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**Formação, Transformação e Evolução da Crosta Continental Inferior:
Investigando o Orógeno Brasília Meridional**

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RESUMO

A *Nappe* Socorro-Guaxupé, localizada na porção mais ao sul do Orógeno Brasília, expõe uma grande fatia da crosta continental inferior neoproterozóica, em que granulitos, migmatitos e granitóides, incluindo charnockitos, fornecem oportunidade ímpar para investigar processos geológicos responsáveis pela sua geração. Nesse trabalho, a investigação desse segmento permitiu reconstruir as condições petrológicas e cronológicas às quais a crosta foi submetida, utilizando geoquímica elemental e isotópica, modelagem termodinâmica, petrocronologia e geologia estrutural. A base do pacote é predominantemente granulítica e atingiu o pico metamórfico em condições de temperatura ultra-alta (UHT), $\sim 1.000\text{ }^{\circ}\text{C}$ e 11 kbar, seguido por decompressão isotérmica até $\sim 9,5$ kbar. Recristalização de núcleos antigos de zircão em *c.* 625 Ma marcam a trajetória progressiva em condições UHT e, em *c.* 615 Ma, neocristais de zircão com alto teor de U registram o início da cristalização do material fundido, o que limita a idade de pico metamórfico entre as duas datas. A trajetória segue caminho de resfriamento isobárico até $\sim 870\text{ }^{\circ}\text{C}$ e 9 kbar, em que fases hidratadas cristalizaram em *c.* 595 Ma, marcado por neocristais de zircão com baixo conteúdo de U e empobrecidos em E RTP. Assim, a história metamórfica dos granulitos mostra que a crosta inferior se manteve aquecida em fácies granulito por pelo menos 30 Myr. Essas idades coincidem com a geração de charnockitos da Suíte Paraguaçu, em *c.* 620 Ma. Os charnockitos são empobrecidos em Th, U, Sr e Ti, enriquecidos em Ba, K, ETRL, P, Zr e Hf, padrão inverso ao observado nos granulitos félsicos, além de possuírem altas razões K/Rb e #Mg. A modelagem termodinâmica mostra que, durante o pico metamórfico nos granulitos félsicos, feldspato potássio, granada, zircão e apatita foram parcialmente consumidos nas reações de anatexia, ao passo que plagioclásio permaneceu no resíduo. Desse modo, o resultado da fusão parcial dos granulitos félsicos teve papel fundamental para a geração da Suíte Paraguaçu. Esses charnockitos estão encaixados em diatexitos que se sobrepõe aos granulitos através de zona de cisalhamento horizontal. Dados estruturais mostram que as geometrias dessas unidades apesar de similares não são coincidentes. Ambas unidades atingiram o pico metamórfico em condições de UHT, contudo as temperaturas de deformação mostram uma diferença de $200\text{ }^{\circ}\text{C}$ em relação ao pico metamórfico nos granulitos e de $350\text{ }^{\circ}\text{C}$ nos diatexitos. Portanto, enquanto os granulitos estavam consolidados, os diatexitos não estavam, assim posteriormente registraram alterações nos regimes de tensões no orógeno. Os dados mostram que o aquecimento a temperaturas extremas como as registradas foi possível pela combinação de dois fatores: a geração de grandes volumes de líquido silicático, com grande concentração de elementos radioativos, principalmente K e a contribuição de fonte externa, provavelmente mantélica, marcado pelo enriquecimento em ^{176}Hf radiogênico na Suíte Paraguaçu e por diques de rocha máfica.

ABSTRACT

The Socorro-Guaxupé Nappe, in the southernmost portion of the Brasília Orogen, exhibits a large slice of a Neoproterozoic lower continental crust, in which granulites, migmatites and granitoids, including charnockites, provide unique opportunity to investigate geological process that was responsible to its generation. In this study, the investigation of this segment allowed to reconstruct petrological and chronological conditions that the crust underwent, using elemental and isotopic geochemistry, thermodynamic modelling, petrochronology and structural geology. The basis of the stack is dominantly granulitic and reached the metamorphic peak at ultra-high temperature (UHT) conditions, $\sim 1,000\text{ }^{\circ}\text{C}$ and 11 kbar, followed by near-isothermal decompression until ~ 9.5 kbar. Recrystallization of older zircon cores at *c.* 625 Ma records the prograde path at UHT conditions and, at *c.* 615 Ma, high-U zircon neocrystals records the onset of melt crystallization, therefore these dates bracket the age of metamorphic peak. After that, the metamorphic path experienced a near-isobaric cooling until $\sim 870\text{ }^{\circ}\text{C}$ and 9 kbar, in which hydrated phases crystallized at *c.* 595 Ma, registered by low-U zircon neocrystals depleted in HREE. Thus, the granulite metamorphic history indicates that the lower continental crust was heated at granulite facies for at least 30 Myr. These ages are coeval to the emplacement of charnockites of the Paraguaçu Suite, at *c.* 620 Ma. The charnockites are depleted in Th, U, Sr and Ti and enriched in Ba, K, LREE, P, Zr and Hf, the opposite pattern observed in the felsic granulites, moreover, they have high K/Rb ratios and #Mg. The thermodynamic modelling shows that, during the felsic granulite metamorphic peak, K-feldspar, garnet, zircon and apatite were partially consumed in the melt-reactions, whereas plagioclase remained in the residuum. Thus, it is likely that the partial melting of the felsic granulite played a major role in the Suite Paraguaçu genesis. These charnockites are emplaced into the diatexites that overlays the granulites by a horizontal shear zone. Structural data show that the geometries of these units, although similar, are not coincident. Both diatexites and granulites achieved the metamorphic peak at UHT conditions, however, the deformation temperature are $200\text{ }^{\circ}\text{C}$ colder than the metamorphic peak in the granulites and $350\text{ }^{\circ}\text{C}$ in the diatexites. Thereby, while the granulites were totally crystallized, the diatexite were not, thus the diatexites were able to record changes in the field stress of the orogen. The data indicate that the heating to elevate to such extreme temperatures, as recorded, was possible by the combination of two features: the generation of large volumes of melt, with high concentration of radioactive elements, mainly K; and a contribution of an external source, probably from the mantle, registered by enrichment radiogenic ^{176}Hf in Paraguaçu Suite and dikes of mafic rocks.

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1 INTRODUÇÃO

Boas exposições da base da crosta continental são relativamente raras e representadas por rochas ígneas ou de alto grau metamórfico (Bohlen & Mezger, 1989; Harley, 1989; Percival *et al.*, 1992; Brown, 2001; Brown & Johnson, 2018). No Brasil, são encontrados alguns bons exemplos dessas exposições (Moraes *et al.*, 2004) e as rochas da *Nappe* Socorro-Guaxupé, MG, são excelente exemplo, em que granulitos, migmatitos, charnockitos e granitos afloram em ampla região em Minas Gerais e São Paulo com exposição que representa seção da crosta continental média a inferior (Campos Neto & Caby, 1999, 2000).

Essas exposições permitem o acesso às rochas formadas em crosta continental profunda, em que deformação dúctil gera as estruturas, em conjunto com metamorfismo de altas temperaturas e fusão parcial, o que possibilita a investigação dos processos e mecanismos envolvidos na geração e transformação da base da crosta continental. Isso permite a investigação das relações entre as condições reológicas, metamórficas, e de anatexia, ajudando na compreensão do quadro do processo deformacional e metamórfico em condições geodinâmicas extremas de temperatura. Os granulitos apresentam especial interesse para a presente investigação, já que sua mineralogia e texturas registram evidências das condições *P-T-t* de sua formação e exumação, sendo que a trajetória gravada permite inferir o quadro tectônico relacionado com sua gênese (Harley, 1989; Brown, 2001; Kelsey & Hand, 2015).

Neste trabalho, foi realizado estudo detalhado das rochas da *Nappe* Socorro-Guaxupé a fim de compreender a geração, transformação e evolução dessa fatia de crosta continental inferior. Para tal, foram utilizadas técnicas modernas de modelagem de equilíbrio de fases (Powell *et al.*, 2005) e petrocronologia, uma nova cadeira dentro das geociências (Engi *et al.*, 2017), aliadas à técnicas mais tradicionais, como geoquímica elemental e isotópica (*e.g.* Frost *et al.*, 2001; White, 2013), geotermobarometria convencional (*e.g.* Newton & Haselton, 1981; Pattison & Bégin, 1994) e geologia estrutural (*e.g.* Ramsay & Huber, 1987). Desse modo, os dados e discussões apresentados contribuem não apenas para a tectônica regional do Orógeno Brasília Meridional, mas também para o entendimento global dos processos petrológicos, cronológicos e geodinâmicos que atuam na crosta continental inferior.

O trabalho foi realizado no Instituto de Geociências da Universidade de São Paulo (IGc-USP), com estágio de um ano no *The Institute of Geoscience Research* (TIGeR) da *Curtin University*, Perth, Austrália. As análises foram realizadas nos parques laboratoriais dos dois institutos citados, além do Laboratório de Microsonda Eletrônica da Universidade Estadual Paulista (UNESP). O estudo contou com recursos financeiros da Fundação de Apoio à Pesquisa do Estado de São Paulo (FAPESP – 13/04007-0 e 16/22627-3) e do *Australian Research Council Discovery Project*

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1.1 Estruturação do volume

Na busca de melhores índices institucionais e pessoais, cada vez mais é incentivado a publicação de artigos científicos nos âmbitos acadêmicos, inclusive dentro das teses de doutoramento. Dentro do Instituto de Geociências da Universidade de São Paulo (IGc-USP) e do Programa de Pós-Graduação Geociências (Mineralogia e Petrologia) a prática já é adotada há algum tempo com sucesso (*e.g.* Florisbal, 2011; Henrique-Pinto, 2012; Vilalva, 2012; Westin, 2015; Almeida, 2016; Andrade, 2016; Cioffi, 2016; Rocha, 2016; Pavan, 2017). As vantagens desse tipo de abordagem em teses de doutoramento são que estas tornam-se mais concisas e pragmáticas, além de artigos em periódicos de alto impacto sempre são revisados por especialistas no assunto, atestando qualidade dos dados e autenticidade das discussões. Seguindo a tendência, a presente tese está estruturada da seguinte forma:

- O Capítulo 1 (Introdução) apresenta a introdução ao texto da tese, justificativas, objetivos, localização e métodos utilizados;
- O Capítulo 2 (Metamorfismo de temperatura ultra-alta) apresenta a revisão bibliográfica sobre rochas de alto grau metamórfico, em especial rochas que atingiram condições de temperatura ultra-alta;
- O Capítulo 3 (Contexto geológico) insere o leitor na configuração geológica da área de estudo;
- O Capítulo 4 (Granulitos do Domínio Guaxupé) apresenta o resumo do artigo que será submetido ao periódico *Journal of Metamorphic Geology*, intitulado “*Recovering P–T–t paths from ultra-high temperature (UHT) rocks that lack diagnostic mineral assemblages: A felsic granulite example from the Southern Brasília Orogen, MG, Brazil*”, que caracteriza os granulitos félsicos da *Nappe Socorro-Guaxupé* que não apresentam associações minerais diagnósticas de temperatura ultra-alta, mas que atingiram tais condições, o que é comprovado com estudo usando modelagem metamórfica, química mineral, datações U-Pb, isotopia Lu-Hf em zircão e geoquímica de rocha total e de elementos traços de zircão (manuscrito completo no ANEXO I);
- O Capítulo 5 (Charnockitos da Suíte Paraguaçu) apresenta o resumo do artigo que será submetido à publicação no periódico *Precambrian Research*, intitulado “*The origin of charnockites from granulite partial melting: an example in Socorro-Guaxupé Nappe, MG, southeastern Brazil*” que atribui a gênese de intrusões anídras de charnockitos da *Nappe Socorro-Guaxupé* à fusão parcial de granulitos, usando

dados de geoquímica de rocha total, modelagem termodinâmica, geocronologia U-Pb e isotopia Lu-Hf em zircão e isotopia Sm-Nd e Rb-Sr em rocha total (manuscrito completo no ANEXO II);

- O Capítulo 6 (Relação entre Granulitos e Migmatitos) apresenta o resumo do manuscrito à ser submetido ao periódico *Journal of South American Earth Sciences*, intitulado “*Relationship between metamorphic peak and deformation temperature in migmatites and granulites and their constrain on exhumation in the Socorro-Guaxupé Nappe, MG, SE Brazil*” que aborda as diferenças de temperatura de pico metamórfico e de deformação através da integração de dados texturais e estruturais e de química mineral em duas unidades da *Nappe Socorro-Guaxupé* (manuscrito completo no ANEXO III);
- O Capítulo 7 (Discussões) apresenta a integração dos dados e interpretações obtidas nos capítulos anteriores;
- O Capítulo 8 (Conclusões) faz um sumário das principais conclusões do presente volume.

1.2 Objetivos

A pesquisa objetiva caracterizar e estudar aspectos petrológicos, geocronológicos e estruturais no Orógeno Brasília Meridional e assim não só avançar no entendimento regional, como contribuir para a crescente “nova” disciplina das Geociências, a Petrocronologia, em que a coleta complementar de dados isotópicos e elementais aliada à dados de cálculos *P-T* tem revolucionado a Petrologia Metamórfica.

Para tal, os seguintes temas são abordados:

- Caracterização dos granulitos, definindo as condições metamórficas e tempo envolvido entre a progressão e a regressão metamórfica;
- Estudo da gênese de intrusões de corpos de charnockitos;
- Estudo das relações estruturais e metamórficas entre os granulitos e migmatitos da região;
- Definição da cronologia dos eventos.

1.3 Área de estudo

A área de estudo está localizada no sul do estado de Minas Gerais e compreende um polígono entre as cidades de Alfenas, Machado, Varginha e Três Pontas (Figura 1-1). A principal rodovia que corta a área é a BR-491 que liga as cidades de Alfenas e Varginha. A grande

concentração das amostras foi coletada às margens do Rio Verde e Rio Dourado, nas proximidades das cidades de Varginha e Paraguaçu, respectivamente.

Figura 1-1 Localização aproximada da área de estudo.

1.4 Métodos

1.4.1 Pesquisa bibliográfica

A pesquisa bibliográfica foi dividida em duas partes: *i*) material para elaboração de Geologia Regional, e *ii*) material para fundamentar os conceitos explorados no trabalho. As pesquisas foram realizadas de modo contínuo utilizando-se o acervo da Biblioteca do Instituto de Geociências da Universidade de São Paulo (IGc-USP), Brasil, ou o portal Periódicos CAPES.

1.4.2 Levantamento de campo

Foram realizados reconhecimentos geológicos regionais, através de perfis transversais às estruturas das rochas. As melhores exposições ocorrem das margens dos rios Verde e Dourado, em que levantamentos de detalhe foram realizados. Os trabalhos foram auxiliados pela utilização de mapas topográficos, bússola, martelo, caderneta de campo, lupa e a coleta de dados estruturais, como foliações, lineações e indicadores cinemáticos envolveu métodos convencionais. Nessa etapa foram coletadas amostras orientadas ou não para análises petrográfica e microestrutural.

1.4.3 Análise petrográfica

As atividades se iniciaram durante os trabalhos de campo, com descrição macroscópica das amostras, identificando os principais minerais, texturas e estruturas. Posteriormente, seções delgadas foram descritas em microscópio ótico de luz transmitida, nas quais foram observadas as associações minerais e texturas, para caracterização metamórfica e microestrutural das rochas. Foram utilizados microscópios óticos do Laboratório de Microscopia Petrográfica do Núcleo de Apoio à Pesquisa (NAP) Geoanalítica (IGc-USP) e *Mineral Separation Facility* da *Curtin University*, Austrália. Algumas amostras foram investigadas através de imagens de elétrons retroespalhados (*back-scattered electron imaging* – BSE), com o intuito de identificar texturas e zoneamentos composicionais entre diversas fases. As imagens foram obtidas no *Microscopy and Microanalysis Facility* (MMF) da *Curtin University*, através de microscópio eletrônico de varredura (MEV) Tescan Mira3 FESEM (*field emission scanning electron microscope* – FESEM), com corrente em torno de 15 kV e distância de trabalho de 15 mm.

1.4.4 Análise de orientação preferencial de eixo-c de quartzo

Medidas de eixo-c de quartzo foram realizadas com uso de platina universal com quatro eixos acoplada a Microscópio Óptico Zeiss. Os eixos medidos foram lançados em redes estereográficas de igual área, na semi-esfera inferior, tendo como referência as direções do elipsoide de deformação finita (X, Y, Z). As curvas de isofrequências geradas foram utilizadas para a quantificação de temperaturas da formação das tramas (Law, 2014; Faleiros *et al.*, 2016).

1.4.5 Química mineral

Análises químicas mineral quantitativa foram realizadas através de espectrometria dispersiva de comprimento de ondas (*wavelength-dispersive spectrometry* – WDS) através de microsonda eletrônica JEOL JXA-8230 Hyperprobe, localizado no Laboratório de Microsonda Eletrônica da Universidade Estadual Paulista (UNESP), utilizando feixe à 15 kV e 20 nA e feixe com diâmetro de 10 µm para todos os minerais, exceto clinopiroxênio, ao qual foi utilizado um feixe com 1 µm de diâmetro devido à presença de pequenas lamelas de exsolução. As análises foram recalculadas utilizando-se o programa “A-X” (<https://www.esc.cam.ac.uk/research/research-groups/holland/ax>). Além de clinopiroxênio, foram analisados ortopiroxênio, granada, biotita, feldspato potássio, plagioclásio e hornblenda. Os dados estão disponíveis no ANEXO IV.

Análises complementares em granada para quantificação de elementos traço foram realizadas no laboratório NAP- Geoanalítica (IGc-USP), utilizando um ICP-MS Thermo Scientific iCAP Q acoplado a um sistema de ablação à laser New Wave UP 213 nm. As análises foram realizadas *in situ* em seção delgada através de raster com ~180 µm e feixe com diâmetro de 65

μm , fluência do laser de 5.52 J/cm^2 e taxa de repetição de 10 Hz. Calibração externa foi realizada relativa ao material de referência NIST610 e o vidro basáltico USGS BCR-2G foi analisado como padrão secundário. As concentrações de Si e Ca, determinadas por microsonda eletrônica, foram utilizadas como padrão interno. A redução dos dados foi realizada no programa GLITTER. O ANEXO V apresenta os dados obtidos.

1.4.6 Geotermobarometria e cálculos de equilíbrio de fases

Os cálculos termobarométricos foram realizados utilizando bancos de dados termodinâmicos internamente consistente, utilizando o programa THERMOCALC, versão 3.45i (Powell & Holland, 1988). Também foram utilizados os geotermobarômetros que considera a quantidade de Al em ortopiroxênio (Pattison *et al.*, 2003), e o barômetro GASP, granada – Al_2SiO_5 – sílica – plagioclásio (Newton & Haselton, 1981; Reche & Martinez, 1996).

Todos os modelos de equilíbrio de fases foram realizados em sistema de dez componentes – NCKFMASHTO (Na_2O - CaO - K_2O - FeO - MgO - Al_2O_3 - SiO_2 - H_2O - TiO_2 - O_2) – utilizando o programa THERMOCALC versão 3.45i (Powell & Holland, 1988). Os modelos de atividades utilizados são os indicados por Holland & Powell, (2003), White *et al.* (2014) e Green *et al.* (2016).

1.4.7 Geoquímica de rocha total

Os elementos maiores, menores e alguns elementos traço foram analisados a partir de pastilhas fundidas e prensadas no Laboratório de Espectrometria de Fluorescência de Raio-X do NAP-Geoanalítica, IGc-USP. Utilizou-se o equipamento da marca Philips, modelo PW2400, seguindo a metodologia do laboratório (Mori *et al.*, 1999). Os elementos terras raras (ETR) e alguns elementos traço foram analisados no Laboratório de Química e ICP-AES/MS do NAP-Geoanalítica, IGc-USP, usando pó de rocha da mesma alíquota da qual foram feitas as pastilhas para as análises de fluorescência. O laboratório utiliza um espectrômetro do tipo quadrupolo, modelo iCAP-Q da Thermo Scientific, seguindo metodologia do laboratório (Navarro *et al.*, 2008). Partes das análises utilizadas nesse trabalho são compilações de trabalhos anteriores (Del Lama *et al.*, 1998; Janasi, 2002; Barros, 2012). Os dados são apresentados no ANEXO VI.

1.4.8 Geoquímica isotópica Rb-Sr e Sm-Nd em rocha total

Análises de geoquímica isotópica foram realizadas em seis amostras para os sistemas Rb-Sr e Sm-Nd no Laboratório de Espectrometria do Centro de Pesquisas Geocronológicas da Universidade de São Paulo (CPGeo-USP). As razões Rb/Sr e Sm/Nd foram calculadas a partir das concentrações determinadas por ICP-MS (item 1.4.7). Cálculos de idade modelo, ϵNd e as razões iniciais Rb/Sr foram calculadas usando o programa GCDkit (Janoušek *et al.*, 2006).

Análises do sistema Rb-Sr foram executadas utilizando espectrômetro de massa por termo-ionização (TIMS – Thermo Triton), sendo as razões isotópicas $^{87}\text{Sr}/^{86}\text{Sr}$ normalizadas para o valor de 0,1194 de $^{86}\text{Sr}/^{88}\text{Sr}$. O padrão NBS-987, utilizado para as análises, forneceu razão $^{87}\text{Sr}/^{86}\text{Sr}$ de $0,710239 \pm 0,000017$ (2σ) durante o período de um ano, ao qual as análises foram submetidas. Análises do sistema Sm-Nd foram realizadas utilizando TIMS (Thermo Triton) em que as razões $^{143}\text{Nd}/^{144}\text{Nd}$ foram normalizadas para o valor de 0,7219 de $^{146}\text{Nd}/^{144}\text{Nd}$. A razão $^{143}\text{Nd}/^{144}\text{Nd}$ média do padrão JNdi-1 para o período do ano em que a análise foi realizada foi de $0,512102 \pm 0,000002$ e o branco analítico forneceu 40 pg. As análises são apresentadas no ANEXO VII.

1.4.9 Geocronologia U-Pb em zircão

1.4.9.1 *Preparação de amostras*

A separação mineral foi realizada no CPGeo-USP. Seguiu-se as etapas de lavagem e retirada de porções muito alteradas da rocha, fragmentação mecânica com marreta, fragmentação com britador de mandíbulas de aço e pulverização em moinho de discos. Após a etapa de desagregação as amostras seguiram para separador de densidade (mesa Wiffley), separação magnética com ímã de mão, separador magnético Frantz e separação de densidade através de líquidos densos – bromofórmio ($2,89 \text{ g/cm}^3$) e iodeto de metileno ($3,32 \text{ g/cm}^3$). Desse modo, zircão concentrou-se em frações densas e não magnéticas.

Os concentrados minerais foram levados para a *Curtin University*, onde foram montados blocos de amostras de 25 mm de diâmetro. Grãos de interesse foram selecionados através de lupa e colados em fita dupla face em placas de vidro. Estes foram organizados em linhas com aproximadamente 12 mm em torno do centro do bloco e cristais de padrões foram adicionados. Resina epóxi é despejada em anéis plásticos, com espessuras entre 5 e 6 mm, e foram curados por no mínimo 48 horas. Após curados foram polidos manualmente até expor os núcleos dos grãos e obter uma superfície plana e lisa.

1.4.9.2 *Imageamento*

As estruturas internas de grãos foram identificadas através de imagens de MEV, usando aparelho Tescan Mira3, realizadas no MMF da *Curtin University*. Foram utilizadas imagens de catodoluminescência (CL), com corrente de 12 kV e distância de trabalho de 16,5 mm.

1.4.9.3 *Datação de zircão*

Datações U-Pb em zircão foram coletadas utilizando microsonda iônica (*Sensitive High Resolution Ion Microprobe* – SHRIMP II), localizada no *John de Laeter Centre* da *Curtin University*. A intensidade do feixe de O_2 primário variou entre 2 e 3 nA com diâmetro de feixe de 20 μm . Um ciclo de 6 leituras e resolução de massa de 5000 foi utilizado durante as análises (Kennedy & de

Laeter, 1994; de Laeter & Kennedy, 1998). Os dados foram processados utilizando os programas SQUID II e Isoplot 3.75 (Ludwig, 2003, 2009). As amostras foram cobertas por película de 40 µm de ouro. Diferentes tipos de padrões foram utilizados (mais informações nos tópicos relacionados a cada análise encontram-se nos ANEXO I e ANEXO II). Os resultados analíticos estão no ANEXO VIII.

1.4.10 Isótopos de Hf e elementos traço em zircão

Dados de elementos traço, isótopos Lu-Hf e datações U-Pb complementares em zircão foram coletados nos mesmos blocos de amostras analisadas via SHRIMP, utilizando ablação a laser com fluxo separado (*Laser Ablation Split-Stream Inductively Coupled Plasma Mass Spectrometry – LASS-ICP-MS*), em que análises elementais e isotópicas são realizadas simultaneamente, em um *ASI Resolution M-50* com laser de Ar-F a um comprimento de onda de 193 nm com espectrômetro de massa Agilent 7700, localizado na *GeoHistory Facility* do *John de Laeter Centre, Curtin University*. Foram utilizados *spot* circulares de 50 µm de diâmetro com fluência de laser de 2 J/cm² e taxa de repetição de 10 Hz. Os dados estão disponibilizados nos ANEXO IX, ANEXO X e ANEXO XI.

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2 METAMORFISMO DE TEMPERATURA ULTRA-ALTA

2.1 Reconhecendo o metamorfismo de temperatura ultra-alta

O termo fácies metamórfica é utilizado para designar as variações das condições de pressão e temperatura que uma rocha metamórfica sofreu durante sua história. Inicialmente foi formulada para rochas de composição basáltica, em que as mudanças das paragéneses minerais definem campos específicos de temperatura e pressão do metamorfismo (Eskola, 1939). Nesse contexto, as rochas portadoras de paragéneses com ortopiroxênio foram consideradas representantes da fácies granulito, condições de temperaturas mais elevadas do metamorfismo (Eskola, 1939). Contudo, fatores como composição e pressão podem condicionar a presença ou ausência de ortopiroxênio nas rochas da fácies granulito (Johnson *et al.*, 2008; O'Brien & Rotzler, 2003). Assim, a partir de estudos experimentais com rochas basálticas (Green & Ringwood, 1967; Hansen, 1981; Ito & Kennedy, 1964), a fácies granulito foi subdivida em campos de alta, média e baixa pressão. Em resposta às subdivisões relativas aos regimes de pressão, tendo a temperatura como condicionante, uma nova terminologia foi proposta por Harley (1998), que definiu o metamorfismo de temperatura ultra-alta (UHT, do inglês *ultra-high temperature*).

Definido como um metamorfismo de fácies granulito em que temperaturas acima de 900 °C em pressões entre 7 e 13 kbar são alcançadas na crosta continental durante o metamorfismo regional (Harley, 1998) (Figura 2-1). O metamorfismo de temperatura ultra-alta foi balizado a partir do ponto invariante espinélio-ausente, marcando o limite de pressão e temperatura inferior da parágênese ortopiroxênio + sillimanita + quartzo + líquido silicático, em estudos experimentais no sistema K_2O -FeO-MgO- Al_2O_3 - SiO_2 - H_2O (KFMASH) (Harley, 1998; Kelsey, 2008; Kelsey & Hand, 2015). Contudo não existe nenhuma lacuna termal entre os granulitos de temperatura ultra-alta (> 900 °C) e os granulitos “comuns” (~750 a 900 °C), havendo distribuição continua das condições de temperatura na fácies granulito a partir de 750 °C até $T > 1100$ °C (Harley, 1998; Pattison *et al.*, 2003). O metamorfismo de temperatura ultra-alta também pode ser definido pelo gradiente termal da crosta, quando atinge razões acima de 75 °C/kbar (Brown, 2014, 2006; Brown & Johnson, 2018; Stüwe, 2007). Assim a análise do gradiente termal é importante para estudar orógenos modernos que não foram completamente exumados, em que a exposição de rochas ligadas ao metamorfismo regional de grau baixo, mas com alto gradiente termal, provavelmente contém rochas de temperatura ultra-alta em profundidade, portanto esse tipo de metamorfismo ainda deve ocorrer atualmente na crosta terrestre (Kelsey & Hand, 2015).

Figura 2-1 UHT – temperatura ultra alta; UHP – pressão ultra alta; HP-g – granulito de alta pressão; Bt – biotita; Qtz, Qz ou q – quartzo; Kfs – feldspato potássio; Opx – ortopiroxênio; Sil – sillimanita; g – granada; osm – ossumilita; cd – cordierita; mt – magnetita. (A) Localização do campo de UHT em diagrama P-T (modificado de Kelsey (2008)). (B) Paragênese diagnóstica de UHT: safirina + quartzo, complexo Anápolis-Itaçu, GO, Brasil (retirado de Moraes et al. (2002)). (C) Paragênese diagnóstica de UHT: ortopiroxênio + sillimanita + quartzo, complexo Napier, Antártica (retirado de Harley (2008)). (D) Paragênese diagnóstica de UHT: ossumilita + granada, Rogaland, Noruega (retirado de Kelsey & Hand (2015)).

As paragêneses diagnósticas de metamorfismo de temperatura ultra-alta são desenvolvidas em rochas muito ricas em Mg e Al, composições raras que podem ser geradas de prováveis protólitos sedimentares ou metassomáticos (Harley, 2008; Kelsey, 2008; Kelsey & Hand, 2015). Assim, o reconhecimento do metamorfismo de temperatura ultra-alta é, em geral, problemático, uma vez que apenas poucas rochas desenvolvem paragêneses diagnósticas. Nesse contexto, três paragêneses são diagnósticas de metamorfismo de temperatura ultra-alta (Figura 2-1): safirina + quartzo, ortopiroxênio (rico em Al) + sillimanita + quartzo e ossumilita + granada (Harley, 2008). Outras associações minerais portadoras de safirina como safirina + cordierita, safirina + ortopiroxênio ± biotita ± cordierita (sem quartzo), safirina + granada e safirina ± espinélio ± coríndon são indicativas de metamorfismo de temperatura ultra-alta desde que líquido silicático esteja presente e a quantidade de ferro férrico nos minerais seja extremamente baixas (Harley, 2008; Kelsey, 2008). Associações minerais sem safirina como espinélio + quartzo, coríndon + quartzo e ortopiroxênio + coríndon também são sugestivas de metamorfismo de ultra-alta temperatura, porém a quantidade de Zn, Cr, Ti e Fe^{3+} incorporado nesses minerais podem influenciar na estabilidade das associações (Clark *et al.*, 2011; Harley, 1998; Kelsey & Hand, 2015), e as condições P-T ainda não são muito bem definidas termodinamicamente.

2.2 Quantificando metamorfismo de temperatura ultra-alta

Tradicionalmente, o metamorfismo de temperatura ultra-alta era apenas identificado pelas paragêneses minerais, uma vez que ocorre grande reequilíbrio químico dos minerais dessas rochas durante o resfriamento, sendo a termobarometria convencional baseada nas trocas Fe-Mg não aplicáveis à rochas de temperatura ultra-alta (Kelsey & Hand, 2015). Entretanto, a difusividade do Al em altas temperaturas é significativamente menor que Fe-Mg (Fitzsimons & Harley, 1994). Assim, uma ótima ferramenta para determinação da temperatura da formação da rocha é conteúdo de Al no ortopiroxênio (Harley, 1998; Kelsey, 2008), que é altamente dependente da temperatura (Fitzsimons & Harley, 1994; Pattison *et al.*, 2003).

Outras técnicas baseadas em química mineral têm sido usadas com grande sucesso para a determinação de condições de temperatura ultra-alta, como a reintegração da composição de feldspato ternário (Fuhrman & Lindsley, 1988; Santosh *et al.*, 2009), os teores de Zr em rutilo (Tomkins *et al.*, 2007; Zack *et al.*, 2004), de Ti em zircão (Watson *et al.*, 2006) e de Ti em quartzo (Thomas *et al.*, 2010).

Cálculos de pseudosseções em que os campos P - T de equilíbrio entre as fases podem ser investigados têm revolucionado o modo de análise de terrenos metamorfizados em temperatura ultra-alta. As pseudosseções são diagramas que funcionam como mapas dos campos de estabilidade P - T das possíveis associações minerais geradas por uma composição de rocha específica. Os cálculos são realizados com programas e banco de dados termodinâmicos internamente consistentes que utilizam dados termodinâmicos dos membros-finais das fases presentes (Powell *et al.*, 2005, 1998). Especialmente modelos de atividade refinados para safirina (Kelsey *et al.*, 2004; Taylor-Jones & Powell, 2010) permitem as construções de pseudosseções para rochas ricas em Mg e Al em sistemas complexos como o KFMASHTO (K_2O - FeO - MgO - Al_2O_3 - SiO_2 - H_2O - TiO_2 - O_2), ou seja, para rochas com composição capaz de gerar paragêneses diagnósticas de temperaturas ultra-altas.

Entretanto, até recentemente, quando se tratava de rochas portadoras de hornblenda, como os protólitos de granulitos félsicos e máficos, um problema se fazia presente: a fusão parcial dessas rochas gera líquido trondhjemítico-tonalítico, isto é, mais rico em Na_2O - CaO e pobre em K_2O e SiO_2 do que os produzidos pela fusão parcial de pelitos. Modelos recentes de atividade que lidam com piroxênios, anfibólios e fusão parcial em temperatura e pressão alta (Green *et al.*, 2016; Palin *et al.*, 2016) trouxeram grandes avanços na modelagem termodinâmica da fusão parcial em condições extremas de metamorfismo de protólitos comuns de granulitos máficos e félsicos, em sistemas complexos, de 10 componentes, NCKFMASHTO (Na_2O - CaO + KFMASHTO).

Desse modo, cálculos de pseudosseção para composições diversas, aquelas que desenvolvem ou não paragêneses diagnósticas, agora são possíveis. Contudo, não se trata de tema trivial e não foi estabelecido na literatura quais são os melhores procedimentos para a construção das pseudosseções para investigação dessas rochas (Kelsey & Hand, 2015). A principal razão para isso é a escolha da composição “ideal” de rocha, pois os sistemas em temperatura alta e ultra alta devem ser considerados abertos, principalmente pela perda de material fundido, processo fundamental para a geração e preservação das associações de fácies granulito, uma vez que a perda de material fundido desidrata e deixa o resíduo com composição refratária, inibindo a retrogressão durante o resfriamento, o que permite a preservação das paragêneses de temperatura ultra alta (*e.g.* Moraes *et al.*, 2002). Um sistema aberto com constante perda de líquido implica em mudança constante no volume de equilíbrio e composição da rocha durante a progressão do metamorfismo, assim a composição residual, isto é, a composição do granulito, não é apropriada para modelar a evolução progressiva do metamorfismo. Porém, a composição do resíduo/granulito é importante para o estabelecimento das condições do pico metamórfico e para modelar a evolução *P-T* retrogressiva (Kelsey, 2008). Recentemente, tentativas de reintegração de líquido nas composições residuais ganham força na literatura com grande potencial para modelar o caminho progressivo das rochas em fácies granulito (*e.g.* Johnson & Brown, 2004; White *et al.*, 2004; Diener *et al.*, 2008; Martinez, 2015).

2.3 Petrocronologia em rochas alto grau metamórfico

A petrocronologia é uma abordagem relativamente recente que tem revolucionado as geociências e consiste na obtenção de dados elementais e/ou isotópicos complementares associados a idades U-Pb (Kylander-Clark *et al.*, 2013).

O zircão se destaca no estudo de rochas de alto grau metamórfico pois:

- possui altas temperaturas de fechamento para o sistema U-Pb (Cherniak & Watson, 2001);
- possui baixa dissolução à baixas atividades de H₂O (Kirkland *et al.*, 2015);
- participa de diversas reações metamórficas (Bingen *et al.*, 2001) – diversas fases incorporam Zr durante o metamorfismo (*e.g.* granada, piroxênio, anfibólio, ilmenita, rutilo, etc);
- são capazes de blindar inclusões de reações metamórficas (Rubatto & Hermann, 2007a) – registrando diferentes estágios da trajetória *P-T*;
- e incorporam HFSE (*high-field strength elements*) (Hoskin & Schaltegger, 2003) – através de substituições simples de Zr por cátions de quadrivalentes (Hf, U, Th e

Ti) ou acopladas por cátions trivalentes (ETR, Y e Sc) associada à cátions pentavalentes (Nb, Ta e P).

A melhor maneira de caracterizar a estrutura interna dos cristais de zircão e suas variações, e assim identificar alguns processos metamórficos aos quais o cristal foi submetido é através de imageamento de catodoluminescência (CL) (Corfu *et al.*, 2003; Kylander-Clark *et al.*, 2013; Rubatto, 2017; Taylor *et al.*, 2016). De modo simplificado, existem apenas dois tipos de zircão metamórfico: aqueles preexistentes nos protólitos que sofreram modificações e aqueles que cresceram como resultado de reações metamórficas (Taylor *et al.*, 2016). Os processos de recristalização e dissolução agem total ou parcialmente em cristais preexistentes, enquanto processos de neocristalização podem ocorrer na borda de cristais ou mesmo formar novos grãos. Os processos de neocristalização são mais comuns pós-pico metamórfico, a partir da cristalização de líquidos formados pela fusão parcial (Taylor *et al.*, 2016), embora migrações locais de pequenos volumes de líquido possam cristalizar zircão durante o pico metamórfico (Harley, 2016; Möller *et al.*, 2003). Reações em estado sólido de formação de zircão, como por exemplo por quebra de rutilo ou granada, em geral estão associadas a descompressão pós-pico metamórfico (Bingen *et al.*, 2001).

Os principais indicadores petrológicos em zircão são as razões Th/U, a solubilidade de Ti, a incorporação de ETR e o sistema Lu-Hf. Em alguns trabalhos razões Th/U <0,1 são consideradas típicas de zircões metamórficos, enquanto as >0,1 são características de cristais ígneos (*e.g.* Hoskin & Schaltegger, 2003; Rubatto, 2017). Contudo, diversos dados na literatura mostram que esse limite não é real, principalmente para rochas de alto grau metamórfico (*e.g.* Kelly & Harley, 2005; Korhonen *et al.*, 2013; Kelsey & Hand, 2015; Yakymchuk *et al.*, 2018), e que os principais fatores que controlam as razões Th/U são a composição química da rocha e quais fases minerais estão presentes e relacionadas com a cristalização do zircão (Yakymchuk *et al.*, 2018). Th é altamente compatível com o líquido gerado em suprasolidus, uma vez que o elemento não se encaixa na estrutura cristalina de fases constituintes de rocha (Bea, 1996). Por outro lado, U, apesar de compatível com o líquido, pode ser alojado em fases constituintes de rocha, por exemplo micas, anfibólios, epídoto e feldspatos (Bea, 1996; Yakymchuk *et al.*, 2018). Desse modo, se zircão coexistir com fases ricas em Th, tais como monazita e allanita, a razão Th/U tende a ser maior a altas temperaturas, diminuindo durante o resfriamento, pois a cristalização das fases ricas em Th sequestram o elemento (Yakymchuk *et al.*, 2018). Contudo, a ausência de fases ricas em Th pode gerar uma relação inversa, uma vez que U será absorvido por fases constituintes de rocha, aumentando a razão Th/U do líquido e, conseqüentemente, as razões de cristais de zircão cristalizados a temperaturas menores. Pela baixa difusividade de Th e U em zircão é possível

registrar as variações das razões Th/U durante o metamorfismo em um único cristal (Cherniak *et al.*, 1997).

O aumento do conteúdo de Ti em zircão é proporcional ao aumento da temperatura (Watson *et al.*, 2006). Para o uso do termômetro Ti-em-zircão (Watson *et al.*, 2006) é necessário o estabelecimento da saturação de Ti e Si na rocha. Como ambos os elementos não ocorrem de forma pura na natureza, petrologicamente é estabelecido que a presença de rutilo (TiO₂) e quartzo (SiO₂) representa a saturação nesses elementos na rocha. A ausência de uma dessas fases na rocha, torna o cálculo de temperatura pouco confiável, uma vez que as atividades de Si ou Ti são escolhidas de forma “arbitrária”. Quando a utilização do termômetro é possível, as temperaturas fornecidas são, em geral, menores que as registradas no pico metamórfico, devido à supracitada cristalização a partir da cristalização da fusão parcial ou ainda pode estar relacionada à difusão de Ti no cristal em alta temperatura (Korhonen *et al.*, 2014).

Em geral, cristais de zircão mostram padrão de enriquecimento em ETR médios (ETRM) e pesados (ETRP) em relação aos ETR leves (ETRL) (Bea, 1996). Ortopiroxênio e, principalmente, granada são fases constituintes em rochas de metamorfismo de alto grau que possuem afinidade geoquímica com ETRM-P (Harley & Kelly, 2007) e portanto, durante a cristalização, competem com zircão pela incorporação desses elementos. Dados experimentais mostram que as altas temperaturas favorecem a partição dos ETRM-P para o granada em relação à zircão (Rubatto & Hermann, 2007b; Taylor *et al.*, 2015). Assim, padrões retilíneos em ETRM-P em zircão sugere que o sistema foi previamente empobrecido nesses elementos. Além disso, os coeficientes de partição de ETRP de zircão e granada tendem a se alinhar se os mesmos estiverem em equilíbrio químico (Taylor *et al.*, 2017). Apesar de não haverem trabalhos experimentais na literatura sobre a relação entre zircão e ortopiroxênio, ambos devem competir pela incorporação de ETRM-P, uma vez que os coeficientes de partição desses elementos também se alinham em certos casos (Taylor *et al.*, 2017).

O sistema isotópico Lu-Hf em zircão é um poderoso marcador de processos de evolução crustal em virtude do alto fracionamento entre Lu e Hf durante a geração de magmas, em que o Lu concentra-se no resíduo, enquanto Hf particiona para o líquido (*e.g.* Blichert-Toft & Albarède, 1997; Kinny & Maas, 2003; Belousova *et al.*, 2010). Porém, durante o metamorfismo o zircão pode registrar informações petrogenéticas importantes. O isótopo ¹⁷⁶Lu decai para ¹⁷⁶Hf, através de um decaimento β. Durante a cristalização o zircão incorpora mínimas quantidades de Lu comparadas às de Hf (Hoskin & Schaltegger, 2003), assim a razão Lu/Hf é próxima de 0, portanto admite-se que as razões ¹⁷⁶Hf/¹⁷⁷Hf são muito próximas ao líquido ao qual o grão foi cristalizado (Kinny & Maas, 2003). O sistema Lu-Hf no zircão praticamente não sofre distúrbios, sendo mais resistente inclusive que o sistema U-Pb (Kinny & Maas, 2003; Kirkland *et al.*, 2011). Assim, com aumento da

temperatura durante o metamorfismo o sistema U-Pb pode ser reiniciado, ao passo que o sistema Lu-Hf continua guardando as informações do protólito. Portanto, cristais que possuam bordas neocristalizadas a partir de líquido anatético e que apresentam a mesma razão $^{176}\text{Hf}/^{177}\text{Hf}$ que núcleos antigos herdados, devem ter sido cristalizadas em sistema isoquímico (Taylor *et al.*, 2016). Por outro lado, se as mesmas bordas são enriquecidas em razões $^{176}\text{Hf}/^{177}\text{Hf}$ comparadas aos núcleos, pode significar a interação com outras fontes mais radiogênicas ou ainda a quebra de granada (Kirkland *et al.*, 2011; Taylor *et al.*, 2016), mineral que é rico em Lu e portanto com altas razões $^{176}\text{Hf}/^{177}\text{Hf}$. Se o oposto ocorrer, bordas empobrecidas nas razões $^{176}\text{Hf}/^{177}\text{Hf}$ com relação ao núcleo, pode ser resultado de mistura de com fontes menos radiogênicas.

2.4 Fontes de calor para geração do metamorfismo das rochas de alto grau

Entender o contexto para a formação de rochas que atingem condições de fácies granulito e, especialmente, temperaturas ultra-altas, não é uma tarefa simples. A caracterização dessas rochas não é possível somente com trabalhos de campo, uma vez que o caráter antigo, as transformações deformacionais e, principalmente, erosão removem o contexto geológico da formação dessas rochas. Os estudos devem ser balizados pela definição de trajetórias metamórficas P - T , com diagramas de equilíbrio de fases e geocronologia. Isso porque as variações na temperatura e pressão são funções da fonte de calor, dos mecanismos de transferência de calor e taxas de erosão (Brown, 1993; England & Thompson, 1984; Harley, 1989; Kelsey & Hand, 2015).

Embora os ambientes tectônicos que atingem condições de temperatura ultra-alta serem relativamente incertos, esse tipo de metamorfismo é, atualmente, aceito como um tipo de metamorfismo regional e, relativamente, comum (Brown, 2007; Clark *et al.*, 2011; Harley, 2008; Kelsey, 2008; Kelsey & Hand, 2015). O metamorfismo de temperatura ultra-alta ocorre através de toda a história do planeta e aparentemente está ligado ao ciclo dos megacontinentes (Brown, 2007). Nesse cenário, os ambientes orogênicos, acrescionários e colisionais, devem ser os principais geradores desse tipo de metamorfismo. Desse modo, uma questão recorrente para o metamorfismo de temperatura ultra-alta é: quais as fontes de calor capazes de gerar temperaturas superiores à 900 °C em ambientes crustais?

Estudos mostram que difusividade térmica das rochas, isto é, a capacidade de difundir o calor entre materiais, diminui com o aumento da temperatura (Whittington *et al.*, 2009). Rochas crustais, a temperaturas variando entre 0 a 575 °C, apresentam valores entre 2 e 0,7 mm²/s (Whittington *et al.*, 2009), os quais correspondem à rochas das crostas superior e média, ao passo que o manto superior, a temperaturas próximas a 1300 °C (Clark *et al.*, 2011), a difusividade é de 0,3 mm²/s (Whittington *et al.*, 2009). Assim, a crosta inferior deve ter extenso campo espacial onde a difusividade térmica é baixa, entre 0,7 e 0,3 mm²/s, tornando-a um enorme isolante

térmico que necessita de alta energia para esquentar, porém, uma vez aquecida, a temperatura tende a manter-se alta por longos períodos (Kelsey & Hand, 2015).

Além da alta energia necessária para o aquecimento da crosta inferior, a geração de fundidos, pelo processo de fusão parcial, é outro fator que dificulta a elevação da temperatura acima de 900 °C, pois muito calor é consumido na produção da fusão (Kelsey & Hand, 2015). O metamorfismo de temperatura ultra-alta ocorre em condições suprasolidus em grande parte das rochas crustais, e as reações de fusão, que são endotérmicas, consomem calor e, conseqüentemente, têm efeito tampão na temperatura, sendo o aumento da temperatura impedido, enquanto grande volume de fundido não é gerado (Stüwe, 1995), ou até que o processo cesse por consumo total de um ou mais dos reagentes. Assim, para atingir condições de temperatura ultra-alta, as rochas devem ser “pré-condicionadas”, ou seja, terem longo período de fusão e extração de fundido, seja policíclico ou em ciclo único e/ou baixa fertilidade, antes de chegar a serem submetidas a tais condições de metamorfismo (Clark *et al.*, 2011).

Diversos trabalhos indicam que gradientes termais elevados na crosta podem ser resultado de advecção mantélica (*e.g.* Bohlen *et al.*, 1983; Sandiford *et al.*, 1987; Harley, 1989; Brown, 1993). Contudo, grande parte dos terrenos granulíticos/UHT não são caracterizados por expressivos volumes sin-metamórficos de intrusões de magmas basálticos (Brown & White, 2008; Clark *et al.*, 2011), sendo os terrenos assim observados exceções e não regra.

Nesse cenário três fontes de calor podem contribuir para o aquecimento da crosta inferior: elevada taxa de calor radiogênico, gerado por espessamento crustal, aquecimento mecânico gerado em zonas de cisalhamento e aquecimento gerado pelo manto astenosférico em zonas de afinamento crustal, em especial zonas de retro-arco (Clark *et al.*, 2011; Jamieson *et al.*, 1998; Jamieson & Beaumont, 2011; Nabelek *et al.*, 2010; Sizova *et al.*, 2014).

O decaimento radioativo de U, Th e, principalmente, K é reconhecido como a fonte geradora de calor mais importante da crosta continental (Clark *et al.*, 2011; Jamieson *et al.*, 1998). A influência do calor gerado por elementos radiogênicos depende da distribuição inicial desses elementos na crosta e como, durante processo de espessamento crustal, esses elementos se redistribuem. Por serem mais félsicas, é assumido que a crosta superior e média possuem maior capacidade de geração de calor do que a crosta inferior (Beaumont *et al.*, 2006; Clark *et al.*, 2011; Jamieson *et al.*, 1998; Jamieson & Beaumont, 2013; Kelsey, 2008). Contudo, estudo recente mostra que mesmo após fusão parcial e posterior extração do material fundido, as taxas de produção de calor no resíduo não são afetadas significativamente (Alessio *et al.*, 2018).

O primeiro e clássico modelo matemático para modelagem do metamorfismo crustal foi proposto por England & Thompson (1984). Nessa modelagem, realizado em uma dimensão (1D),

é considerada crosta radiogênica espessada e os parâmetros controladores do metamorfismo são a condutividade termal, a taxa de erosão, modos de espessamento crustal e do manto, sem deixar de ser avaliado o calor mantélico. Contudo o modelo pode ser considerado simplista, uma vez que os parâmetros são mantidos fixos ao longo do tempo, o espessamento da crosta é feito de maneira instantânea (modelo 1D) e a erosão só é iniciada 20 Ma após o espessamento. Mesmo assim, nunca são atingidas temperaturas superiores a 900 °C dentro da crosta continental em intervalo de tempo condizente com o que é observado nos cinturões metamórficos, ou seja, menores que 100 Ma.

Clark *et al.* (2011) realizaram modelo 1D com as mesmas condições iniciais de England & Thompson (1984). O parâmetro de condutividade térmica foi atualizado e não permanece fixo ao longo da modelagem, porém a erosão inicia-se 20 Ma após o espessamento, como no modelo original. Para atingir condições de temperatura ultra-alta, em profundidade crustais de ~ 40 km ou mais, foi necessária elevada taxa de produção de calor radiogênico na crosta superior, maior que $3,5 \mu\text{Wm}^{-3}$, uma vez que a média é $2 \mu\text{Wm}^{-3}$, baixa taxa de erosão, menor que 0,05 mm/ano, ignorar o efeito da fusão parcial, ou considerar uma rocha “pré-condicionada”, além de considerar ambiente tectônico grande e adequado para a crosta inferior ficar encubada por longos períodos (c. 120 My).

