl eq.) magmatic fluids, mixed with fluids meteoric (T_{ht} $98 - 190\text{ }^{\circ}\text{C}$; lower salinity: $1.39 - 6.74\text{ wt.}\%$ NaCl eq.).

Ore precipitated under considerably high $f\text{S}_2$ and $f\text{O}_2$ conditions, and relatively medium- to high temperatures (e.g., homogenization temperature of t1-a primary fluid inclusion suggest ~ 325 to $380\text{ }^{\circ}\text{C}$).

Isotopic composition of the mineralizing fluid points out to magmatic fluid, with $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ values of $3.11-7.86\text{ }^{\circ}\text{‰}$, indicate magmatic origin with meteoric fluid contribution. And the main stage of ore deposition was triggered by the instability of magmatic fluids with $\delta^{34}\text{S}_{\text{py}}$ values between -1.4 to $+0.1\text{ }^{\circ}\text{‰}$, similar to several regional and world-class magmatic-hydrothermal deposits.

The overall features of the Raimunda deposit are genetically compatible with those of magmatic-hydrothermal systems related to Au-mineralized porphyry formed from oxidized granitic magmas. These findings have implications for mineral exploration of gold and associated metals in the Juruena Mineral Province and Tapajós Province, Amazonian Craton.

Supplementary Materials: The following supporting information can be downloaded at: <https://www.mdpi.com/article/10.3390/min14121185/s1>, Table S1: Electron microprobe analyses results for pyrite crystals to Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil, Table S2: Electron microprobe analyses results for chalcopyrite to Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil, Table S3: Electron microprobe analyses results for gold to Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil, Table S4: Electron microprobe analysis of chalcocite crystals and other copper-rich minerals in samples from Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil, Table S5: Electron microprobe analysis of bismutinita and other copper-rich minerals in samples from Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil, Table S6: Electron microprobe analyses results for chlorite from the Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Juruena Mineral Province, Brazil. Abbreviations: CAT:

Cathelineau (1988) [29]; CAT&N: Cathelineau & Nieva (1985) [28]; Z&F: Zang & Fyfe (1995) [37], Table S7: Electron microprobe analyses results for muscovite to Raimunda Au ($\pm$ Cu) deposit, Novo Mundo, Jurueña Mineral Province, Brazil, Table S8: Summary of microthermometric data of fluid inclusions from the Raimunda deposit, Novo Mundo, Jurueña Mineral Province.

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CAPÍTULO 6

CONSIDERAÇÕES FINAIS

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O complexo granítico Novo Mundo é formado por rochas de composição granítica, como o monzonito, monzogranito, biotita granodiorito e sienogranito, sendo recortado por diques máficos de composição basáltica à andesítica. A idade U–Pb em zircão das rochas félsicas que compõem o complexo mostrou intervalo entre 2,032 e 1,987 Ga e, juntamente com as características geológicas e geoquímicas, foi possível interpretá-lo como representante da fase orogênica do Arco Magmático Cuiú-Cuiú (2,050 a 1,950 Ga).

O modelo alternativo aqui levantado propõe que os granitos do complexo Novo Mundo resultam de um evento de subducção oceano-contidente de longa duração, que produziu arcos magmáticos continentais (entre 2,0 a 1,77 Ga), coerente com os dados obtidos e existentes para o magmatismo cálcio-alcálico na Província Mineral Juruena.

Os dados geoquímicos indicam que os granitos são do tipo I, meta a peraluminosos ($A/CNK = 0,7 - 1,23$) e formados em ambiente de arco vulcânico. O monzonito, de idade 2,023 Ga, tem em sua composição a hornblenda e a biotita primárias. Os dados isotópicos de Sm–Nd mostram idade modelo T_{DM} de 2,156 Ga e $\epsilon_{Nd(t)} = +1.81$, interpretando-o como resultante das fases iniciais do evento plutônico no Arco Magmático Cuiú-Cuiú.

O monzogranito ocorre majoritariamente na área. Apresenta muscovita primária, indicando seu caráter peraluminoso e, com base em dados geoquímicos, de química mineral e isotópicos, foi gerado em ambiente de arco vulcânico com contribuição crustal.

O biotita granodiorito tem a biotita como principal mineral máfico, seguido pela hornblenda. Possui tendência magnésiana e afinidade meta- a peraluminosa, típica de corpos graníticos do tipo I, com composições cálcio-alcálica a alcali-cálcicas provenientes de ambiente de arco vulcânico. A idade T_{DM} ocorre entre 2,20 e 2,48 Ga e o ϵ_{Nd} , entre 0,44 e 1,39, indicando a participação de fonte juvenil ou mistura de magmas félsicos de diferentes composições ou ligados à entrada de magma máfico no sistema.

O sienogranito (U–Pb 1987,7 $\pm$ 7 Ga) tem assinatura cálcio-alcálica e os dados litogeoquímicos indicam um ambiente de arco vulcânico para sua formação. A assinatura T_{DM} de 2,29–2,23 Ga e os valores de $\epsilon_{Nd} = -0,03$ a $-1,12$ indicam forte contribuição crustal. A interação do manto e das fontes crustais neste ambiente acrescionário é potencial gerador de magmas híbridos, condição sustentada pelos valores de ϵ_{Nd} ligeiramente negativos a positivos apresentados pelas rochas do Complexo Novo Mundo.

Em ambiente colisional, a fonte geradora dos granitos seria um manto litosférico proveniente da cristalização fracionada, sob diferentes condições de pressão (crosta continental espessa vs. crosta mais fina) e mistura entre câmaras magmáticas de composição félsica em ambiente de subducção, com contribuição crustal de variadas proporções durante o período de colocação do Arco Magmático Cuiú-Cuiú.

Os dados obtidos, juntamente com estudos recentes encontrados na literatura, permitiu-nos sugerir a existência de um arco magmático contínuo Paleoproterozoico na parte sul do cráton amazônico, responsável pela formação das principais rochas graníticas das províncias minerais Tapajós e Juruena.

Diante deste cenário, hospedado em granitos que foram datados e atribuídos ao Arco Magmático Cuiú-Cuiú, ocorre o depósito Au ($\pm$ Cu) Raimunda, na porção leste da Província Mineral Juruena.

As alterações hidrotermais identificadas no complexo Novo Mundo compreendem o metassomatismo sódico incipiente, microclinização, alteração propilítica e sericítica, silificação, estágio de sulfetação e carbonatação tardia. O depósito de ouro ($\pm$ Cu) Raimunda representa um sistema mineral disposto em dois estágios principais: 1) o minério ocorre de forma disseminada, associada à alteração sericítica (Ser+Ms+Qz2+Au1); 2) e em níveis de sulfeto maciço, referente à principal fase da mineralização (Py3+Ccp+Au2 + sulfetos de cobre e de bismuto).

Foram descritos dois grupos de fluidos no depósito, a partir do estudo de inclusões fluidas em quartzo: 1) fluidos carbônicos e aquoso-carbônicos de origem magmática, possivelmente relacionados ao posicionamento do magma em níveis mais rasos da crosta e o principal aprisionamento ocorreu entre 300 e 360 °C (intervalo médio obtido a partir da temperatura de homogeneização total do tipo t2); 2) fluido aquoso de alta a média temperatura, o qual é subdividido em dois sistemas químicos, com base na relação *Te versus* salinidade: a) fluido do sistema H₂O-NaCl encontrado em cristais de Qz2 e Qz3, com *Te* entre -24,5 e -21°C e salinidade entre 0,87 e 9,72 wt.% NaCl eq., com intervalo de temperatura entre 325 e 380 °C, relacionado à temperatura de homogeneização total de inclusões fluidas do tipo t1-a; b) fluido H₂O-NaCl-CaCl₂ em cristais de Qz3 e Qz4, com *Te* entre -55 e -51,8 °C e salinidade entre 33,02 a 50,25 wt.% NaCl + CaCl₂ eq., com intervalo principal de temperatura de 200–250 °C obtidos a partir das inclusões fluidas do tipo t3 (t3-a e t3-b). A clorita tardia apresentou intervalos de temperatura entre 262 e 280 °C, representando o estágio final do processo hidrotermal no depósito.

Dados isotópicos do fluido hidrotermal relacionados ao processo mineralizador do depósito indicam fonte magmática, com intervalos de $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ entre 3,11 e 7,86 ‰, acusando possível mistura com água meteórica. A principal fase de deposição do minério ocorreu a partir da instabilidade de fluidos magmáticos, com valores de $\delta^{34}\text{S}_{\text{py}}$ entre -1,4 e +0,1 ‰, semelhante a vários depósitos magmático-hidrotermais regionais e globais.

As características gerais do depósito Raimunda sugerem semelhanças genéticas com ambientes magmático-hidrotermais relacionados a magmas graníticos oxidados do tipo Au-pórfito, que têm implicações na exploração de ouro e metais associados na Província Mineral Juruena e na Província Tapajós.

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ANEXOS – ARTIGO 1

Table S1. Electron microprobe analysis table of biotite of Novo Mundo granites.