O calor mecânico gerado em zonas de cisalhamento pode ser um fator que contribua para aquecimento significativo da crosta (Clark *et al.*, 2011; Nabelek *et al.*, 2010). Contudo, a geração de calor por atrito está diretamente relacionada ao comportamento mecânico do material, que por sua vez muda conforme a temperatura. Isso quer dizer que quanto mais quente a rocha, menor o calor gerado por atrito, pois ela tende ao comportamento dúctil e sofrer mais estiramento, o que diminui o atrito. Nabelek *et al.* (2010) mostram que em clinopiroxenito a taxa de deformação plausível ($3 \times 10^{-13} \text{ s}^{-1}$) pode gerar calor suficiente para atingir temperatura ultra-alta em volumes expressivos de crosta. Contudo, típicas rochas crustais, de composição granítica, não têm essa capacidade, assim o calor gerado nas zonas de cisalhamento nas condições de fácies granulito seria apenas uma variável adicional e não um fator fundamental para geração de condições de temperatura ultra-alta (Clark *et al.*, 2011).

Modelos em duas dimensões (2D) que associam a influência do calor radiogênico e o gerado em zonas de cisalhamento têm colaborado muito para o entendimento sobre aquecimento e exumação da crosta (Gerya, 2014; Gorczyk *et al.*, 2015; Jamieson *et al.*, 1998; Jamieson & Beaumont, 2013, 2011; Sizova *et al.*, 2014). Isso porque a acreção de material radiogênico acontece lateralmente nos modelos, assim como ocorre durante a orogênese (Jamieson *et al.*, 1998). Contudo, os modelos para grandes orógenos não preveem a ocorrência de metamorfismo temperatura de ultra-alta (Jamieson *et al.*, 1998; Jamieson & Beaumont, 2013), a exceção de

quando o efeito da fusão é desconsiderado e a crosta inferior é considerada radiogênica (Jamieson & Beaumont, 2011). Mesmo assim, rochas que atingem condições de temperatura ultra-alta são encontradas em grandes orógenos, como por exemplo no Platô Tibetano (Hacker *et al.*, 2000).

As regiões de retro arco são locais de listosfera adelgada, com alto fluxo de calor mantélico e com temperaturas na linha de Moho elevadas (Clark *et al.*, 2011; Hyndman *et al.*, 2005). Segundo Brown (2006), o duplicamento crustal durante a orogênese, nesse ambiente tectônico, pode levar à condições de temperatura ultra-alta, pois a crosta é pré-aquecida, porém os estudos realizados em 1D não conseguem atingir essas condições (Clark *et al.*, 2011; Thompson *et al.*, 2001). Modelos em 2D que duplicam a crosta de regiões de retro arco conseguem prever condições de temperatura acima de 900 °C na crosta continental (Gorczyk *et al.*, 2015; Sizova *et al.*, 2014). Os modelos preveem a ocorrência de “*slab break-off*” raso e magmatismo sin-extensional, com amplo volume de fusão parcial na crosta e baixo volume de fusão parcial do manto.

Em todos os cenários e modelos estudados para a geração de temperaturas acima de 900 °C na crosta continental, um estágio longo de residência mostra-se necessário. As trajetórias *P-T* tendem a ser horárias, ou seja, com pico térmico posterior ao pico bórico, e com resfriamento isobórico, ou quase isobórico, sendo pouco previsíveis as trajetórias de descompressão isotérmica (Clark *et al.*, 2011; Harley, 2008; Jamieson *et al.*, 1998; Kelsey, 2008; Kelsey & Hand, 2015). Porém, exemplos reais com trajetória de descompressão isotérmica não são incomuns (*e.g.* In Ouzal, Argélia; Ghats Oriental, Índia; entre outros) e elas devem refletir os estágios finais da colisão (Harley, 1998).

2.5 Exumação de rochas de alto grau

Dois modelos modernos de exumação de rochas da base da crosta continental, granulito e migmatitos, são amplamente discutidos na literatura, um controlado por fluxo de canal em crosta homogênea (em inglês *channel flow*) e outro controlado por fluxo dúctil em crosta heterogênea (Beaumont *et al.*, 2010, 2004; Godin *et al.*, 2006; Jamieson *et al.*, 2011, 2007, 2004; Jamieson & Beaumont, 2011; Warren *et al.*, 2008).

O modelo de fluxo de canal (Beaumont *et al.*, 2004; Godin *et al.*, 2006; Jamieson *et al.*, 2011; Jamieson & Beaumont, 2011) descreve fluxo prolongado de uma camada menos viscosa da crosta terrestre, viscosidade atingida por fusão parcial, geralmente em crosta média, que flui lateralmente delimitada por camadas mais rígidas, que não atingiram condições de fusão parcial, geralmente a crosta superior e inferior, devido a alto gradiente horizontal de pressão litostática (Figura 2-2). Para o modelo funcionar é necessária crosta lateralmente homogênea. O melhor

exemplo recente de fluxo de canal é o Sistema Himalaia-Tibete (Godin *et al.*, 2006; Jamieson & Beaumont, 2011).

Figura 2-2 Exumação por fluxo de canal. Duas camadas rígidas expulsão lateralmente uma camada viscosa. Modificado de Jamieson & Beaumont (2011).

Os modelos de crosta lateralmente heterogênea e relativamente rígida não atingem viscosidade efetiva baixa o suficiente para o fluxo de canal ser guiado por gravidade (Jamieson *et al.*, 2010, 2007, Jamieson & Beaumont, 2013, 2011). Em vez de fluxo lateral impulsionado tectonicamente, a expulsão da crosta inferior é desencadeada pela colocação de uma cunha rígida e fria. As estruturas dominantes são *nappes* dúcteis que podem ser transportadas centenas de quilômetros em direção ao ante-país (Figura 2-3A). Quando a convergência tectônica termina, ocorre colapso gravitacional que se espalha, em que crosta média e inferior dúctil, que fluem lateralmente, a partir do platô para o ante-país, sendo o núcleo orogênico afinado e ocorrendo cavalgamento nos flancos do orógeno (Figura 2-3B). A geometria resultante pode se assemelhar a produzida pelo fluxo dúctil sin-convergente, exceto para o desenvolvimento de zonas de cisalhamento normais de escala regional na fase pós-convergente (Jamieson *et al.*, 2010, 2007, Jamieson & Beaumont, 2013, 2011). Exemplos dessa tectônica de extrusão são encontrados no Greenville ocidental (Jamieson *et al.*, 2010, 2007) e na porção oriental do Sistema de *Nappes* Andrelândia (Motta & Moraes, 2017).

A comparação entre os dois modelos de exumação sugere que rochas de alto grau metamórfico no núcleo orogênico registram prolongados efeitos de fluxo dúctil sin- e pós-convergência, enquanto as rochas do flanco do orógeno registram, principalmente, cavalgamento e extensão pós-convergência (Jamieson & Beaumont, 2011).

Figura 2-3 Exumação de crosta heterogênea. (A) Estágio sin-convergente mostra empilhamento de nappes e fluxo de canal incipiente. (B) Estágio de colapso orogênico com cisalhamento normal e inverso no flanco orogênico e extensão no núcleo. Modificado de Jamieson & Beaumont (2011).

2.6 Referências bibliográficas

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3 CONTEXTO GEOLÓGICO

3.1 Faixa Brasília

A Faixa Brasília juntamente com as Faixas Araguaia e Paraguai fazem parte do Sistema Orogênico Tocantins, neoproterozóico, resultante da sucessão de convergência entre blocos crustais, cujos núcleos geraram as várias regiões cratônicas, culminando com a formação do megacontinente Gondwana (Campos Neto, 2000; Fuck *et al.*, 1993; Hasui, 2012; Valeriano *et al.*, 2008, 2004).

O Sistema Orogênico Tocantins situa-se entre os crátons Amazônico, a oeste, e São Francisco, a leste, com interferência de um terceiro bloco, o Cráton Paranapanema, a sul (Brito Neves *et al.*, 1999; Valeriano *et al.*, 2008). A Faixa Brasília (Figura 3-1) delimita toda a parte oeste do Cráton São Francisco e é dividida em duas partes (Valeriano *et al.*, 2004): Faixa Brasília Setentrional, de orientação NE, e Faixa Brasília Meridional, de orientação NW. Os segmentos são separados pela Megaflexura dos Pirineus, que funciona como um divisor de tendências estruturais com sobreposição da trama estrutural da porção norte sobre a porção sul, indicando diacronismo da evolução de ambos os segmentos e, portanto, indicando que a Faixa Brasília Meridional é mais jovem (Araújo Filho, 2000; Dardenne, 2000; Uhlein *et al.*, 2012; Valeriano *et al.*, 2008).

3.2 Orógeno Brasília Meridional

Na extremidade sul da Faixa Brasília a colisão das placas Sanfranciscana e Paranapanema culminou com o fechamento do paleo-oceano Goianides, no período Toniano do Neoproterozoico (Campos Neto & Caby, 2000). Atualmente o orógeno é representado por cinturão de dobramento em forma de cunha, sendo estruturado como pilha de *nappes*, com transporte de topo de WSW para ENE, conhecido como Orógeno Brasília Meridional (Campos Neto *et al.*, 2007).

O Orógeno Brasília Meridional foi inicialmente reconhecido como terreno alóctone, correspondendo às zonas internas de alto grau metamórfico localizada ao sul do Cráton São Francisco (Ebert, 1968). A hierarquia das fases de deformação e interpretação da cinemática resultaram em incertezas quanto a proveniência do terreno, mas atualmente a grande maioria dos trabalhos admitem que o terreno pertence à porção meridional do Orógeno Brasília, colocada sobre o Cráton do São Francisco e afetada, ou não posteriormente, pela deformação relacionada à Faixa Ribeira (Campos Neto & Caby, 1999; Ribeiro *et al.*, 1995).

Figura 3-1 Unidades tectônicas da Faixa Brasília (Valeriano et al., 2004). Zona externa de 1 a 4, em que: 1 e 2 – Terrenos granito-greenstone; 3 – Rift paleoproterozóico (Grupo Arai); 4 – Sistema de empurrão Ilícinia-Piumhi. Margem passiva de 5 a 8, em que: 5 – Grupo Paranoá; 6 – Unidades quartzíticas-fílicas (Canastra e Andrelândia); 7 – Grupo Vazante; 8 – Grupo Ibiá. Zona interna de 9 a 14, em que 9a e b – Grupos Araxá e Andrelândia; 10 – Nappes Granulíticas; 11 e 12 – Complexos greenstone arqueanos/paleoproterozóico; 13 – Sequência Vulcano-sedimentar mesoproterozóica (Juscelândia, Indaiaópolis e Palmeirópolis); 14 – Complexos máficos-ultramáficos (Barro Alto, Niquelândia e Cana Brava). Arco magmático de Goiás de 15 a 16, em que: 15 – Sequência supracrustal; 16 – Ortognaisses e granitoides. MR – Arco de Mara Rosa. AA – Arco de Arenópolis. AI – Complexo Anápolis-Itaçu. GD – Domínio Guaxupé (Nappe Socorro-Guaxupé). SD – Domínio Socorro (Nappe Socorro-Guaxupé). O quadrado amarelo indica o Orógeno Brasília Meridional (Figura 3-2).

O estágio colisional na extremidade sul do Orógeno Brasília ocorreu antes de 630 Ma, quando as rochas da *Nappe* Socorro-Guaxupé cavalgaram as da Placa Sanfranciscana, com transporte tectônico de topo para E, e metamorfismo de pressões elevadas. No Orógeno Ribeira, entre de 580 e 560 Ma, ocorreu encurtamento crustal com transporte para NNW, com dobramento em escala regional e metamorfismo de pressão intermediária. Os efeitos da implantação do Orógeno Ribeira funcionaram como uma segunda colisão sobre o Orógeno Brasília, resultando em complexa zona de interferência entre os dois orógenos, segundo alguns autores (Heilbron *et al.*, 2004; Peternel *et al.*, 2005; Trouw *et al.*, 2013, 2000a, 2000b).

Alternativamente, o Orógeno Brasília Meridional é entendido em um único evento colisional contínuo, do tipo himalaiano e formando por complexo sistema de *nappes* (Campos Neto *et al.*, 2011, 2004; Campos Neto & Caby, 1999). Assim, o Orógeno Brasília Meridional consiste em pilha colisional de *nappes*, com cunha orogênica resultante mergulhando para S-SW, em três ambientes principais, de WSW para ENE (Figura 3-2): domínio de arco magmático desenvolvido na margem continental ativa da placa Paranapanema, *Nappe* Socorro-Guaxupé, domínio continental subductado, Sistema de *Nappes* Andrelândia, e domínios com afinidade de margem passiva relacionados à Placa Sanfranciscana, Sistema de *Nappes* Carrancas e Lima Duarte.

3.3 *Nappe* Socorro-Guaxupé

A *Nappe* Socorro-Guaxupé é pilha alóctone, com cerca de 10 km de espessura, que representa seção de crosta continental profunda e parcialmente fundida (Campos Neto & Caby, 2000). A estrutura é segmentada em dois lobos (Del Lama *et al.*, 2000; Janasi, 2002), Domínio Guaxupé a norte, e Domínio Socorro a sul, que são separados por rampas laterais de alto ângulo em que afloram rochas pertencentes ao Sistema de *Nappes* Andrelândia, além de rochas do embasamento cratônico (Cioffi, 2016).

A litoestratigrafia da *Nappe* Socorro-Guaxupé é subdividida em três unidades (Campos Neto *et al.*, 2004; Campos Neto & Caby, 2000):

- Unidade granulítica basal: formada por granulito-gnaisses bandados enderbíticos a charno-enderbíticos caracterizados pela associação granada-clinopiroxênio-ortopiroxênio-plagioclásio-quartzo em rochas máficas e granada-sillimanita-cordierita-espinélio-biotita-plagioclásio-quartzo em rochas metapelíticas. Esses granulitos desenvolvem leucossoma, interpretado como fusão anidra *in situ*, além possuir veios de charnockito hololeucocrático. Ortognaisses migmatíticos gabro-noríticos intercalam-se aos granulitos, principalmente próximo ao topo da unidade;
- Unidade diatexítica intermediária: compreende migmatitos estromáticos descontínuos, em que predominam biotita-hornblenda-clinopiroxênio gnaiss de composição diorítica-

tonalítica. Esses migmatitos são envolvidos por biotita-hornblenda gnaisses/granitos nebulíticos e granitos porfíricos deformados. No topo da unidade ocorrem (biotita)-hornblenda tonalitos a gnaisses granodioríticos com lentes de rochas metabásicas. Ainda podem ocorrer lentes ou enclaves de granulitos metassedimentares com a associação granada-sillimanita-cordierita;

- Unidade metatexítica superior: predominam migmatitos aluminosos de origem metassedimentar com grau de anatexia diminuindo para o topo da unidade, com leucossoma granítico com muscovita. Ocorrem lentes de quartzito, gnaiss calcissilicático, rocha metamáfica e, mais raramente, mármore.

Figura 3-2 Mapa geológico do Orógeno Brasília Meridional (Campos Neto *et al.*, 2011).

Idades em núcleos de zircão espalham-se entre ca. 800 Ma até 630 Ma e são interpretadas como herança magmáticas (Hackspacher *et al.*, 2003; Janasi, 1999; Martins *et al.*, 2009; Mora *et al.*, 2014; Rocha, 2016; Rocha *et al.*, 2017; Tedeschi, 2018).

A *Nappe* Socorro-Guaxupé tem o início da evolução em 670-640 Ma, com subducção inclinada e geração de rochas cálcio-alcálicas, com colisão entre 625-610 Ma, com geração metatexitos e diatexitos, de alto potássio, onde o arco migrou primeiramente para o ante-arco e depois em direção ao interior da placa superior (Mora *et al.*, 2014; Mora, 2013). O gradiente termal observado dentro da *nappe* é análogo a ambientes de margem ativa com manto astenosférico raso (Iyer *et al.*, 1996), podendo, então, ser relacionado à raiz profunda de um arco magmático (Campos Neto *et al.*, 2004). Contudo, rochas juvenis ainda não são reconhecidas, e frequentemente possuem ϵ_{Nd} e ϵ_{Hf} negativos e idades modelo (Nd) entre 1,6 e 1,8 Ga (Janasi, 2002; Mora *et al.*, 2014; Vinagre *et al.*, 2014). A ausência de rochas juvenis têm sido atribuída à presença de manto litosférico enriquecido (Janasi, 1999; Rocha *et al.*, 2018). Porém, recentemente idades Neoarqueanas foram reportadas e sugerem que o processo de assimilação-contaminação também agiu na geração do arco, que teria sido implantado em crosta antiga relacionada às rochas da Placa Paranapanema (Tedeschi, 2018).

O auge do metamorfismo granulítico é concomitante às idades da colisão, entre 625-610 Ma, e o registro de resfriamento e cristalização de fundidos espalha-se até 590 Ma (Rocha *et al.*, 2018, 2017; Tedeschi, 2018).

Os cálculos termobarométricos demonstram que há significativa variação de pressão e temperatura dentro da unidade granulítica com aumento de pressão de sul para norte (Alexandre, 2013; Zanardo *et al.*, 2006). No topo, as condições calculadas para o pico metamórfico são de 820 °C e 4,5 kbar, para a associação granada, cordierita, sillimanita, biotita, plagioclásio, feldspato potássio e quartzo (Vasconcellos *et al.*, 1991); na porção intermediária, condições de 850 °C e 7,5 kbar foram calculadas com a associação granada, cordierita, sillimanita, biotita, plagioclásio, espinélio e feldspato potássio (Campos Neto & Caby, 2000); e, por fim, na base as condições da paragênese granada, clinopiroxênio, plagioclásio e quartzo fornecem condições de 850 °C e 8,5 kbar (Iyer *et al.*, 1996). Outros cálculos realizados com rochas da base da *Nappe* Socorro-Guaxupé sugerem condições ainda mais extremas de temperatura e pressão. A partir da associação granada, clinopiroxênio, hornblenda, biotita, plagioclásio e quartzo condições de 900 °C e 12,5 kbar foram calculadas (Del Lama *et al.*, 1994) e 1040 °C e 14,4 kbar para rochas de região de Guaxupé (Del Lama *et al.*, 2000). Para granulito com granada, clinopiroxênio, ortopiroxênio, plagioclásio e quartzo temperaturas de 890 °C e pressões entre 11,5 a 14 kbar são calculadas em Varginha (Campos Neto & Caby, 2000; Nascimento & Oliveira, 2012). Condições de alta temperatura também são encontradas nas unidades migmatíticas da *nappe*, atingindo ~ 1030 °C (Rocha *et al.*, 2017).

Grandes suítes magmáticas, sin-orogênicas, se alojam nas unidades principais da *nappe*. Ocorrem suítes mangeríticas-charnockíticas, tais como São Pedro de Caldas, Divinópolis e

Paraguaçu (Barros, 2012; Janasi, 2002), que são interpretadas como originadas pela fusão anidra dos granulitos da *nappe*, datados em 622 Ma (Janasi, 2002), enquanto biotita granitos do tipo Pinhal, datados em 625 Ma (Janasi, 1999) e granada-biotita granitos tipo Nazaré Paulista, datados em 625 Ma (Martins *et al.*, 2009), foram gerados a partir de anatexia em presença de água.

O magmatismo tardi-orogênico, com idades entre 610 e 580 Ma, está associado a zonas de transcorrências de direção predominante NE-SW, nas quais ocorrem granitos cálcio-alcalinos potássicos, Plúton Jaguari e Maciço Serra do Lopo, sienogranitos rapakivi, Batólito Itu, e dioritos do Maciço Piracaia (Janasi, 1999).

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4 GRANULITOS DO DOMÍNIO GUAXUPÉ

Esse capítulo apresenta resumo do manuscrito sobre as condições metamórficas e temporais aos quais os granulitos do Domínio Guaxupé foram submetidos. Para tal, dados de termobarometria e modelamento metamórfico através de pseudosseções é combinado com dados petrocronológicos em zircão, que incluem cronologia U–Pb, isotopia Lu–Hf e comportamento de ETR. O manuscrito intitulado “*Recovering P–T–t paths from ultra-high temperature (UHT) rocks that lack diagnostic mineral assemblages: A felsic granulite example from the Southern Brasília Orogen of Brazil*” será submetido ao periódico *Journal of Metamorphic Geology* e é encontrado na íntegra no ANEXO I, sendo sua leitura fundamental para o entendimento dessa tese.

4.1 Resumo

O recente desenvolvimento de modelos termodinâmicos para composições de atividades de líquido trondhjemitico-tonalítico tornou possível, pela primeira vez, quantificar a história metamórfica de terrenos dominados por ortognaisses félsicos e máficos que não apresentam típicas associações minerais diagnósticas de condições de pressão e temperatura (P – T). Nesse capítulo, composição química mineral e de rocha total, modelo termodinâmico de equilíbrio de fases e isotopia U–Pb–Hf são ferramentas utilizadas para inferir a história P – T –tempo de granada granulito félsico e granulitos máficos da *Nappe* Socorro-Guaxupé no Orógeno Brasília Meridional. Essas rochas ocorrem em pilha de *nappes* Neoproterozóicas associadas a formação da parte ocidental do paleocontinente Gondwana. A assembleia de pico metamórfico no granulito félsico, composta por orto e clinopiroxênio, granada, feldspato potássio, plagioclásio, quartzo, ilmenita e leucossoma atingiu o pico metamórfico em ~ 1000 °C à 11 kbar, registrado tanto pelo alto conteúdo de Al no ortopiroxênio, quanto no modelo de equilíbrio de fases. Fases hidratadas não estavam presentes no pico metamórfico e a geração de líquido anatético foi controlada pela quebra de feldspato potássio. O caminho retrogressivo de resfriamento isobárico levou a parcial rehidratação da paragênese de pico, marcado por coroas de hornblenda envolvendo piroxênio. Núcleos de cristais de zircão com zoneamento setorial de granulito félsico fornecem datações U–Pb de *c.* 625 Ma, interpretado como idade do metamorfismo de alta temperatura. Em uma amostra, bordas de zircão que envolvem esses núcleos e cristais *soccer ball* neoformados possuem altos teores de U e fornecem idades de *c.* 615 Ma que marcam o início da cristalização da fusão parcial, enquanto bordas de zircão e cristais *soccer ball* em outras duas amostras fornecem datações de *c.* 595–600 Ma, interpretada como idade final da cristalização de leucossoma. Os dados revelam um longo evento metamórfico, no qual a crosta continental inferior esteve em condições de fácies granulito por pelo menos 30 Myr, consistente com o desenvolvimento regional de condições de temperatura ultra alta (UHT) em ambientes colisionais.

5 CHARNOKITOS DA SUÍTE PARAGUAÇU

Esse capítulo apresenta o resumo do manuscrito em que é discutido a petrogênese de charnockitos da Suíte Paraguaçu através de uma nova abordagem utilizando modelamento termodinâmico da possível fonte para acessar ao produto. O estudo é validado através de geoquímica elemental e isotópica em rocha total e dados isotópicos U–Pb–Hf em zircão. Também são discutidas as fontes de calor envolvidas para a geração de temperaturas alta na crosta, além de propor uma nova discussão sobre a origem da *Nappe* Socorro-Guaxupé. O texto intitulado “*The origin of charnockites from granulite partial melting: an example in Socorro-Guaxupé Nappe, southeastern Brazil*” será submetido ao periódico *Precambrian Research* e pode ser encontrado na íntegra no ANEXO II, sendo sua leitura fundamental para o entendimento dessa tese.

5.1 Resumo

Uma grande exposição de crosta continental inferior é exposta na porção mais sul do Orógeno Brasília, em que rochas da *Nappe* Socorro-Guaxupé foram metamorfizadas em condições de temperatura ultra alta (UHT). O metamorfismo registra uma longa história de anatexia entre 630 e 595 Ma, em que diversos corpos graníticos foram gerados, incluindo charnockitos que apresentam idades de cristalização próximas ao metamorfismo. Uma dessas suítes charnockíticas, Suíte Paraguaçu (c. 630–611 Ma), é subdividida em dois corpos, norte e sul, e apresenta composição de rocha intermediária com alto conteúdo de #Mg, Ba, Zr, Hf e ETR e empobrecimento em Sr, Th e U, sugerindo fonte empobrecida em que feldspato potássio, zircão e granada foram solubilizados na fonte. Modelagem termodinâmico metamórfica de granada granulito félsico, da Unidade Granulítica, mostra que feldspato potássio, zircão e granada estavam quebrando durante o estágio de UHT e o líquido silicático gerado apresenta comportamento de elementos maiores similares aos encontrados na Suíte Paraguaçu. Contudo, as rochas da Suíte Paraguaçu possuem teores maiores de Na₂O e menores de SiO₂ que aqueles previstos no modelamento termodinâmico. Desse modo, é possível que a o granulito félsico serviu como fonte principal para a geração da Suíte Paraguaçu, através da migração e a cristalização do líquido anatético gerado pela fusão parcial, com alguma mistura de outra(s) fonte(s), provavelmente mais radiogênicas, evidenciado pela valor de EHf menos negativo em cristais de zircão do corpo norte, que também é enriquecido em Sr e Ti. A mistura com fontes radiogênicas associado ao decaimento radioativo de grande volume de fusão parcial nesse segmento da *Nappe* Socorro-Guaxupé foram responsáveis para o aquecimento à condições de UHT.

6 RELAÇÃO ENTRE GRANULITOS E MIGMATITOS

Esse capítulo apresenta o resumo do manuscrito sobre as diferenças estruturais entre as rochas das Unidades Granulíticas e Diatexíticas na região estudada. Além da coleta de dados de foliação e lineação em campo, o estudo conta com medição do ângulo de abertura do eixo-c de quartzo e química mineral, que juntos forneceram condições P - T de pico metamórfico e da deformação em estado sólido. O manuscrito intitulado "*Relationship between metamorphic peak and deformation temperature in migmatites and granulites and their constrain on exhumation in the Socorro-Guaxupé Nappe, SE Brazil*" será submetido ao periódico *Journal of South American Earth Sciences* e pode ser encontrado na íntegra no ANEXO III, sendo sua leitura fundamental para o entendimento dessa tese.

6.1 Resumo

No sudeste do Brasil, uma espessa seção Neoproterozóica de crosta continental inferior aflora na *Nappe* Socorro-Guaxupé, porção mais a sul do Orógeno Brasília, Migmatitos, principalmente diatexitos, e granulitos que atingiram condições de temperatura ultra alta durante o pico metamórfico, em torno de 1000 °C. Contudo, ambas as rochas apresentam diferentes padrões estruturais. A foliação nos migmatitos é mais espalhada na direção NW-SE, enquanto que os granulitos a foliação é mais concentrada com direção WNW-ESE. A lineação nos migmatitos é definida por feldspatos e minerais hidratados, tais como hornblenda e biotita, com direção WNW-ESE, ao passo que nos granulitos a lineação é marcada por minerais anidros, como feldspatos e piroxênios, e apresenta direção WSW-ENE. O ângulo de abertura das tramas de eixo-c de quartzo no resíduo dos granulitos apresentam os maiores valores (135-154°), seguido pelo leucossoma dos granulitos (113-130°) e pelo leucossoma dos migmatitos (86-97°). Isso implica que as tramas de eixo-c de quartzo nos granulitos marcam duas temperaturas: uma próxima ao pico metamórfico registrada no resíduo, entre 869 e 923 °C, e outra próxima a temperatura de cristalização do leucossoma, entre 796 e 853 °C. Nos migmatitos a temperatura obtida pela trama de eixo-c de quartzo é inferior, entre 633 e 685 °C. Portanto, a deformação em estado sólido nos granulitos ocorreu em temperaturas mais altas do que nos migmatitos, sendo que o último permaneceu em condições suprasolidus por uma grande intervalo de temperatura (> 300 °C). Assim, quando os granulitos solidificaram, os migmatitos permaneceram inconsolidados, marcando diferentes geometrias de deformação.

7 DISCUSSÕES

7.1 Metamorfismo na Unidade Granulítica

Os dados das amostras VAB01 e PAR03, respectivamente granada granulito félsico e máfico foram combinados para descrever a trajetória *P-T-t* da base da *nappe*. Grãos de biotita com formas arredondadas inclusos em granada contém alto teor de Ti e alto X_{Mg} . Assim, os estágios iniciais de fusão parcial por desidratação deve ter acontecido por reações de consumo de biotita. Contudo, para atingir condições de temperatura ultra-alta e preservar a associação mineral de pico metamórfico é necessário que a rocha tenha perdido esse fundido gerado por desidratação de biotita (Clark *et al.*, 2011; Kelsey & Hand, 2015; Vielzeuf *et al.*, 1990) e, conseqüentemente, ter se tornado menos propensa ao tamponamento termal causado pela fusão parcial. Assim, o modelamento termodinâmico apresentado não reflete a trajetória progressiva do metamorfismo, contudo as condições de pico metamórfico e a trajetória regressiva são contempladas pelo modelo.

De acordo com a modelagem, a fusão foi controlada pela quebra de feldspato potássio e granada, uma vez que minerais hidratados foram consumidos anteriormente. O granada granulito félsico é extremamente empobrecido em Th e U, além de ser empobrecidos também em Rb e Y, o que reforça a ideia perda prévia de fundido. Leves anomalias positivas de Sr e Ti sugerem que plagioclásio e fases portadoras de Ti não foram intensamente envolvidas como reagentes das reações de fusão.

Os campos das associações minerais de pico metamórfico calculados nas pseudosseções para o granada granulito félsico (VAB01) e o granada granulito máfico (PAR03) se sobrepõem e são consistentes com temperaturas em torno de 1000 °C e pressões acima de 9 kbar. Os cálculos termobarométricos no modo *average P-T* do THERMOCALC apresentam condições similares para o pico metamórfico, coincidentes com as estimativas baseadas na quantidade de Al no sítio tetraédrico em ortopiroxênio do granada granulito félsico.

A trajetória retrometamórfica começa com leve queda na pressão até 9 kbar, em que há discreta substituição por ortopiroxênio em bordas de granada e magnetita é exsolvida da ilmenita. Nesse momento, a rocha segue trajetória de resfriamento quase isobárico, no campo de estabilidade da granada, uma vez que coroas ou outras texturas de substituição em granada são raras e quando presentes, mal desenvolvidas. Clinopiroxênio forma coroas nas bordas de ortopiroxênio tanto no granada granulito félsico quanto no máfico. Em temperaturas próximas ao *solidus*, hornblenda cristaliza-se nas bordas dos piroxênios, especialmente no leucossoma, a temperaturas de 870 °C e pressões de 8,5 kbar.

7.2 Ligando trajetória *P-T* ao tempo: petrocronologia de zircão

Os dados geocronológicos revelam pelo menos quatro grupos de idades que podem ser relacionados com períodos distintos de cristalização de zircão: um relacionado ao magmatismo no começo da orogênese; um relacionado à trajetória progressiva e outros dois relacionados à cristalização de fusão parcial. Embora o granada granulito félsico (VAB01 e VAB19) e o diatexitito (RGM55) não tenham o mesmo protólito, algumas comparações podem ser realizadas.

Nas amostras VAB01 e VAB19, cristais de zircão com zoneamento oscilatório exibem alto conteúdo de U e baixas razões Th/U, comparados às suas bordas e cristais *soccer ball*, e apresentam datações mais antigas, entre *c.* 630 até *c.* 715 Ma. Diagramas de partição de ETR entre zircão e granada, propostos por Taylor *et al.* (2017), mostram que esses núcleos de zircão não cresceram em equilíbrio com granada. Núcleos de zircão com zoneamento oscilatório ou setorial da amostra RGM55 também apresentam datações espalhadas entre *c.* 631 até *c.* 660 Ma, embora eles apresentem assinaturas de ETR e Hf diferentes dos anteriores.

Desse modo, essas datações mais antigas nas três amostras devem refletir a idade de cristalização ígnea do protólito e não têm relação com o metamorfismo, como observado em outros estudos (Janasi, 2002, 1999; Martins *et al.*, 2009; Mora *et al.*, 2014; Rocha *et al.*, 2018, 2017; Tedeschi *et al.*, 2018).

No diatexitito (RGM55), cristais de zircão subeudrais, prismáticos e com zoneamento setorial registram datação de $625,9 \pm 3,5$ Ma. Baseado no padrão de enriquecido em ETR leves da rocha, presença de allanita e pouca quantidade de granada, e nos padrões de ETR em zircão, com pronunciadas anomalias positivas de Ce e negativas de Eu, a amostra não tem o mesmo protólito que o granada granulito félsico. Desse modo, a datação de *c.* 625 Ma pode representar a idade da cristalização do protólito, embora existam núcleos mais antigos (*>* 630 Ma) como previamente mencionado. Outra possibilidade é que a datação represente a época de recristalização de material herdado, processo que resulta na expulsão do Pb pré-existente e reiniciam o sistema U-Pb no grão (Taylor *et al.*, 2016). Grãos desse domínio apresentam progressiva perda das feições internas originais, como texturas oscilatórias fantasmas, processo comumente observado em cristais recristalizados em condições de alto grau metamórfico (Taylor *et al.*, 2016). A datação de 625 Ma é similar à idades relacionadas ao pico metamórfico *Nappe* Socorro-Guaxupé (Rocha *et al.*, 2018, 2017; Tedeschi *et al.*, 2018).

As bordas de zircão e cristais sem forma ou *soccer ball* da amostra RGM55 apresentam os maiores conteúdo de U (até 347 ppm) e as menores razões Th/U ($\sim 0,23$). Esse grupo registra idades de $615,8 \pm 3,4$ Ma que representam o início da cristalização dos fundidos gerados na fusão parcial, uma vez que a dissolução de grãos de zircão pré-existent nos protólitos e de outras fases

portadores de U pode, em estágio inicial, enriquecer o conteúdo de U no líquido (Watson & Harrison, 1983; Yakymchuk *et al.*, 2018). Em comparação com seus núcleos, são empobrecidos em Th e possuem maior espalhamento no conteúdo de ETR leves, sugerindo que allanita, mineral portador desses elementos (Hermann, 2002), também foi cristalizada a partir do líquido. Isso é possível, pois a alta abundância de ETR leves na composição total da rocha é alta, o que pode estabilizar allanita em temperaturas tão altas (Hermann, 2002). Idades similares também interpretadas como cristalização de fundido são observadas em estudos prévios (Martins *et al.*, 2009; Mora *et al.*, 2014; Rocha *et al.*, 2018, 2017).

As bordas de cristais de zircão e cristais *soccer ball* das amostras VAB01 e VAB19 marcam os menores teores de U, as maiores razões Th/U e as idades mais jovens. Fases constituintes de rochas são capazes de incorporar mais U que Th (Yakymchuk *et al.*, 2018), portanto, as razões Th/U no fundido tendem a aumentar durante a cristalização se não houver nenhuma fase portadora de Th. Desse modo, esse grupo reflete a cristalização final dos fundidos gerados na fusão parcial, em que a presença de U era baixa.

Cristais de zircão desse domínio na amostra VAB01 cresceram em equilíbrio com granada, como pode ser observado em diagrama D_{REE} entre zircão e granada (Taylor *et al.*, 2017), o que é suportado pelo padrão planar de distribuição de ETR pesados tanto em granada quanto em zircão. Apesar de não haver distinção entre as idades, as análises nas bordas do zircão podem representar grãos que cristalizaram antes, em pressões mais elevadas, uma vez que os dados encaixam com dados experimentais em alta pressão (Rubatto & Hermann, 2007), enquanto os cristais *soccer ball* alinham-se a experimentos em pressões mais baixas (Taylor *et al.*, 2015).

O domínio de baixo U na amostra VAB19 apresenta empobrecimento em ETR médios e pesados em relação à granada. Os dados em diagrama D_{REE} entre zircão e granada mostram deslocamento, o que indica que os minerais não cresceram em equilíbrio. Nesse caso, granada deve ter sido cristalizada previamente, uma vez que ela mostra enriquecimento em ETR pesados, ao passo que zircão apresenta um padrão planar. ETR pesados são fortemente particionados para granada (Chen *et al.*, 2010), assim a granada havia sequestrado previamente esses elementos no momento da cristalização de zircão.

A trajetória de resfriamento isobárico discutida e a modelagem termodinâmica mostram que a proporção modal de granada aumentou durante o resfriamento. Portanto, as datações registradas no grupo de zircão com baixo U (c. 595-600 Ma) representam o final da cristalização do material fundido. Sendo o início da cristalização da fusão parcial em c. 615 Ma, a duração do evento de anatexia durou pelo menos 20 Myr.

7.3 Petrogênese da Suíte Paraguaçu

O modelamento termodinâmico para o granada granulito félsico mostra que feldspato potássio e granada eram as fases constituintes da rocha que estavam sendo consumidas durante o pico metamórfico, assim como zircão e apatita, enquanto plagioclásio e ortopiroxênio aumentavam a proporção modal junto com a geração do fundido. O líquido silicático gerado pelo modelo possui composição similar aos charnockitos da Suíte Paraguaçu, particularmente em FeO, MgO, CaO e K₂O.

Os charnockitos estudados apresentam teores de Fe₂O₃ + MgO + MnO + TiO₂ entre 6 e 10%, superiores aos observados em estudos experimentais em granitos anatéticos típicos (Patiño Douce & Beard, 1995). Assumindo uma fonte totalmente crustal, a quebra da biotita, mineral rico nesses elementos, pode gerar cerca de 4,5% de Fe₂O₃ + MgO + MnO + TiO₂. Devido ao caráter anidro e aos altos valores de Fe₂O₃ + MgO + MnO + TiO₂, parece improvável que esses magmas charnockíticos se formaram apenas através de reações de desidratação da biotita. Desse modo, os altos teores dos elementos citados requerem uma fonte enriquecida, em que, devido as temperaturas altas de formação do magma, possivelmente, biotita foi consumida a temperaturas mais baixas (< 900 °C), deixando resíduos rico em granada, enriquecida em razões Fe/Mg, e então a dissolução de granada, a temperaturas mais altas, seria responsável pelo enriquecimento da razão. As altas razões K/Rb e o notável empobrecimento em Cs, Th e U é atribuído tipicamente à fontes residuais, empobrecidas em Rb por eventos prévios de extração de magmas (Janasi, 2002), como aqui observado.

Existe significativo enriquecimento em Ba nas rochas da Suíte Paraguaçu. Biotita e feldspato potássico são os principais portadores de Ba e, como visto anteriormente, a biotita não é um bom candidato para estar presente na fonte dos magmas, sendo que a contribuição de feldspato potássico aparenta ser importante nas condições de UHT. O conteúdo de Sr, elemento carregado principalmente por plagioclásio, nas rochas da Suíte Paraguaçu, e no granada granulito félsico é complementar, isto é, são observadas anomalias negativas do elemento nos charnockitos e positivas nos granulitos. Isso sugere que plagioclásio não foi totalmente consumido na fonte, dado também mostrado na modelagem termodinâmica e corroborado pelo baixo teor de Ca nos charnockitos. Assim, o enriquecimento em ETR leves e Eu no magma anidro é relacionado à participação de feldspato potássio como reagente nas reações de fusão, fase que é enriquecida em ETR leves e apresenta fortes anomalias positivas de Eu em rochas de fácies granulito (Bea, 1996), dado também apresentado pela modelagem termodinâmica. O corpo norte da Suíte Paraguaçu apresenta leve enriquecimento em Sr sugerindo contaminação ou contribuição de outras fontes, provavelmente ricas em plagioclásio.

Relação inversa é observada entre o comportamento de Zr e Hf nos charnockitos e nos granulitos félsicos, sugerindo que zircão está sendo extraído da fonte e enriquecendo o magma nesses elementos. As rochas da Suíte Paraguaçu ainda apresentam teor elevado de ETR pesados e Y, além de razões La/Yb moderadas. Essas observações reforçam a dissolução de zircão na fonte geradora desses magmas, em que até 50% do zircão foi dissolvido durante o pico metamórfico nos granulitos. Quebra de granada também pode aumentar o conteúdo de ETR pesados, Y e as razões La/Yb no líquido, quebra também observada na modelagem.

As fases minerais portadoras de Ti, como ilmenita, são muito refratárias, portanto suportam ou são formadas em altas temperaturas. As anomalias positivas de Ti no granulito e negativa no charnockito sugerem que as fases de Ti não foram solubilizadas durante a fusão parcial. As principais fases portadoras de Nb são os minerais de Ti, além de hornblenda e zircão. O comportamento inverso de Nb entre os charnockitos e os granulitos não parece possível pela contribuição de minerais de Ti, como visto anteriormente, e deve estar relacionado à quebra de zircão na fonte. A modelagem ainda não consegue prever o comportamento exato das fase de Ti durante a fusão parcial, pois os modelos de atividade para o líquido silicático disponíveis na literatura não incluem Ti.

Dados geocronológicos obtidos em zircão da Suíte Paraguaçu mostram idades máximas de cristalização entre 630 e 619 Ma para o corpo sul e 611 Ma para o corpo norte, com espalhamento de datações para idades mais jovens até *c.* 600 Ma. As idades são contemporâneas ao metamorfismo de UHT da Unidade Granulítica da *Nappe* Socorro-Guaxupé, em que o período entre o pico metamórfico e o final da cristalização do fundido ocorreu entre 630 e 595 Ma. Esse tempo de *c.* 35 Myr corresponde ao tempo de residência magmática, em que idades mais antigas podem estar relacionadas a xenocristais carregados da fonte e/ou autocristais, enquanto idades mais jovens podem estar relacionadas à contínua alimentação de líquido da fonte. A idade de 620 Ma é interpretada como o auge do magmatismo da Suíte Paraguaçu pela grande concentração de dados acerca desse valor.

A diminuição das razões Th/U em zircão em direção as datações mais jovens e a morfologia dos cristais da Suíte Paraguaçu sugerem cristalização lenta do magma com prolongado tempo de residência na crosta inferior (Corfu *et al.*, 2003), corroborado pelo espalhamento nas datações, especialmente em direção as mais jovens. Grande espalhamento em ETR leves nos cristais de zircão reforçam a complexa história de cristalização e a competição com feldspatos por esses elementos.

Assim, o padrão de elementos traço acompanha tendências complementares entre o charnockito e o granulito. Suas idades são correspondentes, incluindo um longo período de cristalização de líquido. Aparentemente fases hidratadas tiveram pouca influência na geração do

magmas anidros, especialmente em condições de UHT. Feldspato potássico foi altamente extraído da fonte, assim como zircão, apatita e granada ao passo que plagioclásio, ortopiroxênio e minerais de Ti ficaram retidos na fonte. Os dados aqui obtidos são compatíveis com as pseudosseções calculadas para o granada granulito félsico. Conteúdo menor de SiO_2 e Na_2O nos charnockitos em relação ao previsto na modelagem sugerem mistura com outras fontes, provavelmente com afinidade mais máfica.

7.4 Diferenças estruturais entre a Unidade Granulítica e a Unidade Diatexítica

A geometria das estruturas está intimamente relacionada à ação dos campos de tensão nos orógenos (Merle, 1986; Tikoff & Fossen, 1999). Os dados estruturais coletados mostram que existem diferenças na direção da foliação e da lineação entre as Unidades Granulítica e Diatexítica da *Nappe* Socorro-Guaxupé.

Na Unidade Diatexítica ocorre um maior espalhamento nas direções de foliação, dispersas no quadrante NW-SE, enquanto o Unidade Granulítica a dispersão é menor, concentradas na direção WNW-ESE. A lineação na Unidade Diatexítica é definida por minerais hidratados e feldspatos na direção WNW-ESE, ao passo que na Unidade Granulítica a direção é WSW-ENE, marcada por minerais anidros. Portanto, o deslocamento das estruturas penetrativas sugere que diferentes campos de tensão agiram nas unidades e em diferentes momentos, já que nos granulitos a lineação é definida pelos minerais anidros do pico metamórfico e nos diatexitos, foi necessário ocorrer a cristalização do leucossoma e das fases hidratadas, após o pico metamórfico, para que ocorresse a sua deformação.

O pico metamórfico desse segmento atingiu condições de temperatura ultra alta (~ 1000 °C), incluindo a Unidade Diatexítica (Rocha *et al.*, 2017). Entretanto, as condições de cristalização do fundido foram diferentes; na Unidade Diatexítica a cristalização do fundido se deu em torno de 720 °C (Rocha *et al.*, 2017) a pressões em próximas de $5,5$ kbar; enquanto que a Unidade Granulítica registra a cristalização do fundido parcial em 800 °C à 9 kbar.

O registro de P - T é refletido no ângulo de abertura do eixo-c de quartzo. As medições realizadas no resíduo da Unidade Granulítica fornecem temperaturas próximas ao pico metamórfico, variando entre 869 e 923 °C, uma lacuna em torno de 100 °C mais baixo em relação ao pico metamórfico. As medidas obtidas no leucossoma da Unidade Granulítica fornecem temperaturas entre 796 e 853 °C, muito próximas à temperatura calculada na modelagem para a cristalização do fundido. Assim, a temperatura fornecida pela abertura do eixo-c de quartzo é interpretada como a temperatura de deformação, em que o quartzo estava cristalizado, portanto capaz de absorver deformação em estado sólido. Por consequência, existe um hiato de

aproximadamente 130 °C entre a deformação registrada no resíduo do granulito e aquela registrada no leucossoma e de 200 °C entre o pico metamórfico e a cristalização do leucossoma.

As temperaturas obtidas pela fábrica de quartzo na Unidade Diatexítica são menores, entre 633 e 685 °C. Essas temperaturas também devem representar a recristalização de quartzo em estado sólido, portanto com grande parte do leucossoma cristalizado. Logo, a diferença de temperatura entre o pico metamórfico e a cristalização do fundido é cerca de 350 °C.

As diferenças entre geometria das estruturas e temperatura de deformação mostram que as unidades tiveram histórias deformacionais distintas. Assim que o leucossoma foi cristalizado na Unidade Granulítica a deformação registrou direção próxima a E-W a temperatura de 800 °C. Pela alta taxa de fusão parcial na Unidade Diatexítica, provavelmente por maior fertilidade do protólito, uma vez que as temperaturas de pico metamórfico são semelhantes, a cristalização do leucossoma ocorreu a menor temperatura, onde os campos de tensão haviam mudado de direção, produzindo planos de direção NW-SE.

A ideia é suportada pela presença de uma zona de cisalhamento de baixo ângulo entre as unidades, em que inclusive se encaixam corpos charnockitoides da Suíte Paraguaçu. A zona de cisalhamento formou-se após a cristalização do fundido na Unidade Granulítica, mudando sua reologia e permitindo que a Unidade Diatexítica, com grande volume do seu fundido ainda não cristalizado, mover-se em direção ao Cráton do São Francisco, mas com cinemática diferente da unidade sotoposta.

7.5 Origem da *Nappe Socorro-Guaxupé*

Os dados isotópicos aqui apresentados e compilados de outros trabalhos (Janasi, 2002; Mora *et al.*, 2014; Rocha *et al.*, 2018; Tedeschi *et al.*, 2018), apontam para longa residência crustal das rochas da *Nappe Socorro-Guaxupé*, marcadas por valores negativos de ϵNd em rocha total e ϵHf em zircão, e idades modelos de Nd e Hf que variam entre 1,2 e 2,3 Ga. O tempo médio de residência crustal, isto é, a idade modelo menos a idade de cristalização, mostram tendências que apontam retrabalhamento de crosta com alguma mistura de material juvenil (Belousova *et al.*, 2009).

Embora a *Nappe Socorro-Guaxupé* seja considerada como formada em ambiente de arco magmático (Campos Neto and Caby, 1999; Mora *et al.*, 2014; Rocha *et al.*, 2018), rochas com assinaturas juvenis nunca foram reconhecidas até o momento. Em geral, dois fatores podem ser responsáveis por essas assinaturas isotópicas: processos de contaminação e assimilação do embasamento crustal (Schoene *et al.*, 2009) e/ou a mistura com manto litosférico enriquecido com componentes crustais (Hagen-Peter *et al.*, 2015).

Apenas recentemente foram encontradas idades antigas na *Nappe* Socorro-Guaxupé, concentrada em 2,5 Ga, que são interpretadas como relacionada ao embasamento da Placa Paranapanema (Tedeschi *et al.*, 2018). Contudo, as idades reconhecidas para o embasamento na região variam do Paleoproterozóico (c. 2,08 Ga) ao Mesoarqueano (c. 3,0 Ga) e são relacionadas ao Cráton do São Francisco (Cioffi *et al.*, 2016a; 2016b). Além disso, as idades modelos, tanto de Nd quanto de Hf, apontam para idades mais antigas que 1,2 Ga, em que a maioria são mais antigas que 2,0 Ga. Se um processo de assimilação/contaminação tivesse ocorrido extensivamente, seria esperado grande participação das rochas do embasamento para a elevação das idades modelos na região. Assim, estatisticamente seria esperado mais dados geocronológicos apontando para idades herdadas relacionadas ao embasamento. Portanto, o processo de assimilação/contaminação do embasamento teria uma papel menor para a formação da *Nappe* Socorro-Guaxupé.

Alguns autores atribuem a ausência de assinaturas juvenis na *Nappe* Socoro-Guaxupé à mistura com manto enriquecido (Carvalho *et al.*, 2014; Janasi, 2002; Mora *et al.*, 2014; Rocha *et al.*, 2018). Essa interpretação é bem suportada pela afinidade shoshonítica das rochas do segmento e pela presença de rochas metassedimentares de alta pressão do Sistema de Nappes Andrelândia (Campos Neto *et al.*, 2010; Campos Neto and Caby, 2000; Frugis *et al.*, 2018; Motta & Moraes, 2017; Tedeschi *et al.*, 2017) que atingiram profundidades mantélicas e podem ter contribuído para o metassomatismo do manto.

Todavia, as longas histórias crustais registradas na *Nappe* Socorro-Guaxupé levanta dúvidas sobre sua origem. Na região tipo do sul do Orógeno Brasília, o domínio de arco magmático, representado pelo Arco de Arenópolis, é composto por plútons cálcio-alcalinos, com graus variáveis de metamorfismo e deformação, e sequências supracrustais, que incluem rochas metavulcânicas e metassedimentares imaturas (Pimentel, 2016; Pimentel *et al.*, 2011). O início do magmatismo é registrado em c. 900 Ma com duração até 640 Ma (Laux *et al.*, 2005), com assinaturas juvenis marcadas por valores positivos a fracamente negativos de ϵNd (Laux *et al.*, 2005; Pimentel *et al.*, 1996; Pimentel and Fuck, 1992). As mesmas características são encontradas no Arco de Mara Rosa (Pimentel, 2016), o domínio de arco magmático do segmento norte do Orógeno Brasília, contudo essas características não são correspondentes na *Nappe* Socorro-Guaxupé.

Por outro lado, na região tipo do sul do Orógeno Brasília, o Complexo Anápolis-Itauçu apresenta similar associação de rochas com a *Nappe* Socorro-Guaxupé, com granulitos derivados de protólitos ígneos, como tonalitos e granodioritos, e rochas metassedimentares com intrusões graníticas associadas (Moraes *et al.*, 2004). As datações U-Pb em zircão estão espalhadas entre 760 e 650 Ma e são relacionadas à atividade magmática (Piuzana *et al.*, 2003a), enquanto o

metamorfismo é registrado entre 650 e 630 Ma (Baldwin & Brown, 2008; Della Giustina *et al.*, 2011; Piuzana *et al.*, 2003b) e atingiu condições de UHT (Baldwin *et al.*, 2005; Moraes *et al.*, 2002). As assinaturas isotópicas mostram valores negativos de ϵNd , entre -9,3 e -1,4, e as idades modelos de Nd apresentam dois intervalos, entre 2,3-1,9 Ga e 1,7-1,4 Ga (Piuzana *et al.*, 2003a). O Complexo Anápolis-Itaçu é interpretado como o núcleo metamórfico do Orógeno Brasília (Della Giustina *et al.*, 2009; Fuck *et al.*, 1994; Pimentel, 2016) e geograficamente está situado entre o Arco de Arenópolis à oeste e a sequência metassedimentar do Grupo Araxá a leste.

A associação de rochas, as idades magmáticas, as idades e condições metamórficas e as assinaturas isotópicas do Complexo Anápolis-Itaçu são muito similares às da *Nappe* Socorro-Guaxupé. A porção oeste da *Nappe* Socorro-Guaxupé é encoberta por espessa bacia sedimentar Paleozóica à Mesozóica, a bacia do Paraná, e à nordeste-leste é delimitada por domínio metassedimentar, o Sistema de *Nappes* Andrelândia (Campos Neto & Caby, 2000). Assim, ao invés da *Nappe* Socorro-Guaxupé estar associada ao domínio de arco magmático, ela pode ser relacionada ao núcleo metamórfico do orógeno. Além disso, nos modelos de evolução da orogenia, a Placa do São Francisco subductou a Placa do Paranapanema (Frugis *et al.*, 2018; Motta & Moraes, 2017), assim o arco magmático deve ter sido estabelecido na Placa Paranapanema. Entretanto, Cioffi *et al.* (2016a, 2016b) relacionam a grande porção de embasamento que separa a *Nappe* Socorro-Guaxupé em Domínios Socorro e Guaxupé ao Cráton São Francisco, desse modo, a *Nappe* Socorro-Guaxupé estaria sobreposta ao cráton. Com certeza, essa discussão não está concluída, ela apenas tenta colocar uma nova ótica ao assunto.

7.6 Geração de temperatura ultra alta no Domínio Guaxupé

Como visto anteriormente minerais hidratados, como biotita e hornblenda, não participaram das reações de fusão no pico metamórfico das rochas da Unidade Granulítica, sendo essas reações controladas pela quebra de feldspato potássio e granada. Também é pouco provável que minerais hidratados estavam presentes na fonte do magma da Suíte Paraguaçu, assim como na fonte geradora das outras suítes charnockíticas no Domínio Guaxupé (Janasi, 2002). Portanto, o líquido silicático gerado pela quebra de minerais hidratados foi previamente retirado do sistema. Esse cenário de baixa fertilidade fez com que as rochas da Unidade Granulítica estivessem “pré-condicionadas” (Clark *et al.*, 2011; Kelsey & Hand, 2015; Vielzeuf *et al.*, 1990) e, consequentemente, propensas ao aquecimento até temperaturas ultra-altas.

Três fatores principais controlam o aumento de temperatura da crosta continental: (1) calor radioativo produzido pelo decaimento de Th, U e, principalmente, K (Alessio *et al.*, 2018; England & Thompson, 1984; Jamieson *et al.*, 1998; Vilà *et al.*, 2010); (2) calor gerado por atrito em

zonas de cisalhamento (Nabelek *et al.*, 2010); (3) advecção mantélica (Brown, 1993; Ellis, 1987; Vielzeuf *et al.*, 1990).

Na *Nappe* Socorro-Guaxupé grande volume de líquido foi gerado durante a orogênese, com extensões de centenas de km² e pelo menos 9 km de espessura (Campos Neto & Caby, 1999), representado pelas Unidades Diatexítica e Metatexítica. Grande parte das rochas da *nappe* possui afinidade shoshonítica (Janasi, 2002, 1999; Mora *et al.*, 2014; Rocha *et al.*, 2018), portanto enriquecida em K. Esse volume expressivo de material radiogênico pode elevar consideravelmente a temperatura da crosta inferior (Beaumont *et al.*, 2004; Gerya *et al.*, 2008; Jamieson *et al.*, 2004). Desse modo, a geração de grandes quantidades de líquido e a migração para níveis intermediários na crosta teve papel fundamental para o aquecimento do orógeno.

O corpo norte da Suíte Paraguaçu apresenta enriquecimento em Sr e valores de ϵHf de até 4 unidades maior em relação ao corpo sul. Esse enriquecimento sugere mistura com fontes mais radiogênicas, ou seja, mais ricas em ¹⁷⁶Hf, além da contribuição de plagioclásio na fonte, o que é compatível com fontes mais máficas. Além disso, comparado com a composição da fusão do granulito félsico modelado a Suíte Paraguaçu é empobrecida em SiO₂ e enriquecida em Na₂O, outra sugestão de contribuição de fontes mais empobrecidas na geração desses magmas. Rocha *et al.* (2018) e Janasi (1999, 2002) também sugerem a participação de manto nessa porção do orógeno. Assim, pequenos corpos de rochas máficas, atualmente expostos como granulitos máficos, alguns ainda preservam contatos intrusivos, indicam contribuição do manto na base da crosta, o que sugere descompressão do manto astenosférico fornecendo calor adicional para o segmento, como sugerido em alguns modelos (Brown, 1993; Ellis, 1987).

7.7 Referências bibliográficas

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8 CONCLUSÕES

Os dados e discussões apresentadas neste trabalho permitem traçar de forma realista a evolução metamórfica dos granulitos da *Nappe* Socorro-Guaxupé, incluindo não apenas informações petrológicas como temporais. Os granulitos iniciaram a trajetória progressiva em *c.* 625 Ma, marcada por núcleos de zircão com zoneamento setorial, interpretados como fruto da recrystalização de núcleos herdados. O pico metamórfico foi atingido em condições de temperatura ultra-alta em $\sim 1000\text{ }^{\circ}\text{C}$ a pressões de 11 kbar. Nos granulitos félsicos, litotipo predominante na Unidade Granulítica na região, a associação de pico metamórfico é marcada pela presença de (granada) – clinopiroxênio – ortopiroxênio – plagioclásio – feldspato potássio – quartzo – ilmenita – leucossoma, enquanto os granulitos máficos apresentam (granada) – clinopiroxênio – ortopiroxênio – plagioclásio – quartzo – ilmenita – (leucossoma) e os granulitos pelíticos apresentam granada – sillimanita – espinélio – rutilo – plagioclásio – feldspato potássio – leucossoma – (quartzo) – (biotita) – (coríndon). Em *c.* 615 Ma, a cristalização de zircão em bordas mais antigas ou como neocristais com alto teor de U, marca o início da cristalização da fusão parcial nos granulitos félsicos. Assim, a idade de pico metamórfico fica restrita entre 625 e 615 Ma. Após o pico, ocorre estágio marcado por descompressão até $\sim 9,5$ kbar, registrada por exsoluções de magnetita em cristais de ilmenita e discretas coroas de ortopiroxênio em granada. Contudo, em geral não há feições de desestabilização de granada e os modelos de equilíbrio de fases calculados apontam para o crescimento modal nas proporções de granada e feldspato potássio durante o resfriamento. A trajetória segue caminho de resfriamento isobárico até $870\text{ }^{\circ}\text{C}$ e 9 kbar, em que fases hidratadas, principalmente hornblenda, começaram a cristalizar em *c.* 595 Ma, período marcado pela cristalização de zircão arredondado de baixo conteúdo de U e empobrecido em ETRP, uma vez que o aumento da moda de granada durante o resfriamento, previamente sequestrou esses elementos do líquido anatético. Desse modo, a história metamórfica dos granulitos mostram que a crosta inferior se manteve aquecida em condições de fácies granulito por pelo menos 30 Myr. A abordagem utilizada, combinando modelos de equilíbrio de fases com petrocronologia, se mostrou eficaz e, o trabalho aqui realizado, pode ser utilizado como guia para novos estudos de natureza similar.

A Suíte Paraguaçu apresenta charnockitos shoshoníticos com idades compatíveis às idades do pico metamórfico na Unidade Granulítica (*c.* 620 Ma) com espalhamento em direção à idades mais novas. Os charnockitos são empobrecidos em Th e U e apresentação altas razões K/Rb, feições tipicamente relacionadas a fontes residuais. Essas rochas também mostram enriquecimento em Ba, K, ETRL, P, Zr e Hf e empobrecimento em Sr e Ti. Padrão inverso é observado nos granulitos. A composição calculada nos modelos de equilíbrios de fases do líquido anatético gerado pelos granulitos félsicos possui semelhanças com a composição química das

rochas da Suíte Paraguaçu. Além disso, os modelos mostram ainda que, durante o pico metamórfico, feldspato potássio, zircão e apatita estavam sendo solubilizados nos granulitos, empobrecendo-os em Ba, K, ETRL, P, Zr e Hf. Ao mesmo tempo, plagioclásio e ilmenita não foram fundidos na fonte. Portanto, através de uma abordagem reversa, isto é, estudando o produto através do protólito foi possível afirmar que o resultado da fusão parcial dos granulitos félsicos teve papel fundamental para a geração da Suíte Paraguaçu.

Através de mapeamento foi possível observar que na região o contato entre a Unidade Granulítica e a Unidade Diatexítica não ocorre de forma gradual e sim por zona de cisalhamento horizontal não muito espessa. Dados estruturais mostram também que as geometrias dessas unidades apesar de similares, direções gerais de foliação em NW-SE e transporte para E, não são coincidentes, o que implica em diferentes campos deformacionais. Cálculos termobarométricos mostram que ambas as unidades atingiram temperaturas de pico metamórfico em fácies granulito. Contudo, as temperaturas de deformação, obtidas pela abertura do eixo-c de quartzo, mostram uma diferença de 200 °C em relação ao pico metamórfico na Unidade Granulítica e de 350 °C na Unidade Diatexítica. Portanto, os dados mostram que a deformação na Unidade Diatexítica ocorreu a temperatura menor que a registrada na Unidade Granulítica e, provavelmente, foi posterior, uma vez que o volume de fusão parcial na Unidade Diatexítica era muito maior e, desse modo, demorando mais para o volume cristalizado absorver as deformações em estado sólido registrada.

Portanto, aliado as discussões propostas na literatura é possível afirmar que o metamorfismo de temperatura ultra-alta ocorreu de forma regional no Domínio Guaxupé. Isso foi possível pois as rochas, da Unidade Granulítica, foram condicionadas, ou seja, eventos prévios de fusão parcial guiados por reações de fusão em presença de água e controlada por minerais hidratados foram extraídos, deixando para trás uma rocha pouco fértil e refratária, condição preponderante para atingir temperaturas extremas como as registradas. O aquecimento foi possível pela combinação de dois fatores: primeiro, a geração de enorme volume de material fundido, representado pelas Unidades Diatexítica e Metatexítica, com grande concentração de elementos incompatíveis como U, Th e, principalmente, K que através do decaimento radioativo podem gerar muito calor; e segundo, a presença de fonte externa, provavelmente mantélica, marcado pelo enriquecimento em Sr e ^{176}Hf radiogênico na Suíte Paraguaçu e em rochas mais primitivas da Suíte Divinolândia.

**ANEXO I – RECOVERING P – T – t PATHS FROM ULTRA-HIGH
TEMPERATURE (UHT) ROCKS THAT LACK DIAGNOSTIC MINERAL
ASSEMBLEGES: A FELSIC GRANULITE EXAMPLE FROM THE SOUTHERN
BRASÍLIA OROGEN OF BRAZIL**

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Recovering P – T – t paths from ultra-high temperature (UHT) rocks that lack diagnostic mineral assemblages: A felsic granulite example from the Southern Brasília Orogen of Brazil

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ABSTRACT

The recent development of thermodynamic models for the activity-composition relationships of trondhjemitic-tonalitic melt composition makes it possible for the first time to quantify the metamorphic histories of terranes dominated by felsic and mafic orthogneiss that typically lack mineral assemblages diagnostic of pressure–temperature (P – T) conditions. In this paper, whole-rock and mineral chemistry, thermodynamic phase equilibrium modelling, and zircon U–Pb–Hf isotope analysis are used to infer the P – T –time history of garnet-bearing felsic and mafic granulites from the Socorro-Guaxupé Nappe of the Southern Brasília Orogen. These rocks occur in a stack of Neoproterozoic nappes that were assembled during the formation of the West Gondwana continent. The peak assemblage in the felsic granulite of ortho- and clinopyroxene, garnet, K-feldspar, plagioclase, quartz, ilmenite, magnetite and silicate melt reached a peak temperature of ~ 1000 °C at 11 kbar, recorded both by the high-Al content in orthopyroxene and the phase equilibrium model. Hydrated phases were not present at the

metamorphic peak, and generation of the melt phase was controlled by K-feldspar breakdown. The retrograde path of near-isobaric cooling led to partial re-hydration of the peak assemblage, revealed by coronas of hornblende around pyroxene. Sector-zoned zircon cores from the felsic granulites yielded U–Pb ages of *c.* 625 Ma, interpreted as dating the onset of high-temperature metamorphism. In one sample, rims around these zircon cores and new soccer ball zircon grains both have high U contents and yield ages of *c.* 615 Ma assumed to mark the onset of melt crystallization, while low-U rims and soccer ball zircons in two other samples yield ages of *c.* 595–600 Ma, interpreted as dating the end of melt crystallization. These data reveal a protracted metamorphic event, in which the lower crust remained at granulite-facies conditions for at least 30 Myr, consistent with the regional-scale development of ultra-high temperature (UHT) conditions in a collisional environment.

Keywords: ultra-high temperature metamorphism, *P–T–t* path, phase equilibrium modelling, petrochronology, Brasília Orogen

INTRODUCTION

Ultrahigh temperature (UHT) metamorphism is the most thermally extreme type of crustal metamorphism that the Earth is able to produce, characterized by temperatures that exceed 900 °C (Harley, 1998). There is widespread evidence these metamorphic conditions have been generated repeatedly in time and space (Brown, 2007; Clark et al., 2011; Harley, 2008; Kelsey, 2008; Kelsey et al., 2008; Korhonen et al., 2014). UHT metamorphism is most commonly recognized via diagnostic mineral assemblages, like sapphirine + quartz, orthopyroxene + sillimanite + quartz, osumilite + garnet, and, in some cases, spinel + quartz (Harley, 2008). However, these diagnostic mineral assemblages occur only in relatively rare MgO- and Al₂O₃-rich protoliths, raising the question of how UHT metamorphism can be identified and characterized in more common protoliths, such as felsic and mafic granulites with granite and basalt bulk-compositions. Partial melting is widespread in most rock compositions under such high-temperature regimes, and the separation and extraction of this melt is critical to the preservation of peak metamorphic mineral assemblages in the lower continental crust (Baldwin et al., 2005; Moraes et al., 2002). Any melt remaining as the crust cools will crystallize and release its volatile content, which is likely to trigger widespread chemical re-equilibration that can destroy peak mineral compositions and promote overprinting of the peak paragenesis by lower temperature equivalents, hampering temperature calculations based on mineral composition (Clark et al., 2011; Fitzsimons and Harley, 1994; Moraes et al., 2002; Pattison and Bégin, 1994; White and Powell, 2002).

Thermodynamic phase equilibrium modelling of bulk-rock compositions can be used to recover peak metamorphic conditions from UHT granulites that no longer preserve peak mineral

compositions, but this is not trivial and no single set of procedures will be successful in every case (Kelsey and Hand, 2015). For example, it was only after a reliable activity-composition model for sapphirine was developed (Kelsey et al., 2005, 2004), that it became possible to accurately quantify P - T conditions from Mg-Al-rich rocks with mineral assemblages diagnostic of UHT conditions (*e.g.* Baldwin et al., 2005). Similarly, the thermodynamic modelling of felsic and mafic granulites under UHT conditions requires robust activity-composition models for complex mineral phases like hornblende, and also for silicate melt compositions appropriate for such rock types. Although there have been activity-composition models for the peraluminous silicate melt produced by metapelites for some time (*e.g.* Holland and Powell, 2001; White et al., 2007), it is only recently that activity-composition models dealing with high-temperature and high-pressure pyroxenes, amphiboles and K_2O -poor CaO -rich melt have become available (Green et al., 2016; Palin et al., 2016), allowing quantitative thermodynamic modelling of partial melting and metamorphism in felsic and mafic granulites under UHT conditions.

Felsic and mafic garnet granulites that crop out in the basal portion of the Socorro-Guaxupé Nappe, Minas Gerais, southeastern Brazil, have consistently yielded UHT conditions when studied via several thermobarometric methods (Campos Neto and Caby, 2000; Del Lama et al., 2000, 1994; Moraes et al., 2015), making them a perfect example to investigate how these bulk compositions record UHT metamorphism. In this study we use the mineral chemistry, whole-rock geochemistry, mineral thermobarometry, phase-equilibrium pseudosection modelling, and U-Pb-Hf-REE zircon compositions to quantify the P - T - t history of the Socorro-Guaxupé granulites during Neoproterozoic partial melting and UHT metamorphism.

GEOLOGICAL SETTING

The Socorro-Guaxupé Nappe is part of a stack of nappes in the Southern Brasília Orogen (Figure 1A), formed during the assembly of West Gondwana by Neoproterozoic collision of a passive margin at the edge of the São Francisco plate and the active margin of the Paranapanema plate (Brito Neves et al., 1999; Brito Neves and Fuck, 2013; Campos Neto, 2000). The southernmost part of the Brasília Orogen is subdivided into three domains: an inner domain consisting of the Socorro-Guaxupé Nappe, which includes high-grade rocks formed in the root of the Paranapanema plate magmatic arc (Campos Neto and Caby, 2000, 1999); a central domain comprising the Andrelândia Nappe System, consisting of subducted high- P continental rocks of the São Francisco plate (Campos Neto et al., 2010; Campos Neto and Caby, 2000, 1999; Motta and Moraes, 2017; Trouw et al., 2000); and an external domain representing the continental passive margin of the São Francisco plate, which includes rocks of the Carrancas Nappe and Lima Duarte Nappe Systems (Campos Neto et al., 2004).

The Socorro-Guaxupé Nappe is an allochthonous high-grade metamorphic terrane dominated by granulites and migmatites (Campos Neto and Caby, 2000; Iyer et al., 1996; Rocha et al., 2017), and comprises two main domains separated by Archean-Paleoproterozoic orthogneisses and Neoproterozoic meta-supracrustal rocks. The northwestern Guaxupé Domain represents the deeper crustal portion where granulites and diatexites are dominant (Figure 1B), whereas metatexites and extensive igneous bodies are more common in the southeastern Socorro Domain, where rocks decrease in grade southwards to amphibolite-facies conditions. In both domains, the peak of metamorphism is dated to be between 625 and 650 Ma (Martins et al., 2009; Mora et al., 2014; Rocha et al., 2018, 2017; Tedeschi, 2018; Tedeschi et al., 2018). Geochemical features of the granulites from Guaxupé Domain are consistent with a magmatic arc origin (Campos Neto et al., 1996; Campos Neto and Caby, 2000). This study is focused on the granulitic base of the Guaxupé Domain, which mainly comprises felsic granulites of acid to intermediate composition, interlayered with ortho- and clinopyroxene-bearing mafic granulite and sillimanite- and spinel-bearing aluminous (metapelitic) granulite, together with intrusive charnockite bodies that are both parallel to, and discordant with, the gneissic banding. Four samples were selected for chemical analysis and/or thermodynamic modelling, comprising two banded garnet-bearing felsic granulites (VAB01 and VAB19), a garnet-bearing felsic diatexite (RGM55) and a garnet-bearing mafic granulite (PAR03).

METHODS

Mineral and whole rock chemistry

Back-scattered electron imaging (BSE) and energy-dispersive X-ray spectroscopy (EDS) were undertaken using a TESCAN MIRA3 Field-Emission Scanning Electron Microscope (FESEM) at the Microscopy & Microanalysis Facility at the John de Laeter Centre, Curtin University, Australia, with a 15 mm working distance and accelerating voltage of 20 kV. Wavelength-dispersive spectrometry (WDS) analyses were performed with a JEOL JXA-8230 Hyperprobe at the Electron Microprobe Facility, Universidade Estadual Paulista (UNESP), Brazil, using a focused beam at 15 kV and 20 nA with 10 µm spot size for all minerals other than clinopyroxene, for which a 1 µm spot size was used because of closely-spaced exsolution lamellae of clinoenstatite. The analyses were recalculated using the A-X software (<https://www.esc.cam.ac.uk/research/research-groups/holland/ax>). Analyses are given in ANEXO IV.

Figure 1. (A) Brasília Orogen and its relation with Neoproterozoic orogens and cratons. (B) Geological Map of the Guaxupé Domain of the Socorro-Guaxupé Nappe, Southern Brasília Orogen (after Campos Neto et al., 2010). Numbers 1 to 4 are the location of the studied samples VAB01, VAB19, RGM55 and PAR03, respectively, two garnet felsic granulites, a diatexite and the last a garnet mafic granulite.

Compositional profiles were made across grains of garnet, pyroxene and feldspar in the two garnet felsic granulite samples (VAB01 and VAB19) and the garnet mafic granulite (PAR03). Biotite inclusions in garnet were analyzed in sample VAB01. Original (*i.e.* pre-exsolution) compositions of clinopyroxene in the felsic granulites were calculated by reintegration of the clinoenstatite lamellae and host clinopyroxene using average mineral compositions for each

phase ($n = 18$ for both) weighted according to their relative proportions estimated from BSE images.

Garnet trace elements analyses were performed using a Thermo Scientific iCAP Q ICP-MS coupled to a New Wave UP 213 nm laser ablation system at the NAP Geoanalítica, Institute of Geosciences of the University of São Paulo (IGc-USP), Brazil. Garnet was analyzed with 65 μm spotsizes, a laser fluence of 5.52 J/cm², at a repetition rate of 10 Hz. External calibration was performed relative to NIST 610 glass and the USGS BCR-2G basaltic glass reference material was analyzed as secondary standard during analyses. Si and Ca determined by electron microprobe were used as internal standards. Results are presented in ANEXO V.

Bulk rock compositions were determined by X-ray fluorescence analysis using a Philips PW2400 at the NAP Geoanalítica, Brazil, following the methodology described by Mori et al. (1999). Trace and rare earth elements (REE) were analyzed by inductively coupled plasma mass spectrometry (ICP-MS) using an iCAP Q by Thermo Scientific also located at IGc-USP, following the methodology described by Navarro et al. (2008). The results are presented in ANEXO VI.

Phase equilibria modelling

Phase equilibrium modelling of garnet felsic granulite sample VAB01 and garnet mafic granulite sample PAR03 was performed in the 10-component NCKFMASHTO system (Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂–O₂) using THERMOCALC (Powell and Holland, 1988) version 3.45i, with the following activity–composition models: silicate melt (L), clinopyroxene (augite – aug) and hornblende (hb) from Green et al. (2016); garnet (g), orthopyroxene (opx), ilmenite (ilm) and biotite (bi) from White et al. (2014); K-feldspar (ksp) and plagioclase (pl) from Holland and Powell (2003); magnetite (mt) from White et al. (2002). Rutile (ru), quartz (q) and aqueous fluid (H₂O) are considered pure phases. Calculations are confined to a P - T window between 7 to 13 kbar and 800 to 1050 °C. Mn has a negligible effect on phase stability at high- T (Johnson et al., 2015) and was excluded from the model system. The bulk compositions used in the pseudosections (expressed as mol. % oxides) are given in Table 1.

Table 1. Bulk composition as molar oxides (mol.%) used in phase equilibria modelling.

Sample	Lithology	Figs	H2O	SiO2	Al2O3	CaO	MgO	FeO	K2O	Na2O	TiO2	O
VAB01	garnet	4, 5, 6,										
	felsic granulite	S1 and S2	1.00	68.34	9.67	5.63	3.88	5.37	1.43	3.11	0.57	1.00
PAR03	garnet	7A	1.00	48.38	7.74	11.28	7.91	14.78	0.1	3.05	2.28	3.48
	mafic granulite	7B	0.01	48.84	7.81	11.39	7.98	14.91	0.1	3.08	2.3	3.58

Granulites are fluid poor rocks, but can contain hydrous phases including melt (represented by leucosome), hornblende and biotite, and the stability of these phases during metamorphism is strongly dependent on the availability of H₂O. Estimation of bulk-rock H₂O contents during high-*T* and UHT metamorphism is problematic, firstly because the present-day H₂O content is typically estimated from loss-on-ignition (LOI) values, which can include other volatile species, and secondly many high-grade metamorphic rocks undergo variable degrees of retrograde open-system hydration leading to bulk H₂O contents that are higher than those during peak metamorphism. In order to estimate the likely water content of felsic granulite VAB01 during high-*T* metamorphism, a *T*-*X*_{H₂O} pseudosection was calculated at a pressure of 10 kbar (approximate peak *P* for this rock – see below), where *X*_{H₂O} ranges from 0.01 to 5 mol. %. The H₂O content of 1 mol. % was chosen such that between 5% and 20% melt might be predicted, consistent with the field observations (Figure S1 - Supplementary material). *P*-*X* pseudosections calculated at a temperature of 900 °C (approximate peak *T* for this rock – see below) were used to constrain Fe₂O₃ contents at the metamorphic peak (expressed as *O* in the pseudosection calculations). An *X*_{Fe₂O₃} value of 1 mol. % was chosen such that ilmenite and magnetite (but not rutile) were stable, which matches the oxide phases observed in the rocks (Figure S2 - Supplementary material).

It was not possible to model an H₂O content for the garnet mafic granulite using *P*- or *T*-*X* diagrams, and therefore two different *P*-*T* pseudosections were constructed, one with the value of the loss-on-ignition value (0.01 wt. %) and another with 1 mol. % of water. Assemblage fields in the *P*-*X*_{Fe₂O₃} pseudosection for this rock do not show significant variability in the oxide phases, thus an Fe³⁺/ΣFe ratio of ~0.2 was assumed, which is similar to the observed value in intrusive igneous rocks associated with arc magmatism (Evans, 2012).

U-Pb geochronology

Zircon grains were extracted from the garnet felsic granulite and diatexite samples (VAB01, VAB19 and RGM55) via traditional methods of disaggregation, magnetic and heavy liquid separation. These were mounted in 25-mm diameter epoxy discs together with NBS-610 glass and primary and secondary zircon standard material, polished to approximately half-grain thickness, carbon-coated, and then imaged by cathodoluminescence (CL) using a TESCAN MIRA3 FESEM. The mounts were then cleaned and covered with a 40 µm gold coat prior to U-Pb isotopic analysis using the Sensitive High Resolution Ion Microprobe (SHRIMP II). All this equipment is housed in the John de Laeter Centre of Mass Spectrometry at Curtin University.

SHRIMP II was operated at a mass resolution of 5,000, with a primary beam current of 2 nA and beam diameter of 20 µm, and a 6-scan cycle was used during analysis (de Laeter and

Kennedy, 1998; Kennedy and de Laeter, 1994). NBS-610 glass was used to calibrate the position of the ^{204}Pb peak. CUYZ (Curtin internal laboratory standard - 568.55 Ma) was used as primary standard and OGC (Stern et al., 2009) as a secondary standard. Data were processed using SQUID II and Isoplot 3.75 (Ludwig, 2009, 2003). External spot-to-spot errors on zircon U–Pb calibration sessions were <1%, which reflects the long-term performance of the SHRIMP II. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of secondary standard determined during the analytical session was 3469.9 ± 7.6 Ma, which is within error of the published age (3465.4 ± 0.6 Ma; Stern et al., 2009). U–Pb data are presented in ANEXO VIII.

Zircon trace element and Lu–Hf isotope analysis

Trace elements and Lu–Hf isotopes and complementary U–Pb data of zircon were collected using Split-Stream Laser Ablation with Inductively Coupled Plasma Mass Spectrometry (LASS-ICP-MS) on an ASI Resolution M-50 with a 193-nm wavelength Ar–F excimer laser with an Agilent 7700 mass spectrometer at the GeoHistory Facility, John de Laeter Centre, Curtin University. Laser fluency was $2\text{J}/\text{cm}^{-2}$, at a repetition rate of 10 Hz and spot size of 50 μm . The same samples from SHRIMP were used to analyze trace elements, Lu and Hf isotopes and, although U–Pb data were acquired by LASS-ICP-MS, and when the laser spot was coincident to the SHRIMP spot or domain, the $^{206}\text{Pb}/^{238}\text{U}$ spot age obtained in SHRIMP was used as reference. Data are presented in ANEXO IX, ANEXO X and ANEXO XI.

RESULTS

Petrography and bulk rock chemistry

Garnet felsic granulite (samples VAB01 and VAB19)

The garnet felsic granulite has a well-developed compositional and textural layering (Figure 2A), comprising medium- to coarse-grained anhydrous leucosome, a fine- to coarse-grained residuum and a poorly developed selvage of hornblende in the edges of the leucosome (Figure 2B). This layering is typically closely-spaced and bulk compositions determined for these rocks will be mixtures of both layer types.

Figure 2. (A) Banded garnet felsic granulite with orthopyroxene porphyroblast (VAB01). (B) Typical mineral assemblage of the banded garnet felsic granulite with two pyroxenes in both the residuum and in the leucosome (VAB19, plane-polarized light). (C) Films of quartz (arrows) surrounding feldspars in the garnet felsic granulite, representing evidence of former anatectic melt (crossed polarized light plus λ plate – VAB01). (D) Films of K-feldspar wrapping plagioclase crystals in the garnet felsic granulite (BSE image – VAB19). (E) Rounded biotite inclusions in garnet and hornblende replacing pyroxene outside garnet in the garnet felsic granulite (VAB01, plane-polarized light). (F) Clinopyroxene (darker grey) in the garnet felsic granulite with fine exsolution lamellae of clinoenstatite (brighter grey) and an adjacent oxide phase in the top right corner oxide comprising exsolved ilmenite (brightest grey) and magnetite (VAB01, BSE image). (G) Biotite and oxide intergrowths surrounding pyroxene grains in the diatexite (RGM55, plane-polarized light). (H) Coarse allanite crystals in leucosome of the diatexite (RGM55, plane-polarized light). (I) Fine to coarse-grained garnet mafic granulite with granoblastic fabric (PAR03, plane-polarized light).

The residuum contains the peak paragenesis of garnet (5–10%), orthopyroxene (10%), clinopyroxene (3–5%), plagioclase, perthite, quartz and ilmenite. Optically-continuous quartz inclusions in garnet, thin quartz films along contacts between garnet and pyroxene (Figure 2C) and cusped grains of orthoclase surrounding plagioclase (Figure 2D) are interpreted as mimicking and crystallizing from trapped anatectic melt (Holness et al., 2011; Waters, 2001). Coronas of clinopyroxene locally surround orthopyroxene. Hornblende is the only hydrous mineral in the rock matrix and replaces pyroxenes (Figure 2E), whereas rounded inclusions of biotite (but not hornblende) occur armored in garnet. The oxide phases comprise exsolution intergrowths of ilmenite and Ti-magnetite with a zebra pattern (Figure 2F).

The leucosome contains garnet, orthopyroxene, clinopyroxene, plagioclase, perthite, quartz and hornblende (Figure 2B). Porphyroblasts or symmetric porphyroclasts of mafic phases

occur locally in this domain (Figure 2A). The volume of leucosome in the rock can reach 15 %. Clinopyroxene in both leucosome and residuum contains exsolution lamellae of clinoenstatite (Figure 2F).

Samples VAB01 and VAB19 have similar chondrite-normalized REE abundances (Figure 3A) with a fractionated trend from more-abundant light REE (LREE) towards less-abundant heavy REE (HREE). This fractionation is stronger in sample VAB19 ($La_N/Yb_N = 19.7$) than the VAB01 ($La_N/Yb_N = 7.9$). Both samples are strongly depleted in Th and U and enriched Ti (Figure 3B), compared to average continental crust (Wedepohl, 1995).

Figure 3. (A) Chondrite-normalized (Sun and McDonough, 1989) REE patterns for the garnet felsic granulites and diatexite; (B) Average crust-normalized (Wedepohl, 1995) trace-element patterns for the garnet felsic granulites and diatexite.

Diatexite (sample RGM55)

The diatexite contains a large volume of leucosome, higher than 60 %, with rafts of residuum. The rafts are fine grained and composed of garnet (<5%), orthopyroxene (5–10%), clinopyroxene (3–5%), plagioclase, perthite, quartz and ilmenite.

The leucosome is composed of medium- to coarse-grained orthopyroxene, clinopyroxene, perthite, plagioclase and interstitial quartz, with a syeno-granite composition. Garnet is scarce in this domain, less than 1 %. Many perthite megacrysts have rounded quartz and subhedral plagioclase inclusions. Pyroxene grains are locally replaced at their rims by late biotite and/or biotite + oxides intergrowths (Figure 2G). Accessory zircon, apatite and allanite stand out due to their large grain size (Figure 2H).

REE pattern indicate pronounced enrichment in LREE (Figure 3A), and a steep fractionation towards HREE ($\text{La}_N/\text{Yb}_N = 149.9$). This rock is enriched in Rb, Th, LREE, Zr and Hf relative to average continental crust (Wedepohl, 1995), but depleted in Sr (Figure 3B), presenting a distinct pattern from the garnet felsic granulite samples.

Garnet mafic granulite (sample PAR03)

The garnet mafic granulite exhibits a peak metamorphic mineral assemblage of garnet, orthopyroxene, clinopyroxene, plagioclase, quartz, ilmenite and magnetite. It has a granoblastic fabric with a poorly-developed foliation and does not bear any hydrous phases (Figure 2I). Rare millimetric patches of leucosome are composed of plagioclase and quartz with a tonalite-diorite composition and represent crystallized anatectic melt. Rare orthopyroxene porphyroblasts contain thin exsolution lamellae of clinopyroxene, and are locally surrounded by a discontinuous corona of the same mineral.

Mineral chemistry

Garnet

Grains of garnet from the garnet felsic granulite show flat major element compositions (Figure 4A and C). X_{Mg} (atomic $\text{Mg}/[\text{Mg}+\text{Fe}^{2+}]$) from VAB01 is higher than in VAB19 (0.37 versus 0.32), but the grossular content ($\text{Ca}/[\text{Ca}+\text{Mg}+\text{Fe}^{2+}+\text{Mn}]$) in both samples is similar ($X_{\text{gr}} = 12\text{--}14$). Garnet in the garnet mafic granulite is homogeneous in composition ($X_{\text{Mg}} \sim 0.24$ and $X_{\text{gr}} \sim 18\text{--}19$), with lower X_{Mg} and higher grossular contents than the garnet felsic granulite (Figure 4E), which is consistent with measured bulk rock compositions.

Garnet from sample VAB01 exhibits three different patterns of HREE (Figure 4B). Crystals larger than 7 mm present cores depleted in HREE, with decreasing towards Lu ($\text{Gd}_N/\text{Yb}_N \sim 7.1$). The intermediate part of crystals larger than 7 mm and crystals core smaller than 7 mm present a flat pattern of HREE ($\text{Gd}_N/\text{Yb}_N \sim 0.9$). The rims of all crystals show a subtle drop of HREE ($\text{Gd}_N/\text{Yb}_N \sim 0.7$), which represents a small decrease in HREE compared to the latter group. Apart of the HREE behaviour, all groups present the same arrangement related to LREE, with a steep pattern of increasing, and MREE, with negative Eu anomalies.

The garnet REE distribution in sample VAB19 does not show significant differences between rims and cores (Figure 4D). They present a steep increasing in LREE, negative Eu anomalies, and increased pattern of HREE ($\text{Gd}_N/\text{Yb}_N \sim 0.2$).

Figure 4. Representative garnet compositions acquire by EMP (figures A, C and E) and LA-ICP-MS (figures B and D). Sample ID is labelled in the figure.

Pyroxenes

Orthopyroxene shows minor variations in X_{Mg} in the garnet felsic granulite (0.57–0.61 in VAB01 and 0.53–0.56 in VAB19) and in the garnet mafic granulite (X_{Mg} 0.49–0.52 in PAR03), again consistent with variations in bulk-rock chemistry. However, major variations in Al^T ($Al-(2-Si)$) are observed, with cores relatively depleted in Al. Sample VAB01 has Al^T between 0.033 and 0.063, whereas sample VAB19 has Al^T between 0.030 and 0.050. Such variation is also observed in the garnet mafic granulite (PAR03), in which Al^T is 0.025–0.044.

Host clinopyroxene within the garnet felsic granulite has an average X_{Mg} of 0.69 and 0.85 apfu of Ca (normalized to 6 oxygens). The exsolution lamellae of clinoenstatite have lower X_{Mg} (~ 0.54), with almost no Ca and $Al^T \sim 0.075$. Based on imaging, the grains are estimated to contain $\sim 10\%$ lamellae, equating to an X_{Mg} of 0.66 and 0.76 apfu Ca in the original high- T clinopyroxene. In the garnet mafic granulite, clinopyroxene has a X_{Mg} of 0.63 and contains 0.84 apfu of Ca.

Plagioclase

In the garnet felsic granulite, plagioclase shows no compositional zoning between core and rims, with average composition An_{58-61} . Conversely, plagioclase in the garnet mafic granulite has cores richer in anorthite (An_{50-70}), than the rims (An_{40-45}) and an intense optical zoning can be observed.

Biotite

Biotite inclusions in garnet from garnet-bearing felsic granulite sample VAB01 have X_{Mg} of 0.7 and Ti contents between 0.133 and 0.286 apfu, based on 11 oxygens.

Thermobarometry

Compositions of coexisting garnet and orthopyroxene were used to estimate P – T conditions using garnet–orthopyroxene thermobarometry corrected for late Fe–Mg exchange (Pattison et al., 2003). The association orthopyroxene–clinopyroxene–garnet–quartz–plagioclase was used to estimate pressure and temperature with the average P – T mode of THERMOCALC (Holland and Powell, 1998) as reported in Table 2.

Table 2. Summary of the thermobarometric results. The uncertainties of THERMOCALC are reported as 1σ .

Sample	Analysis used	THERMOCALC		Pattison et al. (2003)		
		T (°C)	P (kbar)	T (°C)	P (kbar)	T (°C) at fixed P (11 kbar)
VAB01	cpx(reint)*- opx3.7- pl2.5**-grt1.8	1009 ± 72	11.5 ± 1.5	1020	14.4	936
VAB19	cpx2.2*- opx9c- pl4.2**-grt1.5	962 ± 74	12.2 ± 1.5	952	13.5	894
PAR03	cpx7.1*- opx1.3- pl3.3**-grt8.3	936 ± 93	10.2 ± 1.1	919	11.0	921

* clinopyroxene analysis are not used in the Pattinson *et al.* (2003) thermobarometer.

** at fixed pressures plagioclase is not used in the Pattinson *et al.* (2003) thermometer.

Metamorphic conditions estimated for the garnet felsic granulite sample VAB01 using the THERMOCALC average P – T calculations suggest ultra-high temperatures of 1009 ± 72 °C at pressures of 11.5 ± 1.5 kbar. The Pattison et al. (2003) thermobarometer yielded similar temperatures (1020 °C), but significantly higher pressures (14.4 kbar). However, at a fixed pressure of 11 bar, temperatures of 936 °C are calculated, within uncertainties of the THERMOCALC calculations.

Calculated conditions for the garnet mafic granulite are similar to those for the garnet felsic granulite. The THERMOCALC calculations yielded 936 ± 93 °C at 10.2 ± 1.1 kbar, whereas the Al-solubility-based thermobarometer yielded a temperature of 919 °C and pressure of 11.0 kbar.

Phase Equilibria Modelling

The P – T pseudosection for garnet felsic granulite VAB01 is shown in Figure 5. Within the P – T window, the solidus occurs between 840 and 880 °C, from low to high P . Orthopyroxene is

stable at relatively low P , and garnet at relatively high P , with both minerals coexisting within a broad field at pressures of 8 – 12 kbar. Clinopyroxene is not stable at the highest pressures and low temperatures, but otherwise is stable with all other phases in several combinations. The hornblende-out line is strongly dependent on temperature, its stability ranging from ~860 °C at 7 kbar to 930 °C at 13 kbar. Magnetite is only stable at low pressures (<8–10 kbar). K-feldspar stability is predicted up to temperatures of 960 (low P) and 1040 °C (high P). Quartz, plagioclase and ilmenite are present throughout the calculated window, and biotite is not predicted to be stable. The field that contains the inferred peak paragenesis, clinopyroxene–orthopyroxene–garnet–ilmenite–K-feldspar–plagioclase–quartz–melt is predicted to be stable between 880 and 1030 °C at 8.6 to 12.1 kbar. The quantity of melt in this field varies from 5 to 20 mol.% (Figure 5).

Figure 5. Calculated P - T pseudosection for the bulk composition of VAB01, a garnet felsic granulite, using the augite model for clinopyroxene. Contours show the modal proportion of melt (dashed lines), and the solidus is represented by a bold red line.

The changing modal proportion of phases as a function of temperature was calculated at a fixed pressure of 10 kbar (Figure 6). At these conditions, the onset of partial melting occurs around 875 °C, at which point the rock contains 15% hornblende. With increasing temperature, the proportion of melt increases rapidly and the quantity of hornblende decreases until it disappears at 925 °C. The quantity of K-feldspar remains relatively constant (~ 10%) until hornblende is exhausted, thereafter it decreases and disappears around 1000 °C. The amount of

garnet remains constant (~ 8%) prior to the onset of partial melting, thereafter increasing (up to 13%) until the consumption of hornblende when the abundance of garnet decreases until it is exhausted around 1000 °C. The appearance of clinopyroxene broadly coincides with the onset of melting at this pressure, increasing to 4% as hornblende is consumed. Orthopyroxene appears at 900 °C during hornblende breakdown, reaching an abundance of 7%. The proportion of quartz remains constant until the onset of partial melting (around 27%), decreasing slightly towards higher temperatures (18% at 1050 °C). After the onset of the partial melting, plagioclase increases in abundance from 33% up to 38%, thereafter decreasing following the prograde exhaustion of K-feldspar and garnet (Figure 6). At the maximum modelled temperature (1050 °C), the melt fraction has reached ~35%.

Figure 6. Calculated changes in modal proportions of phases at 10 kbar for sample VAB01, a garnet felsic granulite.

The variation of melt composition (in weight %) at fixed pressure (10 kbar) has also been calculated (Figure 7). At lower temperatures, the H₂O content in the melt is 3% decreasing to <1% at 1050 °C, while the content of FeO + MgO increases from 2.5% at 850 °C to 8% at 1050 °C. As temperature increases, K₂O + Na₂O decreases from 8% to 6%, while CaO increases from 2% to 4%. The SiO₂ content decreases towards higher temperature from 66% to 63% and Al₂O₃ contents remain constant.

Figure 7. Calculated melt composition at 10 kbar. The modal proportion of the melt is indicated with a non-linear scale (*) at the top of the figure.

The P - T pseudosection for the garnet mafic granulite with 1% H_2O is shown in Figure 8A. The solidus has a positive slope but switches to negative at pressures above the disappearance of orthopyroxene, at around 9 kbar and between 830 and 880 °C. The hornblende-out line is broadly parallel to the solidus but occurs at higher temperatures, between 870 and 900 °C. Magnetite, garnet and orthopyroxene stabilities are strongly dependent on pressure. Magnetite and orthopyroxene are stable at lower pressures, magnetite below 8.2–9.4 kbar and orthopyroxene below 8.1–10.4 kbar. Garnet is predicted at higher pressures, above 7.6–9.8 kbar. Where hornblende is present also is quartz in the pseudosection, whereas in the hornblende-free region quartz is absent at pressures below ~9.5 kbar. Rutile occurs only at pressures above 12 kbar. Plagioclase, clinopyroxene and ilmenite are present throughout the P - T range considered. The paragenesis quartz–garnet–ilmenite–orthopyroxene–clinopyroxene–plagioclase–leucosome, interpreted to have been present at the metamorphic peak, is calculated to occur between 900 and 1030 °C, at pressures of 9 to 10.4 kbar, within a field no wider than 0.5 kbar. The modal proportion of melt calculated within this field varies from 5 to 20% (Figure 8A).

Figure 8. Calculated P - T pseudosection for a garnet-bearing garnet mafic granulite (PAR03), using an augite model for clinopyroxene. (A) Water content around 1 mol. %. (B) Water content of 0.01 mol. %. Contours for the modal proportion of melt are shown with dashed lines, and the solidus is represented by a bold red line.

The P - T pseudosection for the garnet mafic granulite with minimal water (Figure 8B; $\text{H}_2\text{O} \sim 0.01\%$) presents similarities with the one calculated for 1 mol.% H_2O . However, the solidus and hornblende-out lines coincide and occur at higher temperatures of between 860 and 900 °C. Garnet is stable at pressures above 7.4 kbar and orthopyroxene is stable at pressures below 8.8–10.7 kbar. Magnetite and rutile do not occur in the P - T window considered. Plagioclase, clinopyroxene and ilmenite are present throughout, and quartz is stable at all P - T conditions, except the lowest ones. The peak paragenesis occurs at temperatures above 880 °C and at pressures between 8 and 10.7 kbar, for which melt modal proportions are $<1\%$ (Figure 8B).

Zircon U-Pb Geochronology

Samples VAB01 and VAB19 both contain two types of zircon crystals with different shapes and internal CL domains (Figure 9). One type comprises subhedral prismatic crystals (~2:1 length to width aspect ratio) with rounded terminations, and dark luminescent cores with narrow to broad oscillatory-zoning, surrounded by bright-CL rims that are homogeneous or have sector-zoning. This zircon type yielded $^{206}\text{Pb}/^{238}\text{U}$ dates ranging from c. 631 to 705 Ma (Figure 10A). In both samples, the cores have similar Th/U ratios with an average value of 0.6 and contents of U and Th that are always higher than 100 ppm (Figure 10C). A second group of crystals is rounded and shows sector zoning ('soccer ball'), with bright to moderate CL response, occasionally with darker xenocrystic oscillatory-zoned cores (Figure 9). These bright CL rims and the soccer ball crystals appear to belong to the same age population and yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ dates of 593.8 ± 8.5 Ma in VAB01 and 598.5 ± 4.6 Ma in VAB19 (Figure 10A and B). Their Th/U ratios are

higher than the oscillatory-zoned cores and are often >1.0 (Figure 10C). This group is variably depleted in U compared to the first zircon type and can also be depleted in Th (Figure 10C).

Figure 9. CL images for representative zircon crystals from the three studied samples. Yellow ellipses show the location of SHRIMP spots ($\sim 20 \mu\text{m}$ diameter) and blue circles the location of LASS-ICP-MS spots ($\sim 50 \mu\text{m}$ diameter). Labels present spot ID, $^{206}\text{Pb}/^{238}\text{U}$ date (Ma), U content (ppm) and $\epsilon_{\text{Hf}(i)}$ values, when the latter is available. Labels in yellow show SHRIMP data, while labels in blue LASS-ICP-MS data. (A) Sample VAB01, oscillatory-zoned cores show older ages and higher U than soccer ball crystals. (B) Sample VAB19, oscillatory-zoned cores show older ages and higher U than rims and soccer ball crystals. (C) Sample RGM55, broad oscillatory- and sector zoned cores show the oldest ages, ghost oscillatory- and sector-zoned cores older than rims and soccer ball crystals. Uncertainties are 1σ .

Sample RGM55 contains three morphological types of zircon crystal (Figure 9). These include subhedral prismatic grains with rounded terminations, which are similar to those in VAB01 and VAB19, but with cores that are predominantly sector-zoned with only a few grains exhibiting broad-oscillatory patterns, and rims that are darkly luminescent with relatively high U, rather than bright-CL and low-U. Soccer ball zircons are also present, but have a darker CL

response and higher U content than those in VAB01 and VAB19 (Figure 10C). A third group comprises anhedral zircon grains with sector-zoning. In this sample, the prismatic sector-zoned cores yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of 624.8 ± 3.4 Ma (Figure 10D) with Th/U ratio of 0.56 (Figure 10C), although one spot from a broad oscillatory-zoned core yielded an older $^{206}\text{Pb}/^{238}\text{U}$ date of 639 ± 9 Ma (1σ error). Complementary U-Pb data obtained by LASS-ICP-MS in grains with broad oscillatory-zoned or some sector-zoned revealed oldest $^{206}\text{Pb}/^{238}\text{U}$ dates scattered from 631.4 ± 8.5 to 660.9 ± 7.1 Ma (Figure 10F) and the Th/U ratio is similar to the last group (Figure 10C). The rims, soccer ball and anhedral grains have lower Th/U ratios (0.26 – Figure 10C), with depleted Th contents relative to sector-zone cores and yielded a younger weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of 615.8 ± 3.4 Ma (Figure 10E), although two spots, rejected from this weighted mean age calculation, yielded younger $^{206}\text{Pb}/^{238}\text{U}$ dates of 599 ± 8 Ma and 594 ± 7 Ma (1σ error).

Figure 10. Concordia diagrams with zircon U-Pb isotopic data collected with SHRIMP II (2σ error ellipses) in (A), (B), (D) and (E), in which: (A) Data for sample VAB01 from soccer ball zircon (sb) and rims with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 593.8 ± 8.5 Ma and cores ranging up to c. 700 Ma. (B) Data for sample VAB19 from soccer ball zircon and rims with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 598 ± 4.6 Ma and cores ranging up to c. 700 Ma. (D) Zircon cores from sample RGM55 with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 624.8 ± 3.4 Ma. (E) Soccer ball and shapeless zircon and rims from sample RGM55 with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 615.8 ± 3.4 Ma. Contents of Th and U of different zircon types are shown in (C), in which: 1 – oscillatory-zoned cores from VAB01; 2 – rims and soccer ball crystals from VAB01; 3 – oscillatory-zoned cores from VAB19; 4 – rims and soccer ball crystals from VAB19; 5 – older broad oscillatory-zoned or sector-zoned cores from RGM55, obtained by LASS-ICP-MS; 6 – 625 Ma sector-zoned core from RGM55; and 7 – 615 Ma high U domain from RGM55. (F) Histogram of older dates found with complementary U-Pb data by LASS-ICP-MS for sample RGM55.