Samples ID	Monzonite (F6-10)												Monzonite (F17-07)									
	A1	A2	A3	A4	A5	A11	A12	A13	A14	A15	A16	A1b	A2b	A3b	A4b	A5b	A6	A11	A12	A22	A29	
wt.%																						
SiO2	36.65	37.24	36.73	35.4	37.33	36.75	37.03	37.1	36.84	36.33	37.02	37.21	37.42	37.16	37.21	37.34	37.28	37.39	37.09	36.51	36.76	
TiO2	1.61	1.96	1.87	1.87	2.14	1.62	2.04	1.89	2.09	1.82	1.8	1.53	1.38	1.62	1.67	1.5	1.67	1.66	1.75	2.63	1.47	
Al2O3	16.41	15.61	16.18	16.18	15.83	16.22	16.26	16.39	16.3	16.45	16.4	15.09	15.16	14.88	14.61	15.13	15	14.78	14.9	14.8	14.6	
FeO	19.95	20.44	20.12	20.12	19.61	19.88	19.75	19.44	19.44	19.73	19.95	19.91	19.33	19.44	19.33	18.44	20.33	19.6	19.1	19.99	18.8	
MnO	0.61	0.61	0.39	0.39	0.28	0.59	0.44	0.25	0.39	0.39	0.45	0.52	0.55	0.47	0.46	0.45	0.46	0.55	0.5	0.38	0.55	
MgO	10.15	10.16	9.88	9.88	9.94	9.8	10.24	10.12	10.06	10.3	9.99	12.31	12.45	11.62	11.84	12.11	11.9	12	12.1	11.14	11.99	
CaO	0.04	0.04	0.03	0.03	0.03	0.04	0.03	0.02	0.04	0.03	0.05	0.1	0.05	0.02	0.07	0.03	0	0.02	0.04	0.08	0.01	
BaO	0	0	0	0	0.08	0.15	0.07	0.07	0.11	0.07	0.08	0.11	0.12	0.05	0.08	0.08	0.09	0.12	0.06	0.09	0.11	
Na2O	0.07	0.11	0.13	0.08	0.76	0.15	0.07	0.06	0.1	0.07	0.09	0.05	0.09	0.05	0.08	0.07	0.08	0.12	0.06	0.09	0.11	
K2O	9.7	9.7	10.1	10.1	9.92	9.85	9.68	9.88	9.63	9.66	9.79	9.76	9.84	9.98	9.81	9.9	9.74	9.81	9.95	9.71	9.87	
NiO2	0	0	0.05	0.05	0	0	0	0.35	0	0.28	0	0	0.36	0.36	0	0	0.07	0.12	0	0.06	0	
Cr2O3	0	0	0	0	0	0.11	0	0	0.05	0	0.03	0.07	0	0.05	0	0.09	0	0	0.01	0	0.01	
V2O3	0.05	0.05	0.06	0.06	0	0.04	0.05	0.06	0.1	0.06	0.07	0.06	0.06	0.06	0.05	0.06	0	0.03	0.05	0.02	0.11	
F	0.85	0.85	0.8	0.8	0.72	0.95	0.93	0.8	0.99	0.8	0.89	1.1	1.06	0.8	0.94	1.06	0.86	1.07	1.07	0.83	1.03	
Cl	0.06	0.06	0.08	0.08	0.09	0.06	0.04	0.07	0.07	0.03	0.05	0.12	0.12	0.11	0.07	0.08	0.12	0.11	0.11	0.11	0.09	
H2O	3.53	3.55	2.51	2.51	2.52	2.44	2.48	2.51	2.44	2.5	2.49	2.45	2.47	2.52	2.47	2.44	2.52	2.45	2.44	2.49	2.42	
-O= F2	-0.36	-0.36	0.34	0.34	-0.3	-0.44	-0.39	-0.33	-0.41	-0.34	-0.38	-0.46	-0.45	0.33	0.39	0.45	0.36	0.45	0.45	0.35	0.44	
- O = Cl	-0.01	-0.01	0.02	0.02	-0.02	-0.1	-0.01	-0.02	-0.01	-0.01	-0.01	-0.03	-0.03	0.02	0.02	0.02	0.03	0.02	0	0.02	0.02	
Total	99.31	99.89	99.33	99.33	97.9	97.9	98.64	98.66	98.11	98.1	98.7	99.88	99.9	99.44	99.11	99.11	100.4	100.2	99.67	99.33	98.42	

Structural formula based on 22 oxygen atoms

Site T																					
Si	1.22	2.82	2.79	2.79	2.84	2.8	2.79	2.81	2.79	2.77	2.8	2.78	2.79	2.82	2.82	2.82	2.8	2.81	2.79	2.78	2.81
Al	0.64	1.18	1.21	1.21	1.16	1.2	1.21	1.19	1.21	1.23	1.2	1.22	1.21	1.18	1.18	1.18	1.2	1.19	1.21	1.22	1.19
Site M																					
Al	0.21	0.21	0.24	0.24	0.25	0.26	0.24	0.27	0.25	0.25	0.26	0.11	0.13	0.14	0.13	0.16	0.13	0.12	0.12	0.1	0.13
Ti	0.04	0.11	0.11	0.11	0.12	0.09	0.12	0.11	0.12	0.1	0.1	1.24	0.08	0.09	0.1	0.09	0.09	0.09	0.1	0.15	0.08
Fe2+	0.56	1.29	1.28	1.28	1.25	1.26	1.25	0	1.23	1.26	1.26	0	1.2	1.24	1.22	1.16	1.27	1.23	1.21	0	1.2
Mn	0.02	0.04	0.02	0.02	0.02	0.04	0.03	0.02	0.03	0.03	0.03	0.03	0.04	0.03	0.03	0.03	0.03	0.04	0.03	0.02	0.04
Mg	0.5	1.15	1.12	1.12	1.13	1.11	1.15	1.14	1.14	1.17	1.13	1.37	1.39	1.32	1.34	1.36	1.33	1.35	1.37	1.26	1.37
Ni			0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Cr	0	0	0	0	0	0.01	0	0	0	0	0	0	0	0	0	0.01	0	0	0	0	0
V	0	0	0	0	0	0	0	0	0.01	0	0	0	0	0	0	0	0	0	0	0	0.01
Site I																					
Ca	0	0	0	0	0	0	0	0	0	0	0	0.01	0	0	0.01	0	0	0	0	0.01	0
Ba	0	0.01	0	0	0	0	0	0.01	0	0.01	0	0	0.01	0.01	0	0	0	0	0	0	0
Na	0	0	0.02	0.02	0.01	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.02
K	0.41	0.94	0.98	0.98	0.96	0.96	0.93	0.95	0.93	0.94	0.94	0.93	0.94	0.97	0.95	0.95	0.93	0.94	0.96	0.94	0.96
Site A																					
F	0.2	0.2	0.19	0.19	0.17	0.23	0.22	0.19	0.24	0.19	0.21	0.26	0.25	0.19	0.23	0.25	0.21	0.26	0.26	0.2	0.25
Cl	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0	0.01	0.02	0.02	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01
OH	1.79	1.71	1.8	1.8	1.82	1.76	1.77	1.8	1.76	1.8	1.78	1.72	1.73	1.8	1.77	1.74	1.78	1.73	1.73	1.79	1.74

Granodiorite (F64-10)												Granodiorite (DNFD01-54)			
A33	A34	A35	A36	A37	A38	A49	A53	A54	A55	A60		1	2	3	4
37.02	37.55	37.52	37.46	37.46	37.77	37.21	36.73	36.92	37	36.44		37.19	37.44	37.57	37.98

1.8	1.22	1.5	1.2	1.77	1.59	0.92	1.87	1.8	1.49	1.78	1.86	1.81	1.73	1.7
16.4	15.77	15.35	15.77	15.52	15.71	15.41	16.18	15.29	15.03	15.53	14.77	15.15	14.78	14.22
19.95	16.42	15.9	17.2	16.86	16.6	0	20.12	16.84	16.86	17.6	17.44	17.39	17.21	17.36
0.45	0.61	0.59	0.57	0.6	0.42	16.78	0.39	0.48	0.45	0.6	0.66	0.6	0.68	0.38
9.99	13.1	13.77	13.66	12.85	13.44	0.5	9.88	13.03	13.2	11.91	12.74	12.12	11.74	12.99
0.05	0.04	0	0.07	0.01	0.05	13.04	0.03	0.03	0.02	0.07	0.02	0	0.061	0.01
0.08	0.16	0.09	0.16	0.16	0.12	0.08	0.11	0.09	0.12	0.08	0	0	0	0
0.08	0.15	0.08	0.16	0.16	0.12	0.09	0.11	0.09	0.12	0.08	0.08	0.11	0.14	0.04
9.79	9.8	9.7	9.11	9.68	9.71	9.68	10.1	9.82	9.84	9.64	8.56	9.12	8.88	9.14
0	0	0.24	0.42	0	0	0.36	0	0.3	0	0	0.02	0	0.01	0.01
0.03	0.05	0	0.05	0	0.06	0	0	0.05	0.04	0	0.03	0	0.02	0
0.07	0.03	0.04	0.07	0.02	0.03	0.05	0.06	0.03	0.11	0.07	0.00	0.00	0.00	0.00
0.89	1.06	1	0.87	0.91	0.96	0.99	0.8	0.88	1.05	0.94	0.7	0.66	0.8	0.91
0.05	0.03	0.04	0.09	0.06	0.05	0.04	0.08	0.05	0.05	0.05	0.05	0.06	0.1	0.06
2.49	2.46	2.49	2.53	2.51	2.52	3.47	2.5	2.51	2.44	2.45	2.03	3.41	2.71	3.96
0.38	0.45	0.42	0.37	0.38	0.4	0.42	0.34	0.37	0.44	0.39	0.31	0.29	0.36	0.4
0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.033	0.035	0.01
99.65	98.77	98.77	99.5.46	98.85	99.46	98.96		98.66	98.15	97.61	96.29	96.88	97.17	97.01

Structural formula based on 22 oxygen atoms

2.8	2.82	2.82	2.8	2.81	2.81	2.82	2.79	2.79	2.8	2.79	5.72	5.74	5.81	5.81
1.2	1.18	1.19	1.2	1.19	1.19	1.18	1.21	1.21	1.2	1.21	2.28	2.26	2.19	2.19
0.26	0.2	0.17	0.18	0.18	0.19	0.2	0.24	0.15	0.15	0.19	0.4	0.48	0.5	0.38
0.1	0.07	0.09	0.07	0.1	0.09	0.05	0.11	0.1	0.09	0.1	0.22	0.21	0.2	0.19
1.26	1.03	1	1.08	1.06	1.03	1.07	1.28	0	1.07	1.13	2.24	2.22	2.23	2.22
0.03	0.04	0.04	0.04	0.04	0.03	0.03	0.03	0.03	0.03	0.04	0.08	0.09	0.09	0.05
1.13	1.47	1.53	1.51	1.44	1.49	1.48	1.12	1.47	1.49	1.36	2.92	2.77	2.71	2.96

0	0	0	0	0	0	0	0	0	0	0	0.002	0	0.003	0
0	0	0	0	0	0	0	0	0	0	0	0.004	0	0.002	0
0	0	0	0	0	0	0	0	0	0.01	0	0	0	0	0
0	0	0	0.01	0	0	0	0	0	0	0.01	0.01	0	0.01	0
0	0	0.01	0.01	0	0	0.01	0	0.01	0	0		0	0	0
0.01	0.02	0.01	0.02	0.02	0.02	0.01	0.02	0.01	0.02	0.01	0.02	0.03	0.04	0.01
0.94	0.94	0.94	0.87	0.93	0.92	0.94	0.98	0.95	0.95	0.94	1.68	1.78	1.75	1.78
0.14	0.25	0.24	0.21	0.22	0.23	0.24	0.19	0.21	0.25	0.23	0.341	0.32	0.39	0.44
0.01	0	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.01
1.78	1.75	1.76	1.78	1.78	1.77	1.76	1.8	1.78	1.74	1.77	3.64	3.67	3.58	3.54