REE and Lu-Hf in zircon

The same samples analyzed with the SHRIMP were analyzed for REE abundances and Lu/Hf isotopes by LASS-ICP-MS (VAB01, VAB19 and RGM55). The analyses were made in the five different zircon domains differentiated on the basis of CL and U-Pb analysis: (i) oscillatory-zoned

cores (VAB01 and VAB19), (ii) sector-zoned to broad oscillatory-zoned crystals (RGM55), (iii) sector-zoned cores (RGM55), (iv) soccer ball crystals and rims with high U (RGM55) and (v) soccer ball crystals and rims with low U (VAB01 and VAB19).

Sample VAB01 shows three distinct patterns of zircon REE distribution (Figure 11A). The oscillatory-zoned cores, which yielded the oldest dates up to c. 705 Ma, have moderately enriched in LREE relative to chondrite, with weak or no Eu anomalies and strongly enriched in HREE, up to 10,000 times more than chondrite (Figure 11A). The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios are concentrated between 0.282223 and 0.282251 (Figure 11F), which yielded $\epsilon\text{Hf}(t)$ between -4.73 and -4.11, with an outlier value of 0.282325 ($\epsilon\text{Hf}(t) = -0.87$). The zircon rims are depleted in La with a pronounced positive Ce anomaly and a weakly negative Eu anomaly, and a steeper HREE array, although the enrichment in HREE content are lower than the oscillatory-zoned cores (Figure 11A). The soccer ball crystals, which yielded the same ages as the rims (c. 595), present the same pattern of LREE and MREE than the rims, however, the HREE array is flat or exhibits a slightly enrichment (Figure 11A). Zircon rims and soccer ball grains have similar $^{176}\text{Hf}/^{177}\text{Hf}$ ratios, between 0.282245 and 0.282314 (Figure 11F), with $\epsilon\text{Hf}(t)$ between -6.02 and -3.31.

Figure 11. Zircon REE chondrite-normalized pattern (Sun and McDonough, 1989) in (A) to (E), in which: (A) VAB01; (B) VAB19; (C) Older spot ages from RGM55; (D) 625 Ma sector-zoned cores from RGM55; (E) 615 Ma high-U domains in RGM55. (F) $^{176}\text{Hf}/^{177}\text{Hf}$ against $^{206}\text{Pb}/^{238}\text{U}$ for different zircon types, in which: 1 – oscillatory-zoned cores from VAB01; 2 – soccer ball crystals from VAB01; 3 – rims from VAB01; 4 – oscillatory-zoned cores from VAB19; 5 – soccer ball crystals from VAB19; 6 – rim from VAB19; 7 – older domain from RGM55; 8 – 625 Ma sector-zoned cores from RGM55; 9 – 615 Ma high-U domains from RGM55.

Zircon from sample VAB19 shows similarities with zircon from VAB01 (Figure 11B). Oscillatory-zoned cores (up to c. 700 Ma) exhibit a moderately enriched in L- and MREE, with no Eu anomalies and a strongly enriched in HREE (Figure 11B). The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of this group are similar to the zircon oscillatory-zoned cores from sample VAB01, with $\epsilon\text{Hf}(t)$ between -4.04 and -4.02 (Figure 11F). A single analysis from a rim shows a strongly enriched pattern in LREE,

higher than the oscillatory-zoned cores, and in HREE, lower than the oscillatory-zoned cores, with a negative Eu anomaly (Figure 11B). This rim yielded the lowest $^{176}\text{Hf}/^{177}\text{Hf}$ value of 0.281962 (Figure 11F), hence the most negative $\epsilon\text{Hf}_{(t)}$ (-15.9). Soccer ball crystals, which yielded the same age as the rim (*c.* 600 Ma), show the same pattern as soccer ball grains from VAB01, with depletion in La, strongly positive Ce anomalies, weakly negative Eu anomalies and flat distribution of HREE (Figure 11B). The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios are slightly higher than the oscillatory-zoned cores, between 0.282254 and 0.282271 (Figure 11F) and $\epsilon\text{Hf}_{(t)}$ between -5.92 and -4.12.

In sample RGM55, the sector-zoned to broad oscillatory-zoned domains, which ranging from 631 and 661 Ma, present both enrichment and depletion in LREE, with weakly to strongly positive Ce anomalies and negative Eu anomalies (Figure 11C). There are an enriched pattern in HREE. Zircon crystals from this sample are less radiogenic than the ones from the other samples (VAB01 and VAB19). In this domain, they show a $^{176}\text{Hf}/^{177}\text{Hf}$ ratios between 0.282148 and 0.282207 and the $\epsilon\text{Hf}_{(t)}$ ranges from -8.48 and -6.36 (Figure 11F).

The prismatic crystals from RGM55 with sector-zoned cores and dates of *c.* 625 Ma shows two different pattern of LREE: one with moderately enriched and weakly positive Ce anomalies and another depleted in LREE with strongly positive Ce anomalies. M- and HREE distribution are similar and present negative Eu anomalies and strong enrichment in the heaviest elements (Figure 11D). The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios range from 0.282150 and 0.282181 (Figure 11F) and the $\epsilon\text{Hf}_{(t)}$ from -8.55 and -7.34.

The REE array in the anhedral and soccer ball grains from sample RGM55 (*c.* 615 Ma) shows spread in LREE distribution with a positive Ce anomaly less pronounced than the sector-zoned cores, although an outlier spot is enriched in LREE without Ce anomaly (Figure 11E). The MREE shows negative Eu anomalies and, as the other domains in this sample, there are a steeper enrichment in HREE. The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios, between 0.282133 and 0.282201, and $\epsilon\text{Hf}_{(t)}$, between -9.71 and -7.04, are similar to the other domains (Figure 11F).

DISCUSSION

P-T path

The metamorphic *P-T* path of the base of Socorro-Guaxupé Nappe can be constraint with sample VAB01 due to its particular petrographic features. Biotite grains armored in garnet have rounded shapes, high Ti content and X_{Mg} . Therefore, the earliest stages of the partial melting may have been by reactions consuming biotite as the hydrous mineral. Nevertheless, to reach the UHT conditions and preserve mineral assemblages, the rock is likely to have become depleted in these first batches of melt (Clark et al., 2011; Kelsey and Hand, 2015; Vielzeuf et al., 1990), and hence less prone to thermal buffering. Thus, the metamorphic modelling here calculated cannot reflect

the prograde composition and metamorphic path that the granulites were subjected. However, the metamorphic peak and the retrograde path can be constrained by the pseudosection.

According to the thermodynamic models presented here, melting was driven by breakdown of K-feldspar and garnet once the rock was exhausted of hydrous minerals (Figure 6). The granulite is extremely depleted in Th and U, which highlight the melt loss added with the depletion in Rb and Y (Figure 3). A small enrichment in Sr and Ti suggests that plagioclase and Ti phases were not involved in intensive way as partial melting reactants (Figure 3).

The calculated pseudosection assemblage fields for the inferred metamorphic paragenesis in both the garnet felsic and mafic granulites overlap and are consistent with peak conditions of around 1000 °C at pressures higher than 9 kbar (Figure 12). Calculations using the average P - T mode of THERMOCALC yielded similar conditions, coincident with P - T estimates based on the maximum Al-in-orthopyroxene of the garnet felsic granulite (Figure 12).

Figure 12. Summary of P - T - t estimates for rocks of the Guaxupé Domain. Peak fields for samples VAB01 (felsic granulite) and PAR03 (mafic granulite) are overlain to show the region of overlap. Al^T isopleth in orthopyroxene is at similar conditions to those calculated in the average P - T mode of THERMOCALC. The post-peak evolution leads to the magnetite growth coexisting with garnet, this moment coincides with the crystallization of high U rims and the formation of xenomorphic and soccer ball zircon at c. 615 Ma. The end of the crystallization of the melt happened at 870 °C and 8.5 kbar and it is marked by the crystallization of low U rims and soccer ball zircon depleted in HREE at c. 595 Ma. The prograde evolution is dashed because the uncertainties around the protolith, and partial melt and melt segregation mean that the present-day bulk composition may not be representative of previous conditions, although sector-zoned zircon grains are representative of this period at c. 625 Ma.

The retrograde *P-T* path started with a small decrease in pressure to 9 kbar (Figure 12), by a discrete orthopyroxene replacing garnet and in which magnetite was exsolved from ilmenite (Figure 2). At this point, the rock followed a near-isobaric-cooling path, in the stability field of garnet, since coronas or other textures of garnet replacement are rare, and when present poorly developed. Clinopyroxene formed a partial corona at rims of orthopyroxene and the same feature is observed in the garnet mafic granulite. At temperatures close to the solidus, hornblende began to crystallize at the edge of pyroxene, particularly in leucosome, at 870 °C and 8.5 kbar (Figure 12).

Metamorphic timescales

The geochronologic data obtained reveal at least four groups of ages that can be related to distinct periods of zircon crystallization: one related to early-orogenic magmatism, one related to prograde to peak metamorphism and two related to the melt crystallization (Figure 12). Although the garnet felsic granulite and the diatexite appear not to have the same protolith some comparison can be done.

In samples VAB01 and VAB19, zircon crystals with oscillatory-zoned cores exhibit higher U and lower Th/U ratio than their rims or soccer ball crystals (Figure 10C), and they also display older dates, ranging from *c.* 630 Ma up to 705 Ma. In a zircon/garnet REE partitioning diagram, this domain, in sample VAB01, zircon has higher concentration of LREE and HREE, and similar contents of MREE than any garnet domain (Figure 13A, C and E). Conversely, in sample VAB19 this domain is just enriched in LREE and depleted in M and HREE related to garnet (Figure 13G). However, a novel approach using an array plot of $\log (D_{Yb}/D_{Gd})$ vs. $\log (D_{Yb})$ of every zircon analyses against all different garnet compositions has been proved to deal better with the zircon/garnet equilibrium (Taylor et al., 2017). Thus, no matter how garnet compositional domain is used, zircon cores from VAB01 did not grow in equilibrium with garnet because it shows a negative slope array (Figure 13B, D and F), which is the opposite that is observed in experimental data equilibrium for the pair (Rubatto and Hermann, 2007; Taylor et al., 2015). This domain in sample VAB19, although exhibits a positive slope in the array diagram, shows a displacement from the experiments (Figure 13H), therefore, these zircon cores were not in equilibrium with garnet.

Sample RGM55 also shows older ages scattered from *c.* 631 to 660 Ma in oscillatory-zoned or sector-zoned cores of zircon, although they have distinct REE and Hf signatures. Thereby, these oldest dates from the three samples may reflect the age of igneous crystallization of the protolith and do not have relation to the metamorphism, as observed in others studies (Janasi, 2002, 1999; Martins et al., 2009; Mora et al., 2014; Rocha et al., 2018, 2017; Tedeschi et al., 2018).

Figure 13. Zircon and garnet trace elements data from garnet felsic granulite (VAB01 and VAB19). Traditional D_{REE} (zircon–garnet) plot in (A), (C), (E) and (G). Array plot, purposed by Taylor et al. (2017), showing D_{REE} (zircon–garnet) for all zircon against average garnet domains in (B), (D), (F) and (H). (A) and (B) zircon data plotted against garnet cores depleted in HREE from sample VAB01 (core 1 from Figure 4B). (C) and (D) zircon data plotted against cores/intermediate portion of garnet from sample VAB01 (core 2 from Figure 4B). (E) and (F) zircon data plotted against garnet rims from sample VAB01 (rim from Figure 4B). (G) and (H) zircon data plotted against the single garnet domain from sample VAB19 (Figure 4D). (*) zircon–garnet data from Taylor et al. (2015) performed at 7 kbar. (**) zircon–garnet data from (Rubatto and Hermann (2007) performed at 20 kbar.

In the diatexite (RGM55) subhedral prismatic zircon with sector-zoned cores record a weighted mean date of 625.9 ± 3.5 Ma. Based on its bulk REE pattern (enriched in LREE), mineralogy (allanite-bearing and garnet-depleted), and REE pattern in zircon (with pronounced positive Ce and negative Eu anomalies), this sample may not have the same protolith as the garnet felsic granulite. Thus, this date of c. 625 Ma may represent the age of protolith crystallization, although a previously mentioned oscillatory-zoned core yields older date (> 630 Ma). Another possibility is that the c. 625 Ma date could represent recrystallization of inherited material, since

this process can result in expulsion of the pre-existing Pb, and resetting of the U-Pb system (Taylor *et al.*, 2016). Grains of this domain show a progressive loss of the original internal features, which is commonly observed in high-grade conditions (Taylor *et al.*, 2016), as can be observed in grain #3 (SHRIMP analysis) in Figure 9C that shows a ghost of oscillatory-zoned texture. This age matches the timing of the metamorphic peak in the Socorro-Guaxupé Nappe as determined in other studies (Rocha *et al.*, 2018, 2017; Tedeschi *et al.*, 2018).

The rims of most zircon crystals, soccer ball and shapeless zircon crystals in sample RGM55 register the highest contents in U (up to 347 ppm) and lowest Th/U ratios (~ 0.23) of the studied samples (Figure 10C). They yield a date of 615.8 ± 3.4 Ma, which represents the age of the onset of melt crystallization (Figure 12), since the dissolution of previously crystallized zircon may, at the first stage, provide high U contents to the melt (Watson and Harrison, 1983; Yakymchuk *et al.*, 2018). In comparison to their cores, they are depleted in Th and have a wider variable of LREE contents, suggesting that allanite, a sink for these elements (Hermann, 2002) had also started to crystallize from the melt. This was made possible by the very high abundance of LREE in the bulk composition which will stabilize allanite at such high temperature conditions (Hermann, 2002). Similar ages are also interpreted as the time of melt crystallization in other studies from rocks of the Socorro-Guaxupé Nappe (Martins *et al.*, 2009; Mora *et al.*, 2014; Rocha, 2016; Rocha *et al.*, 2017).

The rims and soccer ball zircon from samples VAB01 and VAB19 record the lowest contents of U, the highest Th/U ratios and the youngest ages (Figure 10C). Major phases are more able to incorporate U rather than Th (Yakymchuk *et al.*, 2018), thus, without the presence of a Th-bearing phase, the Th/U ratios in the melt tend to increase during crystallization. Therefore, this group may reflect the final crystallization of the trapped melt (Figure 12), when the availability of U was low.

The D_{REE} for zircon–garnet from sample VAB01 for soccer ball crystals shows a flat pattern of MREE and HREE (Figure 13C and E), excepted if they were normalized to the core of larger garnet grains (Figure 13A). Other than that, the data from zircon rims show flat pattern of MREE and enrichment in HREE (Figure 13A, C and E). However, in the array plot zircon, although all zircon data show the same positive slope than the experimental data, they are detached from them if compared to this garnet domain (Figure 13B), thus, they were not in equilibrium. The garnet cores from smaller crystals or the intermediate domain of larger crystals and the garnet rims from sample VAB01 fit better the zircon data, which show the same slope and trends than the experimental data (Figure 13D and F), suggesting that they were in equilibrium. A small variation in the garnet composition between cores and rims indicates they were homogenized by the high-grade metamorphic conditions (Figure 4). The zircon rims are more suitable with the high

pressure experimental data from Rubatto and Hermann (2007), whereas the array of soccer ball crystals fits better with the lower pressure experimental data from Taylor et al. (2015). Although, zircon rims and soccer ball crystals yielded the same dates (*c.* 595 Ma), zircon rims could be crystallized earlier at higher pressures, taking advantage of pre-existing crystals, than the truly metamorphic zircon grains, the soccer ball ones. Thus, this behaviour implies that the low U zircon group from VAB01 grew in equilibrium with garnet, which is supported by the flat pattern of HREE in both minerals (Figure 4 and Figure 11), once both were competing to these elements.

Zircon rims and soccer ball crystals D_{REE} for zircon–garnet plot from sample VAB19 show M and HREE depleted in comparison with garnet, although the analyzed rims show the same pattern as their cores (Figure 13G). However, as seen in the array plot (Figure 13H), the data are displaced from the experimental data, which indicates that zircon and garnet were not in equilibrium. In this case, garnet may be crystallized earlier than zircon, since it shows an enriched pattern in HREE (Figure 4), while zircon shows a flat HREE pattern (Figure 11). HREE is strongly partitioned into garnet (Chen et al., 2010), thus garnet had already sequestered HREE when zircon was crystallized.

The isobaric-cooling path earlier discussed shows that the garnet modal proportion increases toward low temperature, as predicted in the pseudosection modelling (Figure 6). It is likely that the date printed in the low U zircon group from VAB01 and VAB19, *c.* 595–600 Ma, represents the age of late trapped melt crystallization. Therefore, with the onset melting crystallization at *c.* 615 Ma, the duration of melt crystallization is inferred to last for *c.* 20 Myr (Figure 12).

Contribution to the regional tectonic framework

Calculated *P-T* conditions predicted UHT metamorphic conditions for all three samples, which are localized thirty kilometers apart from each other. Consistent *T* values higher than 900 °C are calculated with several methods at pressures higher than 10 kbar at the base of Socorro Guaxupé Nappe. These UHT conditions have also been observed in other rocks associated with the Socorro-Guaxupé Nappe (*e.g.* Del Lama et al., 2000; Rocha et al., 2017). Extensive and associated charnockite magmatism is found in the area, with magmas generated at temperatures higher than 900 °C (Janasi, 2002; Mora et al., 2014). Therefore, it is likely that the UHT conditions were widespread within the orogen, although a diagnostic UHT mineral association has never been recognized.

The Guaxupé Domain is composed of different sorts of protoliths, including pelites, mafic and felsic igneous rocks, with felsic igneous rocks being the most voluminous. The magmatism crystallization ages seem to be slightly younger than the metamorphic event (Campos Neto et al.,

1996; Campos Neto and Caby, 2000; Mora et al., 2014; Rocha, 2016; Rocha et al., 2017; Tedeschi, 2018). However, recently zircons of Neoarchean to Paleoproterozoic age were dated in residue and leucosome of the basal granulites (Tedeschi, 2018; Tedeschi et al., 2018).

In this work, the inherited cores of analyzed zircon crystals from garnet felsic granulite (VAB01 and VAB19) have dates between 630 and 705 Ma and their REE pattern is characterized by relative enrichment in LREE and strong enrichment in HREE, which generally is different from patterns of other identified zircon groups. In discrimination diagrams proposed by Belousova et al. (2002), their composition is consistent to that of zircon crystallizing in igneous rocks of TTG affinity. Zircon crystals of a diatexite (RGM55) also show older ages scattered between 630 and 660 Ma and may reflect inherited ages, although their REE patterns and Hf signatures are different from the garnet felsic granulite. These dates suggest a protracted magmatic event, lasting *c.* 85 Myr, although they could also indicate variable Pb loss along the Concordia from a single population of *c.* 705 Ma magmatic zircon during UHT metamorphism. Many authors advocate that these cores are related to the generation of a pre-collisional magmatic arc (*e.g.* Campos Neto et al., 1996; Campos Neto and Caby, 2000; Mora et al., 2014). However, in this study, their Hf signatures show a crustal source for the magmas. Actually, there are no data with juvenile signatures, and probably these rocks were intruded and reworked older continental crust, as indicated by the negative $\epsilon\text{Hf}_{(t)}$ values.

The ages of *c.* 625 Ma, related to the prograde path, and *c.* 615 Ma, related to the onset of melt crystallization, bracket the age of the metamorphic peak. A scenario with a protracted high-grade event and an isobaric-cooling path is extensively discussed in literature (*e.g.* Clark et al., 2015; Harley, 2008; Kelsey and Hand, 2015; Korhonen et al., 2013; Laurent et al., 2018), and seems to fit well with the Guaxupé Domain.

CONCLUSIONS

In the southernmost portion of Brasília Orogen, a stack of nappes from Guaxupé Domain underwent UHT conditions with peak metamorphism at 1000 °C and 11.5 kbar. Hydrated phases were not part of the partial melt reactions at the metamorphic peak, and instead it seems breakdown of K-feldspar and garnet played an important role in melt generation at the highest temperatures. These conditions were preceded by recrystallization of inherited zircon at *c.* 625 Ma. A stage of decompression from the peak conditions to 9 kbar was followed by crystallization of melt and high-U zircon grains and rims at *c.* 615 Ma. A final period of melt crystallization occurred at 870 °C at 8.5 kbar in a field where magnetite and garnet were crystallizing and is marked by soccer ball zircon crystals and rims depleted in U and HREE.

This evolution defines a clockwise P - T path with a short period of decompression that is younger than 625 Ma but no older than 615 Ma, followed by a long period of near isobaric-cooling until 595 Ma, suggesting that the lower crust was heated to high-grade conditions for at least 30 Myr and that the crystallization of partial melt lasted 20 Myr.

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SUPPLEMENTARY MATERIAL

H ₂ O	SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	O	
0.01	69.67	9.86	5.74	3.96	5.48	1.45	3.18	0.59	0.55	x=0
5	69.67	9.86	5.74	3.96	5.48	1.45	3.18	0.59	0.55	x=1

Figure S1. T-X pseudosection, where X varies from 0.01 to 5 mol. % of H₂O. Dashed lines are the modal proportions of melt. The chosen value (red dashed line) was that cover among 5 and 20% of melt at the assemblage peak (aug-opx-g-ilm-ksp-L).

H2O	SiO2	Al2O3	CaO	MgO	FeO	K2O	Na2O	TiO2	O	
1	69.67	9.86	5.74	3.96	5.48	1.45	3.18	0.59	0.2	x=0
1	69.67	9.86	5.74	3.96	5.48	1.45	3.18	0.59	1.75	x=1

Figure S2. P-X pseudosection, where X_{Fe2O3} varies from 0.2 to 1.75 mol. %. An X_{Fe2O3} value of 0.98 mol. % was chosen (blue dashed line) to eliminate rutile from the calculations.

**ANEXO II – THE ORIGIN OF CHARNOKITES FROM GRANULITE
PARTIAL MELTING: AN EXAMPLE IN THE SOCORRO-GUAXUPÉ NAPPE,
SOUTHEASTERN BRAZIL**

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The origin of charnockites from granulite partial melting: an example in the Socorro-Guaxupé Nappe, southeastern Brazil

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ABSTRACT

A large slice of continental lower crust outcrops in the southernmost Brasília Orogen, Brazil, where rocks were metamorphized at UHT conditions in the Socorro-Guaxupé Nappe. The metamorphic rocks there have a long history of melt crystallization between 630 and 595 Ma, and several granitic bodies were generated, including charnockites, which present same ages of crystallization. One of this charnockite suite, Paraguaçu Suite (*c.* 630 to 611 Ma), is subdivided in two bodies, north and south, and presents intermediate composition with high #Mg, Ba, Zr, Hf and REE contents and depleted in Sr, Th and U, suggesting a depleted source that K-feldspar, zircon and garnet were been molten. Metamorphic modelling performed in a felsic granulite, from Granulitic Unit, shows that K-feldspar, zircon and garnet were broken during the UHT stage and the melt generated has similar trends of major elements as those found in Paraguaçu Suite. However, Paraguaçu Suite has higher Na₂O contents and lower SiO₂ than what is predicted in the partial melting granulite model. Thus, it is likely that the migration and crystallization of the felsic granulite partial melting was the major source for the generation of Paraguaçu Suite, and mixture with other source(s), probably more radiogenic, evidenced by a least negative εHf in zircon in the north body, which also is enriched in Sr and Ti. The mixture with radiogenic sources and the radioactive decay of a large volume of partial melt generated in this segment of Socorro-Guaxupé Nappe were the responsables to the heating to UHT conditions.

Keywords: Charnockite source, heat source, phase equilibrium modelling, U-Pb geochronology, Hf isotope, Nd isotope, Brasília Orogen

INTRODUCTION

The granitic magmatism is one of the most important process of formation and recycling of the continental crust, particularly during orogenic events, resulting in the geochemistry differentiation of the crust since the Archean (Brown, 2001). The chemical composition of the resulting magma depends on the composition of sources and the extent of partial melting (Faure, 1998). Therefore, a geochemical study of the igneous rocks provides insights of their sources, being a classical approach to reveal process and sources of the magmas (e.g. Brown et al. 1984; Pearce et al. 1984; Pearce 1996; Janasi 2002; Alves et al. 2016).

A new approach to investigate partial melting and magma formation counts with the recent development of the activity-composition models for tonalite-trondhjemite melt compositions (Green et al., 2016), which combined with thermodynamics of minerals end-member properties (Powell and Holland, 1988), allowing to access and model process of crust formation and differentiation (e.g. Holland et al., 2018; Palin et al., 2016), including how the first continental crust was formed and stabilized (Johnson et al., 2017). Thus, now it is possible do the reverse path, *i.e.*, access the product composition via the thermodynamic modelling of the source.

The present work uses this new approach to retrieve the source of the charnockite magmatism in the Socorro-Guaxupé Nappe, southernmost Brasília Orogen, in Brazil, using thermodynamic equilibria modelling, via THERMOCALC (Powell et al., 1998) coupled with accessory mineral behaviour (Kelsey et al., 2008). It combines isotopic and elemental geochemistry of whole rock and zircon geochronology.

GEOLOGICAL BACKGROUND

The Socorro-Guaxupé Nappe is an allochthonous high-grade metamorphic terrane, with granulites, migmatites and granitic rocks in a horizontal stack of nappes that were formed during the Neoproterozoic (~ 640–600 Ma), in the southernmost Brasília Orogen, at the southwestern border of the São Francisco Craton. It represents a magmatic arc setting that developed in a continental active margin at Paranapanema Plate (Brito Neves et al., 1999; Campos Neto et al., 2011; Campos Neto and Caby, 2000, 1999).

The Socorro-Guaxupé Nappe shows a right-way-up crustal section of hot and partially melted layered crust and is lithologically subdivided into: (1) Metatexite on top; (2) Diatexite in intermediate; and (3) Granulite on its base (Campos Neto and Caby, 2000). The structure is segmented into two domains (Del Lama et al., 2000; Janasi, 2002), which high-angle lateral ramps

are developed: Guaxupé Domain in the northwestern where granulites and diatexites are dominant (Fig. 1), and Socorro Domain in southeastern where metatexites are dominant.

Fig. 1. (A) The southernmost Brasília Orogen and its relation with Neoproterozoic Cratons. (B) Geological framework of the Guaxupé Domain, in which the three charnockite suites are highlighted in shades of red (modified after Campos Neto et al. 2010).

The Guaxupé Domain records high to ultra-high temperature (UHT) of metamorphism. Thermobarometric studies of granulites show a range in T conditions above 800 °C, and sometimes reaching 1000 °C (Campos Neto and Caby, 2000; Del Lama et al., 1998, 1994; Iyer et al., 1996; Rocha et al., 2018, 2017; Tedeschi, 2018). The heat that produced the metamorphism

was able also to generate large volumes of magma. Among several felsic intrusions three suites stand out due to their anhydrous character: Divinolândia, São Pedro de Caldas (Janasi, 2002, 1999, 1997) and Paraguaçu (Barros, 2012).

The Socorro-Guaxupé Nappe has sinuous edges lying above the passive margin assemblage and the Andrelândia Nappe System, which are composed by metasedimentary rocks metamorphosed at high-pressure conditions (Campos Neto et al., 2010; Campos Neto and Cabyl, 1999; Motta and Moraes, 2017).

Charnockites from Guaxupé Domain

The charnockite bodies are intrusive in migmatites, which present different rates of partial melting. Usually they have a semi-tabular shape parallel to the regional foliation. They also show a solid-state foliation consistent to the regional structure. Two large suites, São Pedro de Caldas, more felsic, and Divinolândia, were detailed by Janasi (1999; 2002) and present contrasting features.

São Pedro de Caldas Suite is composed by orthopyroxene monzonite, which has a compositional continuum with leucogranite. This suite has a low color indices (8-15) and quartz contents between 5 and 15 vol.% (Janasi, 2002).

Divinolândia Suite presents coarse-grained orthopyroxene monzonite, with minor amount of orthopyroxene monzogranite and quartz syenite. In the orthopyroxene monzonite is common the presence of enclaves of pyroxene diorites. The Divinolândia Suite has a higher color indices (>12-15), lower quartz contents and slightly more plagioclase than São Pedro de Caldas Suite (Janasi, 2002, 1999).

PARAGUAÇU SUITE

The Paraguaçu Suite presents medium- to coarse-grained green charnockites with a quartz syenite composition. It is composed of two bodies, both elongated N-S, intrusive in the Diatexite Unit (Fig. 2A), although sometimes the contacts are parallelized with the regional foliation (Fig. 2B). Actually, the foliation occurs throughout the bodies, parallel to the regional main foliation, partially obliterating original igneous textures.

Rocks from Paraguaçu Suite has quartz-monzonite to quartz-syenite compositions. Perthitic to mesoperthitic K-feldspar and homogeneous oligoclase are the main felsic minerals. As mentioned, the matrix exhibits deformational fabric, with a granoblastic texture, although some feldspar phenocrysts are preserved (Fig. 2C). The exsolutions in K-feldspar are often concentrated in the core of the crystals, with clean rims (Fig. 2D). The occurrence of orthopyroxene is widespread all over the bodies reaching up to 5% in modal volume (Fig. 2E). Hornblende is the main mafic mineral and occurs as phenocrysts (Fig. 2B), although sometimes it appears

surrounding orthopyroxene (Fig. 2F). Clinopyroxene is subordinate and when present has exsolution lamellae of clinoenstatite (Fig. 2G). Biotite is scarce to absent. The Fe-Ti oxides are ilmenite and magnetite and the most abundant accessory minerals, occurring as irregular-shaped aggregates to prismatic crystals. Zircon is abundant and forms large prismatic crystals, as well as apatite (Fig. 2F).

Fig. 2. (A) Intrusive contact between the charnockite and the diatexite generating some mixture zones (RGM54). (B) Parallelized contacts with the main foliation between the charnockite and diatexite and phenocrysts of hornblende in the charnockite (RGM57) – both in (A) and (B) the rule has 30 cm. (C) Granular to granoblastic fabric in the matrix, and the preservation of K-feldspar phenocrysts in the charnockite (PAR40). (D) Perthitic K-feldspar with coarsed exsolution lamellae of plagioclase in the core and clear rim (PAR32C). (E) Cluster of orthopyroxene parallel to the main foliation surrounding by hornblende (RGM53). (F) Cluster of hornblende wrapping an orthopyroxene grain and huge crystals of accessory phases, such as apatite and zircon (PAR40). (G) Orthopyroxene lamellae hosted in a clinopyroxene crystal (PAR32C).

Elemental geochemistry

Whole-rock major and trace element composition of studied rocks are presented in ANEXO VI. Analytical procedures are described in Appendix A.

The rocks from Paraguaçu Suite present a narrow range of SiO_2 , between 60.33 and 62.75 wt.% (Fig. 3). They show a negative correlation between SiO_2 and FeO, MgO and CaO (Fig. 3A to C), whereas positive with Na_2O and Al_2O_3 (Fig. 3E to F). The #Mg varies between 18 and 25 and show negative correlation with SiO_2 . The contents of major elements are intermediate to the other charnockite suites, which in general present negative correlation in Hacker diagrams against SiO_2 .

Fig. 3. Behaviour of major elements against SiO_2 in the charnockites of Guaxupé Domain. The granulite partial melting composition calculated in THERMOCALC are also shown. (*) São Pedro de Caldas and Divinolândia Suites data extracted from Janasi (2002).

The REE patterns in the Paraguaçu Suite (Fig. 4) present a large range of fractionation ($\text{La}_N/\text{Yb}_N = 10.4 - 35.4$, with 2nd quartile at 12.34), with more degree related to LREE ($\text{La}_N/\text{Sm}_N = 2.37 - 6.40$), which is 40 to 300 times richer than chondrites. The HREE are slightly less fractionated ($\text{Gd}_N/\text{Yb}_N = 2.10 - 3.05$) and are 4 to 30 times richer than chondrite. There is a narrow range of Eu concentration, with anomalies varying from positive to slightly negative ($\text{Eu}/\text{Eu}^* = 0.77 - 2.06$). REE contents, in Paraguaçu Suite, mostly overlap the other charnockite suites, although sometimes the REE contents are slightly depleted.

Fig. 4. Chondrite normalized REE patterns for Paraguaçu Suite (this work), São Pedro de Caldas Suite and Divinolândia Suite (both are data compiled from Janasi, 2002).

U-Pb geochronology and zircon composition

Zircon crystals from four samples were analyzed (RGM02, 04, 53 and 57). Analytical procedures are described in Appendix A. Data are presented in ANEXO VIII and ANEXO X.

In all samples, the zircon crystals present similar shapes. They are prismatic, with elongation ratios ranging 1:3 or 1:4 and, sometimes, doubly terminated. Their cores often have dark CL features showing oscillatory to truncated internal zoning. Convoluted zones give the transitions to the rims, which show no zoning or sector zoning and its response to CL is bright (Fig. 5).

The distribution of dates shows the same pattern for all samples. A histogram of dates shows a peak of data between 630 and 615 Ma, with a skewed pattern toward younger dates, until dates <600 Ma (Fig. 6). Generally, older dates are related to cores and younger to rims, although some grains yielded the same date even bearing different textures (Fig. 5). Younger dates are also direct related to a decrease in the Th/U ratio (Fig. 6), which is always higher than 0.1.

Fig. 5. CL images of representative zircon grains from Paraguaçu Suite. Spot ID are labelled in black; the yellow ellipse shows the location of SHRIMP spot ($\sim 20 \mu\text{m}$) and respective $^{206}\text{Pb}/^{238}\text{U}$ ages (Ma) are also labelled in yellow; the blue circle shows the location of LA-ICP-MS spot ($\sim 50 \mu\text{m}$) and respective $\epsilon\text{Hf}(t)$ are labelled in blue. (A) RGM02; (B) RGM04; (C) RGM53; (D) RGM57.

Sample RGM02 has the largest spreading in $^{206}\text{Pb}/^{238}\text{U}$ spot dates, ranging from 639 ± 10 Ma to 557 ± 28 Ma (1σ uncertainty – Fig. 6A) and a sharp trend of increasing of Th/U ratio to older dates (Fig. 6C). Excluding dates from the skew, this sample yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of 622 ± 6.8 Ma (Fig. 6B).

Fig. 6. U-Th-Pb zircon data. $^{206}\text{Pb}/^{238}\text{U}$ ages distribution for samples RGM02, 04, 53 and 57, respectively in (A), (D), (G) and (J), in which they generally show a long tail toward younger ages. Concordia diagrams (2σ error ellipses) for samples: RGM02 (B), which yielded 622 ± 2.8 Ma (red ellipses were rejected from the calculations); RGM04 (E), which presented two populations (630.3 ± 2.9 Ma in black and 602.5 ± 3.5 Ma in red); RGM53 (H), which yielded 619.6 ± 5.8 Ma; and RGM57 (L), which yielded 611.4 ± 4.8 Ma (red ellipses were rejected from the calculations). Th/U ratios against $^{206}\text{Pb}/^{238}\text{U}$ ages for samples RGM02, 04, 53 and 57, respectively in (C), (F), (I) and (K), in which they generally exhibit a positive slope.

Sample RGM04 shows two peak of $^{206}\text{Pb}/^{238}\text{U}$ spot dates (Fig. 6D). The oldest peak yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ of 630.3 ± 2.9 Ma, whereas the youngest peak provided 602.5 ± 3.5 Ma (Fig. 6E). Overall, the youngest group has a more spread and lower values of Th/U ratios compared to the oldest group of dates (Fig. 6F).

Sample RGM53 has a more symmetric spread in $^{206}\text{Pb}/^{238}\text{U}$ spot dates (Fig. 6G), although they are more concentrated, ranging from 604 ± 11 Ma to 638 ± 12 Ma (1σ uncertainty). Data gave

a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of 619.6 ± 5.8 Ma (Fig. 6H). There is no pattern related to the Th/U ratios (Fig. 6I).

Sample RGM57, the only located in the north body, shows $^{206}\text{Pb}/^{238}\text{U}$ spot dates among 588 ± 9 Ma and 641 ± 9 Ma (1σ uncertainty – Fig. 6J). The zircon rims in this sample are the well developed (Fig. 5). The weighted mean $^{206}\text{Pb}/^{238}\text{U}$ date of the peak in the histogram is the youngest among the samples, yielding 611.4 ± 4.8 Ma (Fig. 6K). The same trend of decreasing of Th/U ratios toward younger dates is observed (Fig. 6L).

The zircon grains from all samples have similar composition of REE (Fig. 7). They present a spread pattern in LREE, with Ce anomalies varying from strongly positive to weakly negative ($\text{Ce}/\text{Ce}^* = 0.74 - 111.67$). The MREE is characterized by enrichment in relation to chondrite, with a negative Eu anomaly. The HREE pattern shows a steep enrichment toward Lu.

Fig. 7. REE patterns of zircon crystals, chondrite normalized (Sun and McDonough, 1989) of Paraguaçu Suite. (A) RGM02, (B) RGM04, (C) RGM53 and (D) RGM57.

Isotopic geochemistry

Nd and Sr signatures

The Sr and Nd isotopic compositions of six samples are reported in ANEXO VII, and analytical procedures are described in Appendix A. All analyzed samples are from southern body. Model ages and ϵNd values were calculated using GCDkit software (Janoušek et al., 2006) and they were set for 620 Ma (further information see discussions below).

The Sr and Nd parameters are very homogeneous among the samples. $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio ranges from 0.7095 to 0.7101 (Fig. 8A), ϵNd_i is negative, ranging between -11.8 and -12.4, and the $\text{Nd } T_{\text{DM}}$ varies from 2.20 to 2.32 Ga (Fig. 8B).

Fig. 8. ϵ_{Nd} versus $^{87}\text{Sr}/^{86}\text{Sr}$ (at 619 Ma) in (A) and versus model ages from depleted mantle (DM) and chondritic uniform reservoir (CHUR) in (B). Triangles in (A) are data from Paraguçu Suite. São Pedro Caldas and Divinolândia Suites from Janasi (2002).

Hf signature in zircon

Analytical techniques are described in Appendix A and the data are available in ANEXO XI. The same samples and spots in zircon grains from SHRIMP were used to analyze Lu and Hf isotopes and, although U-Pb data were acquired by LASS-ICP-MS, the $^{206}\text{Pb}/^{238}\text{U}$ spot age obtained in SHRIMP was used as reference.

Zircon from all samples has similar $^{176}\text{Lu}/^{177}\text{Hf}$ ratios (Fig. 9B). The samples from the south body (RGM02, 04 and 53) have similar ϵ_{Hf} (between -16.1 and -12.6 – Fig. 9A). On the other hand, the sample from the north body (RGM57), presents the least negative ϵ_{Hf} (between -10.6 and -8.0 – Fig. 9A), hence the highest $^{176}\text{Hf}/^{177}\text{Hf}$ ratio values (Fig. 9B).

Fig. 9. Zircon isotopic composition. (A) ϵ_{Hf} against $^{206}\text{Pb}/^{238}\text{U}$ ages (Ma). (B) $^{176}\text{Hf}/^{177}\text{Hf}$ against $\text{Log } ^{176}\text{Lu}/^{177}\text{Hf}$.

GRANULITES FROM GUAXUPÉ DOMAIN

The granulites in Guaxupé Domain have a multiple sorts of protoliths, such as pelites and mafic rocks, although felsic igneous rocks dominate it (Motta et al., n.d.). The ultra-high temperature conditions (UHT) for the metamorphism, $T > 900^\circ\text{C}$, are widespread (Del Lama et

al., 1994; Rocha, 2016; Tedeschi, 2018). Mafic granulites have a little compositional variability (Del Lama et al., 1998), and they are made of clinopyroxene, orthopyroxene, plagioclase, and a variable amounts of quartz, hornblende and biotite, which may be absent. Garnet is usually absent. Pelitic granulite developed a high refractory residuum of garnet, sillimanite, rutile, spinel, biotite and ilmenite, with a little amount of felsic phases, and leucosome of granitic composition is widespread.

The felsic granulites are banded and composed by a monzogranitic leucosome and a residuum of clinopyroxene, orthopyroxene, plagioclase, K-feldspar, quartz, ilmenite and magnetite with variable amount of garnet, hornblende and biotite, which may be absent. This is the dominant rock type among the basal portion of this domain.

Melt composition from partial melting of the garnet felsic granulite

Previous modelled pseudosections are used to constrain the melt composition generated from partial melting of the garnet felsic granulite (Motta et al., n.d.). Calculations were conducted with THERMOCALC, in a ten component chemical system $\text{Na}_2\text{O}-\text{CaO}-\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{O}_2$ (NCKFMASHTO), and for further discussions regarding the pseudosections modelling, check Motta et al. (n.d.). Accessory mineral behaviour is calculated following the methodology of Kelsey et al. (2008). Melt compositions extract from THERMOCALC are combined with the solubility expressions for zircon (Watson and Harrison, 1983) and apatite (Harrison and Watson, 1984) to determine the saturation concentrations of Zr and P_2O_5 . The data obtained were gridded via interpolation and superimposed onto the calculated phase diagram. Stoichiometric values of Zr in zircon were 497,644 ppm (Kelsey et al., 2008) and P_2O_5 in apatite were 41 wt. % (Webster and Piccoli, 2015). It was assumed that all Zr and P_2O_5 are restricted to zircon and apatite, respectively.

The garnet felsic granulite modelled by Motta et al. (n.d.) shows that the metamorphic peak paragenesis, composed of clino- and orthopyroxene, garnet, K-feldspar, plagioclase, quartz, ilmenite and leucosome was stable at temperatures between 880 and 1030 °C at pressures between 9 and 12 kbar (Fig. 10A). The melt proportion, in these *P-T* conditions, is between 5 and 20 % (Fig. 10A). In the metamorphic peak conditions, around 1000 °C, at fixed pressure of 10 kbar, K-feldspar and garnet were not been totally molten, but were decreasing their modal proportions (Fig. 10B) and, at the same time, the residuum was increasing in plagioclase and orthopyroxene proportions.

Fig. 10. Phase equilibria modelling of a garnet felsic granulite from Socorro-Guaxupé Nappe (after Motta et al., n.d.). (A) Calculated P-T pseudosection for the composition shown at the right-top of the figure (values in mol.%), in which the dashed lines represent the modal proportion of melt generated. (B) Changes in the modal proportion of phases at fixed pressure of 10 kbar. (C) Percentage of zircon dissolution relative to the amount at the solidus. (D) Percentage of apatite dissolution relative to the amount at the solidus.

A pressure of 10 kbar was set to constrain the composition of melt generated in a window between 900 and 1050 °C. The melt generated by the model has composition in the threshold between intermediate and acid range, as the content of SiO₂ varies from 67.7 to 66.1% (Fig. 3), being the more acid compositions related to lower temperature (900 °C). The behaviour of CaO, FeO_T, MgO predicted in the model show the same trends and contents than the rocks from Paraguaçu Suite, with a decreasing to more fractionated compositions than observed in the natural rocks (Fig. 3). Similar behaviour and content are observed in K₂O, both show a positive correlation with SiO₂. The major difference between the modelled melt and the observed in

Paraguaçu Suite is the contents of Al_2O_3 and Na_2O , in which both are depleted in the predicted model, although their trends have the same positive slope against SiO_2 (Fig. 3).

The curves of zircon dissolution are straight and strongly dependent on the temperature (Fig. 10C). The proportion of dissolved zircon in the peak paragenesis field can reach 50%. Apatite dissolution has similar behaviour, although the curves show a weakly negative slope in the fields where garnet is not predicted to be stable and sharper temperature dependence in the fields where garnet is predicted to be stable (Fig. 10D). The amount of apatite dissolved in the field of the metamorphic peak assemblage can achieve 20 %. Therefore, even underwent UHT metamorphism, the garnet felsic granulite residuum still had solid zircon and apatite and some of these solid grains could be dragged into the generated melt and worked as xenocrysts and contaminant.

DISCUSSION

Petrogenesis of Paraguaçu Suite

The thermodynamic modelling for the garnet felsic granulite shows that K-feldspar and garnet were the major phases been consumed at metamorphic peak, as zircon and apatite were the accessory phases been dissolved, while plagioclase and orthopyroxene were not. The produced melt have similar composition as the charnockites from Paraguaçu Suite, particularly in relation to FeO , MgO , CaO , and K_2O .

The study charnockites have content of $\text{Fe}_2\text{O}_3 + \text{MgO} + \text{MnO} + \text{TiO}_2$ between 6 and 10%, higher than those observed in experimental studies for typical anatectic granites (Patiño Douce and Beard, 1995). Assuming a crustal source, the breakdown of biotite, a mineral rich in these elements, can generate around 4.5% of $\text{Fe}_2\text{O}_3 + \text{MgO} + \text{MnO} + \text{TiO}_2$. Due to the anhydrous character of the magma and the high values of $\text{Fe}_2\text{O}_3 + \text{MgO} + \text{MnO} + \text{TiO}_2$, it seems unlikely that the charnockitic magma was generated by biotite dehydration reactions alone. Therefore, the high values for the mentioned elements require an enriched source, in which, due to the high temperature of formation, biotite were firstly consumed at lower temperatures ($< 900^\circ\text{C}$), leaving behind a garnet-rich residuum, enriched in Fe/Mg ratios, and then the dissolution of garnet, at higher temperatures was responsible for melt enrichment in these elements. The high K/Rb and the remarkable depletion in Cs, Th and U is typically attributed to residual sources (Fig. 11), depleted in Rb by previous events of magma extraction, highlighting the residual character of the source.

Fig. 11. Continental crust normalized (Wedepohl, 1995) incompatible element patterns.

There is a notable enrichment in Ba in the rocks of the Paraguaçu Suite (Fig. 11). Biotite and K-feldspar are the mainly common Ba-bearing phases and, as previously mentioned, biotite is not a good candidate to be present in the magma source, thus, the contribution of K-feldspar must be important during UHT conditions. The composition of Paraguaçu Suite and the garnet felsic granulite shows complementary content of Sr, element mainly held in calcic plagioclase, with positive anomalies in the garnet felsic granulite and negative in the charnockite, except for the sample located in the north body of the suite. Thus, it is likely that plagioclase was not consumed in the source, which is also observed in the partial melt modeling of the granulite, where plagioclase has been increase in modal proportion in the residuum at the metamorphic peak conditions (Fig. 11), and supported by the low content of Ca in the charnockites. Thereby, the enrichment in LREE and Eu in the anhydrous magma is related to the participation of K-feldspar as a reactant in the melt reactions, a phase in granulite facies enriched patterns of LREE and strongly positives Eu anomalies (Bea, 1996), as it is possible to observe by the decrease of the modal proportion of the mineral in the modelling (Fig. 11). The north body of the Paraguaçu Suite may have been enriched in Sr and Ca by contamination with other sources, probably plagioclase-rich or represent a batch of melt enriched in plagioclase component, facilitated by a more fertile portion of the source.

The behaviour of Zr and Hf between charnockites and granulites shows an inverse relationship (Fig. 11), suggesting that zircon was dissolved in the source and concentrated in the melt. Furthermore, the rocks from Paraguaçu Suite present high contents of HREE and Y and moderated La/Yb ratios. These observations strengthen that zircon was dissolved in the source, which is possible to observe in the modelling of the granulite partial melting, where up to 50% of zircon has been dissolved at metamorphic peak conditions (Fig. 10C). The breakdown of garnet may also increase the content of HREE, Y and the La/Yb ratio in the melt, which occurred during the metamorphic peak in the granulites and their partial melting (Fig. 10B).

Ti-bearing phase, mainly ilmenite, is refractory, hence it tolerates or grows at high temperatures. Positive anomalies in the granulites and negative in the charnockites (Fig. 11) suggests that Ti-phases were not solubilized during the partial melting. The main Nb-bearing phases are zircon and hornblende, apart from Ti-phases. The inverse behaviour of Nb between charnockites and granulites is unlikely to be led by Ti-phases, therefore it may be related to breakdown of hornblende, in an initial stage, and particularly to zircon solubilization in the source. One problem to analyze the Ti-bearing phase behaviour during partial melting with thermodynamic modelling is the melt models do not accept Ti dissolution and so, even if a portion of ilmenite could be melted and dissolved in the melt in nature, it does stay solid in the modelling. Thus, although the geochemical behaviour suggests that Ti-phases were not reactants in the source, some of the conclusion of elements carried must be taken with caution.

The geochronology data obtained in zircon from Paraguaçu Suite shows a maximum crystallization age from *c.* 630 to 619 Ma for the southern body and *c.* 611 Ma for the northern body, with a spread toward younger ages until *c.* 600 Ma, in both bodies (Fig. 6). These dates are coeval to the UHT metamorphism of the Granulitic Unit in the Socorro-Guaxupé Nappe, where the metamorphic peak to the final crystallization of the partial melting happened between 630 and 595 Ma (Rocha et al., 2018, 2017; Tedeschi, 2018). This time span of *c.* 35 Myr corresponds to the magmatic residence time, which the oldest ages must be related to xenocrystals from the source and/or autocrystals and the youngest ages may be related to a continuous input of melt coming from the source. The age of *c.* 620 Ma is interpreted as the peak of the magmatism because the majority of data are concentrated in this date. Therefore, the magmatic age of Paraguaçu Suite fits the time when granulite metamorphism and partial melting were been happening in the Granulitic Unit.

The decreasing of Th/U ratios in zircon toward younger ages (Fig. 6) and morphology of the crystals (Fig. 5), in Paraguaçu Suite, features used to infer a slow crystallization of magma with prolonged residence time in the lower crust (Corfu et al., 2003), as observed by the scatter of dates, particularly toward the youngest ages. A large spread in LREE in zircon grains also suggests a complex crystallization history, and the competition for these elements with feldspars.