Monzogranite (LMN-24)					Monzogranite (LMN-24)										Syenogranite (DNFD03-53)		
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	1	2	3
34.24	34.76	34.74	34.75	34.17	34.38	34.87	34.76	34.48	35.22	35.17	35.07	34.27	34.92	34.87	37.58	37.51	37.51
2.68	2.49	2.36	2.24	2.19	2.07	2.02	2.01	1.98	1.88	1.78	1.74	1.73	1.72	1.66	1.7	1.88	1.96
16.3	16.32	16.03	16.4	15.74	16.38	16.99	16.59	15.98	15.93	15.99	16.05	16.34	16.14	16.15	14.56	13.5	13.01
22.28	22.85	23.19	24.01	23.71	22.72	21.32	21.97	23.42	21.56	21.67	21.15	21.63	22.97	23.94	16.66	14.69	15.34
0.34	0.4	0.44	0.37	0.28	0.36	0.19	0.32	0.37	0.4	0.39	0.46	0.44	0.25	0.45	0.74	0.83	0.8
8.37	8.51	8.08	8.6	7.64	8.15	6.79	8.26	7.66	8.81	8.95	8.91	8.28	8.61	8.18	12.01	15	13.96
0.01	0.01	0.02	0.04	0.05	0.02	0.04	0	0.05	0.02	0.03	0.05	0.01	0.045	0	0.04	0.01	0.021
0.14	0.14	0.14	0.12	0.1	0.15	0.097	0.292	0.072	0.179	0.12	0.28	0.13	0.09	0.12	0	0	0
0.07	0.12	0.08	0.11	0.05	0.08	0.07	0.15	0.08	0.06	0.09	0.09	0.09	0.15	0.09	0.1	0.18	0.26
8.81	8.94	8.69	9.08	8.96	8.93	8.54	8.95	8.86	8.66	8.91	9.04	9.08	8.6	9.04	9.14	9.68	9.37
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.01	0

0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.02
0.6	0.45	0.44	0.43	0.48	0.59	0.39	0.53	0.49	0.72	0.59	0.44	0.38	0.61	0.51	0.83	2.86	2.55
0.03	0.03	0.04	0.04	0.04	0.03	0.04	0.03	0.05	0.04	0.04	0.04	0.04	0.03	0.03	0.06	0.05	0.04
0.26	0.2	0.19	0.19	0.21	0.26	0.17	0.23	0.22	0.31	0.26	0.19	0.17	0.27	0.22	0.36	1.21	1.08
0.02	0.01	0.01	0.01	0.01	0.03	0.067	0.01	0.01	0.01	0.01	0.03	0.02	0.01	0.1	0.05	0.01	0.01
93.61	94.84	94.05	95.99	93.2	93.62	91.19	93.63	93.29	93.19	93.49	93.11	92.27	93.89	94.85	95.84	93.64	95.9
5.44	5.46	5.5	5.42	5.5	5.47	5.62	5.51	5.52	5.59	5.57	5.57	5.51	5.53	5.51	5.83	5.75	5.83
2.56	2.54	2.5	2.58	2.5	2.52	2.38	2.49	2.48	2.41	2.43	2.43	2.49	2.47	2.49	2.17	2.25	2.17
0.49	0.48	0.5	0.43	0.48	0.55	0.85	0.61	0.54	0.57	0.55	0.57	0.61	0.54	0.51	0.49	0.19	0.21
0.32	0.29	0.28	0.26	0.26	0.25	0.24	0.24	0.24	0.23	0.21	0.21	0.21	0.21	0.20	0.20	0.22	0.23
2.96	3	3.07	3.13	3.19	3.03	2.87	2.91	3.14	2.86	2.87	2.81	2.91	3.04	3.16	0.00	0.00	0.00
0.04	0.05	0.06	0.05	0.04	0.05	0.03	0.04	0.05	0.05	0.05	0.06	0.06	0.03	0.06	2.16	1.88	1.99
1.98	1.99	1.91	2	1.83	1.93	1.63	1.95	1.83	2.09	2.11	2.11	1.98	2.03	1.93	0.10	0.11	0.11
0	0	0	0	0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
0	0	0	0	0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	2.78	3.43	3.23
0	0	0	0	0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
0.002	0	0	0.01	0.01	0.00	0.01	0.00	0.01	0.00	0.01	0.01	0.00	0.01	0.00	0.01	0.00	0.00
0.01	0.01	0.01	0.007	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.00	0.00	0.00
0.02	0.04	0.02	0.03	0.01	0.03	0.02	0.05	0.03	0.02	0.03	0.03	0.03	0.05	0.03	0.03	0.05	0.08
1.78	1.79	1.75	1.81	1.83	1.81	1.75	1.81	1.81	1.75	1.80	1.83	1.86	1.74	1.82	1.81	1.89	1.86
0.3	0.22	0.22	0.21	0.24	0.30	0.20	0.27	0.25	0.36	0.29	0.22	0.20	0.31	0.26	0.41	0.14	0.12

0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01
3.69	3.77	3.77	3.78	3.74	3.70	3.79	3.73	3.74	3.63	3.70	3.77	3.80	3.69	3.74	3.58	2.60	2.73

Table S2. Electron microprobe analysis table of primary and secondary muscovite results from Novo Mundo granites.

Sample ID	Monzogranite (primary muscovite)																
	LNM-24		LNM-25				LNM-31							LNM-53			
	2	4	1	2	5	6	1	2	3	4	5	6	7	1	2	3	4
SiO2	44.92	44.97	45.26	45.06	44	44.08	43.57	43.55	44.64	44.01	44.73	44.58	44.1	44.78	43.85	44.64	44.48
TiO2	0.791	0.819	1.093	0.778	1.462	1.126	1.384	1.117	0.977	1.266	0.967	0.925	0.784	1.107	1.265	1.224	0.974
Al2O3	29.76	31.29	30.86	31.67	31.42	30.52	30.79	30.2	31.78	31.28	31.88	31.06	31.42	31.76	31.23	30.9	32.04
FeO	4.195	3.943	4.007	3.859	3.245	4.414	3.352	4.715	3.233	3.621	3.494	3.618	3.497	3.313	3.309	4.075	3.362
MnO	0.047	0.055	0.055	0.014	0.025	0	0.017	0.017	0	0.047	0	0.015	0	0.052	0	0.036	0.085
MgO	1.102	0.704	0.794	0.613	0.588	1.172	0.7	1.743	0.725	0.638	0.737	0.687	0.635	0.803	0.66	0.86	0.653
CaO	0	0.001	0.027	0	0.018	0.067	0.068	0.081	0.036	0.07	0.024	0.001	0	0.024	0.033	0.042	0.016
Na2O	0.205	0.378	0.268	0.409	0.426	0.499	0.516	0.464	0.482	0.507	0.531	0.491	0.429	0.571	0.49	0.437	0.483
K2O	9.726	9.95	10.01	9.476	10.12	9.706	9.899	9.424	10.54	9.942	9.957	9.898	10.1	9.819	9.889	9.95	10.34
BaO	0.126	0.145	0.153	0.127	0.183	0	0.157	0.205	0.112	0.063	0.2	0	0.157	0.106	0.086	0	0.231
Rb2O	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
P2O5	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
V2O5	0.152	0.072	0.051	0	0.043	0.042	0.018	0.039	0.044	0.024	0	0.031	0.086	0.049	0.025	0.086	0.088
F	0.036	0.195	0.058	0.066	0.049	0.021	0.092	0	0	0.006	0	0.001	0.037	0.029	0	0.061	0.083
Cl	0.007	0	0.002	0	0.01	0	0.004	0.008	0.004	0	0.014	0.012	0	0.001	0.01	0.007	0
H2O	4.328	4.423	4.325	5.322	6.901	5.781	4.334	7.124	5.615	4.328	4.436	7.008	4.254	6.173	5.251	4.391	4.39

Total	95.39	96.95	96.96	97.39	98.48	95.88	94.9	98.68	98.19	95.8	96.97	98.43	95.76	98.58	96.1	96.71	97.23
Si	4.148	3.273	3.277	3.301	3.407	3.246	3.253	3.083	3.107	3.1	3.109	3.137	3.115	3.112	3.103	3.122	3.095
Ti	0.042	0.043	0.057	0.041	0.077	0.06	0.074	0.059	0.051	0.067	0.051	0.049	0.042	0.058	0.067	0.064	0.051
Al	1.662	1.719	1.683	1.742	1.703	1.644	1.692	1.602	1.714	1.696	1.721	1.712	1.73	1.713	1.708	1.669	1.723
Fe	0.248	0.23	0.233	0.225	0.191	0.26	0.2	0.279	0.188	0.213	0.203	0.213	0.207	0.193	0.196	0.238	0.196
Mn	0.003	0.003	0.003	0.001	0.001	0	0.001	0.001	0	0.003	0	0.001	0	0.003	0	0.002	0.005
Mg	0.116	0.073	0.082	0.064	0.062	0.123	0.074	0.184	0.075	0.067	0.076	0.072	0.067	0.083	0.07	0.09	0.068
Ca	0	0	0.002	0	0.001	0.005	0.005	0.006	0.003	0.005	0.002	0	0	0.002	0.003	0.003	0.001
Na	0.028	0.051	0.036	0.055	0.058	0.068	0.071	0.064	0.065	0.069	0.072	0.067	0.059	0.077	0.067	0.059	0.065
K	0.878	0.887	0.889	0.843	0.909	0.873	0.9	0.851	0.936	0.893	0.883	0.888	0.909	0.871	0.893	0.888	0.918
Ba	0.003	0.004	0.004	0.003	0.005	0	0.004	0.006	0.003	0.002	0.005	0	0.004	0.003	0.002	0	0.006
Rb	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
P	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
V	0.009	0.004	0.003	0	0.002	0.002	0.001	0.002	0.002	0.001	0	0.002	0.005	0.003	0.001	0.005	0.005
F	0.008	0.042	0	0.012	0	0.005	0	0	0	0.001	0	0.002	0	0.035	0	0.013	0
Cl	0.001	0	0.002	0	0.001	0	0.005	0.001	0	0	0.001	0.001	0	0.006	0.005	0.001	0

Monzogranite (secondary muscovite)																	
LNM-25				LNM-24	LNM-25				F75-05	F70-01 (C3)							F70-03(C2)
3	4	5	3	1	3	4	5	3	9	69	70	71	72	73	74	12	14
45.85	45.15	44.61	45.25	45.398	45.85	45.15	44.61	45.25	45.63	47.67	48.33	48.17	47.28	47.44	46.72	44.83	43.03
0.203	0.691	0.625	0.603	0.571	0.203	0.691	0.625	0.603	0.182	0.015	0	0.299	0.193	0.038	0.15	0	0.259
28.94	29.82	30.9	32.34	30.036	28.94	29.82	30.9	32.34	30.41	30.91	31.74	30.26	29.94	30.79	30.84	30.37	29.92
4.361	4.014	4.434	3.356	4.625	4.361	4.014	4.434	3.356	1.544	4.022	3.835	4.198	4.761	4.558	4.266	1.482	1.751
0.045	0	0	0	0.096	0.045	0	0	0	0.053	0.042	0.091	0	0	0.015	0	0.078	0.001
1.472	1.184	0.904	0.607	1.12	1.472	1.184	0.904	0.607	1.882	1.12	0.854	1.182	1.501	0.963	0.915	1.106	1.277

0.062	0.057	0.048	0.01	0	0.062	0.057	0.048	0.01	0.013	0	0	0	0.009	0	0	0.049	0.017
0.192	0.209	0.263	0.281	0.267	0.192	0.209	0.263	0.281	0.088	0.207	0.265	0.169	0.133	0.225	0.201	0.158	0.178
9.664	9.842	9.936	9.902	9.553	9.664	9.842	9.936	9.902	9.162	9.687	9.851	9.581	9.683	9.799	9.963	10.105	10.152
0.104	0.104	0.219	0.122	0.299	0.104	0.104	0.219	0.122	0.21	0.147	0.094	0.048	0.082	0.12	0.029	0.203	0
0	0	0	0	0	0	0	0	0	0.131	0.027	0	0.121	0.169	0.156	0.034	0.043	0.02
0	0	0	0	0	0	0	0	0	0.012	0	0.05	0.021	0.01	0.028	0	0.02	0.009
0	0.018	0.048	0.005	0.157	0	0.018	0.048	0.005	0	0	0	0	0	0	0	0	0
0.042	0.041	0.082	0.083	0.056	0.042	0.041	0.082	0.083	0.275	0.351	0.35	0.211	0.307	0.105	0.177	0.242	0.243
0	0.012	0	0.014	0.016	0	0.012	0	0.014	0	0.001	0	0.014	0	0.007	0	0.028	0.001
4.263	4.301	7.87	5.65	6.773	4.263	4.301	7.87	5.65	7.422	4.255	5.089	4.956	4.517	4.46	5.062	4.391	4.159
95.35	95.44	99.95	98.22	98.97	95.35	95.44	99.95	98.22	97.02	98.46	100.56	99.23	98.59	98.7	98.36	93.12	91.02
3.281	3.278	3.292	3.274	4.148	3.281	3.278	3.292	3.274	3.234	3.189	3.255	3.572	3.252	3.241	3.258	3.23	3.177
0.011	0.037	0.033	0.031	0.03	0.011	0.037	0.033	0.031	0.01	0.01	0	0.015	0.01	0.002	0.001	0	0.014
1.658	1.672	1.695	1.775	1.664	1.658	1.672	1.695	1.775	1.773	1.725	1.774	1.915	1.679	1.72	1.747	1.81	1.78
0.258	0.237	0.261	0.194	0.271	0.258	0.237	0.261	0.194	0	0.277	0	0.18	0	0	0.23	0.089	0.108
0.003	0	0	0	0.006	0.003	0	0	0	0.091	0.004	0.216	0	0.274	0.26	0.002	0.005	0
0.155	0.125	0.095	0.063	0.117	0.155	0.125	0.095	0.063	0.003	0.005	0.074	0	0.001	0.114	0.119	0.141	0.111
0.005	0.004	0.004	0.001	0	0.005	0.004	0.004	0.001	0.199	0.001	0.086	0	0.154	0.098	0	0.004	0.001
0.026	0.029	0.036	0.038	0.036	0.026	0.029	0.036	0.038	0.002	0.038	0	0.017	0.001	0	0.027	0.022	0.025
0.872	0.887	0.891	0.875	0.854	0.872	0.887	0.891	0.875	0.012	0.856	0.035	0.117	0.018	0.03	0.844	0.929	0.956
0.003	0.003	0.006	0.003	0.008	0.003	0.003	0.006	0.003	0.414	0	0.423	0.008	0.425	0.427	0.004	0.006	0
0	0	0	0	0	0	0	0	0	0.012	0.002	0.005	0.003	0.004	0.006	0.001	0.002	0.001
0	0	0	0	0	0	0	0	0	0.006	0.001	0	0	0.007	0.007	0	0.001	0
0	0.001	0.003	0	0.009	0	0.001	0.003	0	0	0	0	0	0	0	0	0	0
0.009	0	0.01	0	0	0.009	0	0.01	0	0	0.039	0.001	0.041	0	0	0.076	0.055	0
0	0.003	0	0.001	0.002	0	0	0	0.001	0.062	0.002	0.075	0	0.067	0.023	0	0.003	0

Syenogranite (secondary muscovite)							
F70-02 (C1)							
44	45	46	47	48	49	50	51
47.03	47.04	45.95	46.71	45.82	45.76	45.69	46.16
0.248	0.254	0.328	0.389	0.106	0.193	0.098	0.202
32.22	30.33	30.75	31.91	30.71	31.9	31.16	31.14
3.321	4.825	5.032	3.613	4.384	4.504	4.789	4.796
0	0	0.01	0.073	0.042	0	0	0.06
0.632	1.098	0.94	0.702	0.893	0.664	0.89	0.845
0.004	0	0	0.03	0.03	0	0.002	0.014
0.283	0.381	0.403	0.264	0.351	0.301	0.395	0.282
9.615	9.448	9.493	9.387	9.436	9.558	9.462	9.714
0.05	0.182	0	0.048	0.032	0.124	0.113	0
0.122	0.205	0.066	0.031	0.163	0	0.004	0.049
0.028	0.028	0.055	0.075	0.055	0	0.077	0.03
0	0	0	0	0	0	0	0
0.329	0.108	0.24	0.199	0.129	0.275	0.206	0.18
0.026	0.03	0.047	0	0.04	0.013	0.014	0.019
4.444	4.489	4.432	4.337	4.392	8.631	4.398	4.446
98.35	98.42	97.53	97.76	96.59	101.93	97.3	97.78
3.237	3.232	3.212	3.232	3.19	3.2	3.203	3.167
0.01	0.013	0.013	0.013	0.017	0.02	0.006	0.01
1.779	1.687	1.806	1.688	1.707	1.777	1.733	1.768
0.09	0	0.19	0.277	0.292	0.207	0.256	0.261
0.003	0.277	0	0	0.001	0.004	0.002	0
0.119	0	0.064	0.112	0.097	0.072	0.093	0.069
0.001	0.112	0	0	0	0.002	0.002	0
0.012	0	0.037	0.051	0.054	0.035	0.048	0.04

0.829	0.051	0.838	0.828	0.841	0.82	0.841	0.844
0.006	0.414	0.001	0.005	0	0.001	0.001	0.003
0.006	0.01	0.005	0.009	0.003	0.001	0.007	0
0	0.009	0.001	0.001	0.002	0.002	0.002	0
0	0	0	0	0	0	0	0
0.062	0.001	0.071	0.023	0.052	0.043	0.028	0.06
0	0.023	0.003	0.003	0.006	0	0.005	0.002

Table S3. Electron microprobe analysis table of titanite from Novo Mundo granites.

wt.%	Monzonite									Biotite granodiorite			
	Ttn A14	Ttn A15	Ttn A16	Ttn A17	Ttn A18	Ttn A19	Ttn A23	Ttn A31	Ttn A32	Ttn A6	Ttn A46	Ttn A47	Ttn A48
SiO ₂	29.99	30.07	30.39	30.61	29.9	30.5	30.11	30.54	30.03	31.17	30.54	31.19	30.05
TiO ₂	32.9	31.59	34.7	32.4	33.35	34.34	32.36	33.18	35.22	27.9	32.7	32.57	31.41
Al ₂ O ₃	1.53	1.91	1.4	1.34	1.91	1.56	1.57	1.6	1.31	3.25	2.02	2.19	1.8
FeO	1.54	1.78	1.23	1.4	1.79	1.23	1.58	1.31	1.24	2.16	1.44	1.58	1.66
MnO	0.1	0.25	0.23	0.19	0.23	0.17	0.22	0.17	0.2	0.1	0.29	0.19	0.18
MgO	0.02	0.03	0.02	0.01	0	0.02	0.01	0.04	0.03	0.63	0	0	0.02
CaO	25.22	25.41	25.97	25.83	25.18	25.95	25.44	25.55	25.01	24.16	25.58	26.22	24.8
Na ₂ O	0.05	0	0.04	0.01	0	0	0	0.01	0.03	0.03	0.04	0.02	0.02
K ₂ O	0	0.03	0.02	0	0	0	0	0.02	0.02	0.72	0.03	0.02	0.03
BaO	0	0.3	0.35	0.49	0	0.36	0	0.55	0	0.11	0	0	0.2
Cr ₂ O ₃	0.08	0	0.1	0.04	0.1	0.08	0	0	0	0.02	0	0.01	0
V ₂ O ₃	0.11	0.19	0.19	0.19	0.1	0.2	0.17	0.12	0.12	0.2	0.25	0.2	0.23
F	0.67	0.99	0.65	0.59	0.84	0.83	0.62	0.81	0.8	1.06	0.87	1.02	0.6
Cl	0	0	0.01	0.02	0	0.03	0.02	0	0.02	0	0.02	0	0
- O ₂ =F	0.28	0.27	0.27	0.25	0.35	0.35	0.26	0.34	0.33	0.45	0.36	0.43	0.25

- O2=Cl	0	0	0	0	0	0.01	0	0	0	0	0.01	0	0
H2O	7.29	7.56	7.56	7.46	7.07	7.22	7.34	7.14	7.2	6.55	7.05	6.91	7.3
Total	99.21	99.84	102.57	100.32	100.12	102.14	99.19	100.72	93.69	98.51	93.41	94.76	90.77
Structural formula based on 5 oxygen atoms													
Si	1.183	1.186	1.198	1.207	1.179	1.203	1.188	1.204	1.184	1.229	1.204	1.23	1.185
Ti	0.976	0.937	1.03	0.961	0.989	1.019	0.96	0.984	1.045	0.828	0.97	0.966	0.932
Al	0.071	0.133	0.065	0.062	0.089	0.073	0.073	0.074	0.061	0.151	0.094	0.102	0.084
Fe3+	0.046	0.013	0.036	0.042	0.053	0.037	0.047	0.039	0.037	0.064	0.043	0.047	0.049
Mn	0.003	0.008	0.008	0.006	0.008	0.006	0.007	0.006	0.007	0.003	0.01	0.006	0.006
Mg	0.001	0.002	0.001	0.001	0	0.001	0.001	0.002	0.002	0.037	0	0	0.001
Ca	1.065	0.537	1.097	1.091	1.064	1.097	1.075	1.08	1.057	1.021	1.081	1.108	1.048
Na	0.003	0	0.003	0	0	0	0	0.001	0.001	0.002	0.003	0.001	0.001
K	0	0.001	0.001	0	0	0	0	0.001	0.001	0.036	0.001	0.001	0.002
Ba	0	0.005	0.005	0.008	0	0.006	0	0.009	0	0.002	0	0	0.003
Cr	0.003	0	0.003	0.001	0.003	0.003	0	0	0	0.001	0	0	0
V	0.004	0.009	0.006	0.006	0.003	0.006	0.005	0.004	0.004	0.006	0.008	0.006	0.007
F	0.083	0.123	0.08	0.074	0.105	0.103	0.078	0.101	0.099	0.132	0.108	0.128	0.075
Cl	0	0	0.001	0.001	0	0.002	0.001	0	0.001	0	0.002	0	0
Fe/Al	0.642	0.095	0.562	0.67	0.601	0.504	0.643	0.523	0.606	0.424	0.453	0.46	0.588