Thus, the behaviour of major and incompatible elements follows complementary trends between charnockites and granulites. Their ages are correspondent, including a prolonged record of crystallization. It is unlikely that hydrated phases, particularly biotite, were present in the sources of the anhydrous magma, particularly at UHT conditions. K-feldspar was partially extracted from the source, as well as zircon, apatite and garnet, however, Ti-phases, orthopyroxene and plagioclase were retained in the residuum. The geochemical data are compatible with the calculated metamorphic modelling for the garnet felsic granulite. Lower

contents of SiO₂ and Na₂O in the charnockites than the melt generated from the garnet felsic granulite modelling may indicate a mixture of other sources, probably with more mafic affinity. In the field, it is common to observe lenses of mafic granulite, with or without garnet, within the garnet felsic granulite, some of them bearing microscopic to centimetric lenses of leucosome, which present a tonalite composition and would work as a source to deviate the modelled melt composition into what is observed in nature. Another possibility is the mixture with basic magma intruded into the column of granulites with source in the mantle, which is common and observed in several outcrops in the region and coeval or a little younger than granulite facies metamorphism (Rocha et al., 2018, 2017; Tedeschi, 2018).

Comparison between the charnockite suites

The three charnockite suites in the Guaxupé Domain, Divinolândia, São Pedro de Caldas e Paraguaçu, show similar trends of distribution of major elements, being Divinolândia Suite the less fractionated and São Pedro de Caldas the most, with Paraguaçu presenting intermediate composition between them (Fig. 3). Divinolândia Suite has SiO₂ content between 52.8 and 67.3 wt.%, while São Pedro de Caldas presents 61.5 and 78.1 wt.%. All suites are ferroan with shoshonitic affinity (Fig. 12). The highest K/Rb ratio is observed in Paraguaçu Suite, whereas the lowest is observed in São Pedro de Caldas Suite (Fig. 12).

Fig. 12. (A) #Fe (mol. $\text{FeO}_T/(\text{FeO}_T+\text{MgO})$) diagram showing the boundary between ferroan and magnesian plutons. (B) Potassium content based binary diagram showing magmatic series. (C) Ba/Sr ratio against K/Rb. (*) data from Janasi (2002)

The three charnockite suites in the Guaxupé Domain, Divinolândia, São Pedro de Caldas e Paraguaçu, have the same fractionated pattern of REE, enriched in LREE (Fig. 4). Compared to the others suites, Paraguaçu Suite shows depletion in REE, particularly compared to São Pedro de Caldas Suite (Fig. 4).

Although there are no in situ geochronologic data for Divinolândia and São Pedro de Caldas Suite, the ages of crystallization are also similar among them. U-Pb zircon data for Divinolândia Suite indicated crystallization age of 625 ± 7 Ma, by conventional TIMS (Basei et al., 1995). Also by conventional TIMS, Janasi (2002) dated zircon fractions from São Pedro de Caldas Suite and obtained age of 623.3 ± 3.1 Ma, although one fraction of crystals yielded younger Pb-Pb

date of 607 ± 8 Ma. All this dates are similar to obtained in this work for Paraguaçu Suite, which means that the charnockite magmatism in Guaxupé Domain was coeval.

Although there are many similarities among the charnockite suites, it is unlikely that they had the same source. Janasi (2002) suggests that São Pedro de Caldas Suite was formed by melting of a fertile granulite crust, which generated large volumes of magma, whereas Divinolândia Suite may derivate from a more refractory granulite with incorporation of basic magma with calc-alkaline affinities. Paraguaçu Suite is extremely depleted in Rb related to the other suites, and as previously discussed, may originated by a single source, the partial melting of the garnet felsic granulite, with a minor contribution of mafic components.

The origin of the Socorro-Guaxupé Nappe

The whole rock isotopic data point out to a long crustal residence for the three suites (Fig. 8). São Pedro de Caldas has the least negative ϵNd_t values (-3 to -6) and Divinolândia Suite has the most scattered values (-9 to -14). This implies to younger T_{DM} ages in São Pedro de Caldas (1.2 to 1.6 Ga) than Divinolândia Suite (1.8 to 2.3 Ga) (Janasi, 2002). Paraguaçu Suite has the most homogeneous data with ϵNd_t between -11,8 and -12,4 and T_{DM} between 2.2 and 2.3 Ga, representing the oldest model ages.

The isotopic Hf data in zircon crystals from Paraguaçu Suite leads to the same interpretation of long crustal residence. The ϵHf_t is always negative and homogeneous, between -8 and -10.6 in the north body and -12.6 and -16 in the south body. An event signature diagram (Fig. 13) is a useful tool that shows the input of juvenile material relative to crustal reworking (Belousova et al., 2009). The x-axis is the crystallization age (Ma) obtained by U-Pb in SHRIMP and the y-axis represents the crustal residence of the material (Ma), obtained by the crystallization age minus the Lu-Hf model age. The Paraguaçu Suite shows a near horizontal trend with a weak diagonally downward trending, which represents a mixing between reworked crustal segments and juvenile additions (Fig. 13).

Fig. 13. Evolution signature of the charnockites from Paraguaçu Suite, typical from a collision, in which is influenced by a mixed juvenile and reworked sources as oceanic tract closes with volcanic arcs and terranes being accreted.

Although the Socorro-Guaxupé Nappe is considered to be formed in a magmatic arc setting (Campos Neto and Caby, 1999; Mora et al., 2014; Rocha et al., 2018), juvenile rocks have not been recognized here or in other portions of the nappe (Campos Neto et al., 2011; Janasi, 2002; Mora et al., 2014; Rocha et al., 2018; Tedeschi, 2018). In general, two factors can be responsible for this isotopic signature: contamination and assimilation of crustal basement (e.g. Schoene et al., 2009) and/or a mixing between enriched lithospheric mantle and crustal components (e.g. Hagen-Peter et al., 2015).

There are only a few data showing inheritance in zircon cores with ages to the probable basement of the Socorro-Guaxupé Nappe, concentrated in 2.5 Ga and possibly related to rocks of the Paranapanema Craton (Tedeschi et al., 2018), as the ages related to São Francisco Craton are scattered from the Paleoproterozoic, c. 2.08 Ga, (e.g. Cioffi et al. 2016b) to the Mesoarchean, c. 3.0 Ga (e.g. Cioffi et al. 2016a). In addition, the model ages, either Nd or Hf, aim to dates older than 1.2 Ga, in which many of them are older than 2.0 Ga. Thus, if this process of assimilation happened, a great amount of basement would have participated and it would be expected a higher volume of data of inheritance from the basement. Therefore, the contamination/assimilation of basement might have played a minor role in the formation of the Socorro-Guaxupé Nappe.

The absence of juvenile rocks has been attributed to a mixing with enriched mantle (Carvalho et al., 2014; Janasi, 2002; Mora et al., 2014; Rocha et al., 2018). This interpretation is well supported by the shoshonitic affinity of the rocks in the segment and the presence of high-pressure metasedimentary rocks in the Andrelândia Nappes System (Campos Neto et al., 2010; Campos Neto and Caby, 2000; Frugis et al., 2018; Motta and Moraes, 2017; Tedeschi et al., 2017) that reached mantle depths and may have contributed to the metassomatism.

Nevertheless, the long crustal histories printed in the Socorro-Guaxupé Nappe rocks raise doubts about its origin. In the typical region of the southern Brasília Orogen, the magmatic arc domain, represented by Arenópolis Arc (Fig. 14), is composed by calc-alkaline plutonic rocks, with

variable degrees of deformation and metamorphism, and supracrustal sequences, comprising metavolcanic and immature metasedimentary rocks (Pimentel, 2016; Pimentel et al., 2011). The magmatism event started in *c.* 900 Ma lasting until 640 Ma (Laux et al., 2005), with juvenile signatures, marked by positive to weakly negative ϵNd values (Laux et al., 2005; Pimentel et al., 1996; Pimentel and Fuck, 1992). These features agree with data from Mara Rosa Arc (Pimentel, 2016), the magmatic arc domain in the northern Brasília Orogen (Fig. 14), but do not match with those found in Socorro-Guaxupé Nappe.

Other than that, in the typical region of the southern Brasília Orogen, the Anápolis-Itaçu Complex (Fig. 14) presents similar rock association as Socorro-Guaxupé Nappe, with granulites derived from igneous protoliths, such as tonalite and granodiorite, and sedimentary rocks, with granite and layered gabbro intrusions (Moraes et al., 2004). U–Pb zircon ages are spread between 760 and 650 Ma and interpreted as magmatic activity (Piuzana et al., 2003a), while metamorphism is recorded between 650 and 630 Ma (Baldwin and Brown, 2008; Della Giustina et al., 2011; Piuzana et al., 2003b) and reached UHT conditions (Baldwin et al., 2005; Moraes et al., 2002). Isotopic signatures show negative ϵNd values, between -9.3 and -1.4, and Nd model ages fall into two intervals, between 2.3–1.9 Ga and 1.7–1.4 Ga (Piuzana et al., 2003a). Anápolis-Itaçu Complex is interpreted as the metamorphic core of Brasília Orogen (Della Giustina et al., 2009; Fuck et al., 1994; Pimentel, 2016). Geographically, Arenópolis Arc is exposed to the west of Anápolis-Itaçu Complex and eastward Araxá Group, a metasedimentary sequence, outcrops (Fig. 14).

Rock associations, magmatic ages, metamorphic ages and conditions and isotopic signatures of Anápolis-Itaçu Complex are very similar to the Socorro-Guaxupé Nappe. The western region of the Socorro-Guaxupé Nappe is covered by a thick (dozens of kilometer) Paleozoic to Mesozoic sedimentary basin (Campos Neto and Caby, 2000), the Paraná Basin, and in the northeastern-eastern it is limited by a metasedimentary domain, the Andrelândia Nappe System (Fig. 1 and Fig. 14). Thus, instead of Socorro-Guaxupé Nappe being the magmatic arc it seems that it is related to the metamorphic core of the southernmost Brasília Orogen. Moreover, in the evolution models of the orogeny, São Francisco Plate have subducted Paranapanema Plate (Frugis et al., 2018; Motta and Moraes, 2017), thus the magmatic arc must be established in the latter. Cioffi et al. (2016a, 2016b) linked a large portion of basement, which splits the Socorro-Guaxupé Nappe into Guaxupé and Socorro Domain, to the São Francisco Craton, therefore the Socorro-Guaxupé Nappe may be laid down above São Francisco Craton. Of course, this discussion is not closed, and it just tries putting some lights into the issue.

Fig. 14. Overview of Brasília Orogen (Valeriano et al., 2004). External zone (1-4), in which: (1 and 2) Granite-greenstone terrains; 3 – Paleoproterozoic rift (Araí Group); (4) Ilícinia-Piumhi thrust system. Passive margin (5-8), in which: (5) Paranoá Group; (6) Quartzite-phyllite units (Canastra and Andrelândia groups); (7) Vazante Group; (8) Ibiá Group. Internal zone (9-14), in which: (9a and b) Araxá and Andralândia Groups; (10) Granulite nappes; (11 and 12) Archean/Paleoproterozoic gneiss and greenstone complex; (13) Mesoproterozoic volcano-sedimentary sequences (Juscelândia, Indaialópolis, Palmeirópolis); (14) Mafic-ultramafic complexes (Barro Alto, Niquelândia and Cana Brava). Goiás Magmatic Arc (15-16), in which: (15) Supracrustal sequences; (16) Orthogneisses and granitoids. MR – Mara Rosa Arc. AA – Arenópolis Arc. AI – Anápolis—Itaçu Complex. GD – Guaxupé Domain (Socorro-Guaxupé Nappe). SD – Socorro Domain (Socorro-Guaxupé Nappe).

The heat source to generated (ultra)-high temperatures

There is an agreement that UHT conditions occur in many metamorphic orogens, although the heat source of such extreme temperatures ($> 900\text{ }^{\circ}\text{C}$) is not well understood (Brown, 2007;

Clark et al., 2011; Harley, 1998; Kelsey and Hand, 2015). This because the lower continental crust has low thermal diffusivity (Whittington et al., 2009), thus large amounts of energy are required to elevate the temperature, whereas, once they have reached high temperatures, they tend to be heated for a prolonged period (Kelsey and Hand, 2015). Furthermore, the generation of anatectic melt hinder the temperature elevation, because partial melting reactions are endothermic (Lyubetskaya and Ague, 2009), thus buffering the temperature (Stüwe, 1995). A great amount of melt must be generated and drained from its source, thus the temperature can rise, and therefore the rock is “pre-conditioned” to achieve such temperatures (Clark et al., 2011; Kelsey and Hand, 2015; Vielzeuf et al., 1990).

As previously discussed, it is unlikely biotite had a significant contribution to the generation of the charnockites. Therefore, the partial melt generated by hydrate-phases breakdown was formerly extracted. In addition, the partial melting in the Granulitic Unit of the Socorro-Guaxupé Nappe was controlled by the breakdown of K-feldspar and garnet. Thus, it is likely that rocks from the Granulitic Unit had reduced fertility (*i.e.* pre-conditioned), hence more prone to heat at UHT conditions.

Three main factors are responsible to increase the temperature of the continental crust: (1) radioactive heat production by decay of Th, U and, mainly, K (Alessio et al., 2018; England and Thompson, 1984; Jamieson et al., 1998; Vilà et al., 2010); (2) mechanical heating in shear zones (Nabelek et al., 2010); (3) advective mantle heat (Brown, 1993; Ellis, 1987; Vielzeuf et al., 1990).

The Diatexite Unit lies down the Granulite Unit in the Socorro-Guaxupé Nappe, in which significant amount of silicate melt were generated during the orogenesis, with an extension of hundreds of km² and ~ 9 km thick (Campos Neto and Caby, 1999). The rocks from both Diatexite and Granulite Units present shoshonitic compositions (Janasi, 2002, 1999; Mora et al., 2014; Rocha et al., 2018), hence enriched in potassium. Such expressive volume of radiogenic material in the mid crust can elevate considerably the temperature in the lower crust (Gerya et al., 2008; Jamieson et al., 2004). Accordingly, the very generation of partial melt and the subsequent migration to higher levels in the crust played a major role to the heating of the orogen.

Compare to the south body, the north body of Paraguaçu Suite presents enrichment in Sr, and higher values of ϵ_{Hf} , up to + 4 ϵ units greater. These enrichments suggest a mix with other source(s), more radiogenic (*i.e.* richer in ¹⁷⁶Hf), in which plagioclase and Ti-phases were being molten in the source, which is compatible with mafic sources. Moreover, compare to the modelled melt composition, the Paraguaçu Suite is depleted in SiO₂ and enriched in Na₂O, another suggestion that depleted sources have contributed in the generation of these magmas, and hence to the heating in the Socorro-Guaxupé Nappe, as suggested before due to interactions with mantle derived material (Janasi, 2002; Rocha et al., 2018). Thus, small bodies of mafic rocks, now

recognized as mafic granulites, some preserving intrusive contacts, indicates a mantle contribution to the base of this crust, and this may implies decompression of asthenospheric mantle, which might worked as an additional heat source for metamorphism and some of these basic bodies, as suggested in some models (Brown, 1993; Ellis, 1987).

CONCLUSIONS

The combination of data and methods used in this work have allowed concluding that:

- Paraguaçu Suite had a source in which K-feldspar, garnet, apatite and zircon were partially solubilized;
- The garnet felsic granulite modelling shows K-feldspar, garnet and zircon were been broken during the UHT stage;
- The composition of the modelled partial melt of a garnet felsic granulite show complementarities with the composition of charnockites in Paraguaçu Suite;
- The ages of the charnockitic magmatism of Paraguaçu Suite and the UHT conditions in the Granulitic Unit of Guaxupé Domain were the same, from *c.* 630 Ma scattered until *c.* 600 Ma, with concentration in *c.* 620 Ma.

Thus, the partial melting of the garnet felsic granulite may have migrated, collected, concentrated and then generated the Paraguaçu Suite.

Futhermore, the charnockite magmatism in Guaxupé Domain are coeval, although the three identified suites, Divinolândia, São Pedro de Caldas and Paraguaçu Suites, have different sources or degrees of mixture/fractionation.

The heat source to generate UHT conditions in Guaxupé Domain is related to a radioactive decay of a thick layer of molten rock that occurs above the Granulitic Unit add up with an input of heat from the asthenospheric mantle.

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APPENDIX A. Analytical methods

Elemental geochemistry

For whole-rock geochemical analyses, unweathered samples were crushed in a steel jaw-crusher and subsequently in an agate mill. Major elements were obtained by X-ray fluorescence (XRF) spectrometry after lithium metaborate/tetraborate fusion, carried out at the Núcleo de Apoio à Pesquisa Geoanalítica (NAP Geoanalítica), from the Institute of Geosciences of the University of São Paulo (IGc-USP), Brazil, following the methods described by Mori et al. (1999). Trace elements concentrations, including REE, were also obtained at NAP Geoanalítica, by inductively coupled mass spectrometry (ICP-MS) using a Perkin Elmer Plasma Quadrupole MS Elan 6100DRC (further details in Navarro et al., 2008).

Sm-Nd and Rb-Sr isotopes

Sr and Nd isotopic compositions were determined using a Neptune multi-collector ICP-MS at Centro de Pesquisas Geocronológicas (CPGeo) from IGc-USP, Brazil. The same powders used for whole-rock elemental analyses were dissolved in acid, and the elements of interest were separated in ion-exchange columns following the procedures described by Sato et al., (1995). The Sr and Nd isotopic ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ isotopic ratios for mass fractionation correction. The NBS987 standard yielded a long-term average $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.710239 ± 0.0000017 (2σ error). The JNdi standard (Tanaka et al., 2000) yielded a mean $^{143}\text{Nd}/^{144}\text{Nd}$ value of 0.512102 ± 0.000002 (2σ error).

U-Pb, REE and Lu-Hf isotopes in zircon

Grains of zircon of Paraguaçu Suite were prepared via traditional methods of disaggregation, magnetic and heavy liquid separation. These grains were mounted in 25 mm diameter epoxy discs, polished and coated with 40 μm of gold. The internal texture of the grains was characterized using CL imaging with a Tescan Mira3 FESEM from Microscopy and Microanalysis Facility, at John de Laeter Centre, Curtin University, Australia.

U-Pb isotopic data were collected using the Sensitive High Resolution Ion Microprobe (SHRIMP II) also based at John de Laeter Centre. The mass filtered O_2 primary beam strength was among 2 and 3 nA with a 20 μm spot size. A 6-scan cycle and a mass resolution of $\sim 5,000$ were used during analysis (de Laeter and Kennedy, 1998; Kennedy and de Laeter, 1994). Data were processed using SQUID II and Isoplot 3.75 (Ludwig, 2009, 2003) with the correction of measured isotopic ratios for common Pb based on Stacey and Kramers (1975). Standards were located in a separate mount for the analysis and this mount was Au-coated at the same time as the unknowns mount. NBS610 glass was used to calibrate the position of the ^{204}Pb peak. As a primary standard CUYZ (Curtin internal laboratory standard, 568.55 Ma) was used, and Pleišovice (Sláma et al.,

2008) was used as a secondary standard. The secondary standard was within error of reported age (337.13 ± 0.37 Ma), yielding 332.3 ± 4.5 Ma during the session. A minimum error of 1.5% was assigned to the external spot-to-spot error to reflect the long-term performance of the SHRIMP.

Trace elements and Lu-Hf isotopes and complementary U-Pb data of zircon were collected from the same samples analyzed in the SHRIMP, using Split-Stream Laser Ablation with Inductively Coupled Plasma Mass Spectrometry (LASS-ICP-MS) on an ASI Resolution M-50 with a 193-nm wavelength Ar-F excimer laser with an Agilent 7700 mass spectrometer at the GeoHistory Facility, John de Laeter Centre, Curtin University. Laser fluency was $2\text{J}/\text{cm}^2$, at a repetition rate of 10 Hz and spot size of 50 μm .

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**ANEXO III – RELATIONSHIP BETWEEN METAMORPHIC PEAK AND
DEFORMATION TEMPERATURE IN MIGMATITES AND GRANULITES
AND THEIR CONSTRAIN ON EXHUMATION IN THE SOCORRO-GUAXUPÉ
NAPPE, SE BRAZIL**

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Relationship between metamorphic peak and deformation temperature in migmatites and granulites and their constrain on exhumation in the Socorro-Guaxupé Nappe, SE Brazil

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ABSTRACT

In the southeastern of Brazil, a thick section of a lower continental crust outcrops in the Socorro-Guaxupé Nappe, part of a neoproterozoic orogenic collage in the southernmost part of the Brasília Orogen. Migmatites, mostly diatexites, and granulites reached ultra-high temperature conditions during their metamorphic peak, around 1000 °C. However, both rocks present different structural patterns. Foliation in the migmatites is more scattered and present NW-SE direction, while the granulites is more concentrated with WNW-ESE direction. Lineation in migmatites is defined by feldspars and hydrate minerals, such as biotite and hornblende with WNW-ESE direction, whereas in granulites the lineation is marked by anhydrous minerals, such as pyroxenes and feldspars, and present WSW-ENE direction. The quartz c-axis fabric opening-angle for the residuum of granulite shows the highest values (135-154°), followed by the leucosome of granulites (113-130°) and then leucosome of migmatites (86-97°). These implies that quartz c-axis fabric in granulites records two temperatures: the residuum is close to the metamorphic peak, between 869 and 923 °C, and another close to the temperature crystallization of the leucosome, between 796 and 853 °C. The migmatites record quartz c-axis fabric temperature between 633 and 685 °C. Therefore, the solid-state deformation in the granulites happened at higher temperature than the migmatites, which were in suprasolidus conditions during a larger temperature range (> 300 °C). Thus, when the granulites solidified, the migmatites still were molten, printing different deformation geometries.

Keywords: Quartz c-axis fabric, deformation temperature, Socorro-Guaxupé Nappe

INTRODUCTION

Partial melting modifies rock rheology and density, leading to a generation of and segregation of two phases, a low density melt and a high density solid residue (Kohlstedt, 2002;

Sawyer, 1994; Vanderhaeghe, 2009). During active deformation, the melt moves by pervasive flow parallel to the principal finite elongation into foliation-parallel sheets, fold hinges and dilation sites, such as inter-boudins (Brown and Solar, 1998; Hasalová et al., 2008; Sawyer, 2001; Weinberg, 1999). At same time, granulite terrains develops characteristic horizontal structures during the metamorphic climax (Sandiford, 1989) in which melt will be allocated.

The distinction among how and when these paralleled horizontal structures were generated linked to temperature-pressure paths may help to understand the structural-tectonic evolution of the middle and lower continental crust. With this purpose, quantitative structural analysis was used to constrain the relationship between metamorphism and deformation (e.g. Díaz Azpiroz and Fernández, 2003; Faleiros et al., 2016a; Fazio et al., 2016; Hasalová et al., 2008; Lister and Dornsiepen, 1982). However, when partial melting is involved, after its segregation from the site that it was generated, the melt moves to dilation sites and is adjusted to any space or structure imposed by the solid residue layers and only after temperature drops, its crystallization starts and a network of solid-touching crystals are developed that solid-state deformation might act over the crystallized leucosome. In regions where granulites and migmatites occur either as interlayered or pillared rocks, it brings the possibility to compare deformation and its kinematics along prograde and retrograde paths, investigating residues and respective leucosomes. Within this context, the progress made in recent years with quartz c-axis thermometer brings another interesting tool (Faleiros et al., 2016b; Law, 2014), which allows to not only compare deformation texture patterns in quartz, but also temperature along prograde and retrograde paths and with this, evaluate how deformation kinetics change along heating and cooling stages at the lower continental crust.

In southeastern of Brazil, a medium- to high-metamorphic stack of nappes occur in the southernmost portion of the neoproterozoic Brasília Orogen (Campos Neto et al., 2007), exhibiting a diversity in migmatites, granulites and igneous bodies. Several works have constrained timescales, exhumation and causes of high-temperature metamorphism in the Brasília Southern Orogen (e.g. Campos Neto et al., 2010; Campos Neto and Caby, 2000; Coelho et al., 2017; Janasi, 2002; Motta and Moraes, 2017; Rocha et al., 2018, 2017; Tedeschi et al., 2017; Trouw et al., 2013) and, although they present robust data, none investigate if deformation histories of diatexites and granulites, most voluminous rocks at the base of the stack, are compatible. This paper combines structural and metamorphic data in order to enhance the understanding of deformation history of rocks formed at the lower continental crust.

GEOLOGICAL SETTING

During the Neoproterozoic, Sanfranciscan and Paranapanema plates collided and as a result, part of Gondwana continent was formed (Brito Neves et al., 2014, 1999). The collision

generated a pile of sub-horizontal nappes with a southwest-dipping tectonic wedge, and transport towards the east-northeast. Rocks from three different tectonic settings are recognized within these nappes, from WSW to ENE and bottom to top: (1) the Socorro-Guaxupé Nappe is the magmatic arc domain developed at the active margin of Paranapanema Plate, and includes granulites, assorted migmatites and intrusive bodies (2) the Andrelândia Nappe System is a subducted metasedimentary sequence and (3) the Carrancas and Lima Duarte Nappes represent a passive continental margin, related to the São Francisco continent (Campos Neto et al., 2011, 2010; Campos Neto and Caby, 2000; Trouw et al., 2013). These nappes record the migration of the orogenic front towards the cratonic area, with older ages recorded by the upper nappes, in the west, and younger ages recorded by the lower nappes, in the east (Campos Neto et al., 2011, 2010, 2004).

The Socorro-Guaxupé Nappe is a thick section of the lower continental crust (~10 km) and is divided in three distinct units (Figure - 1): (1) a basal Granulite Unit, with mafic and felsic interlayered garnet granulites with hornblende, orthopyroxene and garnet-bearing leucosome, (2) intermediate Diatexite Unit, dominated by granite to granodiorite leucosome-bearing garnet and hornblende, which may present pyroxenes core, biotite schlieren and schollen of biotite or hornblende stromatic residue, and (3) a Metatexite Unit derived from metasedimentary rocks (Campos Neto and Caby, 2000, 1999). An window where Archean-Paleoproterozoic orthogneisses crops out and divides the Socorro-Guaxupé Nappe in two main domains: Socorro Domain, in the southeastern and Guaxupé Domain, in the northwestern (Campos Neto and Caby, 2000; Janasi, 2002). This work is focused in eastern portion of Guaxupé Domain (Figure - 1).

The high temperature metamorphic conditions lasted for 30-50 Ma, making it a long-lived event (Motta et al., n.d.; Rocha et al., 2018, 2017; Tedeschi et al., 2018). EPMA dating of monazite from diatexite unit indicates metamorphic peak was reached at 631 ± 4 Ma (Rocha et al., 2017). Similar ages (*ca.* 630-620 Ma) are recorded in charnockite and high-K calc-alkaline suites intrusive suites into the Socorro-Guaxupé Nappe (Basei et al., 1995; Janasi, 2002, 1999; Martins et al., 2009; Mora et al., 2014; Reno et al., 2009). While the Granulite Unit records initial stages of retrograde cooling at 622 Ma (Reno et al., 2009), the migmatites units record ages *ca.* 600 Ma for the final stages of melt crystallization (Rocha et al., 2017).

Figure - 1 Geological framework of the southmost part of Brasília Orogen. The yellow polygon indicates the Guaxupé Domain, while the red polygon indicates the study area and the location of Figure - 2.

METHODS

Structural measures were performed using standard procedures involving identification and description of geometric and kinematic relationships among tectonic structures. Structural data was collected with compass and stereoplots were elaborated using the software Stereo32 version 1.03 (Röller and Trepmann, 2011), projected into lower hemisphere Schmidt-Lambert equal area nets.

Quartz c-axis orientations were measured using a universal stage in thin sections cut perpendicular to the foliation and parallel to the mineral lineation. Data were plotted into lower hemisphere Schmidt-Lambert equal area nets, also using the software Stereo32 version 1.03 (Röller and Trepmann, 2011).

Pressure calibration of the Diatexite Unit was made to constraint the pressure achieved in the metamorphic peak using the barometer GASP (Holdaway, 2001) with minerals from a residual sample that bears garnet, sillimanite, quartz and plagioclase apart from K-feldspar and rutile. Mineral compositions were obtained by profiles across garnet and plagioclase with a wavelength-dispersive spectrometry (WDS) in a JEOL JXA-8230 Hyperprobe at the Electron Microprobe Facility at Universidade Estadual Paulista (UNESP), Brazil, using a focused beam at 15 kV and 20 nA with 10 μm spot size. Natural minerals were used as standards.

STRUCTURAL GEOLOGY

Along the study area of basal portion of the Socorro-Guaxupé Nappe, the main metamorphic foliation has a low to sub-horizontal dip-angle, generated by a non-coaxial deformation developed during the displacement of the nappes. These horizontal structures are quite common in high-grade metamorphic terrains (Sandiford, 1989). In the northeastern portion, the Socorro-Guaxupé Nappe is limited by Três Pontas-Varginha Nappe, marking the boundary between the magmatic arc domain and the metasedimentary sequence (Figure - 2). Their contact is defined by a thrust fault with significant lateral component.

The Granulite Unit has a stromatic structure defined by composition and texture/grain size variations (Figure - 3A), parallel to the main foliation. Their texture is typical granoblastic, where quartz and feldspars present interlobate to polygonal grain boundary geometries and in layers with higher deformation ribbons of quartz are developed (Figure - 3B). The foliation is concentrated around the WNW-ESE direction and strike to NNE (Figure - 4), with average of N297/17°NE.

Figure - 2 Geological map of the studied area. Representative foliation (blue) and lineation (red) data are shown with their respective dip. The stars show the location of samples that were measured quartz c-axis open angle, this work in yellow and Faleiros et al. (2016b) in black.

Figure - 3 Banded felsic granulite, in the middle of the picture a boundin with residual mafic composition surrounding by leucosome. B. Granoblastic fabric with quartz ribbon in leucosome of granulite. C. Shear zone showing milonitic fabric of banded migmatite near to the contact between the units. D. Sillimanite-garnet milonite located near the contact between the Diatexite and Granulite Units.

The absence of good outcrops is a factor that makes difficult to identify the contact relations between the Diatexite and Granulite Units. However, the intrusion of a charnockite body of the Paraguaçu Suite (Barros, 2012) (Figure - 2) and some outcrops with weathered rocks suggest that the contact is made by a low angle shear zone, where the leucosome was strongly stretched and residuum underwent for boudinage process (Figure - 3C). In this zone the rocks have a milonite fabric marked by decreasing of granulometry and formation of kinematic indicators (Figure - 3D).

Figure - 4 A. Poles of foliation density contour for the Diatexite Unit. B. Poles of foliation density contour for the Granulite Unit.

The Diatexite Unit presents variable degree of partial melting, including schollen diatexite (Figure - 5A) and stromatic metatexite, with subordinate layers of mafic granulite and voluminous leucosome. With the increasing of melt fractions, the foliation elements gradually lose their preferred orientation due to flow perturbations and diatexites are then developed, marked by overall complex geometries of the field foliation (Figure - 5B) and the presence of abundant melt pockets, sometimes with boudinaged leucosome as competent layer (Figure - 5C), in which, sometimes, the textural aspect is of a plutonic rock (Figure - 5D). Nevertheless, field foliations can be tracked with overall consistency in the metatexites where the stromatic layered are parallel to the main foliation. In general, granoblastic fabric is recognized in the leucosome matrix with development of quartz ribbons (Figure - 5E). The plutonic-like pockets into diatexites also show deformational fabric with recrystallization of mineral rims and undulose extinction (Figure - 5F). This rocks exhibit more scattered and slightly different foliation direction (NW-SE) with higher strikes than the granulites, with average of N335/27°NE (Figure - 4).

Figure - 5 A. Schollen diatexite with mafic rafts. B. Schollen diatexite with random distribution of the foliation. C. Metatexite exhibiting leucosome forming pinch-and-swell structures. D. Huge pocket of leucosome with plutonic texture in a diatexite. E. Fine-coarsed granoblastic fabric with quartz ribbons defining the foliation in a leucosome of a metatexite. F. Plagioclase showing undolose extinction in a leucosome pocket with igneous texture. G. Mineral lineation marked by K-feldspar in a granulite. H. Asymmetric fold showing transport to E. I. Pegmatoid aspect of the diatexite with centimetric phenocrysts of hornblende.

Mineral lineation in the Granulite Unit is mainly given by orientation of pyroxenes and feldspars (Figure - 5G) and is concentrated in the WSW-ENE quadrants, with average of $N255/7^\circ$ (Figure - 6). In the Diatexite Unit, the mineral lineation is given by feldspars and hydrated minerals, such as hornblende and biotite. The directions are concentrated in the WNW-ESE quadrants, with average of $N284/7^\circ$ (Figure - 6). In both units, the kinematic indicators show transport to E (Figure - 5H).

Figure - 6 A. Lineation density contours for Diatexite Unit, showing direction WNW-ESE. B. Lineation density contours for Granulite Unit, showing direction WSW-ENE.

P-T CONDITIONS OF METAMORPHIC PEAK

Granulite Unit

The Granulite Unit is mainly formed by felsic granulite, with acid to intermediate compositions. In addition, there are lenses of mafic granulite interlayered. The peak mineral assemblage is composed by (garnet) – orthopyroxene – clinopyroxene – K-feldspar – plagioclase – quartz – ilmenite, in which biotite and hornblende are the phases that replaced anhydrous minerals after metamorphic peak. Rarely, there are bands of pelitic granulite, with the paragenesis composed by rutile – sillimanite – spinel – garnet – K-feldspar – plagioclase – quartz.

Conditions of metamorphic peak were first calculated among 890-900 °C at 11.5-14 kbar (Campos Neto and Caby, 2000; Del Lama et al., 1994), but higher conditions are found for felsic granulites varying from 940-1040 °C and 11.5-14.4 kbar (Del Lama et al., 2000; Rocha et al., 2018; Tedeschi, 2018). Phase equilibria model for felsic granulite showed that the melt crystallization must have occurred at temperatures no lower than 800 °C at pressure around 9 kbar (Motta et al., n.d.). Thus, the Granulite Unit achieved ultra-high temperature (UHT) conditions and the melt was crystallized at granulite facies.

Diatexite Unit

In the area of study, the Diatexite Unit is dominantly metasedimentary (Campos Neto and Caby, 2000; Rocha et al., 2017) and the degree of partial melting is variable, although the diatexite

are the dominant rock-type. The residual part of the migmatites consists of residual (garnet)-biotite stromatic scholen or garnet mafic granulite. The leucosome is coarse-grained, sometimes pegmatite-like (Figure - 5I), and may contain euhedral peritectic garnet, hornblende, clinopyroxene and orthopyroxene, although the latter is scarce. There are few lenses of milonitized pelitic diatexite with rutile – sillimanite – garnet – K-feldspar – plagioclase – quartz assemblage.

Garnet and plagioclase in the pelitic migmatite (sample RGM25) show flat major element distribution (Figure - 7). The dominant end-member in garnet is almandine ($\text{Fe}^{2+}/[\text{Fe}^{2+}+\text{Mg}+\text{Ca}+\text{Mn}]$), which yielded an average of 0.59, with a discrete increase in the extreme edges of the grains where can reach 0.61. The same pattern, although mirrored, is observed in the X_{Mg} ($\text{Mg}/[\text{Fe}^{2+}+\text{Mg}]$) with average between 0.36 and 0.37, but the extreme edges may show a small decrease yielding 0.34. Grossular ($\text{Ca}/[\text{Fe}^{2+}+\text{Mg}+\text{Ca}+\text{Mn}]$) is depleted in the core (~ 0.03) and shows enrichment toward the rims (~ 0.04). Thus, the garnet cores have average composition of $\text{alm}_{0.59}\text{prp}_{0.36}\text{grs}_{0.03}\text{sps}_{0.02}$, while the garnet edges is $\text{alm}_{0.61}\text{prp}_{0.32}\text{grs}_{0.04}\text{sps}_{0.02}$.

Figure - 7 Distribution of end-members along representative grains of garnet and plagioclase from the pelitic diatexite (RGM25).

Albite ($\text{Na}/[\text{Na}+\text{Ca}+\text{K}]$) is the dominant end member in the plagioclase (Figure - 7). The average composition is An_{29} with a subtle tendency to be anortite-rich in the core, although the variation is no higher than 2.

Rocha et al. (2017) applied the compositions of coexisting garnet and orthopyroxene in leucosome of diatexite to estimate the *P-T* conditions of migmatites in the area. The peak conditions vary from 918-1139 °C and 10.3-13 kbar, and these rocks also underwent UHT metamorphic conditions. Although rutile is not present in the rock, Ti contents in zircon yielded temperatures of ca. 690-734 °C, and were interpreted as the record of melt crystallization (Rocha et al., 2017).

Pressure calculated with the geobarometer GASP (Holdaway, 2001), for a sillimanite-bearing sample (RGM25), yielded 10.6 kbar, using the cores composition of the pair garnet/plagioclase, assuming that they were in equilibrium during metamorphic peak, around 1,000 °C. However, using the edges composition of the pair garnet/plagioclase, the geobarometer yielded lower pressures, ~ 5.5 kbar, assuming that the edges were in equilibrium during melting crystallization, around 720 °C. The temperatures of calculations were based in the results obtained by Rocha et al. (2017), both for metamorphic peak and partial melting crystallization.

TEMPERATURE OF DEFORMATION

The opening-angle of quartz c-axis fabrics (OA) is strongly temperature dependent and has proven to be a powerful deformation thermometer (Faleiros et al., 2016b; Kruhl, 1996; Law, 2014). Eight samples were measured, being four from Diatexite Unit and four from Granulite Unit. The location of the samples is shown in the map of Figure - 2. Temperatures were calculated using OA thermometer equations (1), (2) and (3) from Faleiros et al. (2016b).

$$T(^{\circ}\text{C}) = 410.44 \ln \text{OA} + 14.22 P(\text{kbar}) - 1272 \quad (1)$$

$$T(^{\circ}\text{C}) = 6.9 \text{OA} + 48 \text{ (OA} \leq 87^{\circ}\text{)} \quad (2)$$

$$T(^{\circ}\text{C}) = 4.6 \text{OA} + 258 \text{ (OA} \geq 87^{\circ}\text{)} \quad (3)$$

The Diatexite Unit set of samples was measured in thin sections composed by two hornblende-biotite diatexite, with peritectic garnet, clinopyroxene and hornblende (RGM18 and RGM23), one rutile-sillimanite-garnet diatexite (RGM25) and one hornblende-biotite metatexite with peritectic hornblende (RGM54). They show OA between 86 and 97 ° (Figure - 8) with temperatures ranging from 633 to 685 °C (Table - 1) at a fixed pressure of 5.5 kbar, estimated pressure obtained in GASP (see topic above), using the equation (1).

Figure - 8 Optically measured quartz c-axis fabrics; lower hemisphere equal area projection. Foliation oriented left-right and vertical, stretching lineation horizontal. Red arcs show the maximum opening-angles, using for discussion. A to D are samples for Diatexite Unit and E to H are samples for Granulite Unit.

The Granulite Unit set of samples is composed by two leucosome of felsic granulite showing peak mineral assemblages with two pyroxenes, two feldspars, garnet, quartz and ilmenite (RGM28 and RGM29), one residuum of a mafic granulite (RGM30) and one leucosome of a pelitic diatexite, in which the residuum shows spinel-rutile-sillimanite-garnet (RGM48). The OA vary from 113 to 135 ° (Figure - 8), yielding temperatures of 796-869 °C, using equation (1) at pressures of 9 kbar, pressure estimated for the partial melting crystallization in this unit (Motta et al., n.d.). Faleiros et al. (2016b) presented OA even higher in the same portion of the Guaxupé

Domain for residual parts of granulites (samples VAB-1, VAB-9b and VAB-19, also located into the map of Figure - 2), varying from 136 to 154 ° and temperatures from 872 to 923 °C (Table - 1).

Table - 1 Quartz c-axis fabric opening-angle and their correspondent deformation temperatures. Pressure of 5.5 kbar from the GASP results in this work; pressure of 9 kbar extracted from Motta *et al.* (n.d.); both values represent the final melt crystallization pressure.

Sample	Unit	Rock type	Analyzed domain	Pressure (kbar)	Larger OA	Temperature	
						Linear correlation	Pressure effect
RGM18	Migmatite	Opx diatexite	leucosome	5.5	89	667	648
RGM23	Migmatite	Opx diatexite	leucosome	5.5	91	678	659
RGM25	Migmatite	Sill diatexite	leucosome	5.5	97	705	685
RGM54	Migmatite	Opx diatexite	leucosome	5.5	86	639	633
RGM28	Granulite	felsic granulite	leucosome	9.0	130	855	853
RGM29	Granulite	felsic granulite	leucosome	9.0	130	855	853
RGM30	Granulite	mafic granulite	residuum	9.0	135	879	869
RGM48	Granulite	aluminous granulite	leucosome	9.0	113	777	796
VAB-1	Granulite	felsic granulite	residuum	9.0	154*	966	923
VAB-9b	Granulite	felsic granulite	residuum	9.0	136*	884	872
VAB-19	Granulite	felsic granulite	residuum	9.0	139*	897	881

* AO extracted from Faleiros *et al.* (2016b).

DISCUSSION

The geometry of the structures are straightly related to the action of strain fields (Merle, 1986; Tikoff and Fossen, 1999). Structural data obtained here shows that there are differences in the direction of foliation and lineation between Diatexite and Granulite Units.

The Diatexite Unit shows a more scattered foliation direction, mainly disperse in the NW-SE quadrant, whereas the Granulite Unit has a more concentrated dispersion of foliation in the WNW-ESE direction. The lineation in Diatexite Unit is given by hydrated minerals and feldspars, with directions concentrated in the WNW-ESE quadrants, while the Granulite Unit presents the

direction WSW-ENE made up by anhydrous minerals, mainly pyroxenes and feldspars. The offset in the penetrative structures suggests that the strain fields were different between the units.

The metamorphic peak in both Granulite and Diatexite Units were achieved at similar conditions of UHT, $\sim 1000\text{ }^{\circ}\text{C}$ and 10-12 kbar (Motta et al., n.d.; Rocha et al., 2018, 2017). However, the conditions for the crystallization of the trapped melting in each unit was different; in the Diatexite Unit the crystallization of melt is recorded around $720\text{ }^{\circ}\text{C}$ (Rocha et al., 2017) at pressures around 5.5 kbar, whereas in the Granulite Unit these conditions were $800\text{ }^{\circ}\text{C}$ at 9.0 kbar (Motta et al., n.d.).

These P - T settings are printed in the quartz fabrics. The measurements of c-axis quartz opening-angle in the residuum of rocks from Granulite Unit yielded temperatures near to the metamorphic peak, ranging from 869 to $923\text{ }^{\circ}\text{C}$, around $100\text{ }^{\circ}\text{C}$ lower than observed in the metamorphic peak ($\sim 1,000\text{ }^{\circ}\text{C}$). The Granulite Unit also bears the highest quartz OA temperatures in the leucosome, 796 to $853\text{ }^{\circ}\text{C}$, similar temperature that is observed for the crystallization of the partial melting (Motta et al., n.d.). These quartz c-axis temperatures are interpreted as deformation temperatures, which the quartz was recrystallized, thus able to absorb a solid-state deformation. Therefore, there is a hiatus of $130\text{ }^{\circ}\text{C}$ between the deformation temperature recorded in the residuum and in the leucosome of Granulite Unit, and a gap of $200\text{ }^{\circ}\text{C}$ between metamorphic peak and the melting crystallization.

The quartz fabric in the Diatexite Unit exhibits the lowest temperature, between 633 and $685\text{ }^{\circ}\text{C}$. These temperatures also represent the recrystallization of the fabric in a solid-state deformation, which means that an enough amount of leucosome had been crystallized to form a network of solid-touching crystals. Hence, the gap between metamorphic peak and deformation of a crystallized fabric is near to $350\text{ }^{\circ}\text{C}$.

Thus, the difference in the structural directions and the deformation temperatures show that the units had different deformational histories (Figure - 9). As soon as the trapped melt crystallized in the Granulite Unit, the deformation recorded a direction near to the E-W at a higher temperature and transport to ENE, around $800\text{ }^{\circ}\text{C}$. Through the high volume of melt in the Diatexite Unit, probably by the fertility of the protolith, once the metamorphic peak conditions were the same as the Granulite Unit, the crystallization of the melt happened at a much lower temperature, where the strain fields had undergone a variation producing plans with NW-SE directions and ESE lineation (Figure - 9).

Figure - 9 Structural evolution of the northeastern portion of Socorro-Guaxupé Nappe. During metamorphic peak (t_1) structures were not record in the rock due to the high content of melt. The crystallization of melt in the Granulite Unit (t_2) printed the foliation and the lineation marked by anhydrous minerals with transport to ENE. Foliation and lineation in the Diatexite Unit were recorded at lower temperature, in which the lineation is marked by hydrous minerals (t_3).

This idea is supported by the occurrence of a low angle shear zone in the contact between the units, where some charnockite bodies are emplaced. This shear zone was formed after the crystallization of the leucosome in the Granulite Unit, changing the rheology and allowing the Diatexite Unit, not yet crystallized, to move toward the craton in a channel-flow like movement.

CONCLUSIONS

Regardless of the formation and exhumation model chosen for these rocks some constrain about the Socorro-Guaxupé Nappe are important to the reconstruction of the tectonic framework of the southernmost part of Brasília:

- The Granulite and Diatexite Units show metamorphic peak at granulite facies, probably at widespread UHT conditions;
- Although subtle, the geometries of the units are different, either the foliation and lineation are displaced, which imply different strain fields;
- The contact among these units, in the studied area, is not transitional or gradational. There is a low angle shear zone that, even, charnockite body are intruded;
- The deformation temperature in the residuum of the Granulite Unit is near to its metamorphic peak;
- There is a gap, around 200 °C, between the temperature of the metamorphic peak and the temperature of deformation in the Granulite Unit and a gap of 350 °C in the Diatexite Unit.

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ANEXO IV – ANÁLISES DE QUÍMICA MINERAL POR MICROSSONDA ELETRÔNICA

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
	GrtI.1	GrtI.2	GrtI.3	GrtI.4	GrtI.5	GrtI.6	GrtI.7	GrtI.8	GrtI.9	GrtI.10	GrtI.11	GrtI.12	GrtI.13	GrtI.14	GrtI.15
SiO2	38.48	38.44	38.72	38.38	38.64	38.49	37.14	38.5	38.58	38.62	39.22	38.49	38.67	38.4	38.31
TiO2	0.06	0.08	0.07	0.09	0.06	0.04	0.04	0.04	0.06	0.06	0.06	0.03	0.05	0.07	0.08
Al2O3	21.83	21.65	21.87	21.56	21.72	21.74	21.67	21.65	21.69	21.74	21.97	21.72	21.58	21.87	21.58
Cr2O3	0.04	0.05	0.03	0.02	0.03	0.02	0.03	0.03	0.02	0.02	0.03	0.01	0.01	0.01	0.03
Fe2O3	1.25	2.06	1.77	2.11	1.36	1.65	1.42	2.63	1.78	2.09	1.18	1.78	1.78	2.19	2
FeO	23.52	23.06	23.15	22.86	23.27	22.75	22	22.2	22.67	22.69	23.53	22.75	22.85	22.68	22.66
MnO	0.93	0.95	0.99	0.99	0.93	0.91	0.86	0.94	0.95	0.86	0.95	0.93	0.95	0.95	0.98
MgO	7.5	7.78	7.77	7.74	7.8	7.68	7.46	8.02	7.83	7.95	7.81	7.79	7.98	7.89	7.85
CaO	6.34	6.34	6.43	6.4	6.34	6.73	6.37	6.73	6.68	6.63	6.65	6.59	6.38	6.45	6.4
Na2O	0.01	0	0.02	0.02	0	0.02	0.01	0	0.01	0	0	0	0.02	0	0
K2O	0.02	0	0.01	0.01	0.01	0	0.03	0.01	0	0	0	0.01	0	0	0.01
Totals	99.97	100.43	100.83	100.17	100.15	100.03	97.03	100.73	100.26	100.66	101.4	100.1	100.27	100.51	99.89
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12
Si	2.969	2.954	2.96	2.957	2.972	2.964	2.944	2.945	2.963	2.955	2.979	2.962	2.97	2.944	2.956
Ti	0.003	0.005	0.004	0.005	0.003	0.002	0.002	0.002	0.003	0.003	0.003	0.002	0.003	0.004	0.004
Al	1.986	1.962	1.972	1.958	1.969	1.973	2.025	1.953	1.964	1.961	1.967	1.97	1.954	1.977	1.963
Cr	0.002	0.003	0.002	0.001	0.002	0.001	0.002	0.002	0.001	0.001	0.002	0.001	0.001	0.001	0.002
Fe3	0.072	0.119	0.102	0.122	0.079	0.095	0.084	0.151	0.103	0.12	0.068	0.103	0.103	0.127	0.116
Fe2	1.517	1.482	1.48	1.473	1.497	1.465	1.459	1.42	1.457	1.452	1.495	1.464	1.467	1.454	1.462
Mn	0.061	0.062	0.064	0.065	0.06	0.06	0.058	0.061	0.061	0.056	0.061	0.061	0.062	0.062	0.064
Mg	0.862	0.891	0.886	0.888	0.894	0.882	0.881	0.914	0.897	0.907	0.884	0.893	0.914	0.902	0.903
Ca	0.524	0.522	0.527	0.528	0.522	0.555	0.541	0.551	0.55	0.544	0.541	0.544	0.525	0.53	0.529
Na	0.001	0	0.003	0.003	0	0.002	0.002	0	0.001	0	0.001	0	0.002	0	0
K	0.002	0	0.001	0.001	0.001	0	0.003	0.001	0	0	0	0.001	0	0	0.001
Sum	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8
X(Mg)	0.362	0.375	0.374	0.376	0.374	0.376	0.376	0.392	0.381	0.384	0.372	0.379	0		

Sample	VAB01 Gr2.1	VAB01 Gr2.2	VAB01 Gr2.3	VAB01 Gr2.4	VAB01 Gr2.5	VAB01 Gr2.6	VAB01 Gr2.7	VAB01 Gr2.8	VAB01 Gr2.9	VAB01 Gr2.10	VAB01 Gr2.11	VAB01 Gr2.12	VAB01 Gr2.13	VAB01 Gr2.14	VAB01 Gr2.15
SiO2	38.97	38.68	38.53	38.77	38.58	39.11	38.63	38.61	38.4	38.83	38.77	38.66	38.57	38.57	38.26
TiO2	0.09	0.08	0.06	0.08	0.06	0.04	0.06	0.08	0.06	0.05	0.06	0.07	0.06	0.07	0.1
Al2O3	21.81	21.49	21.48	21.63	21.52	21.42	21.61	21.66	21.66	21.34	21.5	21.64	21.56	21.55	21.52
Cr2O3	0.03	0.02	0.02	0.03	0.03	0.03	0.02	0.03	0.01	0.03	0.04	0.04	0.03	0.03	0.04
Fe2O3	1.15	1.78	2.15	1.71	2.26	1.22	2.11	1.91	2.21	1.84	1.81	1.87	1.7	1.9	2.38
FeO	23.89	23.48	22.92	23.27	22.81	23.49	22.86	23.08	22.64	23.28	23.11	23.14	23.43	23.15	23.29
MnO	0.94	0.98	0.95	0.98	0.96	0.93	0.97	0.93	0.91	0.96	0.95	0.99	0.96	1.02	1.01
MgO	7.71	7.69	7.85	7.91	7.89	7.93	8	7.82	7.91	7.85	7.86	7.88	7.63	7.73	7.42
CaO	6.23	6.33	6.42	6.28	6.49	6.41	6.36	6.46	6.44	6.38	6.49	6.33	6.34	6.36	6.4
Na2O	0.01	0.01	0	0	0.01	0	0	0	0.01	0.01	0	0	0.01	0.01	0.01
K2O	0.01	0	0.01	0.01	0	0	0	0	0	0	0	0	0	0.01	0.01
Totals	100.84	100.52	100.38	100.66	100.6	100.59	100.63	100.58	100.26	100.58	100.59	100.61	100.28	100.4	100.45
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12
Si	2.98	2.971	2.961	2.969	2.958	2.995	2.959	2.961	2.952	2.979	2.972	2.964	2.969	2.965	2.948
Ti	0.005	0.004	0.004	0.004	0.003	0.002	0.003	0.004	0.004	0.003	0.004	0.004	0.003	0.004	0.006
Al	1.966	1.946	1.946	1.953	1.946	1.934	1.951	1.958	1.963	1.93	1.943	1.955	1.956	1.953	1.955
Cr	0.002	0.001	0.001	0.002	0.002	0.002	0.001	0.002	0	0.002	0.002	0.002	0.002	0.002	0.002
Fe3	0.066	0.103	0.124	0.099	0.131	0.071	0.122	0.11	0.128	0.106	0.104	0.108	0.098	0.11	0.138
Fe2	1.528	1.508	1.473	1.49	1.462	1.504	1.465	1.48	1.455	1.494	1.482	1.483	1.508	1.488	1.501
Mn	0.061	0.064	0.062	0.064	0.062	0.06	0.063	0.06	0.059	0.062	0.062	0.064	0.063	0.066	0.066
Mg	0.879	0.88	0.899	0.903	0.901	0.905	0.913	0.894	0.906	0.898	0.898	0.9	0.876	0.886	0.852
Ca	0.511	0.521	0.528	0.516	0.533	0.526	0.522	0.531	0.531	0.525	0.533	0.52	0.523	0.524	0.529
Na	0.002	0.001	0.001	0	0.001	0.001	0	0	0.002	0.001	0	0	0.002	0.002	0.002
K	0.001	0	0.001	0.001	0	0	0	0	0	0	0	0	0	0.001	0.001
Sum	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8
X(Mg)	0.365	0.369	0.379	0.377	0.381	0.376	0.384	0.377	0.384	0.375	0.377	0.378	0.367	0.373	0.362
alm	0.513	0.507	0.												

Continuação - Análise química de granada (VAB01)

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
Min	Gr3.1	Gr3.2	Gr3.3	Gr3.4	Gr3.5	Gr3.6	Gr3.7	Gr3.8	Gr3.9	Gr3.10	Gr3.11	Gr3.12	Gr3.13	Gr3.14	Gr3.15
SiO2	38.48	38.66	38.51	38.72	38.88	38.61	38.41	38.88	38.69	38.93	38.54	38.39	38.35	38.63	39.06
TiO2	0.02	0.06	0.08	0.08	0.08	0.09	0.06	0.06	0.06	0.09	0.07	0.06	0.07	0.07	0.02
Al2O3	21.65	21.52	21.72	21.43	21.66	21.64	21.55	21.26	21.47	21.68	21.54	21.74	21.7	21.73	22.26
Cr2O3	0	0.03	0.03	0.02	0.03	0.04	0.02	0.03	0.02	0.03	0.04	0.03	0.04	0.03	0.03
Fe2O3	2.02	1.73	1.64	1.5	1.16	2.02	2.14	1.25	1.32	1.91	1.59	1.96	2.16	1.99	1.02
FeO	23.32	23.56	23.28	23.61	23.58	22.97	23.07	23.88	23.47	23.26	23.31	23.09	23.23	23.27	23.86
MnO	1.02	1.02	0.99	0.99	0.97	0.98	1.01	0.97	0.95	0.97	0.96	0.98	0.9	0.95	0.99
MgO	7.62	7.57	7.62	7.78	7.79	7.83	7.71	7.59	7.7	7.96	7.71	7.8	7.62	7.71	8.01
CaO	6.31	6.4	6.42	6.15	6.31	6.38	6.31	6.35	6.38	6.36	6.34	6.21	6.38	6.41	5.88
Na2O	0	0	0	0	0.01	0.03	0.01	0	0	0.01	0	0	0	0.01	0.01
K2O	0	0	0.01	0	0	0	0	0	0	0	0	0.01	0	0.01	0.01
Totals	100.44	100.55	100.3	100.29	100.46	100.6	100.31	100.25	100.06	101.2	100.1	100.26	100.45	100.81	101.14
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12
Si	2.959	2.971	2.963	2.979	2.982	2.96	2.957	2.995	2.982	2.966	2.97	2.954	2.949	2.958	2.971
Ti	0.001	0.003	0.005	0.005	0.005	0.005	0.004	0.003	0.003	0.005	0.004	0.003	0.004	0.004	0.001
Al	1.963	1.95	1.969	1.944	1.959	1.956	1.956	1.93	1.951	1.948	1.957	1.972	1.967	1.962	1.996
Cr	0	0.002	0.002	0.001	0.002	0.002	0.001	0.002	0.001	0.002	0.002	0.002	0.002	0.002	0.002
Fe3	0.117	0.1	0.095	0.087	0.067	0.117	0.124	0.072	0.077	0.109	0.092	0.113	0.125	0.115	0.058
Fe2	1.5	1.514	1.498	1.519	1.513	1.473	1.485	1.538	1.513	1.482	1.502	1.485	1.494	1.49	1.518
Mn	0.067	0.066	0.064	0.065	0.063	0.064	0.066	0.063	0.062	0.062	0.063	0.064	0.059	0.062	0.064
Mg	0.873	0.867	0.874	0.893	0.89	0.895	0.884	0.871	0.884	0.904	0.886	0.894	0.874	0.879	0.908
Ca	0.52	0.527	0.529	0.507	0.519	0.524	0.521	0.524	0.527	0.519	0.523	0.512	0.526	0.526	0.48
Na	0	0	0.001	0	0.001	0.005	0.002	0	0	0.001	0	0	0	0.001	0.001
K	0	0	0.001	0	0	0	0	0	0	0	0	0.001	0	0.001	0.001
Sum	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8
X(Mg)	0.368	0.364	0.368	0.370	0.370	0.378	0.373	0.362	0.369	0.379	0.371	0.376	0.369	0.371	0.374
alm	0.507	0.509	0.505	0.509	0.507	0.498	0.502	0.513	0.507	0.499	0.505	0.503	0.506	0.504	0.511
prp	0.295	0.292	0.295	0.299	0.298	0.303	0.299	0.291	0.296	0.305	0.298	0.303	0.296	0.297	0.306
grs	0.119	0.128	0.132	0.127	0.141	0.121	0.117	0.139	0.139	0.122	0.131	0.119	0.118	0.123	0.133
sps	0.023	0.022	0.022	0.022	0.021	0.022	0.022	0.021	0.021	0.021	0.021	0.022	0.020	0.021	0.022

Análise química de plagioclásio (VAB01)

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
Min	Pl2.1	Pl2.2	Pl2.3	Pl2.4	Pl2.5	Pl2.6	Pl2.7	Pl2.8	Pl2.9	Pl2.10	Pl2.11	Pl4.1	Pl4.2						
SiO2	58.44	58.44	58.42	59.14	58.17	58.29	58.4	59.09	58.91	59.16	58.54	58.53	58.89						
TiO2	0	0	0	0	0	0	0	0	0	0	0	0	0						
Al2O3	26.85	26.95	26.83	26.85	27.1	26.87	26.84	26.75	26.63	26.42	26.68	26.88	26.61						
Cr2O3	0	0	0	0	0	0	0	0	0	0	0	0	0						
Fe2O3	0.07	0.05	0.08	0.06	0.09	0.08	0.11	0.08	0.12	0.1	0.2	0.09	0.09						
FeO	0	0	0	0	0	0	0	0	0	0	0	0	0						
MnO	0	0	0	0	0	0	0	0	0	0	0	0	0						
MgO	0	0	0	0	0	0	0	0	0	0	0	0	0						
CaO	8.18	8.16	8.18	7.9	8.2	8.21	8	7.9	7.66	7.65	7.86	7.87	7.86						
Na2O	7.1	6.9	6.96	6.95	6.84	6.86	6.92	7.1	7.08	7.16	6.91	6.86	6.8						
K2O	0.16	0.14	0.17	0.2	0.11	0.14	0.1	0.1	0.29	0.28	0.27	0.28	0.34						
Totals	100.79	100.65	100.64	101.09	100.51	100.45	100.37	101.02	100.7	100.77	100.47	100.51	100.58						
Oxygens	8	8	8	8	8	8	8	8	8	8	8	8	8						
Si	2.595	2.595	2.597	2.612	2.588	2.595	2.599	2.612	2.614	2.622	2.605	2.602	2.615						
Ti	0	0	0	0	0	0	0	0	0	0	0	0	0						
Al	1.406	1.411	1.406	1.398	1.421	1.41	1.409	1.394	1.393	1.381	1.4	1.409	1.393						
Cr	0	0	0	0	0	0	0	0	0	0	0	0	0						
Fe3	0.002	0.002	0.003	0.002	0.003	0.003	0.004	0.003	0.004	0.003	0.007	0.003	0.003						
Fe2	0	0	0	0	0	0	0	0	0	0	0	0	0						
Mn	0	0	0	0	0	0	0	0	0	0	0	0	0						
Mg	0	0	0	0	0	0	0	0	0	0	0	0	0						
Ca	0.389	0.389	0.389	0.374	0.391	0.392	0.382	0.374	0.364	0.364	0.375	0.375	0.374						
Na	0.611	0.594	0.599	0.595	0.59	0.592	0.597	0.609	0.609	0.616	0.596	0.591	0.585						
K	0.009	0.008	0.01	0.011	0.006	0.008	0.005	0.006	0.016	0.016	0.015	0.016	0.019						
Sum	5.012	4.999	5.004	4.992	4.999	4.999	4.996	4.997	5.001	5.001	4.997	4.996	4.989						
an	38.6	39.3	39.0	38.2	39.6	39.5	38.8	37.8	36.8	36.5	38.0	38.2	38.2						
ab	60.6	59.9	60.0	60.7	59.8	59.7	60.7	61.6	61.6	61.8	60.4	60.2	59.8						
or	0.9	0.8	1.0	1.1	0.6	0.8	0.5	0.6	1.6	1.6	1.5	1.6	1.9						

Continuação - Análise química de plagioclásio (VAB01)

Sample	VAB01 P 4.3	VAB01 P 4.4	VAB01 P 4.5	VAB01 P 4.6	VAB01 P 4.7	VAB01 P 4.8	VAB01 P 4.9	VAB01 P 4.10	VAB01 P 5.1	VAB01 P 5.2	VAB01 P 5.3	VAB01 P 5.4	VAB01 P 5.5
Min													
SiO2	58.61	59.67	59.06	59.05	58.9	58.95	58.38	58.69	58.62	58.89	58.85	58.31	58.87
TiO2	0	0	0	0	0	0	0	0	0	0	0	0	0
Al2O3	26.48	26.98	26.53	26.6	26.7	26.67	26.44	26.86	26.98	26.61	26.72	26.54	27.2
Cr2O3	0	0	0	0	0	0	0	0	0	0	0	0	0
Fe2O3	0.09	0.1	0.1	0.11	0.11	0.16	0.18	0.12	0.05	0.08	0.22	0.29	0.02
FeO	0	0	0	0	0	0	0	0	0	0	0	0	0
MnO	0	0	0	0	0	0	0	0	0	0	0	0	0
MgO	0	0	0	0	0	0	0	0	0	0	0	0	0
CaO	7.79	7.86	7.67	7.72	7.72	7.93	7.79	7.83	8.14	7.62	7.86	7.93	8.29
Na2O	7.03	7.01	6.85	7.05	6.88	6.86	6.9	6.76	7	6.95	7.17	7.03	6.88
K2O	0.35	0.35	0.41	0.42	0.44	0.38	0.44	0.38	0.13	0.24	0.18	0.14	0.08
Totals	100.36	101.97	100.62	100.96	100.75	100.95	100.12	100.64	100.92	100.38	100.99	100.24	101.34
Oxygens	8	8	8	8	8	8	8	8	8	8	8	8	8
Si	2.612	2.614	2.621	2.615	2.612	2.611	2.609	2.606	2.597	2.618	2.606	2.602	2.595
Ti	0	0	0	0	0	0	0	0	0	0	0	0	0
Al	1.391	1.394	1.388	1.389	1.396	1.393	1.393	1.406	1.409	1.394	1.395	1.396	1.414
Cr	0	0	0	0	0	0	0	0	0	0	0	0	0
Fe3	0.003	0.003	0.003	0.004	0.004	0.005	0.006	0.004	0.002	0.003	0.007	0.01	0.001
Fe2	0	0	0	0	0	0	0	0	0	0	0	0	0
Mn	0	0	0	0	0	0	0	0	0	0	0	0	0
Mg	0	0	0	0	0	0	0	0	0	0	0	0	0
Ca	0.372	0.369	0.365	0.367	0.367	0.376	0.373	0.373	0.386	0.363	0.373	0.379	0.392
Na	0.608	0.596	0.589	0.606	0.592	0.589	0.598	0.582	0.602	0.599	0.615	0.608	0.588
K	0.02	0.019	0.023	0.024	0.025	0.021	0.025	0.022	0.007	0.013	0.01	0.008	0.005
Sum	5.005	4.995	4.99	5.003	4.996	4.996	5.003	4.991	5.002	4.99	5.006	5.003	4.994
an	37.2	37.5	37.4	36.8	37.3	38.1	37.4	38.2	38.8	37.2	37.4	38.1	39.8
ab	60.8	60.6	60.3	60.8	60.2	59.7	60.0	59.6	60.5	61.4	61.6	61.1	59.7
or	2.0	1.9	2.4	2.4	2.5	2.1	2.5	2.3	0.7	1.3	1.0	0.8	0.5

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
Min	Opx1.1	Opx1.2	Opx1.3	Opx1.4	Opx1.5	Opx1.6	Opx1.7	Opx1.8	Opx1.9	Opx1.10	Opx2.1	Opx2.2	Opx2.3	Opx2.4	Opx2.5	Opx3.1	Opx3.2	Opx3.3	Opx3.4
SiO2	51.86	51.15	50.71	50.6	50.68	50.56	50.38	50.22	50.31	50.86	51.37	50.79	50.8	50.46	51.14	50.95	50.61	50.43	50.41
TiO2	0.06	0.09	0.08	0.09	0.12	0.11	0.11	0.1	0.1	0.09	0.08	0.1	0.09	0.09	0.12	0.12	0.09	0.11	0.11
Al2O3	1.62	2.44	2.55	2.58	2.79	2.95	2.74	2.75	2.81	2.45	3	2.69	2.57	3.07	2.51	2.34	2.45	2.9	2.72
Cr2O3	0.02	0.01	0.01	0.03	0.02	0.02	0.02	0	0.01	0.02	0.01	0.02	0	0.03	0.01	0.01	0.01	0.02	0.01
Fe2O3	0	0.01	0.65	0.99	0.25	0.66	1.07	1.33	0.13	0.43	0.56	1.04	0.76	0.66	0.21	0.78	0.9	0.47	0.76
FeO	24.13	24.17	23.89	23.71	24.21	23.98	23.63	23.5	24.32	24.35	23.77	23.06	23.62	23.64	23.37	23.81	24.04	24.52	24.12
MnO	0.36	0.34	0.37	0.37	0.35	0.35	0.33	0.36	0.34	0.36	0.39	0.36	0.39	0.36	0.34	0.39	0.41	0.4	0.39
MgO	20.49	20.2	20	19.88	19.62	19.8	19.7	19.66	19.44	19.76	20.57	20.27	20.05	19.88	20.43	20.1	19.63	19.35	19.49
CaO	0.48	0.51	0.51	0.73	0.82	0.67	0.81	0.85	0.69	0.65	0.47	0.88	0.78	0.69	0.73	0.64	0.77	0.63	0.76
Na2O	0.01	0.02	0.03	0.03	0.03	0.02	0.05	0.03	0.01	0.03	0.01	0.03	0.01	0.02	0.04	0.04	0.03	0.03	0.03
K2O	0	0	0	0	0	0	0	0	0.01	0	0	0.01	0	0	0	0	0.01	0.01	0.01
Totals	99.02	98.94	98.79	99.01	98.89	99.11	98.85	98.8	98.14	99	100.23	99.24	99.08	98.9	98.91	99.17	98.94	98.86	98.81
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6
Si	1.967	1.943	1.932	1.926	1.931	1.922	1.921	1.917	1.932	1.937	1.924	1.923	1.93	1.919	1.939	1.934	1.931	1.926	1.926
Ti	0.002	0.003	0.002	0.003	0.003	0.003	0.003	0.003	0.003	0.003	0.002	0.003	0.003	0.003	0.003	0.003	0.003	0.003	0.003
Al	0.072	0.109	0.115	0.116	0.125	0.132	0.123	0.124	0.127	0.11	0.132	0.12	0.115	0.138	0.112	0.105	0.11	0.13	0.122
Cr	0.001	0	0	0.001	0.001	0	0.001	0	0	0	0	0	0	0.001	0	0	0	0.001	0
Fe3	0	0	0.019	0.028	0.007	0.019	0.031	0.038	0.004	0.012	0.016	0.03	0.022	0.019	0.006	0.022	0.026	0.013	0.022
Fe2	0.765	0.768	0.761	0.755	0.772	0.762	0.754	0.75	0.781	0.776	0.745	0.73	0.75	0.752	0.741	0.756	0.767	0.783	0.771
Mn	0.012	0.011	0.012	0.012	0.011	0.011	0.011	0.012	0.011	0.012	0.012	0.012	0.013	0.012	0.011	0.013	0.013	0.013	0.013
Mg	1.158	1.144	1.136	1.128	1.114	1.121	1.12	1.118	1.113										

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
Min	Opx3.5	Opx3.6	Opx3.7	Opx3.8	Opx3.9	Opx3.10	Opx4.1	Opx4.2	Opx4.3	Opx4.4	Opx6.1	Opx6.2	Opx6.3	Opx6.4	Opx6.5	Opx9.1	Opx9.2	Opx9.3	Opx9.4
SiO2	50.26	50.54	50.21	50.55	50.78	50.35	50.84	50.63	50.23	50.43	50.72	50.81	51.18	50.97	51.15	50.63	50.63	48.44	50.63
TiO2	0.1	0.08	0.12	0.09	0.09	0.09	0.11	0.12	0.11	0.09	0.1	0.08	0.09	0.09	0.08	0.1	0.11	0.12	0.09
Al2O3	2.86	2.74	3.08	2.86	2.69	2.25	2.51	2.74	2.66	2.38	2.38	2.59	2.31	2.49	2.37	2.43	2.6	2.66	2.58
Cr2O3	0.01	0.01	0.02	0.02	0.02	0	0.01	0.01	0.02	0.02	0.02	0.01	0.02	0.01	0.01	0.01	0	0.01	0.02
Fe2O3	1.01	0.79	1.09	0	0.21	1.11	0.63	0.39	1.55	0.91	0.91	0.14	0.13	0.78	0.59	0	0.48	1.07	0.92
FeO	23.88	24.35	24.01	24.92	24.86	24.54	23.74	24	23.17	24.11	23.86	24.01	24.6	23.15	23.69	24.45	24.63	24.15	24.14
MnO	0.38	0.37	0.43	0.36	0.41	0.39	0.42	0.41	0.39	0.41	0.38	0.4	0.36	0.32	0.36	0.41	0.37	0.4	0.38
MgO	19.68	19.6	19.3	19.12	19.28	19.36	20	19.63	19.74	19.65	19.94	19.89	19.95	20.05	20.45	19.67	19.53	18.19	19.82
CaO	0.58	0.62	0.88	0.78	0.81	0.63	0.75	0.86	0.98	0.6	0.62	0.68	0.57	1.19	0.53	0.48	0.57	0.67	0.48
Na2O	0.03	0.01	0.04	0.03	0.03	0.01	0.03	0.03	0.04	0.01	0.04	0.03	0.01	0.06	0.01	0.01	0.02	0.04	0.04
K2O	0	0.01	0	0	0	0	0	0.01	0	0	0	0	0	0	0	0	0	0.01	0
Totals	98.78	99.12	99.18	98.72	99.18	98.74	99.04	98.82	98.88	98.62	98.95	98.63	99.21	99.12	99.25	98.2	98.94	95.76	99.08
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6
Si	1.919	1.925	1.913	1.934	1.935	1.931	1.932	1.93	1.916	1.931	1.932	1.938	1.944	1.933	1.937	1.943	1.932	1.92	1.928
Ti	0.003	0.002	0.004	0.003	0.003	0.003	0.003	0.003	0.003	0.002	0.003	0.002	0.003	0.003	0.002	0.003	0.003	0.003	0.002
Al	0.129	0.123	0.138	0.129	0.121	0.102	0.112	0.123	0.119	0.107	0.107	0.117	0.103	0.111	0.106	0.11	0.117	0.124	0.116
Cr	0	0	0	0.001	0.001	0	0	0	0	0.001	0	0	0.001	0	0	0	0	0	0.001
Fe3	0.029	0.023	0.031	0	0.006	0.032	0.018	0.011	0.045	0.026	0.026	0.004	0.004	0.022	0.017	0	0.014	0.032	0.026
Fe2	0.762	0.776	0.765	0.797	0.792	0.787	0.755	0.765	0.739	0.772	0.76	0.766	0.781	0.734	0.75	0.785	0.786	0.801	0.769
Mn	0.012	0.012	0.014	0.012	0.013	0.013	0.014	0.013	0.012	0.013	0.012	0.013	0.011	0.01	0.012	0.013	0.012	0.014	0.012
Mg	1.12	1.113	1.096	1.09	1.095	1.107	1.133	1.116	1.122	1.1									

Análise química de clonopiroxênio (VAB01)

Sample	Min	SiO2	TiO2	Al2O3	Cr2O3	Fe2O3	FeO	MnO	MgO	CaO	Na2O	K2O	Totals	Oxygens	Si	Ti	Al	Cr	Fe3	Fe2	Mn	Mg	Ca	Na	K	Sum
cp5.1	cp5.2	cp5.3	cp5.4	cp5.5	cp5.6	cp7.1	cp7.2	cp7.3	cp7.4	cp7.5	cp7.6	cp7.7	cp7.8	cp7.9	cp7.10	cp7.11	cp7.12	cp8.1	cp8.2	cp8.3	cp8.4	cp8.5	cp8.6	cp8.7	cp8.8	
49.87	50.01	50.24	49.7	49.63	49.77	49.59	49.71	49.45	49.61	49.67	49.84	49.86	50.36	49.49	49.68	50.02	50.06	49.92	49.82	49.4						
0.41	0.41	0.42	0.43	0.45	0.44	0.42	0.44	0.47	0.48	0.39	0.39	0.43	0.38	0.45	0.46	0.37	0.42	0.43	0.46	0.41						
4.25	4.69	4.7	4.72	4.85	4.49	4.76	4.66	4.88	4.87	4.5	4.57	4.73	4.48	4.83	4.76	4.28	4.06	4.83	4.86	4.7						
0.03	0.02	0.01	0.03	0.02	0.01	0.01	0.02	0.04	0.02	0.02	0.03	0.02	0.02	0.03	0.01	0.02	0.03	0.02	0.02	0.02						
2.35	2.94	2.45	2.75	2.76	3.22	2.43	2.14	2.26	2.36	1.68	2.56	2.72	1.83	1.61	1.98	2.53	2.34	2.69	2.05	2.19	2.31					
8.3	8.18	8.62	8.94	7.62	8.89	10.57	10.32	9.8	9.79	11.91	9.31	9.68	10.09	9.7	8.55	8.17	8.64	10.11	9.67	11.23						
0.17	0.17	0.2	0.2	0.17	0.21	0.21	0.19	0.21	0.18	0.25	0.21	0.21	0.22	0.22	0.21	0.18	0.2	0.21	0.21	0.22						
11.86	11.53	11.38	11.58	11.46	11.27	11.72	11.7	11.3	10.91	12.34	11.58	11.18	11.71	11.29	11.17	11.65	11.77	11.45	11.16	11.96						
21.28	21.06	20.35	20.3	21.55	20.8	19.03	19.35	19.87	20.54	17.3	20.03	20.42	19.78	20.11	20.91	21.16	20.78	19.84	20.38	18.05						
0.8	0.88	0.88	0.9	0.87	0.84	0.88	0.78	0.79	0.85	0.87	0.74	0.89	0.87	0.82	0.94	0.84	0.82	0.85	0.88	0.76						
0	0.02	0.01	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.01	0.01						
99.81	99.78	99.7	100.3	99.66	99.58	99.18	99.25	99.44	99.24	98.95	99.69	99.21	99.52	98.91	99.21	99.04	99.46	99.7	99.66	99.07						
6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6						
1.89	1.875	1.882	1.878	1.87	1.885	1.88	1.88	1.875	1.885	1.879	1.881	1.888	1.898	1.88	1.879	1.891	1.889	1.882	1.88	1.878						
0.012	0.012	0.011	0.012	0.012	0.013	0.013	0.012	0.012	0.013	0.014	0.011	0.012	0.011	0.013	0.013	0.011	0.012	0.012	0.013	0.012						
0.188	0.208	0.209	0.208	0.216	0.205	0.201	0.213	0.208	0.218	0.201	0.203	0.211	0.199	0.216	0.212	0.191	0.18	0.215	0.216	0.211						
0.001	0.001	0	0.001	0.001	0.001	0	0	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0	0.001	0.001	0.001	0.001	0.001						
0.066	0.083	0.07	0.077	0.078	0.091	0.069	0.061	0.064	0.067	0.048	0.073	0.077	0.052	0.046	0.057	0.072	0.066	0.076	0.058	0.066						
0.26	0.257	0.271	0.276	0.282	0.24	0.282	0.335	0.326	0.311	0.311	0.377	0.294	0.307	0.318	0.308	0.271	0.258	0.273	0.319	0.305						
0.005	0.006	0.005	0.006	0.006	0.005	0.007	0.007	0.006	0.007	0.006	0.008	0.007	0.007	0.007	0.007	0.007	0.006	0.006	0.007	0.007						
0.663	0.646	0.638	0.662	0.65	0.644	0.636	0.662	0.66	0.638	0.618	0.696	0.651	0.631	0.658	0.639	0.629	0.657	0.662	0.643	0.628						
0.856	0.848	0.849	0.815	0.819	0.87	0.844	0.773	0.784	0.807	0.836	0.701	0.81	0.828	0.799	0.818	0.847	0.857	0.84	0.801	0.824						
0.058	0.064	0.064	0.065	0.063	0.062	0.064	0.057	0.058	0.062	0.064	0.054	0.065	0.064	0.063	0.06	0.069	0.062	0.06	0.062	0.064						
0	0.001	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.001						
4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4						
Sum	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4						

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
Min	Lam1	Lam2	Lam3	Lam4	Lam5	Lam6	Lam7	Lam8	Lam9	Lam10	Lam11	Lam12	Lam13	Lam14	Lam15	Lam16	Lam17	Lam18
SiO2	49.86	49.46	50.12	49.78	49.71	49.9	50.37	50.17	50.39	50.5	50.53	50.92	50.49	50.4	49.52	50.2	50.23	49.8
TiO2	0.06	0.16	0.08	0.05	0.03	0.04	0.05	0.06	0.04	0.05	0.05	0.04	0.05	0.06	0.22	0.06	0.08	0.13
Al2O3	3.4	3.73	3.58	3.61	3.44	3.31	3.05	3.11	3.16	3.15	2.87	2.8	2.85	2.87	3.65	3.18	3.48	3.68
Cr2O3	0.01	0.02	0.01	0	0.01	0.02	0.02	0.02	0.01	0	0.03	0.02	0.01	0.01	0.02	0.02	0.02	0.01
Fe2O3	0.36	0.58	0	0	0.16	0	0	0	0	0	0	0	0	0	0.53	0	0	0.22
FeO	26.64	22.7	26.78	26.94	26.76	26.79	26.74	26.73	27.18	26.38	26.61	26.71	26.57	26.46	19.5	26.83	26.11	24.67
MnO	0.43	0.37	0.45	0.45	0.44	0.43	0.46	0.45	0.42	0.48	0.45	0.42	0.43	0.46	0.34	0.47	0.46	0.42
MgO	17.32	15.91	17.32	17.47	17.73	17.65	17.83	17.8	17.6	17.79	18.11	17.86	17.79	17.49	14.72	17.47	16.98	16.38
CaO	1.21	5.36	1.15	0.47	0.52	0.55	0.54	0.66	0.68	0.79	0.75	0.85	0.77	1.27	9.07	0.96	1.88	3.78
Na2O	0.04	0.21	0.04	0.01	0	0	0.01	0.02	0	0.04	0.01	0.03	0	0.04	0.37	0.03	0.07	0.11
K2O	0	0	0	0	0	0	0	0.01	0	0.02	0	0	0	0.01	0	0	0	0
Totals	99.34	98.51	99.52	98.78	98.8	98.69	99.06	99.04	99.48	99.18	99.39	99.65	98.96	99.07	97.95	99.23	99.29	99.21
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6
Si	1.917	1.91	1.921	1.921	1.918	1.927	1.936	1.93	1.932	1.936	1.935	1.945	1.941	1.939	1.916	1.93	1.928	1.914
Ti	0.002	0.005	0.002	0.001	0.001	0.001	0.001	0.002	0.001	0.001	0.001	0.001	0.001	0.002	0.006	0.002	0.002	0.004
Al	0.154	0.17	0.162	0.164	0.156	0.151	0.138	0.141	0.143	0.143	0.129	0.126	0.129	0.13	0.166	0.144	0.157	0.167
Cr	0	0.001	0	0	0	0.001	0.001	0.001	0	0	0.001	0.001	0	0	0.001	0.001	0.001	0
Fe3	0.01	0.017	0	0	0.005	0	0	0	0	0	0	0	0	0	0.016	0	0	0.007
Fe2	0.857	0.733	0.858	0.869	0.864	0.865	0.859	0.86	0.872	0.846	0.852	0.853	0.855	0.851	0.631	0.863	0.838	0.793
Mn	0.014	0.012	0.015	0.015	0.014	0.014	0.015	0.015	0.014	0.015	0.015	0.014	0.014	0.015	0.011	0.015	0.015	0.014
Mg	0.993	0.916	0.989	1.005	1.02	1.016	1.021	1.021	1.006	1.016	1.034	1.017	1.02	1.003	0.849	1.001	0.971	0.938
Ca	0.05	0.222	0.047	0.019	0.021	0.023	0.022	0.027	0.028	0.033	0.031	0.035	0.032	0.052	0.376	0.04	0.077	0.155
Na	0.003																	

Sample	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01	VAB01
Min	Hosp1	Hosp2	Hosp3	Hosp4	Hosp5	Hosp6	Hosp7	Hosp8	Hosp9	Hosp10	Hosp11	Hosp12	Hosp13	Hosp14	Hosp15	Hosp16	Hosp17	Hosp18
SiO2	50.05	49.8	50.03	50.04	50.02	49.91	50.07	49.78	50.03	51.22	50.53	50.13	50.26	50.31	50.45	49.95	50.22	50.16
TiO2	0.51	0.45	0.46	0.45	0.44	0.43	0.44	0.48	0.45	0.34	0.45	0.42	0.43	0.47	0.47	0.47	0.46	0.45
Al2O3	4.88	4.84	4.65	4.75	4.97	4.76	4.88	4.98	4.83	3.76	4.86	4.83	4.59	5.06	5.03	4.99	5.17	4.87
Cr2O3	0.01	0.02	0.03	0.01	0.03	0.03	0.03	0.03	0.02	0.04	0.03	0.02	0.02	0.02	0.03	0.02	0.02	0.01
Fe2O3	2.01	1.94	2.16	1.74	1.39	2.19	1.6	2.31	2.16	1.69	2.27	2.29	1.53	2.37	1.77	2.39	1.87	1.72
FeO	8.89	9.1	8.79	8.81	15.93	8.6	9.14	8.31	8.67	8.82	8.72	8.24	10.27	8.5	8.9	8.69	8.77	9.14
MnO	0.21	0.2	0.18	0.18	0.22	0.18	0.17	0.18	0.19	0.2	0.2	0.22	0.19	0.19	0.19	0.17	0.2	0.21
MgO	10.96	11.07	11.13	11.17	14.07	11.1	11	11.18	11.1	11.77	10.99	11.1	11.33	11.14	11.15	11.11	11.05	11.03
CaO	21.44	20.72	21.26	21.37	12.98	21.54	21.3	21.46	21.43	21.72	21.64	21.4	20.04	21.39	21.16	21.18	21.35	21.09
Na2O	0.91	0.95	0.9	0.86	0.51	0.84	0.87	0.87	0.89	0.81	0.99	1	0.89	1	1	0.94	0.95	0.94
K2O	0	0	0.01	0	0	0	0.01	0	0	0	0.01	0.01	0.01	0	0.01	0	0	0
Totals	99.86	99.08	99.59	99.38	100.55	99.59	99.5	99.59	99.75	100.36	100.68	99.66	99.54	100.43	100.16	99.9	100.06	99.63
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6
Si	1.882	1.886	1.886	1.888	1.876	1.881	1.888	1.875	1.882	1.912	1.884	1.885	1.897	1.878	1.887	1.876	1.881	1.889
Ti	0.014	0.013	0.013	0.013	0.012	0.012	0.013	0.014	0.013	0.01	0.013	0.012	0.012	0.013	0.013	0.013	0.013	0.013
Al	0.216	0.216	0.207	0.211	0.22	0.212	0.217	0.221	0.214	0.165	0.214	0.214	0.204	0.223	0.222	0.221	0.228	0.216
Cr	0	0.001	0.001	0	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0	0.001	0.001	0	0
Fe3	0.057	0.055	0.061	0.049	0.039	0.062	0.045	0.065	0.061	0.048	0.064	0.065	0.043	0.066	0.05	0.068	0.053	0.049
Fe2	0.279	0.288	0.277	0.278	0.5	0.271	0.288	0.262	0.273	0.275	0.272	0.259	0.324	0.265	0.278	0.273	0.275	0.288
Mn	0.007	0.006	0.006	0.006	0.007	0.006	0.005	0.006	0.006	0.006	0.006	0.007	0.006	0.006	0.006	0.005	0.006	0.007
Mg	0.614	0.624	0.625	0.628	0.786	0.623	0.618	0.627	0.622	0.655	0.611	0.622	0.637	0.62	0.622	0.622	0.617	0.619
Ca	0.864	0.841	0.858	0.864	0.522	0.87	0.861	0.										

Análise química de biotita (VAB01)

Sample	VAB01_bio_1	VAB01_bio_2	VAB01_bio_3	VAB01_bio_4	VAB01_bio_5	VAB01_bio_6	VAB01_bio_7	VAB01_bio_8
Min	bi	bi	bi	bi	bi	bi	bi	bi
SiO2	37.87	37.83	37.52	37.38	37.47	38.25	38.11	37.52
TiO2	5.18	4.2	2.67	4.15	3.72	5.11	4.28	2.4
Al2O3	15.26	14.57	17.57	15.35	15.32	14.03	14.93	17.5
Cr2O3	0	0.02	0	0.01	0.01	0.02	0.01	0
Fe2O3	0	0	0.03	0	0	0	0	1.58
FeO	11.51	12.74	11.54	12.71	11.97	12.22	12.57	10.03
MnO	0.02	0.01	0.03	0.01	0	0.01	0.03	0.03
MgO	15.66	15.8	16.1	15.04	15.79	15.47	15.7	16.4
CaO	0	0	0	0	0	0	0	0
Na2O	0.2	0.11	0.55	0.07	0.08	0.08	0.06	0.27
K2O	9.32	9.02	8.44	9.18	9.31	9.24	9.39	8.4
Totals	95.01	94.3	94.44	93.89	93.66	94.43	95.07	94.13
Oxygens	11	11	11	11	11	11	11	11
Si	2.784	2.815	2.756	2.795	2.8	2.837	2.813	2.754
Ti	0.286	0.235	0.147	0.233	0.209	0.285	0.237	0.133
Al	1.323	1.279	1.522	1.353	1.35	1.227	1.299	1.515
Cr	0	0.001	0	0.001	0.001	0.001	0.001	0
Fe3	0	0	0.002	0	0	0	0	0.087
Fe2	0.707	0.793	0.709	0.795	0.748	0.758	0.776	0.616
Mn	0.001	0.001	0.002	0	0	0	0.002	0.002
Mg	1.716	1.752	1.763	1.675	1.758	1.711	1.727	1.794
Ca	0	0	0	0	0	0	0	0
Na	0.028	0.016	0.079	0.01	0.011	0.011	0.008	0.038
K	0.874	0.857	0.791	0.876	0.888	0.875	0.884	0.786
Sum	7.72	7.748	7.77	7.739	7.765	7.707	7.747	7.725
X(Mg)	0.708	0.688	0.713	0.678	0.702	0.693	0.690	0.744

Sample	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
Min	grt.1.1	grt.1.2	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.1.1	grt.1.2	grt.1.3	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.1.4	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.1.5	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.1.6	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.1.7	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.1.8	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.1	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.2	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.3	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.4	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.5	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.6	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.7	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.8	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.9	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
	grt.2.10	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
Sio2	38.82	38.43	38.52	38.72	38.64	38.59	38.7	38.55	39.06	38.42	38.56	38.77	38.69	38.77	38.5	38.66	38.45	38.25	38.25
TiO2	0.07	0.05	0.08	0.08	0.08	0.09	0.07	0.09	0.06	0.09	0.09	0.06	0.08	0.07	0.06	0.08	0.08	0.06	0.06
Al2O3	21.33	21.19	21.45	21.08	21.29	21.29	21.4	21.13	21.52	21.3	21.49	21.21	21.33	21.43	21.48	21.07	21.23	21.19	21.19
Cr2O3	0.01	0	0	0.02	0	0.01	0.01	0	0	0.02	0	0.01	0.01	0	0.01	0.01	0.03	0.01	0.01
Fe2O3	0.98	1.37	1.71	0.96	1.37	1.12	0.51	1.4	0.72	1.19	0.96	0.92	0.92	0.81	1.21	0.95	1.46	1.14	1.14
FeO	25.32	24.97	24.71	25.15	24.53	24.94	25.35	25.03	25.14	24.7	24.65	24.85	24.76	24.97	25.01	25.82	25.91	26.1	26.1
MnO	1.48	1.																	

Análise química de plagioclásio (VAB19)

[illegible]

Sample	VAB19 Opx6.1	VAB19 Opx6.2	VAB19 Opx6.3	VAB19 Opx6.4	VAB19 Opx7.1	VAB19 Opx7.2	VAB19 Opx7.3	VAB19 Opx7.4	VAB19 Opx9c	VAB19 Opx9b
Min										
SiO2	50.96	50.46	49.83	50.33	50.27	50.77	50.66	51.07	50.67	50.47
TiO2	0.08	0.07	0.07	0.06	0.06	0.06	0.07	0.07	0.09	0.09
Al2O3	1.92	2.01	2	1.9	2.16	1.69	1.94	1.98	2.32	1.95
Cr2O3	0.02	0	0.01	0.01	0	0	0.01	0.02	0.01	0
Fe2O3	0.13	0.7	0.09	0.47	0.51	0.53	0.57	0	0.44	0.41
FeO	27.32	26.94	27.15	26.71	27.04	26.8	26.72	26.79	26.16	25.82
MnO	0.68	0.72	0.71	0.69	0.71	0.69	0.65	0.65	0.68	0.65
MgO	17.97	17.82	17.31	17.97	17.73	18.24	18.08	18.13	18.29	18.6
CaO	0.63	0.62	0.65	0.55	0.53	0.57	0.72	0.54	0.77	0.58
Na2O	0.04	0.03	0.02	0.02	0.03	0.01	0.03	0.02	0.05	0.01
K2O	0	0.01	0	0	0	0	0	0.01	0	0.01
Totals	99.75	99.39	97.84	98.71	99.03	99.35	99.44	99.27	99.47	98.58
Oxygens	6	6	6	6	6	6	6	6	6	6
Si	1.954	1.943	1.951	1.949	1.943	1.953	1.947	1.96	1.941	1.948
Ti	0.002	0.002	0.002	0.002	0.002	0.002	0.002	0.002	0.003	0.003
Al	0.087	0.091	0.093	0.087	0.098	0.077	0.088	0.09	0.105	0.089
Cr	0.001	0	0	0	0	0	0	0	0	0
Fe3	0.004	0.02	0.003	0.014	0.015	0.015	0.016	0	0.013	0.012
Fe2	0.876	0.868	0.889	0.865	0.874	0.862	0.859	0.86	0.838	0.833
Mn	0.022	0.023	0.023	0.023	0.023	0.022	0.021	0.021	0.022	0.021
Mg	1.027	1.023	1.01	1.037	1.021	1.046	1.035	1.037	1.044	1.07
Ca	0.026	0.026	0.027	0.023	0.022	0.023	0.03	0.022	0.032	0.024
Na	0.003	0.002	0.002	0.002	0.002	0.001	0.002	0.001	0.004	0.001
K	0	0	0	0	0	0	0	0	0	0
Sum	4	4	4	4	4	4	4	3.994	4	4
X(Mg)	0.539674	0.540984	0.531859	0.545216	0.538786	0.548218	0.546463	0.546653	0.554729	0.56227
xAl	0.041	0.034	0.044	0.036	0.041	0.03	0.035	0.05	0.046	0.037

Sample	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
Min	cpx1.1	cpx1.2	cpx1.3	cpx2.1	cpx2.2	cpx2.3	cpx3.1	cpx3.2	cpx3.3	cpx3.4	cpx5.1	cpx5.2	cpx5.3	VAB19	VAB19	VAB19	VAB19	VAB19	VAB19
SiO2	51.41	50.31	50.7	50.49	50.79	50.96	49.69	50.34	50.4	50.11	50.46	50.21	50.88	50.72	50.88	50.21	50.88	50.72	50.96
TiO2	0.25	0.4	0.33	0.27	0.31	0.27	0.35	0.36	0.36	0.34	0.36	0.33	0.33	0.33	0.33	0.33	0.33	0.33	0.29
Al2O3	3.19	4.27	4.02	3.44	3.98	3.39	4.27	4.36	4.42	4.14	4.19	4.08	3.84	3.85	3.84	4.08	3.84	3.85	3.59
Cr2O3	0.01	0.03	0	0.04	0.01	0.02	0	0.02	0.01	0.03	0.02	0.02	0.01	0.02	0.01	0.02	0.01	0.02	0.02
Fe2O3	1.67	2.2	2.37	1.92	2.11	1.84	3.34	2.48	2.22	2.62	2.37	1.78	2.08	2.32	2.08	1.78	2.08	2.32	1.73
FeO	9.19	9.13	8.9	9.58	11.05	9.57	9.67	10.16	10.1	9.85	9.5	12.32	10.01	9.6	10.01	12.32	10.01	9.6	9.66
MnO	0.3	0.3	0.33	0.32	0.35	0.33	0.37	0.35	0.34	0.35	0.35	0.37	0.34	0.31	0.34	0.37	0.34	0.31	0.29
MgO	11.85	10.88	11.15	11.24	11.2	11.29	11.16	11.08	10.97	11.09	11.15	11.42	11.66	11.39	11.66	11.42	11.66	11.39	11.36
CaO	21.12	20.71	21.18	20.81	19.57	21	19.65	19.68	20.15	20.39	20.6	18.1	20.25	20.64	20.25	18.1	20.25	20.64	20.98
Na2O	0.86	1.12	1.01	0.87	0.98	0.91	0.99	1.09	1.03	0.89	0.98	0.88	0.88	0.92	0.88	0.88	0.88	0.92	0.88
K2O	0.01	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.01
Totals	99.87	99.34	99.99	98.98	100.36	99.57	99.5	99.91	99.99	99.81	99.98	99.52	100.28	100.1	100.28	99.52	100.28	100.1	99.78
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6
Si	1.93	1.903	1.905	1.919	1.909	1.924	1.883	1.897	1.898	1.893	1.899	1.906	1.908	1.906	1.908	1.906	1.908	1.906	1.92
Ti	0.007	0.011	0.009	0.008	0.009	0.008	0.01	0.01	0.01	0.01	0.01	0.009	0.009	0.009	0.009	0.009	0.009	0.009	0.008
Al	0.141	0.19	0.178	0.154	0.176	0.151	0.191	0.194	0.196	0.184	0.186	0.183	0.17	0.171	0.17	0.183	0.17	0.171	0.16
Cr	0	0.001	0	0.001	0	0	0	0	0	0.001	0.001	0.001	0	0.001	0	0.001	0	0.001	0.001
Fe3	0.047	0.063	0.067	0.055	0.06	0.052	0.095	0.07	0.063	0.074	0.067	0.051	0.059	0.066	0.059	0.051	0.059	0.066	0.049
Fe2	0.289	0.289	0.28	0.304	0.348	0.302	0.307	0.32	0.318	0.311	0.299	0.391	0.314	0.302	0.314	0.391	0.314	0.302	0.304
Mn	0.009	0.01	0.01	0.01	0.011	0.011	0.012	0.011	0.011	0.011	0.011	0.012	0.011	0.01	0.011	0.012	0.011	0.01	0.009
Mg	0.663	0.613	0.624	0.637	0.627	0.635	0.631	0.622</											

Análise química de granada (PAR03)

Sample	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03
Min	grt1-1	grt1-2	grt1-3	grt1-4	grt3-1	grt3-2	grt3-3	grt3-4	grt4-1	grt4-2	grt4-3	grt4-4	grt5_core	grt5_rim	grt6-1	grt6-1	grt6-1	grt6-1	grt6-1
SiO2	37.59	37.53	38.02	37.98	37.82	38.06	37.75	37.92	37.88	38.12	37.67	37.34	37.93	37.89	37.86	37.86	37.86	37.86	37.86
TiO2	0.04	0.05	0.08	0.06	0	0.05	0.06	0.06	0.07	0.05	0.05	0.02	0	0.03	0.05	0.05	0.05	0.05	0.05
Al2O3	21.71	21.4	21.42	21.46	21.66	21.65	21.72	21.87	21.55	21.5	21.44	21.62	21.69	21.66	21.95	21.95	21.95	21.95	21.95
Cr2O3	0	0.01	0.01	0	0.03	0.02	0	0.01	0.02	0.01	0	0.02	0	0.01	0.01	0.01	0.01	0.01	0.01
Fe2O3	2.15	2.01	1.82	1.74	1.33	1.9	1.62	1.07	1.59	1.59	1.78	2.17	1.33	0.84	1.61	1.61	1.61	1.61	1.61
FeO	25.9	25.86	26.44	26.43	26.39	26.32	26.33	26.25	26.52	26.84	26.32	25.68	26.25	26.59	26.62	26.62	26.62	26.62	26.62
MnO	2.3	2.32	2.45	2.38	2.49	2.58	2.41	2.59	2.35	2.33	2.37	2.39	2.51	2.48	2.42	2.42	2.42	2.42	2.42
MgO	4.6	4.61	4.53	4.45	4.26	4.49	4.34	4.26	4.43	4.5	4.38	4.35	4.27	4.28	4.26	4.26	4.26	4.26	4.26
CaO	6.68	6.63	6.67	6.79	6.81	6.73	6.78	6.97	6.68	6.56	6.68	6.88	6.99	6.71	6.75	6.75	6.75	6.75	6.75
Na2O	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
K2O	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Totals	100.97	100.43	101.44	101.3	100.78	101.8	101.01	100.99	101.08	101.49	100.7	100.47	100.96	100.49	101.53	101.53	101.53	101.53	101.53
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12
Si	2.935	2.947	2.959	2.959	2.96	2.951	2.949	2.959	2.957	2.965	2.953	2.933	2.962	2.972	2.944	2.944	2.944	2.944	2.944
Ti	0.002	0.003	0.005	0.004	0	0.003	0.004	0.003	0.004	0.003	0.003	0.001	0	0.002	0.003	0.003	0.003	0.003	0.003
Al	1.999	1.981	1.965	1.971	1.999	1.979	2	2.012	1.983	1.971	1.982	2.002	1.997	2.003	2.012	2.012	2.012	2.012	2.012
Cr	0	0.001	0.001	0	0.002	0.001	0	0	0.001	0	0	0.001	0	0.001	0.001	0.001	0.001	0.001	0.001
Fe3	0.126	0.119	0.106	0.102	0.078	0.111	0.095	0.063	0.093	0.093	0.105	0.128	0.078	0.05	0.094	0.094	0.094	0.094	0.094
Fe2	1.691	1.698	1.721	1.722	1.727	1.707	1.72	1.713	1.732	1.746	1.726	1.687	1.714	1.744	1.731	1.731	1.731	1.731	1.731
Mn	0.152	0.154	0.162	0.157	0.165	0.169	0.159	0.171	0.155	0.154	0.157	0.159	0.166	0.165	0.159	0.159	0.159	0.159	0.159
Mg	0.535	0.539	0.525	0.517	0.497	0.519	0.505	0.495	0.515	0.522	0.512	0.509	0.497	0.5	0.494	0.494	0.494	0.494	0.494
Ca	0.559	0.558	0.556	0.567	0.571	0.559	0.567	0.583	0.559	0.547	0.561	0.579	0.585	0.564	0.562	0.562	0.562	0.562	0.562
Na	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
K	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Sum	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8
XMg	0.240	0.241	0.234	0.231	0.223	0.233	0.227	0.224	0.229	0.230	0.229	0.232	0.225	0.223	0.222	0.222	0.222	0.222	0.222
alm	0.576	0.576	0.581	0.581	0.583	0.578	0.583	0.578	0.585	0.588	0.584	0.575	0.579	0.587	0.588	0.588	0.588	0.588	0.588
prp	0.182	0.183	0.177	0.174	0.168	0.176	0.171	0.167	0.174	0.176	0.173	0.173	0.168	0.168	0.168	0.168	0.168	0.168	0.168
grs	0.190	0.189	0.188	0.191	0.193	0.189	0.192	0.197	0.189	0.184	0.190	0.197	0.198	0.190	0.191	0.191	0.191	0.191	0.191
sps	0.052	0.052	0.055	0.053	0.056	0.057	0.054	0.058	0.052	0.052	0.053	0.054	0.056	0.055	0.054	0.054	0.054	0.054	0.054