Table S4. Electron microprobe analysis table of amphibole from Novo Mundo granites.

wt.%	Monzonite						Biotite granodiorite								
	A8	A9	A10	A20	A21	A30	A43	A44	A50	A51	A52	A61	A62	A63	A64
SiO2	43.38	43.78	43.62	42.9	42.74	42.7	42.77	43.26	42.89	44.72	42.47	43.15	44.73	44.9	43.83
TiO2	0.99	0.64	0.89	1.03	0.97	1.19	0.76	1.07	0.64	1.53	0.46	0.56	0.94	0.64	0.9
Al2O3	9.84	9.08	9.45	10.12	9.89	10.24	10.64	10.04	10.33	9.22	10.88	10.2	9.25	9	9.73

FeO*	19.64	18.54	18.85	19.78	19.25	19.62	18.14	18.01	18.13	16.25	18.08	18.33	16.85	16.85	17.41
MnO	0.68	0.85	0.79	0.65	0.89	0.72	0.89	0.53	0.56	0.76	0.88	0.9	0.98	0.77	0.94
MgO	9.13	9.73	9.58	9.5	9.36	9.6	9.88	10.05	9.68	11	9.83	9.74	11.17	10.75	10.52
CaO	10.59	10.44	10.68	10.78	10.36	10.66	10.72	10.92	10.76	10.62	10.66	10.88	10.56	10.99	11.04
Na ₂ O	1.41	1.4	1.45	1.42	1.46	1.42	1.16	1.21	1.12	1.31	1.28	1.01	1.23	1	0.97
BaO	0.68	0.9	1.32	1.26	1.32	1.28	1.34	1.25	1.33	1.18	1.29	1.14	1.14	1.03	1.11
K ₂ O	1.41	1.32	0	0.02	0.01	0	0.05	0.02	0	0	0.04	0	0.02	0.04	0
NiO	0	0.03	0.02	0	0.24	0	0.17	0	0	0	0	0	0.67	0	0
Cr ₂ O ₃	0	0.3	0	0	0	0.06	0	0.01	0	0	0	0.05	0.03	0.06	0.04
P ₂ O ₅	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
V ₂ O ₃	0.06	0.02	0.06	0.02	0.02	0.02	0.03	0.08	0.07	0.04	0.01	0.03	0.07	0.03	0.08
F	0.49	0.08	0.48	0.51	0.55	0.33	0.46	0.4	0.37	0.49	0.48	0.33	0.4	0.31	0.45
Cl	0.1	0.56	0.06	0.1	0.1	0.12	0.08	0.09	0.08	0.09	0.13	0.1	0.04	0.07	0.08
Cor F	0.21	0.23	0.2	0.21	0.23	0.14	0.2	0.17	0.16	0.21	0.2	0.14	0.17	0.13	0.19
Cor Cl	0.02	0.02	0.01	0.02	0.02	0.03	0.02	0.02	0.02	0.02	0.03	0.02	0.01	0.02	0.02
Total	97.72	96.77	97.24	98.06	97.14	97.95	97.1	96.94	95.97	97.2	96.47	96.4	98.08	96.42	97.11
Structural formule															
Site T															
Si	6.6	6.71	6.64	6.48	6.54	6.44	6.48	6.56	6.57	6.72	6.44	6.54	6.69	6.78	6.6
Al	1.4	1.29	1.36	1.52	1.46	1.56	1.52	1.44	1.43	1.28	1.56	1.46	1.31	1.22	1.4
Site C															
Al	0.36	0.35	0.34	0.28	0.32	0.27	0.39	0.36	0.43	0.36	0.39	0.37	0.32	0.38	0.33
Ti	0.11	0.07	0.1	0.12	0.11	0.13	0.09	0.12	0.07	0.17	0.05	0.06	0.11	0.07	0.1
Cr	0	0	0	0	0	0.01	0	0	0	0	0	0.01	0	0.01	0
V	0.01	0.01	0.01	0	0	0	0	0.01	0.01	0	0	0	0.01	0	0.01
Fe ³⁺	0.34	0.29	0.13	0.33	0.22	0.43	0.34	0.22	0.29	0	0.56	0.55	0.21	0.22	0.32
P	0	0													
Mn	0.09	0.11	0.1	0.08	0.11	0.09	0.11	0.07	0.07	0.1	0.11	0.12	0.12	0.1	0.12

Ni	0	0	0	0	0	0	0.01	0	0	0	0	0	0	0	0
Ba	0	0.02	0	0	0.01	0	0.01	0	0	0	0	0	0.04	0	0
Fe2+	2.15	2.09	2.27	2.17	2.24	2.04	1.96	2.06	2.04	2.05	1.74	1.78	1.9	1.9	1.87
Mg	1.93	2.05	2.05	2.01	1.97	2.02	2.09	2.15	2.09	2.32	2.15	2.12	2.29	2.3	2.24
Site B															
Mg	0.14	0.17	0.13	0.12	0.16	0.14	0.14	0.12	0.12	0.14	0.08	0.08	0.2	0.12	0.12
Ca	1.73	1.71	1.74	1.75	1.7	1.72	1.74	1.77	1.77	1.71	1.73	1.77	1.69	1.78	1.78
Na	0.14	0.12	0.13	0.13	0.14	0.14	0.12	0.11	0.12	0.15	0.19	0.15	0.11	0.1	0.1
Site A															
Na	0.28	0.3	0.3	0.28	0.29	0.28	0.22	0.25	0.22	0.23	0.18	0.15	0.25	0.19	0.18
K	0.01	0.01	0.01	0	0	0	0	0.01	0.01	0	0	0	0.01	0	0.01
Site W															
F	0.24	0.27	0.23	0.25	0.26	0.16	0.23	0.23	0.18	0.24	0.23	0.16	0.2	0.15	0.22
Cl	0.03	0.03	0.02	0.03	0.03	0.03	0.02	0.02	0.02	0.02	0.03	0.03	0.01	0.02	0.02
OH	1.74	1.71	1.75	1.72	1.71	1.81	1.75	1.75	1.8	1.74	1.74	1.81	1.79	1.83	1.76

Table S5. Electron microprobe analysis table of amphibole from Novo Mundo mafic dikes.

wt. %	Andesitic basalt														Basalt			
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
SiO ₂	42.82	40.74	40.4	41.3	41.45	42.32	41.94	41.05	41.03	45.47	49.3	42.28	50.75	47.49	36.08	36.23	37.25	37.67
TiO ₂	1.07	1.5	1.65	1.42	1.83	1.53	1.99	1.95	1.92	0.61	0.9	2.75	0.17	1.35	0.67	0.89	0.86	1.29
Al ₂ O ₃	11.86	12.93	13.21	13.08	13.36	11.94	12	12.88	13.64	9.59	5.92	12.5	5.95	8.12	8.26	15.48	16.09	16
FeO*	13.21	12.47	12.27	12.34	10.8	12.96	10.04	11.55	10.03	12.99	11.96	9.83	11.12	11.28	20.31	15.66	15.94	15.77

MnO	0.25	0.09	0.14	0.2	0.09	0.2	0.17	0.11	0.21	0.22	0.19	0	0.37	0.07	0.41	0.22	0.23	0.24
MgO	14.41	13.81	15.41	14.21	15.32	14.01	15.57	13.88	15.01	14.64	15.31	15.31	15.85	15.46	7.5	16.3	15.82	15.12
CaO	11.29	11.39	10.72	11.65	11.65	11.12	11.73	11.44	11.78	9.64	11.25	11.13	11.77	11.76	10.9	0	0	0.01
Na2O	2.2	2.41	1.63	2.28	1.83	2.32	2.35	2.33	2.29	0.72	0.7	2.3	0.59	1.29	0.81	0.04	0.09	0.14
K2O	0.65	0.69	0.58	0.77	0.68	0.62	0.78	0.72	0.81	0.32	0.32	0.66	0.2	0.4	0.98	8.39	8.36	9.46
NiO	0	0	0.02	0.03	0.05	0.01	0.01	0	0.16	0.01	0.06	0.06	0.06	0.08	0	0.14	0.07	0.12
Cr2O3	0.06	0.07	0	0	0	0.12	0	0	0.13	0.16	0	0.06	0.07	0	0	0.83	0.59	0.89
P2O5	0.06	0.08	0.11	0.06	0	0.08	0.06	0.06	0.04	0.04	0.03	0.07	0.02	0	0.06	0	0	0
F	0.06	0	0.19	0.34	0.33	0	0.24	0.3	0.1	0.24	0.1	0.06	0.01	0	0.17	1.3	0.99	1.62
Cl	0.01	0.01	0.02	0	0.01	0	0	0.02	0.02	0	0	0	0.08	0.02	0.02	0.01	0.02	0.03
H2O	2	1.99	1.91	1.87	1.88	2	1.92	1.86	1.98	1.89	2	2.02	2.05	2.06	1.62	2	1.47	1.2
O2 = F	0.03	0	0.08	0.14	0.14	0	0.1	0.13	0.04	0.1	0.04	0.02	0	0	0.07	0.03	0.42	0.68
O2 = Cl	0	0	0	0	0	0	0	0	0	0	0	0	0.02	0	0	0	0.01	0.01
Total	99.96	98.17	98.33	99.67	99.43	99.22	98.89	98.28	99.17	96.64	98.08	99.04	99.08	99.39	87.86	97.51	98.21	100.23
Site T																		
Si	6.21	6.04	5.87	6.04	5.99	6.2	6.14	6.09	5.98	6.67	7.19	6.15	7.28	6.86	6.26	6.21	5.52	5.6
Al	1.79	1.96	2.13	1.96	2.01	1.8	1.86	1.91	2.02	1.34	0.81	1.85	0.72	1.15	1.74	1.79	2.48	2.4
Site C																		
Al	0.24	0.3	0.13	0.29	0.27	0.26	0.21	0.35	0.33	0.32	0.2	0.29	0.29	0.24	0	0.24	0.33	0.4
Ti	0.12	0.17	0.18	0.16	0.2	0.17	0.22	0.22	0.21	0.07	0.1	0.3	0.02	0.15	0.09	0.12	0.1	0.15
Cr	0.01	0.01	0	0	0	0.01	0	0	0.01	0.02	0	0	0	0	0	0.01	0.07	0.11
Fe 3+	0.81	0.67	1.39	0.73	0.9	0.66	0.56	0.48	0.58	1.1	0.4	0.43	0.33	0.35	1.07	0.81	2.26	1.74
P	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Mn	0.03	0.01	0.02	0.03	0.01	0.02	0.02	0.01	0.03	0.03	0.02	0	0.05	0.01	0.06	0.03	0.03	0.03
Ni	0	0	0	0	0.01	0	0	0	0.02	0	0.01	0.01	0.01	0.01	0	0	0.01	0.01
Fe 2+	0.8	0.88	0.1	0.78	0.41	0.93	0.67	0.95	0.64	0.5	1.06	0.77	1	1.01	1.88	0.8	0.28	0.22
Mg	3	2.97	3.18	3.02	3.21	2.94	3.32	2.99	3.18	2.97	3.21	3.2	3.3	3.24	1.96	3	1.92	2.34