Sample	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03
Min	grt6-2	grt6-3	grt6-4	grt6-5	grt7-1	grt7-2	grt7-3	grt7-4	grt7-5	grt8-1	grt8-2	grt8-3	grt8-4	grt8-5	grt8-6	grt8-7	grt8-8
SiO2	38.02	38.05	37.85	37.79	37.99	37.93	37.88	37.82	37.65	38.28	37.91	37.7	37.84	37.78			
TiO2	0.05	0.08	0.08	0.14	0.07	0.03	0.04	0.06	0.07	0.03	0.05	0.07	0.06	0.04			
Al2O3	21.84	21.92	21.96	21.82	21.92	21.66	22	21.77	21.65	21.85	21.35	21.59	21.42	21.61			
Cr2O3	0.01	0	0	0	0	0	0	0.02	0	0	0	0	0	0.01			
Fe2O3	1.33	1.27	2.15	1.72	1.68	1.59	1.24	1.34	1.38	1.01	1.87	2.51	1.96	2.04			
FeO	26.89	26.64	26.04	26.27	26.46	26.11	26.29	26.26	25.88	26.78	26.01	25.74	25.83	25.7			
MnO	2.27	2.42	2.46	2.51	2.4	2.41	2.45	2.39	2.46	2.57	2.5	2.43	2.4	2.4			
MgO	4.43	4.54	4.52	4.27	4.52	4.66	4.65	4.63	4.56	4.37	4.71	4.67	4.74	4.68			
CaO	6.57	6.55	6.82	6.94	6.67	6.66	6.46	6.51	6.7	6.73	6.59	6.73	6.71	6.82			
Na2O	0	0	0	0	0	0	0	0	0	0	0	0	0	0			
K2O	0	0	0	0	0	0	0	0	0	0	0	0	0	0			
Totals	101.41	101.47	101.88	101.46	101.71	101.05	101	100.81	100.35	101.62	100.99	101.43	100.96	101.09			
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12			
Si	2.957	2.955	2.931	2.94	2.945	2.956	2.951	2.954	2.954	2.97	2.96	2.933	2.953	2.945			
Ti	0.003	0.005	0.004	0.008	0.004	0.002	0.002	0.003	0.004	0.002	0.003	0.004	0.004	0.002			
Al	2.002	2.007	2.005	2.002	2.003	1.99	2.021	2.005	2.003	1.998	1.965	1.98	1.971	1.986			
Cr	0.001	0	0	0	0	0	0	0.001	0	0	0	0	0	0.001			
Fe3	0.078	0.074	0.126	0.101	0.098	0.093	0.073	0.079	0.081	0.059	0.11	0.147	0.115	0.12			
Fe2	1.749	1.73	1.686	1.71	1.715	1.702	1.713	1.716	1.698	1.738	1.698	1.674	1.686	1.675			
Mn	0.15	0.159	0.161	0.165	0.158	0.159	0.162	0.158	0.163	0.169	0.165	0.16	0.159	0.158			
Mg	0.513	0.525	0.522	0.495	0.522	0.541	0.54	0.539	0.533	0.505	0.548	0.541	0.551	0.544			
Ca	0.547	0.545	0.566	0.579	0.554	0.556	0.539	0.545	0.563	0.559	0.551	0.561	0.561	0.57			
Na	0	0	0	0	0	0	0	0	0	0	0	0	0	0			
K	0	0	0	0	0	0	0	0	0	0	0	0	0	0			
Sum	8	8	8	8	8	8	8	8	8	8	8	8	8	8			
XMg	0.227	0.233	0.236	0.224	0.233	0.241	0.240	0.239	0.239	0.225	0.244	0.244	0.246	0.245			
alm	0.591	0.585															

Análise química de plagioclásio (PAR03)

Sample	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3	PARO3
Min	plg.1.1	plg.1.2	plg.1.3	plg.1.4	plg.1.5	plg.2.1	plg.2.2	plg.2.3	plg.2.4	plg.3.1	plg.3.2	plg.3.3	plg.4.1	plg.4.2	plg.4.3	plg.4.4	plg.5.1	plg.5.2	plg.5.3	plg.5.4	plg.5.5	plg.6.1	plg.6.2	plg.6.3	plg.6.4	
SiO2	57.9	57.27	57.8	57.65	57.82	57.76	50.13	55.39	58.38	57.2	55.7	57.85	57.71	57.17	57.33	58.08	58	56.61	58.23	51.22	58.18	57.78	55.75	55.05	58.23	
TiO2	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	
Al2O3	27.54	27.85	27.77	27.59	27.44	27.58	32.55	29.17	27.23	27.69	28.98	27.61	27	27.91	27.82	27.19	27.72	28.45	27.17	32.27	27.18	27.6	29.03	29.31	27.38	
Cr2O3	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	
Fe2O3	0.14	0.23	0.1	0.1	0.08	0.1	0.12	0.12	0.07	0.39	0.11	0.11	0.88	0.09	0.12	0.25	0.2	0.16	0.09	0.14	0.2	0.13	0.11	0.4	0.13	
FeO	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	
MnO	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	
MgO	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	
CaO	8.7	9.36	9.04	8.77	8.66	8.82	14.74	10.74	8.6	9.13	10.39	8.74	8.07	9.17	9.08	8.59	8.9	9.72	8.62	14.07	8.62	8.85	10.44	10.68	8.75	
Na2O	6.47	6.04	6.18	6.51	6.41	6.43	3.09	5.46	6.64	6	5.51	6.4	6.47	6.05	6.13	6.44	6.13	5.92	6.71	3.39	6.58	6.43	5.51	5.21	6.52	
K2O	0.19	0.19	0.19	0.2	0.19	0.17	0.07	0.17	0.19	0.19	0.16	0.24	0.29	0.19	0.2	0.26	0.21	0.2	0.18	0.1	0.23	0.21	0.16	0.15	0.15	
Totals	100.9	100.9	101.1	100.8	100.6	100.9	100.7	101.1	101.1	100.6	100.8	101	100.4	100.6	100.7	100.8	101.2	101.1	101	101.2	101	101	101	100.8	101.2	
Oxygens	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	
Si	2.568	2.545	2.56	2.562	2.572	2.564	2.269	2.47	2.584	2.549	2.485	2.566	2.575	2.546	2.551	2.579	2.565	2.516	2.581	2.301	2.58	2.563	2.483	2.461	2.576	
Ti	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	
Al	1.44	1.459	1.45	1.445	1.439	1.443	1.736	1.533	1.421	1.455	1.524	1.444	1.421	1.465	1.459	1.424	1.446	1.491	1.42	1.709	1.421	1.443	1.525	1.545	1.428	
Cr	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	
Fe3	0.005	0.008	0.003	0.003	0.003	0.003	0.004	0.004	0.002	0.013	0.004	0.004	0.029	0.003	0.004	0.008	0.007	0.005	0.003	0.005	0.007	0.004	0.004	0.014	0.004	
Fe2	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	
Mn	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	
Mg	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	
Ca	0.413	0.446	0.429	0.417	0.413	0.42	0.714	0.513	0.408	0.436	0.497	0.415	0.386	0.438	0.433	0.409	0.422	0.463	0.409	0.677	0.41	0.421	0.498	0.511	0.415	
Na	0.556	0.52	0.531	0.561	0.553	0.553	0.272	0.472	0.57	0.519	0.477	0.55	0.56	0.522	0.529	0.554	0.525	0.51	0.577	0.295	0.566	0.553	0.476	0.452	0.559	
K	0.011	0.011	0.011	0.011	0.011	0.01	0.004	0.01	0.011	0.011	0.009	0.014	0.016	0.011	0.012	0.015	0.012	0.011	0.01	0.006	0.013	0.012	0.009	0.008	0.008	
Sum	4.993	4.988	4.984	5	4.99	4.994	4.999	5.002	4.995	4.982	4.994	4.993	4.988	4.986	4.988	4.989	4.977	4.997	5.001	4.993	4.996	4.996	4.995	4.99	4.991	
an	42.14	45.65	44.18	42.16	42.27	42.73	72.12	51.56	41.25	45.13	50.56	42.39	40.12	45.11	44.46	41.82	44.00	47.05	41.06	69.22	41.46	42.70	50.66	52.63	42.26	
ab	56.73	53.22	54.69	56.72	56.60	56.26	27.47	47.44	57.63	53.73	48.52	56.18	58.21	53.76	54.31	56.65	54.74	51.83	57.93	30.16	57.23	56.09	48.42	46.55	56.92	
or	1.12	1.13	1.13	1.11	1.13	1.02	0.40	1.01	1.11	1.14	0.92	1.43	1.66	1.13	1.23	1.53	1.25	1.12	1.00	0.61	1.31	1.22	0.92	0.82	0.81	

Análise química de ortopiroxênio (PAR03)

Sample	PAR03 opx1.1	PAR03 opx1.2	PAR03 opx1.3	PAR03 opx1.4	PAR03 opx3.1	PAR03 opx3.2	PAR03 opx3.3	PAR03 opx6.1	PAR03 opx6.2	PAR03 opx6.3	PAR03 opx6.4	PAR03 opx8.1	PAR03 opx8.2	PAR03 opx8.3
Min														
SiO2	50.76	50.83	50.53	50.69	50.62	50.3	50.92	50.3	50.23	50.27	50.44	50.77	50.94	50.33
TiO2	0.07	0.1	0.09	0.05	0.05	0.1	0.16	0.08	0.09	0.08	0.05	0.08	0.04	0.07
Al2O3	1.61	1.37	1.51	1.5	1.62	1.65	1.52	1.81	1.85	1.73	1.38	1.41	1.23	1.63
Cr2O3	0	0.01	0	0.01	0.01	0.01	0.02	0	0.01	0.01	0.01	0.02	0	0
Fe2O3	0	0	0	0	0.35	0.71	0	0	0.28	0.39	0	0	0	0
FeO	28.4	28.81	28.4	28.8	28.03	28.33	27.97	29.32	29.39	29.11	29.02	28.39	28.44	28.64
MnO	0.91	0.93	0.97	0.97	0.97	0.97	0.97	0.98	0.99	0.99	1.02	0.96	0.95	0.99
MgO	17	16.73	16.31	16.69	17.2	16.67	17.36	16.12	16.03	16.21	16.32	16.92	17.22	16.66
CaO	0.64	0.64	0.8	0.55	0.62	0.84	0.57	0.68	0.85	0.84	0.63	0.6	0.51	0.56
Na2O	0.02	0.04	0.02	0.03	0.02	0.03	0.04	0.01	0.02	0.03	0.02	0	0	0.03
K2O	0	0	0	0	0	0	0	0.01	0	0	0.01	0.01	0	0
Totals	99.4	99.45	98.62	99.27	99.48	99.63	99.53	99.3	99.74	99.65	98.89	99.15	99.33	98.91
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6
Si	1.965	1.971	1.974	1.969	1.957	1.95	1.964	1.959	1.951	1.953	1.971	1.97	1.973	1.962
Ti	0.002	0.003	0.003	0.001	0.001	0.003	0.005	0.002	0.003	0.002	0.001	0.002	0.001	0.002
Al	0.073	0.063	0.07	0.069	0.074	0.075	0.069	0.083	0.085	0.079	0.064	0.065	0.056	0.075
Cr	0	0	0	0	0	0	0.001	0	0	0	0	0.001	0	0
Fe3	0	0	0	0	0.01	0.021	0	0	0.008	0.012	0	0	0	0
Fe2	0.919	0.934	0.928	0.936	0.907	0.918	0.902	0.955	0.955	0.946	0.948	0.921	0.921	0.934
Mn	0.03	0.03	0.032	0.032	0.032	0.032	0.032	0.032	0.033	0.033	0.034	0.032	0.031	0.033
Mg	0.98	0.966	0.949	0.966	0.991	0.963	0.998	0.936	0.928	0.938	0.95	0.979	0.994	0.968
Ca	0.027	0.026	0.033	0.023	0.026	0.035	0.024	0.028	0.035	0.035	0.026	0.025	0.021	0.023
Na	0.001	0.003	0.001	0.002	0.002	0.002	0.003	0.001	0.002	0.002	0.002	0	0	0.002
K	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Sum	3.997	3.997	3.99	3.997	4	4	3.998	3.998	4	4	3.997	3.995	3.998	3.999
X(Mg)	0.516	0.508	0.506	0.508	0.522	0.512	0.525	0.495	0.493	0.498	0.501	0.515	0.519	0.509
xAl	0.038	0.034	0.044	0.038	0.031	0.025	0.033	0.042	0.036	0.032	0.035	0.035	0.029	0.037

Análise química de clinopiroxênio (PAR03)

Sample	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03	PAR03
Min	cpx2.1	cpx2.2	cpx2.3	cpx2.4	cpx2.5	cpx4.1	cpx4.2	cpx4.3	cpx4.4	cpx5.1	cpx5.2	cpx5.3	cpx7.1	cpx7.2	cpx7.3				
SiO2	51.11	50.38	50.22	50.32	50.08	50.91	50.97	50.8	50.94	50.56	50.54	50.54	50.87	50.54	50.92				
TiO2	0.25	0.35	0.35	0.31	0.31	0.25	0.27	0.27	0.26	0.32	0.35	0.27	0.27	0.34	0.27				
Al2O3	2.69	3.61	3.47	3.23	3.42	2.84	2.76	2.93	2.63	3.34	3.48	2.97	3.05	3.56	3.03				
Cr2O3	0	0	0.01	0	0.02	0	0.01	0.01	0	0	0	0	0.01	0.02	0				
Fe2O3	1.07	1.65	2.55	2.21	2.66	1.89	1.63	1.24	1.54	1.77	1.7	2.49	1.71	1.88	0.95				
FeO	11.32	11.47	11.65	11.55	10.58	11.28	11.11	11.98	11.7	11.09	10.85	10.26	11.43	10.95	11.37				
MnO	0.41	0.45	0.47	0.48	0.43	0.43	0.43	0.43	0.44	0.47	0.46	0.46	0.44	0.45	0.41				
MgO	11.1	10.46	10.78	10.94	10.76	11.12	11.14	10.78	11.12	11	10.83	11.4	11.08	10.76	11.05				
CaO	21.12	20.89	20.09	20.21	21.15	20.7	20.88	20.51	20.58	20.66	20.88	20.75	20.41	20.9	20.98				
Na2O	0.6	0.7	0.7	0.65	0.62	0.65	0.65	0.67	0.61	0.67	0.73	0.65	0.71	0.73	0.6				
K2O	0	0	0.01	0.01	0	0.01	0	0	0	0	0	0.01	0	0	0				
Totals	99.66	99.95	100.29	99.9	100.03	100.08	99.85	99.6	99.8	99.88	99.81	99.8	99.99	100.12	99.59				
Oxygens	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6				
Si	1.939	1.911	1.902	1.911	1.899	1.927	1.931	1.933	1.934	1.916	1.915	1.915	1.926	1.911	1.933				
Ti	0.007	0.01	0.01	0.009	0.009	0.007	0.008	0.008	0.007	0.009	0.01	0.008	0.008	0.01	0.008				
Al	0.121	0.161	0.155	0.144	0.153	0.126	0.123	0.131	0.118	0.149	0.155	0.132	0.136	0.159	0.136				
Cr	0	0	0	0	0.001	0	0	0	0	0	0	0	0	0.001	0				
Fe3	0.03	0.047	0.073	0.063	0.076	0.054	0.046	0.036	0.044	0.05	0.049	0.071	0.049	0.053	0.027				
Fe2	0.359	0.364	0.369	0.367	0.336	0.357	0.352	0.381	0.371	0.351	0.344	0.325	0.362	0.346	0.361				
Mn	0.013	0.014	0.015	0.015	0.014	0.014	0.014	0.014	0.014	0.015	0.015	0.015	0.014	0.014	0.013				
Mg	0.628	0.591	0.609	0.619	0.608	0.627	0.629	0.611	0.629	0.621	0.612	0.644	0.625	0.606	0.625				
Ca	0.859	0.849	0.815	0.823	0.859	0.839	0.848	0.836	0.837	0.839	0.848	0.842	0.828	0.847	0.853				
Na	0.044	0.051	0.052	0.048	0.046	0.048	0.048	0.049	0.045	0.049	0.053	0.048	0.052	0.053	0.044				
K	0	0	0	0	0	0.001	0	0	0	0	0	0	0	0	0				
Sum	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4				
X(Mg)	0.636	0.619	0.623	0.628	0.644	0.637	0.641	0.616	0.629	0.639	0.640	0.665	0.633	0.637	0.634				

Sample	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25
Min	grt.1.1	grt.1.2	grt.1.3	grt.1.4	grt.1.5	grt.1.6	grt.1.7	grt.1.8	grt.1.9	grt.1.10	grt.2.1	grt.2.2	grt.2.3	grt.2.4	grt.2.5	grt.2.6	grt.2.7	grt.2.8	grt.2.9	grt.2.10
SiO ₂	38.48	38.32	38.34	38.31	38.31	38.46	38.23	38.22	38.26	38.43	38.28	38.27	38.1	38.43	38.18					
TiO ₂	0.03	0.02	0	0.01	0.03	0.02	0	0	0.02	0.02	0.03	0.02	0.01	0.01	0					
Al ₂ O ₃	21.85	22.21	22.17	21.72	22.01	21.91	22.27	21.84	21.79	21.9	21.89	21.84	22.04	21.87	22.08					
Cr ₂ O ₃	0.05	0.03	0.01	0.02	0.02	0.02	0.01	0.01	0.03	0.05	0.06	0.05	0.04	0.03	0.02					
Fe ₂ O ₃	0.85	0.64	0.99	0.87	1.29	0.86	1.16	1.52	1.05	0.57	1.22	1.47	1.41	0.99	0.88					
FeO	27.51	27.42	27.31	27.56	27.28	27.38	27.36	27.03	27.4	27.65	27.14	27.2	27.02	27.43	27.37					
MnO	0.75	0.73	0.75	0.73	0.75	0.76	0.72	0.72	0.76	0.77	0.74	0.78	0.74	0.74	0.74					
MgO	8.92	8.93	9.04	8.94	9.12	9.12	8.97	9.08	8.87	8.79	9.01	8.88	8.97	8.95	8.95					
CaO	1.42	1.36	1.27	1.2	1.21	1.21	1.24	1.32	1.36	1.38	1.41	1.49	1.4	1.31	1.19					
Na ₂ O	0.01	0	0.01	0.01	0	0.01	0.01	0	0.01	0.01	0.01	0.01	0	0.03	0.01					
K ₂ O	0	0	0	0	0	0.01	0	0.02	0	0.01	0	0	0	0.01	0					
Totals	99.87	99.68	99.88	99.37	100.02	99.76	99.97	99.77	99.56	99.58	99.8	100.01	99.72	99.79	99.43					
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12					
Si	2.977	2.966	2.962	2.979	2.958	2.975	2.953	2.96	2.971	2.981	2.963	2.96	2.951	2.975	2.964					
Ti	0.002	0.001	0	0.001	0.002	0.001	0	0	0.001	0.001	0.002	0.001	0.001	0	0					
Al	1.993	2.027	2.019	1.991	2.004	1.998	2.028	1.993	1.995	2.003	1.997	1.991	2.012	1.996	2.021					
Cr	0.003	0.002	0.001	0.001	0.001	0.001	0.001	0.001	0.002	0.003	0.004	0.003	0.002	0.002	0.001					
Fe ₃	0.05	0.037	0.058	0.051	0.075	0.05	0.067	0.089	0.062	0.033	0.071	0.085	0.082	0.057	0.051					
Fe ₂	1.779	1.775	1.764	1.792	1.762	1.771	1.767	1.751	1.779	1.794	1.757	1.76	1.751	1.775	1.777					
Mn	0.049	0.048	0.049	0.048	0.049	0.05	0.047	0.047	0.05	0.051	0.049	0.051	0.048	0.048	0.049					
Mg	1.028	1.03	1.041	1.036	1.049	1.052	1.033	1.048	1.027	1.017	1.04	1.024	1.036	1.033	1.035					
Ca	0.118	0.113	0.105	0.1	0.1	0.1	0.103	0.109												

Continuação - Análise química de granada (RGM25)

Sample	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25
Min	grt.2.6	grt.2.7	grt.2.8	grt.2.9	grt.2.10	grt.3.1	grt.3.2	grt.3.3	grt.3.4	grt.3.5	grt.3.6	grt.3.7	grt.3.8	grt.3.9	grt.3.10	grt.3.11	grt.3.12	grt.3.13	grt.3.14	grt.3.15
SiO2	38.33	38.2	38.43	37.93	38.15	38.26	38.03	38.41	38.04	38.07	38.13	38.51	38.28	38	38.06					
TiO2	0.01	0	0.01	0.02	0.04	0.02	0.03	0	0.01	0	0	0	0	0	0.04					
Al2O3	22.11	22.07	21.95	21.92	21.72	21.99	21.77	21.97	21.96	22.12	22.11	21.84	22	21.81	21.76					
Cr2O3	0.01	0.02	0.03	0.03	0.07	0.08	0.04	0.03	0.02	0.04	0.02	0.04	0.02	0.03	0.05					
Fe2O3	1.1	1.04	0.64	1.55	1.57	1.29	1.09	0.88	1.14	1.25	1.84	0.52	0.71	1.04	1.03					
FeO	27.23	27.33	27.47	27.05	27.04	28.25	27.16	27.56	27.24	27.09	26.76	27.64	27.32	27.02	26.98					
MnO	0.71	0.74	0.76	0.76	0.74	0.88	0.76	0.77	0.81	0.76	0.79	0.76	0.77	0.81	0.79					
MgO	9.15	8.99	9.03	8.88	9.01	8.32	8.92	8.97	8.96	9.07	9.29	9.12	9.08	8.93	8.88					
CaO	1.23	1.15	1.22	1.29	1.41	1.35	1.29	1.16	1.11	1.11	1.1	1.06	1.16	1.27	1.39					
Na2O	0	0.01	0.01	0.01	0	0.02	0.01	0.02	0.01	0.01	0.01	0	0	0.02	0.03					
K2O	0	0.01	0	0	0	0	0	0.01	0	0.01	0.01	0.01	0	0	0.01					
Totals	99.87	99.56	99.56	99.46	99.76	100.45	99.09	99.78	99.3	99.52	100.06	99.51	99.34	98.95	99.02					
Oxygens	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12					
Si	2.961	2.962	2.978	2.949	2.957	2.958	2.965	2.973	2.959	2.952	2.941	2.986	2.972	2.965	2.968					
Ti	0	0	0	0.001	0.003	0.001	0.002	0	0	0	0	0	0	0.001	0.002					
Al	2.013	2.017	2.006	2.009	1.985	2.004	2.001	2.005	2.014	2.022	2.011	1.996	2.014	2.007	2.001					
Cr	0.001	0.001	0.002	0.002	0.004	0.005	0.002	0.002	0.001	0.002	0.001	0.003	0.001	0.002	0.003					
Fe3	0.064	0.061	0.037	0.09	0.092	0.075	0.064	0.052	0.067	0.073	0.107	0.031	0.041	0.061	0.06					
Fe2	1.759	1.772	1.781	1.759	1.753	1.827	1.771	1.784	1.772	1.757	1.727	1.792	1.774	1.764	1.76					
Mn	0.046	0.048	0.05	0.05	0.049	0.057	0.05	0.051	0.053	0.05	0.052	0.05	0.051	0.054	0.052					
Mg	1.053	1.039	1.043	1.029	1.041	0.958	1.036	1.035	1.038	1.049	1.068	1.054	1.051	1.038	1.033					
Ca	0.102	0.096	0.101	0.108	0.117	0.112	0.108	0.097	0.093	0.093	0.091	0.088	0.096	0.107	0.116					
Na	0	0.002	0.002	0.002	0	0.002	0.001	0.002	0.001	0.001	0.001	0	0	0.002	0.004					
K	0	0.001	0	0	0	0	0	0.001	0	0.001	0.001	0.001	0	0	0.001					
Sum	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8					
X(Mg)	0.374	0.370	0.369	0.369	0.373	0.344	0.369	0.367	0.369	0.374	0.382	0.370	0.372	0.370	0.370					
alm	0.594	0.600	0.599	0.597	0.592	0.618	0.597	0.601	0.599	0.596	0.588	0.601	0.597	0.595	0.594					
prp	0.356	0.352	0.351	0.349	0.352	0.324	0.349	0.349	0.351	0.356	0.364	0.353	0.354	0.350	0.349					
grs	0.034	0.032	0.034	0.037	0.040	0.038	0.036	0.033	0.031	0.032	0.031	0.029	0.032	0.036	0.039					
spss	0.016	0.016	0.017	0.017	0.017	0.019	0.017	0.017	0.018	0.017	0.018	0.017	0.017	0.018	0.018					

Análise química de plagioclásio (RGM25)

Sample	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25	RGM25
Min	pIlg.1.1	pIlg.1.2	pIlg.1.3	pIlg.3.1	pIlg.3.2	pIlg.3.3	pIlg.3.4	pIlg.5.1	pIlg.5.2	pIlg.5.3	pIlg.8.1	pIlg.8.2	pIlg.8.3	pIlg.9.1	pIlg.9.2	pIlg.9.3	pIlg.9.4	pIlg.9.4	pIlg.9.4	pIlg.9.4
SiO2	60.59	59.79	60.24	60.3	60.23	60.44	60.07	60.75	59.87	60.16	60.01	60.02	60.42	60.3	59.82	59.76	59.7	59.7	59.7	59.7
TiO2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Al2O3	25.31	25.4	25.3	25.34	25.36	25.23	25.47	25.36	25.35	25.39	25.22	25.45	25.46	25.43	25.23	25.45	25.33	25.33	25.33	25.33
Cr2O3	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Fe2O3	0.03	0.02	0.02	0.02	0.01	0.03	0	0.13	0.05	0.01	0.02	0.02	0.03	0.07	0.05	0.02	0.05	0.05	0.05	0.05
FeO	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
MnO	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
MgO	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
CaO	6	6.11	6.1	6.14	6.18	6.17	6.16	6.07	6.18	6.11	6.05	6.18	6.05	6.13	6.14	6.19	6.24	6.24	6.24	6.24
Na2O	8.26	7.91	8.08	7.91	8.05	8.19	8.1	8.08	7.97	7.9	7.99	7.97	7.93	7.99	7.85	7.81	7.92	7.92	7.92	7.92
K2O	0.11	0.16	0.14	0.13	0.22	0.12	0.08	0.11	0.21	0.17	0.14	0.15	0.17	0.09	0.15	0.17	0.11	0.11	0.11	0.11
Totals	100.29	99.4	99.88	99.84	100.05	100.17	99.88	100.49	99.62	99.74	99.44	99.78	100.06	100.01	99.24	99.4	99.36	99.36	99.36	99.36
Oxygens	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8	8
Si	2.686	2.674	2.682	2.683	2.679	2.684	2.675	2.687	2.674	2.681	2.682	2.675	2.683	2.68	2.679	2.673	2.673	2.673	2.673	2.673
Ti	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Al	1.323	1.34	1.328	1.33	1.33	1.321	1.337	1.322	1.335	1.334	1.329	1.337	1.333	1.332	1.332	1.342	1.337	1.337	1.337	1.337
Cr	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Fe3	0.001	0.001	0.001	0.001	0	0.001	0	0.004	0.002	0	0.001	0.001	0.001	0.002	0.002	0.001	0.002	0.002	0.002	0.002
Fe2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Mn	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Mg	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Ca	0.285	0.293	0.291	0.293	0.295	0.294	0.294	0.288	0.296	0.292	0.29	0.295	0.288	0.292	0.295	0.297	0.299	0.299	0.299	0.299
Na	0.71	0.686	0.697	0.683	0.694	0.705	0.699	0.693	0.69	0.683	0.693	0.689	0.683	0.689	0.682	0.677	0.688	0.688	0.688	0.688
K	0.006	0.009	0.008	0.007	0.012	0.007	0.004	0.006	0.012	0.009	0.008	0.008	0.009	0.005	0.008	0.01	0.006	0.006	0.006	0.006
Sum	5.01	5.003	5.006	4.997	5.01	5.011	5.009	5	5.008	4.999	5.003	5.005	4.997	5	4.999	4.999	5.005	5.005	5.005	5.005
an	28.5	29.7	29.2	29.8	29.5	29.2	29.5	29.2	29.7	29.7	29.3	29.7	29.4	29.6	29.9	30.2	30.1	30.1	30.1	30.1
ab	70.9	69.4	70.0	69.5	69.3	70.1	70.1	70.2	69.1	69.4	69.9	69.5	69.7	69.9	69.2	68.8	69.3	69.3	69.3	69.3
or	0.6	0.9	0.8	0.7	1.2	0.7	0.4	0.6	1.2	0.9	0.8	0.8	0.9	0.5	0.8	1.0	0.6	0.6	0.6	0.6

ANEXO V – ANÁLISES DE ELEMENTOS TRAÇO EM GRANADA VIA LA-ICP- MS

Dados de química mineral em granada (LA-ICP-MS)

Sample	Spot/Domain	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Er	Yb	Lu	(Gd/Yb) _N
VAB01	grt1-centro1	0.065	1.191	0.616	8.21	10.03	2.36	16.81	2.87	13.22	2.94	1.63	0.198	8.51
VAB01	grt1-borda1	<0.081	1.23	0.633	7.76	10.2	2.35	18.78	4.6	30.44	17.07	20.53	2.62	0.75
VAB01	grt1-borda2	0.062	1.187	0.581	7.81	9	2.13	18.18	3.98	25.37	11.35	9.12	1.078	1.65
VAB01	grt1-centro2	0.082	1.318	0.529	6.09	8	2.18	13.94	2.59	12.66	3.26	1.99	0.211	5.78
VAB01	grt2-borda1	0.0447	1.041	0.525	7.39	8.62	1.95	16.13	3.59	24.1	13.43	13.59	1.786	0.98
VAB01	grt2-borda2	0.066	1.205	0.513	7.02	8.46	1.95	15.78	3.23	22.42	12.47	12.09	1.718	1.08
VAB01	grt2-centro1	0.056	1.154	0.648	9.17	10.83	2.38	18.84	4.17	28.65	16.06	19.55	2.62	0.80
VAB01	grt2-centro2	0.065	1.059	0.617	8.58	10.45	2.22	16.79	3.43	22.79	13.02	13.73	1.86	1.01
VAB01	grt3-borda1	0.0349	0.987	0.5	9.78	9.4	2	15.62	3.33	21.73	13.24	14.99	2.09	0.86
VAB01	grt3-borda2	0.051	1.08	0.565	7.93	9.46	2.21	17.18	3.67	25.76	16.45	17.93	2.57	0.79
VAB01	grt3-centro1	0.0457	0.961	0.539	8.28	9.87	2.21	18.79	4.09	30.22	20.91	24	3.29	0.65
VAB01	grt3-centro2	0.048	1.003	0.548	7.87	9.58	2.11	16.68	4.08	28.38	18.41	21.47	3.19	0.64
VAB01	grt1-borda3	14.14	122.91	<-0.00	73.49	564.1	130.43	476.77	186.38	1692.35	284.05	<-0.00	14.29	-
VAB01	grt1-centro3	0.088	1.34	0.513	6.19	8.72	2.24	14.38	2.25	10.36	2.39	1.61	0.204	7.37
VAB19	grt2-borda1	<0.0152	0.546	0.309	5.1	8.72	2.3	27.46	9.15	86.48	79.57	96.69	13.79	0.23
VAB19	grt2-centro1	0.056	0.392	0.26	4.28	8.3	2.14	25.14	7.64	69.39	61.57	80.44	11.59	0.26
VAB19	grt2-borda2	<0.0145	0.309	0.232	4.06	6.42	2.08	23.91	8.15	74.17	68.02	86.14	12.54	0.23
VAB19	grt5-borda1	0.085	0.613	0.271	4.44	8.4	2.24	27.33	8.64	79.82	71.94	94.12	13.84	0.24
VAB19	grt5-centro1	<0.0164	0.284	0.245	4.62	9.72	2.11	29.94	9.68	89.35	77.89	96.88	14.55	0.26
VAB19	grt5-centro2	0.0164	0.37	0.268	4.98	9.17	2.17	32.02	10.1	88.47	82.03	101.35	15.03	0.26
VAB19	grt5-centro3	<0.0160	0.287	0.233	4.21	8.69	2.09	29.3	9.85	91.18	81.59	102.47	14.84	0.24
VAB19	grt5-borda2	0.0303	0.437	0.287	4.7	8.4	2.17	28.23	8.97	86.92	80.79	99.8	15.24	0.23
VAB19	grt5-borda3	0.087	0.623	0.333	4.56	7.89	2.15	26.59	8.61	79.66	67.85	80.24	11.54	0.27
VAB19	grt6-centro1	0.015	0.483	0.292	5.22	8.95	2.21	28.73	9.44	85.03	78.6	107.15	15.3	0.22
VAB19	grt6-centro2	0.0309	0.413	0.281	4.62	8.9	2.25	30.04	9.75	94.26	87.69	112.71	16.73	0.22
VAB19	grt6-borda1	<0.0203	0.447	0.292	5.08	9.37	2.17	31.01	9.87	95.06	90.47	117.78	17.13	0.22
VAB19	grt6-borda2	0.082	0.654	0.36	5.63	9.32	2.1	30.1	9.92	94.87	84.09	104.58	15.23	0.24
VAB19	grt7-centro1	<0.0162	0.374	0.219	4.27	7.98	2.18	26.94	8.02	71.29	50.18	55.25	7.2	0.40
VAB19	grt7-centro2	0.0456	0.37	0.217	4.2	7.71	2.06	25.39	8.31	77	61.45	68.36	9.38	0.31
VAB19	grt7-borda1	0.0331	0.599	0.306	4.63	7.9	2.16	27.42	8.74	82.66	73.36	81.3	11.14	0.28
VAB19	grt7-borda2	0.053	0.518	0.317	4.36	7.76	2.34	25.13	8.23	75.69	58.21	58.3	7.7	0.36

ANEXO VI – ANÁLISES QUÍMICAS DE ROCHA TOTAL

Química total de rocha. (*) FRX. (**) ICP-AES/MS

Sample	PAR-32A	PAR-32C	PAR-33	PAR-36A	PAR-36C	PAR-40	RGM02	RGM04	RGM53	RGM57	VAB-1	VAB19	RGM55	PAR03
Rock	cham	cham	cham	cham	cham	cham	cham	cham	cham	cham	Grt-gran	Grt-gran	diatex	gran maf
SiO2*	60.91	60.79	61.63	61.27	60.33	62.75	-	-	-	-	63.49	-	-	46.19
TiO2*	1.173	1.147	0.802	1.14	1.182	0.86	-	-	-	-	0.735	-	-	2.907
Al2O3*	15.88	16.15	17.52	15.97	16.01	16.55	-	-	-	-	15.89	-	-	12.62
Fe2O3*	6.71	6.6	4.71	6.42	6.76	5.58	-	-	-	-	6.61	-	-	19.51
MnO*	0.095	0.102	0.089	0.088	0.088	0.083	-	-	-	-	0.116	-	-	0.266
MgO*	1.13	1.06	0.64	1.07	1.12	0.65	-	-	-	-	2.4	-	-	5.1
CaO*	3.3	3.15	2.38	3.14	3.36	2.61	-	-	-	-	5.05	-	-	10.12
Na2O*	4.26	4.4	4.3	4.1	3.98	4.63	-	-	-	-	2.96	-	-	3.03
K2O*	5.27	5.34	6.66	5.36	5.64	5.66	-	-	-	-	2.04	-	-	0.14
P2O5*	0.347	0.343	0.171	0.326	0.351	0.216	-	-	-	-	0.132	-	-	0.267
LOI	0.08	0.08	0.16	0.12	0.14	0.02	-	-	-	-	0.02	-	-	0
Total*	99.16	99.16	99.06	99.00	98.96	99.61	-	-	-	-	99.44	-	-	100.15
Rb**	66.43	68.47	95.46	78.65	82.83	67.72	91.60	109.00	76.90	133.00	36.25	44.10	97.50	-
Cs**	0.10	0.11	0.19	0.32	0.08	0.02	0.10	0.20	0.07	0.19	0.07	0.01	0.23	-
Ba**	2055.1	2102.5	2172.0	2013.9	1906.5	2695.0	1406.0	861.0	1823.0	2134.0	522.4	678.0	546.0	-
Sr**	327.87	337.03	275.22	320.81	329.34	320.53	191.00	103.00	302.00	409.00	282.49	350.00	49.00	-
Nb**	41.76	41.43	30.45	46.74	38.95	37.47	55.80	19.40	69.20	3.23	10.54	11.20	15.50	-
Hf**	17.58	18.53	18.33	19.27	17.54	13.32	24.80	14.10	15.10	5.57	2.57	3.12	9.67	-
Zr**	858.3	927.4	916.6	948.7	868.0	682.2	1211.0	697.0	786.0	257.0	87.3	120.0	478.0	-
Y**	27.98	28.42	19.35	27.91	30.29	23.60	44.60	10.40	42.20	8.58	15.14	7.71	16.30	-
Th**	0.64	0.72	0.31	1.71	0.42	0.26	0.51	0.33	0.46	0.76	0.15	0.24	19.10	-
U**	0.26	0.29	0.25	0.49	0.25	0.12	0.42	0.24	0.17	0.22	0.06	0.09	0.95	-
La**	42.86	42.18	26.13	41.83	42.41	30.06	72.30	16.30	73.30	34.50	18.45	21.40	370.00	-
Ce**	94.79	92.37	54.91	92.15	95.91	66.57	161.00	33.30	170.00	53.90	32.50	42.80	622.00	-
Pr**	12.33	11.96	7.23	12.22	13.07	9.25	20.50	4.09	22.90	5.43	3.78	4.75	62.70	-
Nd**	50.71	48.93	30.46	50.42	54.53	39.46	80.40	18.60	94.40	21.40	14.36	19.30	206.00	-
Sm**	9.98	9.52	6.23	10.06	11.54	8.07	14.20	3.59	15.80	3.48	3.05	3.44	18.80	-
Eu**	2.89	2.93	3.19	2.83	2.81	3.56	3.05	2.14	3.24	1.87	1.00	1.08	1.40	-
Gd**	7.81	7.44	4.82	7.76	8.90	6.45	10.30	2.85	11.10	2.22	2.98	2.23	3.37	-
Tb**	1.06	1.03	0.70	1.08	1.23	0.90	1.56	0.39	1.56	0.33	0.46	0.31	0.85	-
Dy**	5.77	5.66	3.83	5.76	6.54	4.97	8.59	2.10	8.22	1.71	2.70	1.58	4.03	-
Ho**	1.13	1.13	0.78	1.13	1.23	0.98	1.68	0.42	1.54	0.33	0.59	0.30	0.68	-
Er**	2.86	2.88	1.98	2.84	2.98	2.47	4.46	1.16	4.04	0.85	1.61	0.82	1.88	-
Yb**	2.46	2.55	1.79	2.46	2.40	2.12	3.93	1.12	3.18	0.70	1.67	0.78	1.77	-
Lu**	0.38	0.40	0.29	0.38	0.37	0.33	0.59	0.19	0.46	0.10	0.27	0.12	0.31	-

ANEXO VII – ANÁLISES ISOTÓPICAS (Rb-Sr E Sm-Nd) DE ROCHA TOTAL

Dados isotópicos Rb-Sr e Sm-Nd em rocha total para charnockitos da Suite Paraguaçu

SAMPLE	Age	Rb	Sr	$^{87}\text{Sr}/^{86}\text{Sr}$	erro (2s)	$^{7}\text{Sr}/^{86}\text{Sr}$	(i)	Sm	Nd	$^{47}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	Ni erro (2s)	$\epsilon_{\text{Nd}(0)}$	$\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}(620)}$	TDM
PAR-32A	620	66.43	327.87	0.714666	0.000018	0.7095		9.98	50.71	0.1190	0.511695	0.000005	-18.40	0.51121	-12.3	2.21
PAR-32C	620	68.47	337.03	0.714758	0.000020	0.7096		9.52	48.93	0.1176	0.511681	0.000003	-18.67	0.5112	-12.4	2.20
PAR-33	620	95.46	275.22	0.719586	0.000019	0.7107		6.23	30.46	0.1236	0.511714	0.000003	-18.02	0.51121	-12.3	2.29
PAR-36A	620	78.65	320.81	0.716093	0.000018	0.7098		10.06	50.42	0.1206	0.511710	0.000006	-18.11	0.51122	-12.1	2.22
PAR-36C	620	82.83	329.34	0.716551	0.000022	0.7101		11.54	54.53	0.1279	0.511756	0.000002	-17.20	0.51124	-11.8	2.32
PAR-40	620	67.72	320.53	0.715523	0.000018	0.7101		8.07	39.46	0.1236	0.511707	0.000004	-18.16	0.51121	-12.4	2.3

ANEXO VIII – DATAÇÃO U-Pb EM ZIRCÃO VIA SHRIMP

Dados U-Pb em zircão (SHRIMP) - Grt-granulito fêlsico e diatexito - Unidade Granulítica

Sample-spot	Texture	U ppm	Th ppm	Th/U	206Pb	207Pb/206Pb	2σ%	207Pb/235U	2σ%	206Pb/238U	2σ%	Age 206Pb/238U	1σ
VAB01-1.1	sball	16.5	60.3	3.64	1.59	0.0544	16.2	0.7296	16.4	0.0973	2.7	598.8	15.3
VAB01-2.1	sball	23.5	92.1	3.92	-0.49	0.0678	9.0	0.9469	9.3	0.1014	2.3	622.4	13.8
VAB01-3.1	sball	70.1	64.9	0.93	0.60	0.0595	4.7	0.7996	4.9	0.0974	1.4	599.2	8.0
VAB01-4.1	sball	53.7	160.0	2.98	1.22	0.0586	7.1	0.7770	7.2	0.0962	1.5	592.2	8.7
VAB01-5.1	sball	17.7	31.9	1.81	4.11	0.0467	28.6	0.6037	28.8	0.0938	2.9	578.2	16.1
VAB01-6.1	sball	68.0	74.7	1.10	1.52	0.0577	7.8	0.7468	7.9	0.0939	1.5	578.4	8.2
VAB01-7.1	sball	32.9	103.7	3.16	3.24	0.0455	20.5	0.5991	20.6	0.0955	2.1	588.1	11.8
VAB01-8.1	sball	36.0	37.9	1.05	-1.26	0.0763	8.1	1.0729	8.4	0.1019	2.2	625.7	12.9
VAB01-9.1	sball	158.6	287.9	1.82	0.00	0.0615	1.3	0.8165	1.7	0.0963	1.0	592.4	5.7
VAB01-10.1	sball	21.4	35.0	1.64	2.97	0.0465	23.6	0.6105	23.7	0.0953	2.6	586.7	14.6
VAB01-11.1	sball	14.4	19.6	1.36	-4.21	0.1068	15.5	1.5057	15.9	0.1023	3.5	627.7	21.2
VAB01-12.1	sball	15.8	20.5	1.30	7.09	0.0204	121.2	0.2554	121.2	0.0910	4.0	561.3	21.5
VAB01-13.1	sball	12.9	16.8	1.31	0.00	0.0863	7.2	1.1334	7.8	0.0952	3.0	586.4	16.9
VAB01-14.1	rim	16.3	52.7	3.23	8.90	0.0022	976.0	0.0270	976.0	0.0876	3.7	541.6	19.3
VAB01-14.2	osc-core	134.4	99.2	0.74	0.07	0.0627	1.6	0.9285	2.0	0.1074	1.1	657.4	6.8
VAB01-15.1	osc-core	255.0	170.1	0.67	0.40	0.0601	1.9	0.9524	2.3	0.1150	1.4	701.7	9.6
VAB01-16.1	osc-core	177.7	100.1	0.56	0.45	0.0583	2.6	0.7982	2.8	0.0994	1.0	610.6	5.8
VAB01-17.1	osc-core	144.1	72.0	0.50	0.06	0.0657	1.5	0.9818	1.8	0.1083	1.1	663.0	6.7
VAB01-18.1	osc-core	175.8	94.4	0.54	0.24	0.0632	1.8	0.9826	2.3	0.1128	1.4	688.9	9.4
VAB01-19.1	osc-core	198.3	136.2	0.69	0.14	0.0639	2.1	0.9333	2.3	0.1060	1.0	649.2	5.9
VAB19-1.1	sball	24.6	29.8	1.21	3.08	0.0492	25.6	0.6359	25.7	0.0937	2.7	577.4	14.9
VAB19-2.1	sball	29.9	25.9	0.87	3.18	0.0496	23.9	0.6632	24.0	0.0969	2.4	596.3	13.7
VAB19-3.1	osc-core	226.1	131.8	0.58	0.34	0.0611	1.9	0.9630	2.1	0.1142	0.9	697.2	6.2
VAB19-4.1	sball	15.4	21.0	1.36	0.87	0.0834	16.1	1.1340	16.4	0.0987	3.1	606.6	18.2

Continuação - Dados U-Pb em zircão (SHRIMP) - Grt-granulito fêlsico e diatextito - Unidade Granulítica													
Sample-spot	Texture	U ppm	Th ppm	Th/U	²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ%	²⁰⁷ Pb/ ²³⁵ U	2σ%	²⁰⁶ Pb/ ²³⁸ U	2σ%	Age ²⁰⁶ Pb/ ²³⁸ U	1σ
VAB19-5.1	rim	38.4	43.8	1.14	1.79	0.0723	10.8	0.9723	10.9	0.0975	1.9	600.0	11.1
VAB19-6.1	rim	77.5	48.0	0.62	1.85	0.0564	7.6	0.7670	7.7	0.0986	1.6	606.3	9.0
VAB19-7.1	osc-core	197.7	120.4	0.61	0.46	0.0610	2.3	0.9421	2.5	0.1121	1.0	684.9	6.3
VAB19-8.1	rim	23.0	8.2	0.36	3.85	0.0634	24.7	0.8436	24.8	0.0966	2.7	594.2	15.2
VAB19-9.1	osc-core	95.2	54.3	0.57	0.17	0.0613	2.2	0.8534	2.6	0.1010	1.5	620.3	8.7
VAB19-10.1	rim	30.3	69.0	2.28	1.35	0.0524	10.2	0.7066	10.4	0.0977	2.1	601.1	12.2
VAB19-11.1	rim	13.1	32.6	2.49	6.45	0.0833	26.8	1.1055	27.0	0.0962	3.5	592.2	19.8
RGM55-1.1	sect-core	242.7	144.5	0.60	0.07	0.0609	1.2	0.8637	1.7	0.1029	1.3	631.4	7.6
RGM55-2.1	rim	311.4	38.7	0.12	0.22	0.0599	1.4	0.8045	2.0	0.0974	1.5	599.2	8.3
RGM55-3.1	sball	85.0	58.9	0.69	-0.10	0.0632	2.0	0.8677	2.5	0.0996	1.5	612.0	8.8
RGM55-4.1	rim	272.2	31.3	0.12	0.15	0.0598	1.3	0.8256	1.8	0.1001	1.2	615.1	7.3
RGM55-5.1	rim	346.2	55.8	0.16	0.10	0.0603	1.1	0.8030	1.6	0.0966	1.2	594.5	6.9
RGM55-6.1	sect-core	187.2	46.2	0.25	0.00	0.0622	1.1	0.8743	1.7	0.1020	1.3	626.2	7.7
RGM55-7.1	sball	289.9	16.3	0.06	-0.03	0.0615	1.0	0.8421	1.8	0.0994	1.5	610.7	8.7
RGM55-8.1	rim	293.4	13.8	0.05	0.00	0.0607	0.9	0.8426	1.5	0.1006	1.2	617.9	7.2
RGM55-9.1	sball	244.4	37.1	0.15	0.24	0.0600	1.6	0.8272	2.0	0.1001	1.3	614.7	7.4
RGM55-10.1	sect-core	277.7	134.8	0.49	0.00	0.0613	0.9	0.8590	1.5	0.1016	1.2	623.7	7.3
RGM55-11.1	rim	347.7	41.3	0.12	-0.10	0.0622	1.1	0.8423	1.8	0.0982	1.4	603.6	8.3
RGM55-11.2	sect-core	105.6	83.1	0.79	0.00	0.0627	1.5	0.9010	2.1	0.1042	1.5	639.1	8.9
RGM55-12.1	sect-core	181.1	94.2	0.52	0.30	0.0596	2.1	0.8368	2.4	0.1018	1.3	624.8	7.9
RGM55-13.1	rim	176.3	104.7	0.59	0.23	0.0625	1.7	0.8667	2.2	0.1006	1.3	617.6	7.7
RGM55-13.2	sect-core	210.3	164.0	0.78	0.39	0.0585	2.0	0.8202	2.4	0.1017	1.3	624.3	7.6
RGM55-14.1	rim	238.6	64.4	0.27	0.10	0.0608	1.3	0.8472	2.1	0.1011	1.6	620.8	9.6
RGM55-14.2	sect-core	126.6	66.1	0.52	0.32	0.0596	2.4	0.8214	2.8	0.1000	1.4	614.4	8.1

sball = soccerball zircon

osc-core = oscillatory-zoned core

sect-core = sectorial-zoned core

Dados U-Pb em zircão (SHRIMP) - charnockito - Suíte Paraguaçu

Sample-spot	$^{232}\text{Th}/^{238}\text{U}$	^{206}Pb	$^{207}\text{Pb}/^{206}\text{Pb}$	$2\sigma\%$	$^{207}\text{Pb}/^{235}\text{U}$	$2\sigma\%$	$^{206}\text{Pb}/^{238}\text{U}$	$2\sigma\%$	Age $^{206}\text{Pb}/^{238}\text{U}$	1σ
RGM02-1.1	0.64	7.96	0.0185	273.5	0.2300	273.6	0.0903	5.2	557.2	27.8
RGM02-2.1	0.99	-0.26	0.0628	1.8	0.8840	2.6	0.1021	1.8	626.8	10.8
RGM02-2.2	1.22	0.39	0.0585	1.7	0.8215	2.3	0.1019	1.6	625.7	9.3
RGM02-3.1	0.83	0.10	0.0586	1.8	0.7858	2.7	0.0973	2.0	598.5	11.4
RGM02-4.1	0.76	-2.18	0.0762	13.6	1.0550	13.9	0.1005	3.2	617.2	18.9
RGM02-4.2	0.89	-0.09	0.0627	1.5	0.8664	2.2	0.1002	1.6	615.7	9.5
RGM02-5.1	0.25	4.19	0.0285	56.7	0.3976	56.8	0.1012	3.6	621.7	21.6
RGM02-5.2	0.62	1.64	0.0473	20.7	0.6084	20.9	0.0933	2.8	575.1	15.5
RGM02-6.1	0.15	0.34	0.0543	6.4	0.7092	6.8	0.0947	2.3	583.5	12.6
RGM02-6.2	0.56	-0.06	0.0617	1.6	0.8726	2.3	0.1025	1.7	629.0	9.9
RGM02-7.1	0.39	0.21	0.0586	4.1	0.7423	4.6	0.0919	2.0	566.8	10.8
RGM02-8.1	0.69	0.00	0.0633	1.4	0.8363	2.2	0.0958	1.6	589.7	9.3
RGM02-9.1	0.37	0.03	0.0611	1.1	0.8700	1.9	0.1034	1.6	634.1	9.4
RGM02-9.2	1.36	0.19	0.0596	1.1	0.8291	1.9	0.1010	1.5	620.1	9.0
RGM02-10.1	0.23	1.16	0.0520	8.2	0.7053	8.5	0.0985	2.1	605.4	11.9
RGM02-11.1	0.66	0.12	0.0587	4.5	0.8049	4.8	0.0994	1.8	611.0	10.8
RGM02-12.1	1.31	0.15	0.0603	1.5	0.8658	2.2	0.1042	1.6	639.1	9.7
RGM04-1.1	0.85	0.00	0.0613	0.5	0.8769	1.6	0.1037	1.5	635.9	9.0
RGM04-2.1	1.02	-0.09	0.0620	0.9	0.8363	1.8	0.0978	1.6	601.5	9.3
RGM04-2.2	1.05	0.04	0.0609	0.5	0.8529	1.7	0.1016	1.6	623.6	9.3
RGM04-3.1	0.33	0.57	0.0548	4.0	0.7493	4.4	0.0991	1.8	609.1	10.3
RGM04-3.2	0.70	-0.08	0.0611	1.0	0.8653	1.8	0.1027	1.5	630.4	9.2
RGM04-4.1	1.05	-0.23	0.0623	1.8	0.8809	2.4	0.1026	1.6	629.5	9.7
RGM04-4.2	0.63	-0.16	0.0610	2.2	0.8510	2.8	0.1011	1.7	620.9	10.1
RGM04-5.1	0.93	0.47	0.0573	2.5	0.8160	3.0	0.1034	1.6	634.1	9.8
RGM04-6.1	1.00	-0.10	0.0616	1.6	0.8752	2.2	0.1031	1.6	632.7	9.8
RGM04-7.1	0.94	-0.31	0.0635	1.8	0.9065	2.4	0.1035	1.6	635.1	9.7
RGM04-8.1	0.23	-0.13	0.0614	3.8	0.8187	4.2	0.0967	1.8	595.0	10.3
RGM04-9.1	0.81	0.10	0.0609	2.3	0.8228	2.8	0.0979	1.6	602.4	9.3
RGM04-8.2	0.91	-0.05	0.0615	2.1	0.8430	2.7	0.0994	1.6	611.0	9.3
RGM04-10.1	0.85	0.13	0.0590	1.1	0.8431	1.9	0.1036	1.5	635.8	9.3
RGM04-10.2	0.42	0.04	0.0596	1.1	0.7996	1.8	0.0973	1.5	598.4	8.5
RGM04-11.1	1.01	0.07	0.0613	0.8	0.8582	1.7	0.1016	1.5	623.6	9.0
RGM04-12.1	0.50	0.36	0.0586	2.8	0.7894	3.3	0.0977	1.7	600.6	9.7