Site B																		
Mg	0.12	0.09	0.15	0.08	0.09	0.12	0.07	0.08	0.08	0.23	0.12	0.13	0.09	0.09	0	0.12	1.57	1.01
Ca	1.76	1.81	1.67	1.82	1.81	1.75	1.84	1.82	1.84	1.51	1.76	1.74	1.81	1.82	2	1.76	0	0
Na	0.13	0.11	0.18	0.09	0.1	0.14	0.09		0.08	0.26	0.13	0.14	0.1	0.09	0	0.13	0.43	0.99
Site A																		
Na	0.49	0.59	0.28	0.55	0.41	0.52	0.58	0.57	0.57	0	0.07	0.51	0.07	0.27	0.28	0.49	0	0
K	1.12	0.13	0.11	0.14	0.13	0.12	0.14	0.14	0.15	0.06	0.06	0.12	0.04	0.07	0.22	0.12	1.58	1.79
Site W																		
F	0.03	0	0.09	0.16	0.16	0	0.11	0.14	0.05	0.11	0.05	0.03	0	0	0.07	0.03	0.51	0.8
Cl	0	0	0.01	0	0	0	0	0	0.01	0	0	0	0.02	0.01	0	0	0.01	0.01
OH	1.97	2	1.91	1.84	1.84	2	1.89	1.86	1.95	1.88	1.95	1.97	1.98	2	1.93	1.97	1.49	1.2

Table S6. Lithogeochemistry data for Novo Mundo mafic rocks.

Wt. (%)	Basalt andesite/ andesite					Basalt	
	F06-12	F17-02	F48-10	F65-06	F70-07	F06-03	DNFD09
SiO ₂	51.8	57.7	49.8	49.6	48.9	46.1	47.4
TiO ₂	0.83	0.49	0.94	0.82	0.81	0.9	2.73
Al ₂ O ₃	17.55	13.25	15.15	15.85	15.75	12.1	14.3
Fe ₂ O ₃	10.6	6.72	8.06	10.8	10	10.75	15.15
MnO	0.19	0.16	0.16	0.24	0.2	0.3	0.21
MgO	4.85	6.71	8.51	4.94	5	12.55	5.3
CaO	7.23	4.5	7.99	5.38	5.96	5.92	7.4
Na ₂ O	2.7	2.11	3.54	2.81	2.88	0.78	2.78
K ₂ O	1.9	2.47	1.82	2.04	2.25	0.81	1.76

P ₂ O ₅	0.35	0.18	0.37	0.2	0.18	0.21	0.64
SrO	0.1	0.03	0.09	0.04	0.03	0.02	0.04
BaO	0.08	0.1	0.07	0.04	0.03	0.02	0.08
LOI	3.67	6.78	3.44	6.79	8.21	10.15	0.85
Total	101.85	101.28	100.01	99.55	100.2	100.75	98.65
ppm							
Ba	771	1010	720	420	310	214	744
Sr	777	271	731	313	249	140.5	319
Rb	97.3	91.8	93.7	166.5	208	53.7	50.7
Zr	96	171	146	130	129	106	214
Y	15.3	14.6	15.1	20.6	21.3	14.5	41.5
Nb	3.3	9.3	6.7	5	4.8	8.2	10.4
Ga	21.2	16.5	17.7	18.1	18.3	22.3	24.9
Th	2.73	8.23	4.74	3.89	3.98	3.3	3.75
Cs	0.73	2.11	0.94	5.73	4.91	2.99	1.52
Hf	2.8	4.7	3.9	3.6	4.3	2.7	5.3
Ta	0.3	0.8	0.5	0.4	0.7	0.6	0.8
U	0.69	2.78	1.31	1.04	1.24	0.86	0.95
V	249	112	155	182	179	228	268
C	0.15	0.8	0.29	0.85	1.21	1.17	0.07
S	0.06	0.01	0.08	0.02	0.03	0.02	0.11
Ge	5	5	5	5	5	5	5
Sn	2	2	2	2	2	1	2
In	0.012	0.026	0.02	0.03	0.025	0.07	0.034
Re	0.001	0.002	0.001	0.003	0.002	0.001	0.002
Sb	0.12	0.05	0.19	0.08	0.09	0.05	0.05
Se	0.2	0.2	0.2	0.2	0.2	0.2	0.5
Te	0.06	0.01	0.04	0.01	0.01	0.05	0.01

Tl	0.06	0.11	0.06	0.72	0.4	0.28	0.16
Ag	0.5	0.5	0.5	0.5	0.5	0.5	0.5
As	0.7	0.3	4.2	0.5	1.3	1.2	0.6
Bi	0.07	0.05	0.13	0.07	0.07	0.12	0.04
Cd	0.5	0.5	0.5	0.5	0.5	0.5	0.5
Cu	82	6	3	112	49	3	53
Hg	0.005	0.011	0.005	0.005	0.005	0.005	0.005
Mo	1	1	1	1	1	1	1
Ni	22	130	194	98	90	288	85
Pb	17	12	11	8	6	9	15
Li	50	70	30	60	50	130	10
Sc	23	15	20	24	23	27	27
Zn	156	177	84	132	114	151	162
La	21.5	31.7	32.2	19.9	20.3	23.9	32
Ce	44.9	67.3	69.5	41.6	41	47.3	69
Pr	5.22	7.39	8.27	4.76	4.94	5.46	8.56
Nd	21.8	28	34.1	20.3	20.7	23.1	36.7
Sm	4.38	4.68	6.24	3.91	5.16	4.12	7.95
Eu	1.29	1.03	1.61	1.08	1.52	0.96	2.46
Gd	4.25	4.24	4.83	4.23	5.09	3.28	8.3
Tb	0.64	0.55	0.61	0.74	0.94	0.56	1.28
Dy	2.97	2.69	3.4	3.98	4.87	3.08	7.66
Ho	0.62	0.6	0.61	0.82	0.95	0.53	1.49
Er	1.78	1.64	1.39	2.54	3.69	1.55	4.15
Tm	0.24	0.21	0.19	0.33	0.58	0.2	0.6
Yb	1.3	1.61	1.22	2.34	3.05	1.3	4.03
Lu	0.19	0.21	0.13	0.32	0.51	0.16	0.53

Table S7. LA-ICP-MS U-Pb analytical data for zircon crystals from Novo Mundo granites.

Monzonite (F06-10)														
Concentration			Radiogenic (correct) ratios					Calculated age $\pm \sigma$ (Ma)						
Sample ID	U	Th/U	$^{206}\text{Pb}/^{238}\text{U}$	2 σ (%)	$^{207}\text{Pb}/^{206}\text{Pb}$	1 σ %	Rho	$^{207}\text{Pb}/^{206}\text{Pb}$	2 σ abs	$^{207}\text{Pb}/^{235}\text{U}$	2 σ abs	$^{206}\text{Pb}/^{238}\text{U}$	2 σ abs	Conc (%)
SMPABC010	323.59	0.283	0.371	2.42	0.126	0.986	0.93	2041	18	2037	25	2033	42	99.8
SMPABC033	71.88	1.121	0.371	2.1	0.126	1.559	0.8	2037	28	2037	25	2036	37	100
SMPABC034	117.62	1.09	0.372	2.04	0.126	0.958	0.9	2037	17	2037	22	2038	36	100
SMPABC052	125.28	0.651	0.371	2.17	0.126	1.103	0.89	2036	20	2036	23	2036	38	100
SMPABC059	160.79	0.574	0.365	2.12	0.123	1.086	0.89	2004	19	2004	23	2004	37	100
SMPABC069	200.08	0.981	0.371	2.22	0.126	1.236	0.87	2038	22	2036	24	2034	39	99.9
SMPABC072	166.71	0.975	0.372	1.94	0.125	0.844	0.92	2031	15	2035	21	2038	34	100.2
SMPABC073	226.02	0.82	0.371	1.91	0.126	0.842	0.92	2036	15	2036	20	2036	34	100

Monzogranite (LNM-25)																
Concentration				Radiogenic (correct) ratios						Calculated age $\pm \sigma$ (Ma)						
Sample ID	^{204}Pb cps	Th/U	$^{206}\text{Pb}/^{204}\text{Pb}$	1 σ %	$^{206}\text{Pb}/^{238}\text{U}$	1 σ %	$^{207}\text{Pb}/^{206}\text{Pb}$	1 σ %	Rho	$^{207}\text{Pb}/^{206}\text{Pb}$	2 σ abs	$^{207}\text{Pb}/^{235}\text{U}$	2 σ abs	$^{206}\text{Pb}/^{238}\text{U}$	2 σ abs	Conc (%)
045-ZR30	15	0.257	121666	20.08	0.3782	2.15	0.12426	0.58	0.95	2018	21	2068	76	2043	39	101.2
044-ZR29	13	0.151	315549	30.2	0.389	1.35	0.1251	0.37	0.93	2030	13	2118	48	2074	25	102.2
043-ZR28	30	0.215	92113	31.23	0.3921	1.02	0.12491	0.4	0.88	2028	14	2133	37	2080	20	102.6
040-ZR27	74	0.284	8802	24.98	0.3903	0.92	0.12476	0.52	0.82	2025	18	2124	33	2074	20	102.4
039-ZR26	15	0.206	175203	18.61	0.3891	1.04	0.12488	0.47	0.87	2027	16	2119	38	2073	21	102.3
038-ZR25	19	0.086	120028	22.2	0.3896	0.91	0.12513	0.42	0.85	2031	15	2121	33	2076	19	102.2

037-ZR24	10	0.21	141727	15.9	0.3868	0.79	0.12464	0.36	0.84	2024	13	2108	28	2066	17	102.1
036-ZR23	102	0.075	8421	18.77	0.3771	1.7	0.12388	0.45	0.95	2013	16	2063	60	2038	31	101.2
035-ZR22	21	0.12	296707	58.72	0.3893	0.97	0.12467	0.49	0.84	2024	17	2120	35	2072	20	102.4
034-ZR21	26	0.108	74927	29.16	0.3856	1.02	0.12518	0.44	0.87	2031	15	2102	37	2067	21	101.8
030-ZR19	15	0.082	148126	21.4	0.3801	1.15	0.12587	0.56	0.86	2041	20	2077	41	2059	23	100.9
029-ZR18	15	0.108	192255	46.09	0.3801	1.04	0.12512	0.53	0.85	2030	19	2077	37	2054	22	101.2
028-ZR17	14	0.124	133023	17.43	0.3879	1.36	0.12464	0.46	0.92	2024	16	2113	49	2068	26	102.2
027-ZR16	17	0.096	165083	23.12	0.3814	1.01	0.12553	0.38	0.89	2036	13	2083	36	2059	20	101.1
026-ZR15	47	0.173	11714	15.98	0.3888	1.13	0.1254	0.42	0.9	2034	15	2117	41	2076	22	102.1
024-ZR14N	25	0.194	78437	19.98	0.2957	1.83	0.12611	0.44	0.95	2044	15	1670	54	1843	32	90.2
023-ZR13	24	0.087	132770	49.68	0.3797	1.61	0.12506	0.51	0.93	2030	18	2075	57	2052	30	101.1
020-ZR12	10	0.31	152987	16.48	0.4082	1.58	0.12542	0.53	0.93	2035	19	2207	59	2119	30	104.1
019-ZR11B	41	0.006	88475	24.37	0.3264	2	0.12198	1.79	0.74	1985	63	1821	63	1899	46	95.7
017-ZR10	10	0.079	227829	16.95	0.3886	1.75	0.12552	0.39	0.96	2036	14	2116	63	2076	32	102
013-ZR07N	14	0.159	164432	24.98	0.3805	2.5	0.1258	0.5	0.97	2040	18	2079	88	2059	45	100.9
010-ZR06	15	0.164	90595	21.6	0.3928	2.06	0.12461	0.57	0.95	2023	20	2136	75	2079	38	102.8
008-ZR04	16	0.21	95519	22.85	0.3967	1.93	0.12569	0.37	0.97	2039	13	2154	71	2096	35	102.8
006-ZR03N	11	0.361	113596	19.84	0.3892	2.17	0.1247	0.4	0.97	2025	14	2119	78	2072	39	102.3
004-ZR02N	11	0.473	208423	14.15	0.3983	1.12	0.1249	0.38	0.9	2027	14	2161	41	2093	22	103.3
025-ZR14B	10	0.012	362604	20.9	0.3446	1.74	0.12793	0.52	0.94	2070	18	1987	32	1909	57	92.2
014-ZR07B	14	0.009	346175	17.95	0.3686	2.23	0.13077	0.57	0.96	2108	20	2066	41	2023	77	96
009-ZR05	38	0.186	37245	33.77	0.3991	2.09	0.12979	0.49	0.96	2095	17	2129	38	2165	77	103.3
007-ZR03B	38	0.015	124041	21.82	0.3112	4.58	0.13398	1.01	0.97	2151	35	1939	80	1747	139	81.2

Syenogranite (LNM-46)

Concentration

Radiogenic (correct) ratios

Calculated age $\pm \sigma$ (Ma)

Sample ID	²⁰⁴ Pb cps	Th/U	²⁰⁶ Pb/ ²⁰⁴ Pb	1σ %	²⁰⁶ Pb/ ²³⁸ U	1σ %	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ %	Rho	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ abs	²⁰⁷ Pb/ ²³⁵ U	2σ abs	²⁰⁶ Pb/ ²³⁸ U	2σ abs	Conc (%)
055-ZR29	15	0.8	174418	17.67	0.3728	0.71	0.12313	0.58	0.72	2002	20	2023	17	2043	25	101
054-ZR28	17	0.639	313799	15.9	0.3626	0.93	0.12325	0.44	0.85	2004	16	1999	19	1995	32	99.8
053-ZR27	16	0.65	222009	15.05	0.3799	0.62	0.12223	0.43	0.74	1989	15	2033	15	2076	22	102.1
051-ZR25	22	0.661	161329	14.18	0.3786	0.76	0.12272	0.72	0.68	1996	26	2033	19	2070	27	101.8
048-ZR24	38	0.706	144908	20.1	0.3877	1.26	0.1211	0.62	0.87	1973	22	2042	25	2112	45	103.4
047-ZR23B2	184	0.558	126314	34.15	0.2913	0.82	0.12214	0.4	0.83	1988	14	1803	17	1648	24	91.4
045-ZR23N	13	1.074	185398	13.37	0.3682	0.72	0.12273	0.64	0.7	1996	23	2009	18	2021	25	100.6
042-ZR22	13	0.553	560035	15.71	0.3876	0.58	0.12211	0.57	0.65	1987	20	2049	16	2112	21	103.1
040-ZR20	66	0.607	18723	62.2	0.3033	1.38	0.12384	0.48	0.92	2012	17	1849	26	1708	41	92.4
036-ZR18	96	0.258	14898	15.02	0.3371	0.81	0.12196	0.56	0.77	1985	20	1927	18	1873	26	97.2
035-ZR17	71	0.748	16572	30.77	0.3808	0.89	0.12369	0.7	0.75	2010	25	2045	21	2080	32	101.7
033-ZR16N	83	1.202	41741	51.08	0.3828	0.93	0.12297	0.45	0.85	2000	16	2045	19	2090	33	102.2
029-ZR15N	7	0.527	254505	12.71	0.3771	0.87	0.12295	0.44	0.83	1999	16	2031	18	2063	31	101.6
028-ZR14	47	0.424	150987	31.12	0.3426	0.83	0.12374	0.52	0.79	2011	18	1953	18	1899	27	97.2
027-ZR13	9	0.664	429581	22.84	0.3832	0.79	0.12261	0.5	0.79	1995	18	2043	18	2091	28	102.3
024-ZR12B	120	0.493	162684	33.13	0.278	1.58	0.12318	0.82	0.87	2003	29	1771	30	1581	44	89.3
022-ZR11	12	0.529	549171	13.46	0.3947	1.17	0.12024	1.32	0.65	1960	47	2052	31	2145	43	104.5
021-ZR10	15	0.636	1146416	17.1	0.401	1.66	0.11784	1.66	0.7	1924	59	2048	41	2173	61	106.1
018-ZR9	17	0.615	409627	39.42	0.3782	1.15	0.1206	1.87	0.52	1965	66	2017	39	2068	41	102.5
017-ZR8	20	0.882	174686	20.71	0.3767	0.89	0.12116	1.03	0.63	1973	37	2018	25	2061	31	102.1
016-ZR7B	15	0.5	865782	19.35	0.382	0.75	0.11936	0.63	0.71	1947	23	2017	18	2086	27	103.4
015-ZR7N	28	0.455	217581	18.97	0.3648	0.65	0.12152	0.61	0.67	1979	22	1992	17	2005	22	100.7
012-ZR6	287	0.532	5697	8.06	0.3602	0.83	0.12306	0.51	0.8	2001	18	1992	18	1983	28	99.5
011-ZR5	38	0.401	56490	20.48	0.3827	0.93	0.12183	0.4	0.86	1983	14	2036	19	2089	33	102.6
006-ZR3N	15	0.931	258061	13.58	0.3687	0.54	0.12206	0.44	0.69	1987	16	2005	14	2023	19	100.9
005-ZR2B	16	0.482	558264	38.64	0.3753	1.02	0.12179	0.37	0.89	1983	13	2019	20	2054	36	101.7

004-ZR2N	10	0.768	186641	14.15	0.3663	0.6	0.12348	0.45	0.71	2007	16	2010	15	2012	21	100.1
003-ZR1	86	0.348	213270	38.84	0.3842	1.71	0.12136	0.48	0.94	1976	17	2036	32	2096	61	102.9
052-ZR26	107	0.429	198803	24.62	0.3902	0.98	0.11743	1.31	0.58	1918	47	2021	29	2124	35	105.1
009-ZR3B	99	0.212	172703	54.57	0.3588	0.96	0.12036	0.42	0.86	1962	15	1969	19	1976	33	100.4
046-ZR23B1	195	0.468	3403	7.47	0.3121	1.56	0.12699	0.48	0.93	2057	17	1895	29	1751	48	92.4
041-ZR21	39	1.098	99340	28.26	0.3651	2.09	0.12643	1.11	0.87	2049	39	2027	42	2006	72	99
034-ZR16B	18	0.512	401533	17.63	0.3962	0.64	0.12488	0.74	0.61	2027	26	2089	18	2151	23	103
030-ZR15B	100	0.458	10241	40.67	0.3724	1.74	0.12958	1.95	0.66	2092	68	2067	46	2041	61	98.8

Granodiorite (F64-10)

Sample ID	Concentration		Radiogenic (correct) ratios					Calculated age ± σ (Ma)						
	U	Th/U	²⁰⁶ Pb/ ²³⁸ U	2σ (%0	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ %	Rho	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ abs	²⁰⁷ Pb/ ²³⁵ U	2σ abs	²⁰⁶ Pb/ ²³⁸ U	2σ abs	Conc (%)
SMPABC091	470.62	0.207	0.362	2.298	0.123	1.211	0.88	1998	22	1995	25	1991	39	99.8
SMPABC092	202.24	0.639	0.364	2.156	0.122	0.931	0.92	1983	17	1991	22	1999	37	100.4
SMPABC093	435.03	0.616	0.363	2.369	0.122	1.389	0.86	1992	25	1993	26	1994	41	100.1
SMPABC094	236.85	1.032	0.362	2.287	0.123	1.188	0.89	1993	21	1993	25	1993	39	100
SMPABC095	315.39	0.917	0.361	2.257	0.122	1.152	0.89	1989	21	1989	24	1989	39	100
SMPABC096	256.17	0.675	0.363	2.053	0.123	0.835	0.93	1996	15	1996	21	1996	35	100
SMPABC097	304.75	0.71	0.364	2.119	0.122	0.811	0.93	1978	15	1991	22	2002	37	100.6
SMPABC098	250.41	0.582	0.365	1.971	0.121	0.824	0.92	1976	15	1990	21	2004	34	100.7
SMPABC099	462.39	0.393	0.362	2.308	0.122	1.592	0.82	1992	29	1992	27	1991	40	100
SMPABC107	243.65	0.61	0.362	2.964	0.122	1.483	0.89	1992	27	1992	31	1992	51	100
SMPABC109	655.03	0.122	0.362	3.061	0.122	2.436	0.78	1991	44	1991	36	1991	53	100

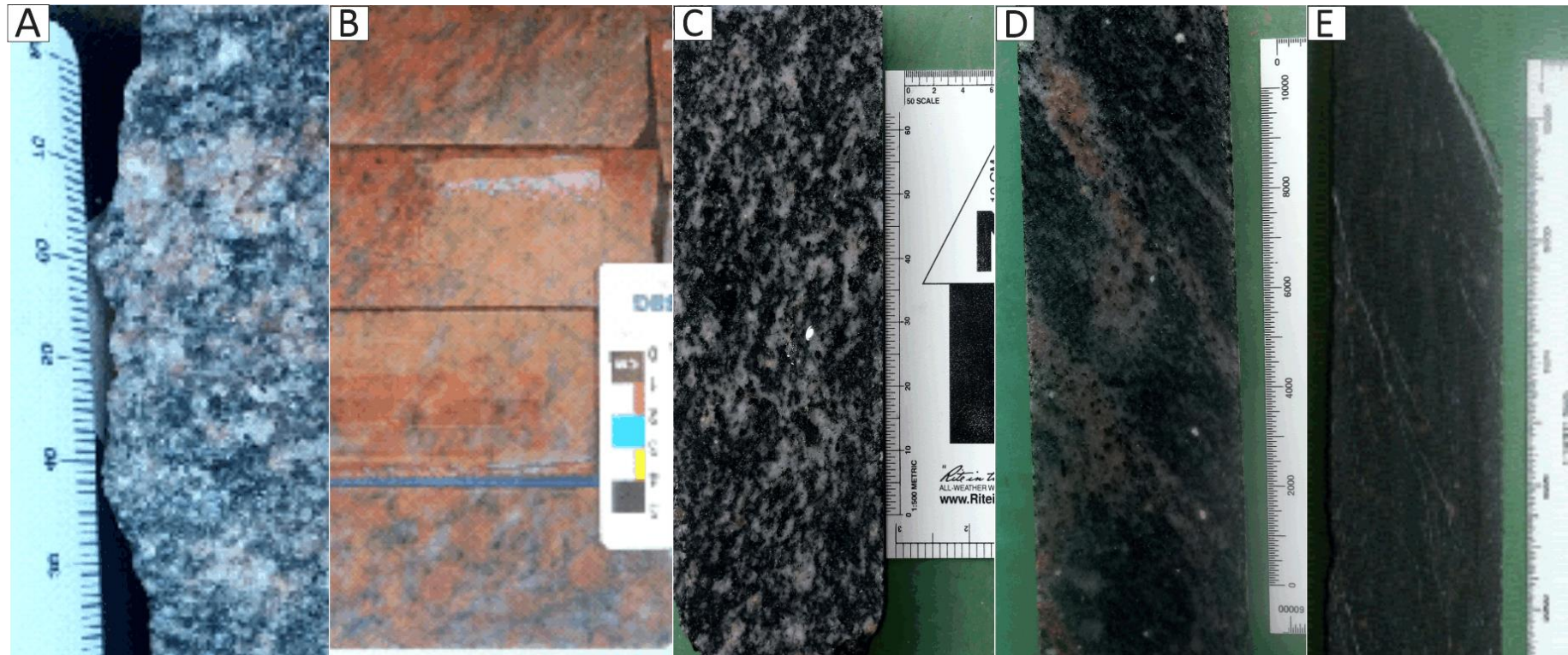


Fig. S1. Macroscopic characteristics of the Novo Mundo rocks. A) Monzogranite: inequigranular phaneritic texture, locally containing plagioclase fenocrysts. B) Syenogranite: inequigranular phaneritic texture. C) Granodiorite: inequigranular texture. D) Monzonite: inequigranular texture, mafic-rich rock. E) Mafic dike: fine-grained, locally containing plagioclase porphyries.

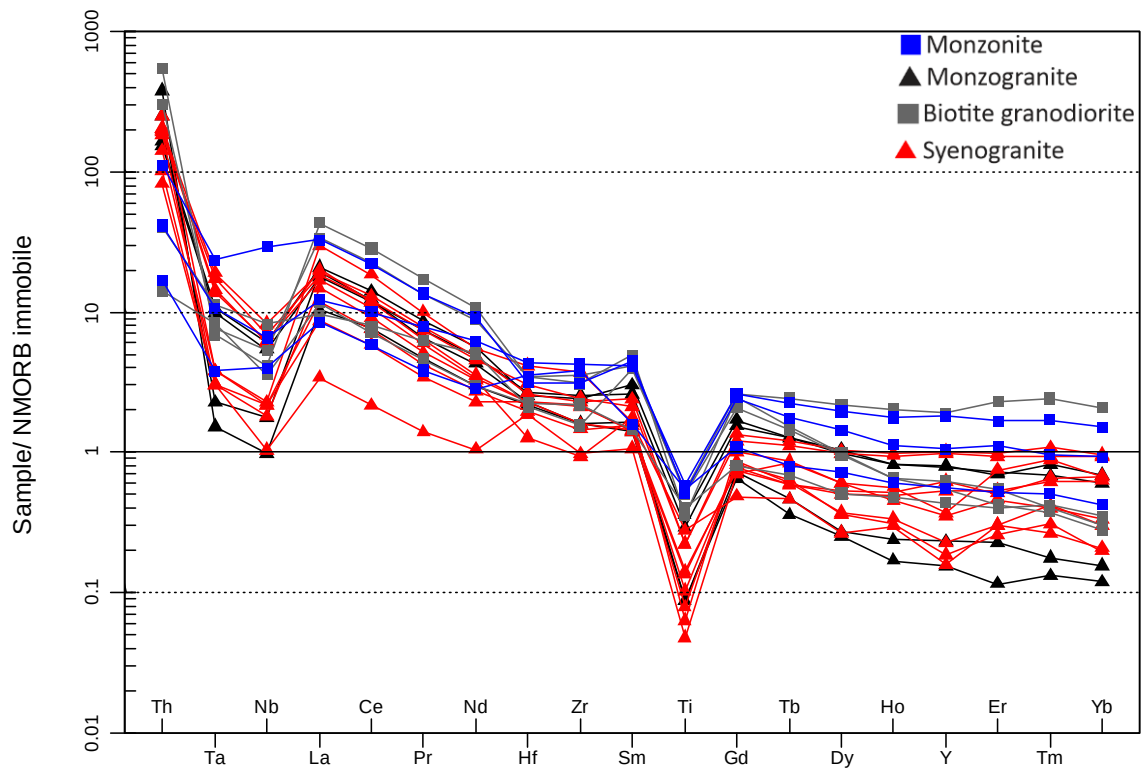


Fig. S2. Multielement distribution pattern of Novo Mundo Complex facies (Sun and McDonough 1989).

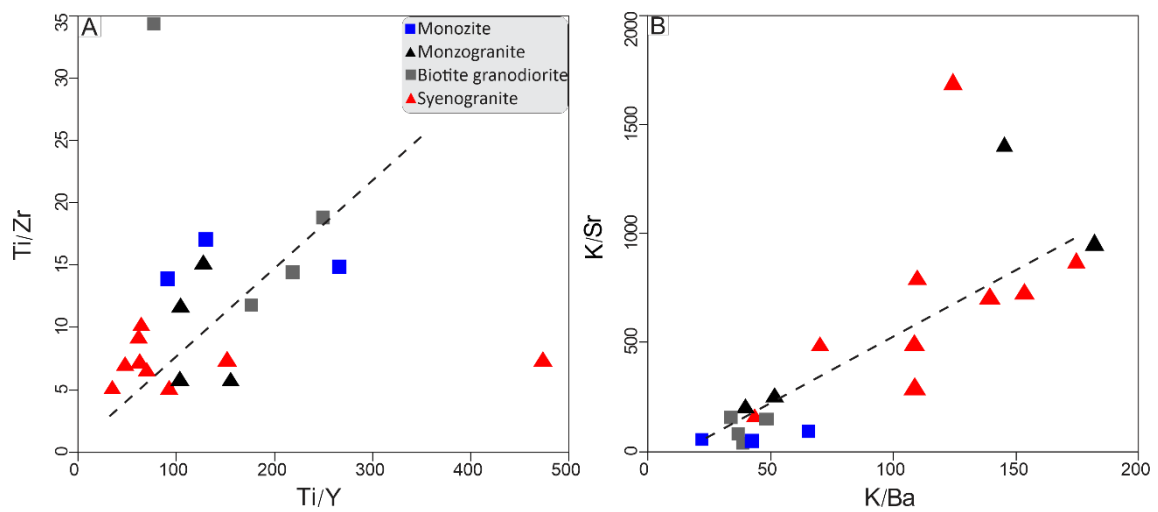


Fig. S3. Geochemical models from incompatible/compatible elements ratios from Novo Mundo granites. The main trend source that originated the granitic facies of the Novo Mundo complex is visible in the straight lines.

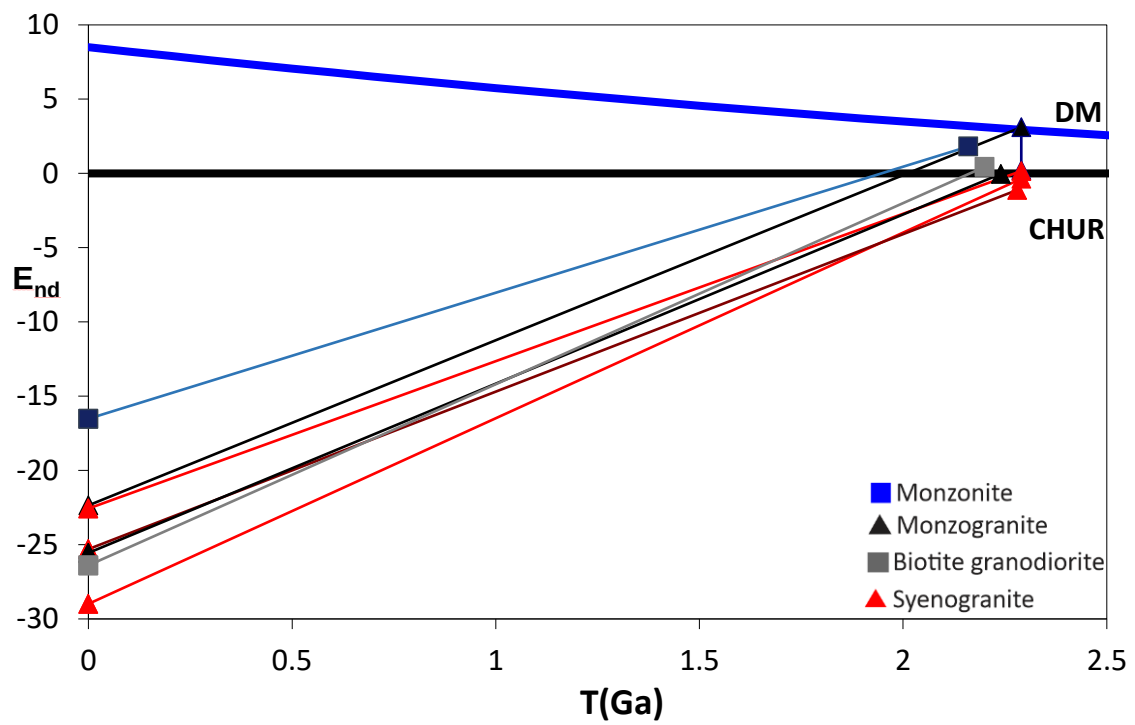


Fig. S4. $\epsilon_{Nd(t)}$ vs magmatic age plot for the Novo Mundo granites, based on depleted mantle array after [De Paolo \(1981\)](#).