Sample-spot	²³² Th/ ²³⁸ U	²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ%	²⁰⁷ Pb/ ²³⁵ U	2σ%	²⁰⁶ Pb/ ²³⁸ U	2σ%	Age ²⁰⁶ Pb/ ²³⁸ U	1σ
RGM53-1.1	1.08	0.15	0.0598	1.1	0.8360	1.9	0.1014	1.5	622.8	9.0
RGM53-1.2	0.75	0.46	0.0561	4.7	0.7772	5.0	0.1005	1.9	617.6	11.4
RGM53-2.1	1.19	0.42	0.0604	4.9	0.8272	5.3	0.0993	2.1	610.5	12.3
RGM53-3.1	1.01	0.28	0.0602	3.6	0.8152	4.0	0.0983	1.9	604.2	10.8
RGM53-4.1	1.13	0.09	0.0626	3.1	0.8556	3.6	0.0991	1.7	609.2	10.1
RGM53-5.1	1.13	0.22	0.0605	4.1	0.8269	4.6	0.0992	2.1	609.5	12.1
RGM53-6.1	0.29	0.42	0.0558	5.2	0.7820	5.6	0.1016	2.1	623.5	12.2
RGM53-6.2	1.05	-0.61	0.0656	5.0	0.9399	5.4	0.1040	2.1	637.5	12.5
RGM53-7.1	0.67	1.90	0.0466	21.0	0.6634	21.2	0.1032	3.0	633.1	18.0
RGM53-8.1	1.13	0.40	0.0587	3.4	0.8212	3.8	0.1015	1.8	623.3	10.6
RGM53-9.1	0.44	-0.08	0.0609	2.0	0.8361	2.6	0.0996	1.7	612.3	10.0
RGM53-10.1	1.21	0.00	0.0622	2.8	0.8764	3.5	0.1023	2.1	627.6	12.4
RGM53-11.1	1.21	-0.19	0.0645	3.4	0.9164	3.9	0.1030	2.0	632.2	11.9
RGM53-12.1	1.09	0.57	0.0588	5.4	0.7983	5.7	0.0984	2.0	605.0	11.6
RGM53-13.1	1.12	1.50	0.0516	8.6	0.7087	8.9	0.0996	2.0	612.1	11.7
RGM53-14.1	0.89	-1.07	0.0674	5.6	0.9682	6.0	0.1041	2.0	638.5	12.3
RGM53-15.1	0.43	0.09	0.0611	2.2	0.8591	2.8	0.1020	1.7	626.3	10.4
RGM57-1.1	-0.36	0.95	0.0639	3.3	0.8666	3.7	0.0984	1.8	605.1	10.6
RGM57-2.1	0.40	0.96	0.0592	4.9	0.7801	5.3	0.0955	1.9	588.1	10.5
RGM57-3.1	-0.37	0.84	0.0646	4.1	0.9002	4.5	0.1010	2.0	620.4	11.6
RGM57-3.2	-0.06	2.29	0.0619	1.0	0.8530	1.8	0.0999	1.5	614.1	9.0
RGM57-4.1	0.42	0.88	0.0551	5.3	0.7726	5.7	0.1017	2.1	624.4	12.5
RGM57-5.1	0.05	1.29	0.0601	0.6	0.8346	1.6	0.1007	1.5	618.6	8.8
RGM57-5.2	-0.43	0.89	0.0638	3.2	0.8819	3.7	0.1003	1.8	615.9	10.5
RGM57-6.1	0.15	0.95	0.0605	3.1	0.8123	3.6	0.0974	1.9	599.2	10.8
RGM57-7.1	0.04	1.05	0.0603	0.5	0.8268	1.5	0.0994	1.5	610.7	8.6
RGM57-8.1	0.10	1.65	0.0606	1.0	0.8394	1.9	0.1004	1.5	617.0	9.0
RGM57-9.1	0.01	0.51	0.0604	0.7	0.8189	1.7	0.0984	1.5	605.0	8.7
RGM57-9.2	-0.34	0.82	0.0630	3.1	0.8331	3.6	0.0959	1.8	590.3	10.2
RGM57-10.1	0.00	0.49	0.0614	1.6	0.8079	2.3	0.0955	1.7	587.7	9.4
RGM57-10.2	0.12	0.91	0.0598	0.8	0.8625	1.7	0.1046	1.5	641.2	9.2
RGM57-11.1	-0.06	1.80	0.0620	0.9	0.8354	1.8	0.0978	1.5	601.3	8.8
RGM57-12.1	0.06	0.72	0.0596	0.8	0.8198	1.7	0.0998	1.5	613.3	8.9
RGM57-13.1	0.50	0.85	0.0571	6.0	0.7728	6.3	0.0982	1.9	603.9	11.0

ANEXO IX – DATAÇÃO COMPLEMENTAR U-Pb EM ZIRCÃO VIA LASS-ICP- MS

Análise complementar U-Pb via LASS-ICP-MS (granulitos - Nappe Socorro-Guaxupé). Em negrito dados utilizados nas discussões.

Sample-spot	SHRIMP ID	U ppm	Th ppm	Th/U	207Pb/206Pb	2σ	207Pb/235U	2σ	206Pb/238U	2σ	Age 206Pb/238U (Ma)	2σ
VAB01 - 1	1.1	30.7	138.1	4.498371	0.0609	0.0063	0.843	0.088	0.1004	0.0026	617	15
VAB01 - 2	3.1	85.3	86	1.008206	0.0589	0.0033	0.815	0.045	0.1003	0.0016	616.3	9.3
VAB01 - 3	2.1	24.3	111.9	4.604938	0.0567	0.0064	0.79	0.088	0.1013	0.0029	622	17
VAB01 - 4	13.1	14.49	19.29	1.331263	0.0548	0.0081	0.76	0.11	0.1011	0.0036	621	21
VAB01 - 5	4.1	94.6	87.8	0.928118	0.0587	0.0036	0.835	0.051	0.1034	0.0017	634	10
VAB01 - 6	5.1	60.7	104	1.713344	0.0595	0.0041	0.836	0.058	0.1018	0.0017	624.8	9.9
VAB01 - 7	11.1	15.05	41.5	2.757475	0.063	0.01	0.86	0.14	0.099	0.0042	608	25
VAB01 - 8	7.1	34.09	90.4	2.651804	0.0583	0.0055	0.798	0.072	0.0999	0.0027	614	16
VAB01 - 9	8.1	48	67	1.395833	0.0565	0.0068	0.653	0.066	0.0958	0.0034	590	20
VAB01 - 10	9.1	58	222	3.827586	0.0586	0.0052	0.823	0.075	0.1017	0.0029	624	17
VAB01 - 11	10.1	18.4	44	2.391304	0.0533	0.0083	0.74	0.12	0.1006	0.0038	618	22
VAB01 - 12	15.1	298	211	0.708054	0.0615	0.0015	1.006	0.028	0.1184	0.0012	721.4	6.9
VAB01 - 13	-	13.65	29.9	2.190476	0.056	0.011	0.78	0.16	0.1024	0.0036	628	21
VAB01 - 14	16.1	264	127	0.481061	0.0614	0.0022	0.929	0.035	0.1092	0.0018	668	10
VAB01 - 15	17.1	197	116	0.588832	0.0612	0.002	0.949	0.038	0.1106	0.0012	676.3	6.8
VAB01 - 16	18.1	226	74.5	0.329646	0.062	0.0023	0.963	0.033	0.1126	0.0016	687.9	9.2
VAB01 - 17	19.1	275	277	1.007273	0.0612	0.0016	0.9	0.022	0.1066	0.0013	653	7.5
VAB01 - 18	6.1	156	98	0.628205	0.0596	0.0023	0.912	0.038	0.1108	0.0019	677	11
VAB01 - 19	-	102.2	66.4	0.649706	0.0617	0.0028	0.783	0.04	0.1017	0.0024	624	14
VAB19 - 1	5.1	158.3	109.9	0.694251	0.0613	0.0025	0.879	0.034	0.1044	0.0017	639.9	9.7
VAB19 - 2	4.1	19.7	34.2	1.736041	0.0568	0.0055	0.782	0.095	0.1021	0.0039	626	23
VAB19 - 3	3.1	261.6	184.3	0.704511	0.0613	0.0019	0.987	0.03	0.1171	0.002	714	11
VAB19 - 4	2.1	43	47.6	1.106977	0.0569	0.0055	0.776	0.072	0.0997	0.0016	612.4	9.6
VAB19 - 5	7.1	219	147	0.671233	0.0621	0.0024	0.988	0.043	0.1168	0.0022	712	13
VAB19 - 8	1.1	14.12	29.22	2.069405	0.066	0.01	0.93	0.14	0.1022	0.0037	627	22
VAB19 - 9	8.1	132.3	76.4	0.577475	0.0592	0.0022	0.745	0.028	0.0962	0.0015	592.3	8.9
VAB19 - 10	10.1	164	155	0.945122	0.0626	0.0025	0.945	0.046	0.1095	0.0029	670	17
RGM55 - 1	13.2	172	134.6	0.782558	0.058	0.0025	0.844	0.037	0.1059	0.0016	648.6	9.2
RGM55 - 2	14.1	147.7	118.5	0.802302	0.0604	0.0024	0.857	0.036	0.1032	0.001	633.3	5.8
RGM55 - 4	12.1	176.7	99.3	0.561969	0.0591	0.0023	0.848	0.032	0.10468	0.0007	641.8	4.1
RGM55 - 5	11.1	372	55.9	0.150269	0.0579	0.0015	0.801	0.019	0.10066	0.00084	618.3	4.9
RGM55 - 6	9.1	414	158	0.381643	0.0593	0.0013	0.87	0.022	0.1067	0.0014	653.8	8.1
RGM55 - 7	8.1	420	24.5	0.058333	0.06	0.0015	0.84	0.021	0.10191	0.0009	625.6	5.2
RGM55 - 8	7.1	371	45.5	0.122642	0.0602	0.0015	0.854	0.021	0.10308	0.00085	632.4	4.9
RGM55 - 10	3.1	143	125.6	0.878322	0.062	0.0021	0.924	0.033	0.1083	0.0012	662.6	7.1
RGM55 - 11	4.1	348	36.4	0.104598	0.0593	0.0014	0.846	0.021	0.10358	0.00088	635.4	5.1
RGM55 - 12	1.1	293	80.7	0.275427	0.0596	0.0016	0.854	0.023	0.1042	0.0011	639.2	6.1
RGM55 - 14	5.1	426	68.9	0.161737	0.0599	0.0012	0.824	0.015	0.10006	0.00093	614.7	5.5
RGM55 - 15	-	481	41.48	0.086237	0.0598	0.0016	0.841	0.02	0.10227	0.00092	627.7	5.4
RGM55 - 17	10.1	540	166	0.307407	0.06	0.0017	0.857	0.025	0.1036	0.0011	635.3	6.1
RGM55 - 18	-	452	74	0.163717	0.0597	0.0011	0.718	0.014	0.09707	0.00096	597.2	5.6
RGM55 - 20	2.1	332	102	0.307229	0.0594	0.0017	0.854	0.019	0.10374	0.0009	636.2	5.3
RGM55 - 3	-	200.3	152.2	0.75986	0.0612	0.0017	0.899	0.028	0.1068	0.0017	654	10
RGM55 - 16	-	403	161.2	0.4	0.0588	0.0013	0.868	0.017	0.108	0.0012	660.9	7.1
RGM55 - 13	-	240	172.5	0.71875	0.0602	0.0015	0.882	0.025	0.1064	0.0015	651.5	8.5
RGM55 - 9	-	378.8	53.1	0.14018	0.0604	0.0014	0.862	0.017	0.10444	0.00097	640.3	5.7
RGM55 - 19	-	316	166.8	0.52785	0.0612	0.0019	0.868	0.03	0.1029	0.0015	631.4	8.5

Análise complementar U-Pb via LASS-ICP-MS (charnockito - Suíte Paraguaçu)

Sample-spot	SHRIMP ID	U ppm	Th ppm	Th/U	207Pb/206Pb	2σ	207Pb/235U	2σ	206Pb/238U	2σ	Age 206Pb/238U (Ma)	2σ
RGM02 - 1	6.2	161	84	0.521739	0.0657	0.0033	0.887	0.043	0.098	0.0017	602.5	9.9
RGM02 - 2	4.2	166.1	119.3	0.718242	0.0636	0.0027	0.887	0.043	0.1008	0.0014	619.2	8.2
RGM02 - 3	3.1	112	119	1.0625	0.0627	0.0035	0.949	0.053	0.1099	0.0018	672	11
RGM02 - 4	9.2	172	117	0.680233	0.06	0.0025	0.856	0.034	0.1036	0.0012	635.2	7.2
RGM02 - 5	2.2	189	153.2	0.810582	0.0622	0.0016	0.926	0.023	0.1079	0.0013	660.3	7.7
RGM02 - 6	2.1	200	210	1.05	0.0609	0.0023	0.879	0.032	0.1046	0.0013	641.3	7.7
RGM02 - 7	-	264	256	0.969697	0.0618	0.0015	0.894	0.022	0.1049	0.0013	642.9	7.4
RGM02 - 8	-	191	159	0.832461	0.0611	0.0026	0.844	0.033	0.101	0.0018	620	11
RGM02 - 9	11.1	150	75.2	0.501333	0.0635	0.0026	0.822	0.034	0.0949	0.0014	584.2	8.5
RGM02 - 10	12.1	195	199	1.020513	0.0606	0.0018	0.84	0.026	0.1004	0.0013	616.4	7.7
RGM02 - 11	10.1	183.5	94.1	0.512807	0.0615	0.0026	0.874	0.035	0.1031	0.0011	632.2	6.7
RGM04 - 1	11.1	454	479	1.055066	0.0605	0.0011	0.848	0.017	0.1016	0.0013	624	7.5
RGM04 - 2	10.2	91.9	99.4	1.08161	0.0602	0.0029	0.719	0.037	0.0961	0.0018	592	10
RGM04 - 3	-	129	139	1.077519	0.0601	0.0028	0.865	0.039	0.1045	0.0013	640.6	7.7
RGM04 - 4	-	299	302	1.010033	0.0589	0.0015	0.854	0.021	0.1051	0.0011	644	6.5
RGM04 - 5	2.2	409	380	0.929095	0.06	0.0015	0.877	0.022	0.10626	0.00086	651	5
RGM04 - 6	1.1	1098	735	0.669399	0.0597	0.001	0.869	0.028	0.1054	0.0029	646	17
RGM04 - 7	8.2	147.8	145	0.981055	0.0645	0.0039	0.91	0.06	0.1021	0.0011	626.7	6.3
RGM04 - 8	5.1	188	202	1.074468	0.0616	0.0022	0.855	0.03	0.1005	0.0012	617.3	7.3
RGM04 - 9	6.1	113.6	104.8	0.922535	0.0598	0.0023	0.868	0.038	0.105	0.0014	643.7	8.2
RGM04 - 10	4.1	70.8	42.1	0.594633	0.0592	0.0038	0.841	0.057	0.1029	0.002	631	12
RGM53 - 1	6.1	42.7	18.7	0.437939	0.0555	0.0053	0.767	0.071	0.1007	0.0017	618.2	9.9
RGM53 - 2	15.1	61.4	49.6	0.807818	0.0583	0.0037	0.823	0.054	0.1019	0.002	625	11
RGM53 - 3	-	186	196	1.053763	0.0598	0.0021	0.819	0.029	0.0995	0.0014	611.3	8.2
RGM53 - 4	2.1	56.7	66.6	1.174603	0.0623	0.0042	0.884	0.059	0.103	0.0019	632	11
RGM53 - 5	10.1	48.7	56	1.149897	0.0676	0.0043	0.958	0.064	0.1028	0.0025	630	14
RGM53 - 6	9.1	103.6	54.4	0.525097	0.0615	0.0028	0.86	0.042	0.1014	0.0012	622.3	7.2
RGM53 - 7	1.1	459	479	1.043573	0.0591	0.0016	0.846	0.022	0.10391	0.00077	637.3	4.5
RGM53 - 8	8.1	125.4	84.3	0.672249	0.0579	0.0028	0.807	0.038	0.1013	0.0013	621.9	7.7
RGM53 - 9	13.1	141	174	1.234043	0.0581	0.0024	0.713	0.033	0.0976	0.0018	600	10
RGM53 - 10	12.1	44.6	47.6	1.067265	0.0628	0.0045	0.873	0.066	0.1009	0.0018	619	11
RGM53 - 11	5.1	58.4	67.1	1.148973	0.0597	0.0041	0.841	0.058	0.1021	0.0019	627	11
RGM57 - 1	6.1	69.7	72.4	1.038737	0.075	0.0068	1.034	0.097	0.0999	0.0017	613.7	9.8
RGM57 - 2	5.1	676	789	1.16716	0.0608	0.0011	0.871	0.018	0.10389	0.00078	637.2	4.6
RGM57 - 3	13.1	53.8	56.6	1.052045	0.0605	0.004	0.846	0.053	0.1017	0.0019	625	11
RGM57 - 4	7.1	376	403	1.071809	0.0604	0.0014	0.867	0.022	0.1041	0.0011	638.1	6.7
RGM57 - 5	12.1	641	697	1.087363	0.0601	0.0012	0.848	0.016	0.10226	0.00063	627.6	3.7
RGM57 - 6	1.1	69	69.3	1.004348	0.0586	0.005	0.806	0.066	0.1001	0.0016	614.7	9.6
RGM57 - 7	2.1	182.6	90.5	0.495619	0.0598	0.002	0.828	0.03	0.1001	0.001	614.9	5.9
RGM57 - 8	3.2	328	145	0.442073	0.0588	0.0018	0.828	0.023	0.1022	0.001	628.8	5.4
RGM57 - 9	9.1	199	297	1.492462	0.0682	0.0033	1.035	0.053	0.11	0.0019	673	11

ANEXO X – ANÁLISES DE ELEMENTOS TRAÇO EM ZIRCÃO VIA LASS-ICP- MS

Análise de elementos traço em zircão (granulitos - *Nappe Socorro-Guaxupé*).

Sample/spot	SHRIMP ID	Texture	Ti	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Er	Yb	Lu	(Gd/Yb)N	Eu/Eu*	Ce/Ce*
VAB01 - 1	1.1	soccer ball	21.4	0.047	53.8	0.296	5.59	11.2	2.32	35.3	5.96	36.3	14.6	11.25	1.53	2.589	0.357	111.834
VAB01 - 2	3.1	soccer ball	16.7	-	35.9	0.158	2.09	4.4	0.76	11.8	2.49	19.3	16.71	21.5	3.86	0.453	0.323	-
VAB01 - 3	2.1	soccer ball	16.8	0.044	51.6	0.216	2.64	4.17	0.76	14.1	3.31	28.2	21	22.5	4.07	0.517	0.303	129.772
VAB01 - 4	13.1	soccer ball	24.2	0.057	30.91	0.148	2.22	2.62	0.87	9.3	1.92	14.7	9.74	8.87	1.41	0.865	0.539	82.512
VAB01 - 6	5.1	soccer ball	28.2	0.091	50.6	0.317	6.1	9.2	2.46	32.2	6.13	46.3	27.3	27.2	4.25	0.977	0.437	73.044
VAB01 - 7	11.1	soccer ball	18	0.001	37.3	0.156	2.08	4.31	0.98	17.2	4.28	38.6	36.2	46	8.82	0.309	0.348	732.202
VAB01 - 8	7.1	soccer ball	19.7	-	40.26	0.152	3.27	5.56	1.4	17.2	3.71	29.2	22.8	25.4	4.46	0.559	0.438	-
VAB01 - 9	8.1	soccer ball	24.3	0.035	34.4	0.144	3.2	3.92	1.1	11.8	2.46	21.5	22.1	42	8.7	0.232	0.495	118.803
VAB01 - 10	9.1	soccer ball	18.1	0.02	43.3	0.163	2.67	3.93	1.03	14.3	3.27	23.5	19.6	23.1	3.73	0.511	0.421	185.936
VAB01 - 11	10.1	soccer ball	20.5	0.033	41.1	0.172	3.94	8	1.78	24.9	4.24	27.1	12.07	9.63	1.54	2.134	0.386	133.754
VAB01 - 13	-	soccer ball	18.2	0.009	37.8	0.126	1.72	4.41	0.81	14.8	3.53	30.2	28	35.6	6.48	0.343	0.307	275.214
VAB01 - 14	16.1	soccer ball	22.1	3.9	25.3	0.9	4	3.52	0.56	14.8	5.18	68.5	154.1	386.1	95.1	0.032	0.237	3.311
VAB01 - 18	6.1	rim	19.8	0.166	29.7	0.374	3.96	4.53	1.39	15.7	4.93	52.3	91.6	213	50.2	0.061	0.505	29.225
VAB01 - 5	4.1	rim	19.7	0.046	35.4	0.104	2.1	4.25	1.13	15.6	4.71	53.7	91	178	38.6	0.072	0.425	125.485
VAB01 - 19	-	rim	21.3	0.02	33.6	0.124	2.06	3.43	0.96	10.9	2.7	24.9	35.2	60	13.09	0.150	0.481	165.424
VAB01 - 16	18.1	core	16.1	1.56	23.7	1.68	14.1	7.7	3.57	21.2	6.6	79.8	157.5	377	89.2	0.046	0.855	3.589
VAB01 - 15	17.1	core	21.5	7.3	34.3	1.61	7.7	6.7	2.11	36.8	12.6	147.7	284	593	130.3	0.051	0.411	2.453
VAB01 - 17	19.1	core	40	2.39	36.7	1.02	9.5	10.5	3.57	40.9	12.64	155.7	313.6	755	175	0.045	0.527	5.763
VAB01 - 12	15.1	core	13.1	3.5	32.6	0.92	5.8	5.24	1.2	26.7	9.98	121.1	252.8	549	122.4	0.040	0.311	4.454
VAB19 - 1	5.1	soccer ball	18.8	0.073	26.5	0.15	3.35	6.18	1.33	25.3	7.82	82.8	137	289	64.3	0.072	0.326	62.090
VAB19 - 2	4.1	soccer ball	22.4	0.014	33.7	0.183	3.51	5.9	1.56	19.1	3.22	18.6	7.95	6.91	1.03	2.281	0.450	163.239
VAB19 - 4	2.1	soccer ball	24.6	0.001	35.2	0.155	3.63	6.2	1.29	19	4.07	29.1	19.5	20.2	3.41	0.776	0.364	693.204
VAB19 - 8	1.1	soccer ball	20.5	-	27.74	0.08	1.57	2.11	0.74	7.28	1.41	9.08	3.85	3.74	0.568	1.606	0.578	-
VAB19 - 10	10.1	soccer ball	21	3.6	37	1.58	10.3	6.92	2.65	25.8	8.14	91	159	344	75.5	0.062	0.607	3.804
VAB19 - 9	8.1	rim	11.8	64	16.8	16.4	66	11.7	1.17	20	4.84	56.2	104.2	242.7	52.7	0.068	0.234	1.271
VAB19 - 5	7.1	core	520	27	87	7.8	34	8.2	1.42	22.5	7.07	87.4	188.8	457	110.2	0.041	0.320	1.470
VAB19 - 3	3.1	core	15.4	13.7	54	3.5	14.2	6.18	1.53	28.3	9.44	118.4	234.4	528.5	121	0.044	0.354	1.912
RGM55 - 1	13.2	sector-zoned	14	33.4	108	8.8	44	15	0.42	41	11.29	116.9	183.3	355.6	75.9	0.095	0.052	1.545
RGM55 - 4	12.1	sector-zoned	11.2	-	28.37	0.093	2.41	5.3	0.132	26.7	8.99	109.9	194.4	389.5	84.1	0.057	0.034	-
RGM55 - 12	1.1	sector-zoned	10.1	0.025	24.2	0.092	2	4.49	0.143	28.3	9.11	103	182.8	385	86.8	0.061	0.039	123.719
RGM55 - 15	-	sector-zoned	4.7	0.03	21.5	0.069	0.95	2.33	0.192	14.2	4.98	61.6	125.6	321	80.3	0.037	0.102	115.861
RGM55 - 17	10.1	sector-zoned	129	23	91	9.1	46	16.9	0.32	57	17.6	203	329	626	134.8	0.075	0.032	1.542
RGM55 - 5	11.1	high U	12.3	0.28	23.6	0.099	1.69	3.04	0.161	17	6.06	69.3	130.3	313.3	78.6	0.045	0.069	34.754
RGM55 - 6	9.1	high U	18.9	8	50.8	2.39	13.2	12.5	0.204	49.7	15.01	170.7	281	523	111.8	0.078	0.025	2.848
RGM55 - 2	14.1	high U	9.3	1.35	37.15	0.69	12.5	14.9	0.65	54.7	15.3	157.1	240.8	460.3	97.5	0.098	0.070	9.437
RGM55 - 7	8.1	high U	6	0.43	16.3	0.061	1.41	2.07	0.146	16.6	6.48	82	164.9	398.9	100.7	0.034	0.076	24.676
RGM55 - 8	7.1	high U	12.1	0.89	20.5	0.3	2.35	3.1	0.143	18.5	6.41	75.4	144.4	356	85.9	0.043	0.058	9.727
RGM55 - 10	3.1	high U	15.3	1.24	36.9	0.61	5.22	6.5	0.4	34.5	9.78	104.9	167.9	320.6	68.2	0.089	0.082	10.402
RGM55 - 11	4.1	high U	7	0.03	17.62	0.056	1.45	2.14	0.127	15.4	5.22	67.3	130.4	324.2	79.1	0.039	0.068	105.399
RGM55 - 14	5.1	high U	11.7	0.356	40	0.285	1.91	3.48	0.121	19.4	5.79	69.7	133.3	321.2	79	0.050	0.045	30.789
RGM55 - 18	-	high U	8.2	0.249	23.5	0.221	3.4	6.3	0.239	29.9	9.5	109	188	419	100.1	0.059	0.053	24.562
RGM55 - 20	2.1	high U	11.8	148	252	32.5	120	22.5	0.52	47.5	13.1	133	222	434	99	0.090	0.049	0.891
RGM55 - 3	-	core	15.8	0.041	40	0.373	7.51	13.2	0.5	59.6	16.5	181	279	515	108.2	0.096	0.055	79.305
RGM55 - 16	-	core	12.3	0.02	29.7	0.161	2.95	7.9	0.161	45.8	15.19	182.3	305	582	124.4	0.065	0.026	128.326
RGM55 - 13	-	core	5.1	0.028	34.3	0.212	5.07	10.4	0.245	55.3	15.94	176.1	281.3	508	105.4	0.090	0.031	109.152
RGM55 - 9	-	core	7.5	2.79	24.2	0.47	3.7	5.39	0.27	26.1	8.23	95.8	168.8	370.2	90.6	0.058	0.070	5.181
RGM55 - 19	-	core	11.2	5.2	44.9	1.39	15.5	15.7	0.65	63.4	18.86	201.2	327.8	619	136.5	0.085	0.063	4.095

Eu/Eu* = $\text{Eu}_N/(\text{Sm}_N * \text{Gd}_N)^{0.5}$ e Ce/Ce* = $\text{Ce}_N/(\text{La}_N * \text{Pr}_N)^{0.5}$

Análise de elementos traço em zircão (charnockito - Suíte Paraguacú).

Sample/spot	SHRIMP ID	Ti	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Er	Yb	Lu	(Gd/Yb) _N	Eu/Eu*	Ce/Ce*
RGM02 - 1	6.2	10.4	9.5	34.6	3.51	18.6	8.8	1.93	21.4	7.39	90	163	369	76.8	0.048	1.469	0.430
RGM02 - 2	4.2	430	52	127	11	44	10.9	1.82	18.8	5.79	60.2	105.5	236	48.4	0.066	1.302	0.389
RGM02 - 3	3.1	8.6	0.77	23.4	0.4	6.7	9.4	1.07	38.1	12.1	128	200	376	76.4	0.084	10.338	0.173
RGM02 - 4	9.2	12.9	0.105	21.4	0.53	6.8	11.9	0.71	43.5	12.9	147	232	432	84.5	0.083	22.242	0.096
RGM02 - 5	2.2	12.3	0.76	30.2	0.97	11.2	16.6	1.4	55.8	16.8	188	279	510	99.7	0.090	8.624	0.141
RGM02 - 6	2.1	11.5	0.48	36.8	1.27	17.3	22.8	2.02	92.3	26.7	286	411	709	135.7	0.107	11.556	0.135
RGM02 - 7	-	20.7	0.82	35.9	1.3	20.2	25	1.74	96	27.5	301	465	850	167.4	0.093	8.525	0.109
RGM02 - 8	-	110	1.25	26.4	0.79	10.1	14.1	1.34	57	17.5	194	305	584	116	0.081	6.514	0.145
RGM02 - 9	11.1	11.2	0.41	22.78	0.322	3.8	5	1	23.4	7.4	85	148	322	68.7	0.060	15.372	0.283
RGM02 - 10	12.1	8.4	1.01	41.5	0.67	8.9	13.1	1.25	56.2	17.7	198	300	548	109.5	0.085	12.369	0.141
RGM02 - 11	10.1	218	8.3	36	3	16.2	10.5	1.1	36.7	10.23	114.3	179.8	354.8	71.5	0.085	1.769	0.172
RGM04 - 1	11.1	9.5	0.197	51	1.01	16.3	28.2	1.4	143.8	44.5	498	763	1258	240	0.094	28.032	0.067
RGM04 - 2	10.2	9.9	1.04	21.3	1.51	17	23.2	2.21	98.9	26.84	277.2	383.1	637	121.5	0.128	4.167	0.141
RGM04 - 3	-	6.2	0.15	24.2	0.71	13.1	18.9	1.72	80.8	24.5	255	366	617	119	0.108	18.181	0.135
RGM04 - 4	-	57	5.6	48.8	2.94	26.6	28.9	1.41	137.3	41.5	445	648	1083	204.4	0.105	2.949	0.069
RGM04 - 5	2.2	11.7	10	63	4.2	30	24.5	1.62	107	33.2	373	587	1038	199	0.085	2.383	0.097
RGM04 - 6	1.1	27.5	4.8	59.6	2.66	20.9	26.8	1.75	129	42.8	491	846	1640	327	0.065	4.089	0.091
RGM04 - 7	8.2	28	6.4	37.6	2.28	15	11	0.93	43.8	13.52	147.4	234	425	87.1	0.085	2.413	0.130
RGM04 - 8	5.1	7.8	0.99	30	1.28	16.4	25.1	1.37	114.8	35.8	375	539	906	173.1	0.105	6.534	0.078
RGM04 - 9	6.1	7.7	16.8	60	5.6	34.6	22	1.4	84.3	23.3	257	366	617	119.3	0.113	1.517	0.100
RGM04 - 10	4.1	9.7	0	9.91	0.099	1.42	3.37	0.29	17.3	6.23	79.3	170.3	364	77	0.039	-	0.116
RGM53 - 1	6.1	7.4	0.051	16.23	0.131	1.65	3.4	0.55	16.9	5.54	67.2	118.8	258.8	53.4	0.054	48.684	0.222
RGM53 - 2	15.1	20.9	39	86	8.8	40	15.2	1.71	44	12.5	126	182	323	65.5	0.112	1.138	0.202
RGM53 - 3	-	8.5	24.7	92.1	6.97	31.4	15	1.25	57.4	16.95	195.1	301	527	104.2	0.090	1.721	0.130
RGM53 - 4	2.1	17.6	11.5	42.8	3.66	25.2	21.2	2.72	77.4	21.4	222.3	303.2	495	97	0.129	1.617	0.206
RGM53 - 5	10.1	17.5	0.022	16.57	0.33	8.1	15.3	1.99	64.8	18.8	192	268	456	87.8	0.117	47.680	0.193
RGM53 - 6	9.1	2.6	-	16.66	0.157	2.81	4.26	0.64	19.2	6.12	67	109.1	224.1	47.8	0.071	-	0.217
RGM53 - 7	1.1	11	0.132	56.1	1.24	18.5	31.1	1.29	157.4	49.7	549.1	825	1378	260.8	0.094	33.998	0.056
RGM53 - 8	8.1	16.5	0.016	21.71	0.142	3.38	6.49	0.37	32.2	10.29	111.8	185	342.7	70.5	0.078	111.671	0.078
RGM53 - 9	13.1	18.3	0.63	38.6	0.53	10.4	16.6	1.39	67.4	20	219	322	551	108.5	0.101	16.378	0.127
RGM53 - 10	12.1	21.4	57	153	18	84	25.3	2.85	65.2	16.6	168.1	237.2	409	78.9	0.132	1.171	0.215
RGM53 - 11	5.1	17.3	11.1	41.8	3.5	21.6	18.1	2.11	64.7	18.4	190.8	265.3	440.8	86	0.121	1.644	0.189
RGM57 - 1	6.1	14.5	43	65.1	10.7	68	26.5	5.4	35.2	8.47	82.6	123.4	240.1	48.8	0.121	0.744	0.541
RGM57 - 2	5.1	14.3	4.9	235.8	3.19	26.4	30.3	5.18	126	40.5	468	789	1485	302.1	0.070	14.623	0.257
RGM57 - 3	13.1	10.4	-	37.71	0.091	1.91	2.54	0.97	14.9	5.39	60.9	100.9	213.9	45.04	0.057	-	0.483
RGM57 - 4	7.1	16.6	3.47	196.5	5.5	47.5	30.2	5.06	84	26.34	301.6	484.8	897	180.3	0.077	11.028	0.308
RGM57 - 5	12.1	8	6.4	176.6	3.63	28.5	25.8	4.11	113.5	39.3	461	756	1358	261.8	0.069	8.983	0.232
RGM57 - 6	1.1	17.3	-	39.1	0.091	1.54	2.88	0.93	15.8	4.72	55.5	93.9	191.7	40.4	0.068	-	0.422
RGM57 - 7	2.1	6.7	-	48.1	0.064	1.66	2.87	0.74	14.3	4.66	54.1	97.8	226.7	53.1	0.052	-	0.354
RGM57 - 8	3.2	11.8	0.55	85.4	0.62	6	7.8	1.53	32.9	10.56	112.7	191	448	104.3	0.061	35.856	0.292
RGM57 - 9	9.1	138	97	198	11.6	52	21.2	4.43	60.9	17.9	192	303	575	119.7	0.087	1.447	0.377

Eu/Eu* = $\text{Eu}_N / (\text{Sm}_N * \text{Gd}_N)^{0.5}$ e Ce/Ce* = $\text{Ce}_N / (\text{La}_N * \text{Pr}_N)^{0.5}$

ANEXO XI – ANÁLISES ISOTÓPICAS (Lu-Hf) EM ZIRCÃO VIA LASS-ICP-MS

Análise isotópica Lu-Hf em zircão (Granulitos - Nappe Socorro-Guaxupé) * Datações via IASS-ICP-MS, as demais são utilizadas as datações via SHRIMP como referência.

Sample/Spot_ ID	SHRIMP ID	Domain	$^{176}\text{Lu}/^{177}\text{Lu}$	2σ	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	ϵHf (o)	$^{207}\text{Pb}/^{206}\text{Pb}$ age (Ma)	$^{176}\text{Hf}/^{177}\text{Hf}$ (o)	ϵHf (o)	TDMc (Ma)
VAB01 - 1	1.1	soccer ball	0.282288	0.000034	0.00001002	0.0000003	-17.58	599	0.282288	-4.22	1799.2
VAB01 - 2	3.1	soccer ball	0.282314	0.000025	0.00002468	0.0000003	-16.66	599	0.282314	-3.31	1741.3
VAB01 - 3	2.1	soccer ball	0.282297	0.000030	0.00002277	0.0000003	-17.26	622	0.282297	-3.39	1765.0
VAB01 - 4	13.1	soccer ball	0.282268	0.000036	0.00000916	0.0000003	-18.28	586	0.282268	-5.22	1852.0
VAB01 - 6	5.1	soccer ball	0.282311	0.000033	0.00002820	0.0000010	-16.76	578	0.282311	-3.89	1761.2
VAB01 - 7	11.1	soccer ball	0.282246	0.000028	0.00005430	0.0000027	-19.06	628	0.282245	-5.08	1876.3
VAB01 - 8	7.1	soccer ball	0.282277	0.000030	0.00002600	0.0000013	-17.96	561	0.282277	-5.47	1847.8
VAB01 - 9	8.1	soccer ball	0.282287	0.000032	0.00006500	0.0000300	-17.61	626	0.282286	-3.67	1786.1
VAB01 - 10	9.1	soccer ball	0.282272	0.000029	0.00002253	0.0000003	-18.14	592	0.282272	-4.95	1839.7
VAB01 - 11	10.1	soccer ball	0.282274	0.000031	0.00000902	0.0000003	-18.07	628	0.282274	-4.07	1812.5
VAB01 - 13	-	soccer ball	0.282272	0.000034	0.00004100	0.0000043	-18.14	628*	0.282272	-4.15	1817.8
VAB01 - 14	16.1	soccer ball	0.282234	0.000027	0.00059470	0.0000050	-19.48	611	0.282227	-6.10	1927.5
VAB01 - 18	6.1	rim	0.282254	0.000026	0.00032200	0.0000170	-18.78	578	0.282251	-6.02	1895.8
VAB01 - 5	4.1	rim	0.282259	0.000032	0.00024200	0.0000400	-18.60	592	0.282256	-5.50	1874.2
VAB01 - 19	-	rim	0.282262	0.000026	0.00007980	0.0000029	-18.49	624*	0.282261	-4.61	1843.7
VAB01 - 16	18.1	core	0.282333	0.000035	0.00059400	0.0000230	-15.98	689	0.282325	-0.87	1658.8
VAB01 - 15	17.1	core	0.282263	0.000033	0.00100200	0.00000510	-18.46	663	0.282251	-4.11	1842.9
VAB01 - 17	19.1	core	0.282265	0.000030	0.00190800	0.00000920	-18.39	649	0.282242	-4.73	1871.3
VAB01 - 12	15.1	core	0.282235	0.000029	0.00091600	0.0000180	-19.45	702	0.282223	-4.21	1880.3
VAB19 - 1	5.1	soccer ball	0.282276	0.000032	0.00040600	0.0000290	-18.00	600	0.282271	-4.78	1835.4
VAB19 - 2	4.1	soccer ball	0.282255	0.000031	0.00006060	0.0000004	-18.74	607	0.282255	-5.21	1868.0
VAB19 - 4	2.1	soccer ball	0.282293	0.000028	0.00011997	0.0000007	-17.40	596	0.282293	-4.12	1790.1
VAB19 - 8	1.1	soccer ball	0.282254	0.000026	0.00000355	0.0000003	-18.78	577	0.282254	-5.92	1888.7
VAB19 - 10	10.1	soccer ball	0.282261	0.000023	0.00051500	0.0000450	-18.53	601	0.282255	-5.34	1871.1
VAB19 - 9	8.1	rim	0.281966	0.000033	0.00035850	0.0000015	-28.96	594	0.281962	-15.87	2526.7
VAB19 - 5	7.1	core	0.282249	0.000036	0.00077400	0.0000250	-18.95	685	0.282239	-4.02	1854.9
VAB19 - 3	3.1	core	0.282242	0.000028	0.00086000	0.0000170	-19.20	697	0.282231	-4.04	1865.9
RGM55 - 1	13.2	sector-zoned	0.282189	0.000034	0.00069980	0.0000051	-21.08	624	0.282181	-7.45	2022.9
RGM55 - 4	12.1	sector-zoned	0.282174	0.000033	0.00068250	0.0000037	-21.61	625	0.282166	-7.96	2055.3
RGM55 - 12	1.1	sector-zoned	0.282162	0.000031	0.00070550	0.0000073	-22.03	631	0.282154	-8.26	2079.2
RGM55 - 15	-	sector-zoned	0.282189	0.000027	0.00064050	0.0000027	-21.08	628*	0.282181	-7.34	2019.0
RGM55 - 17	10.1	sector-zoned	0.282162	0.000037	0.00104900	0.0000640	-22.03	624	0.282150	-8.55	2092.2
RGM55 - 5	11.1	high U	0.282196	0.000027	0.00062850	0.0000056	-20.83	604	0.282189	-7.62	2017.3
RGM55 - 6	9.1	high U	0.282201	0.000034	0.00088000	0.0000160	-20.65	615	0.282191	-7.30	2006.1
RGM55 - 2	14.1	high U	0.282191	0.000037	0.00090500	0.0000160	-21.01	614	0.282181	-7.69	2029.6
RGM55 - 7	8.1	high U	0.282190	0.000036	0.00075800	0.0000170	-21.04	618	0.282181	-7.57	2025.8
RGM55 - 8	7.1	high U	0.282208	0.000034	0.00064600	0.0000150	-20.40	611	0.282201	-7.04	1986.8
RGM55 - 10	3.1	high U	0.282164	0.000031	0.00063560	0.0000062	-21.96	612	0.282157	-8.58	2084.1
RGM55 - 11	4.1	high U	0.282151	0.000027	0.00060380	0.0000067	-22.42	615	0.282144	-8.96	2110.4
RGM55 - 14	5.1	high U	0.282187	0.000030	0.00061291	0.0000010	-21.15	594	0.282180	-8.15	2042.9
RGM55 - 18	-	high U	0.282173	0.000032	0.00082900	0.0000470	-21.64	597*	0.282164	-8.66	2077.7
RGM55 - 20	2.1	high U	0.282142	0.000034	0.00080100	0.0000620	-22.74	599	0.282133	-9.71	2144.8
RGM55 - 3	-	core	0.282159	0.000036	0.00093000	0.0000460	-22.14	654*	0.282148	-7.96	2078.5
RGM55 - 16	-	core	0.282169	0.000037	0.00094400	0.0000280	-21.78	661*	0.282157	-7.45	2052.5
RGM55 - 13	-	core	0.282145	0.000031	0.00089500	0.0000190	-22.63	652*	0.282134	-8.48	2109.9
RGM55 - 9	-	core	0.282183	0.000039	0.00068900	0.0000160	-21.29	640*	0.282175	-7.31	2026.6
RGM55 - 19	-	core	0.282221	0.000045	0.00115700	0.0000440	-19.94	631*	0.282207	-6.36	1959.5

Análise isotópica Lu-Hf em zircão (charnockitos - Suíte Paraguaçu) * Datações via LASS-ICP-MS, as demais são utilizadas as datações via SHRIMP como referência.

Sample/Spot_ ID	SHRIMP ID	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	$^{176}\text{Lu}/^{177}\text{Hf}$	2σ	$\epsilon\text{Hf}_{(0)}$	$^{207}\text{Pb}/^{206}\text{Pb}$ age (Ma)	$^{176}\text{Hf}/^{177}\text{Hf}_{(0)}$	$^{176}\text{Hf}/^{177}\text{Hf}_{\text{CHUR}}$	$\epsilon\text{Hf}_{(0)}$	TDMc (Ma)
RGM02 - 1	6.2	0.282016	0.000037	0.0005340	0.0000270	-27.19	629	0.282010	0.282388	-13.40	2400.1
RGM02 - 2	4.2	0.281988	0.000032	0.0003820	0.0000200	-28.18	616	0.281984	0.282396	-14.62	2465.8
RGM02 - 3	3.1	0.282019	0.000030	0.0006080	0.0000340	-27.09	599	0.282012	0.282407	-13.99	2412.8
RGM02 - 4	9.2	0.281969	0.000030	0.0007900	0.0001600	-28.86	620	0.281960	0.282394	-15.37	2515.9
RGM02 - 5	2.2	0.281998	0.000038	0.0007570	0.0000550	-27.83	626	0.281989	0.282390	-14.20	2447.5
RGM02 - 6	2.1	0.281984	0.000034	0.0012900	0.0001000	-28.33	627	0.281969	0.282389	-14.89	2491.8
RGM02 - 7	-	0.282025	0.000045	0.0015100	0.0000610	-26.88	642*	0.282007	0.282380	-13.21	2398.7
RGM02 - 8	-	0.281995	0.000044	0.0010800	0.0001200	-27.94	620*	0.281982	0.282394	-14.57	2465.9
RGM02 - 9	11.1	0.281999	0.000041	0.0005750	0.0000460	-27.79	611	0.281992	0.282400	-14.42	2449.3
RGM02 - 10	12.1	0.281975	0.000034	0.0010490	0.0000980	-28.64	639	0.281962	0.282382	-14.85	2498.7
RGM02 - 11	10.1	0.282004	0.000039	0.0005550	0.0000160	-27.62	605	0.281998	0.282403	-14.36	2441.2
RGM04 - 1	11.1	0.281986	0.000027	0.0018180	0.0000650	-28.25	624	0.281965	0.282391	-15.11	2502.7
RGM04 - 2	10.2	0.282054	0.000026	0.0010940	0.0000240	-25.85	598	0.282042	0.282408	-12.96	2348.0
RGM04 - 3	-	0.282036	0.000037	0.0010440	0.0000710	-26.49	640*	0.282023	0.282381	-12.67	2363.0
RGM04 - 4	-	0.282015	0.000032	0.0016730	0.0000350	-27.23	644*	0.281995	0.282379	-13.59	2424.1
RGM04 - 5	2.2	0.281967	0.000027	0.0014950	0.0000790	-28.93	624	0.281949	0.282391	-15.64	2536.3
RGM04 - 6	1.1	0.281975	0.000033	0.0022700	0.0001200	-28.64	636	0.281948	0.282384	-15.43	2532.6
RGM04 - 7	8.2	0.282029	0.000035	0.0007630	0.0000260	-26.73	611	0.282020	0.282400	-13.43	2387.7
RGM04 - 8	5.1	0.281993	0.000036	0.0013950	0.0000760	-28.01	634	0.281976	0.282385	-14.47	2470.8
RGM04 - 9	6.1	0.282013	0.000030	0.0009750	0.0000420	-27.30	633	0.282001	0.282386	-13.60	2416.0
RGM04 - 10	4.1	0.282022	0.000032	0.0007180	0.0000210	-26.98	630	0.282014	0.282387	-13.24	2391.1
RGM53 - 1	6.1	0.281973	0.000028	0.0004355	0.0000054	-28.71	624	0.281968	0.282391	-14.99	2495.6
RGM53 - 2	15.1	0.282041	0.000045	0.0005160	0.0000950	-26.31	626	0.282035	0.282390	-12.57	2346.0
RGM53 - 3	-	0.282037	0.000034	0.0007870	0.0000270	-26.45	611*	0.282028	0.282400	-13.16	2370.6
RGM53 - 4	2.1	0.281997	0.000043	0.0008959	0.0000074	-27.87	610	0.281987	0.282400	-14.64	2462.4
RGM53 - 5	10.1	0.281986	0.000035	0.0007580	0.0000250	-28.25	628	0.281977	0.282389	-14.58	2472.9
RGM53 - 6	9.1	0.282026	0.000031	0.0004150	0.0000130	-26.84	612	0.282021	0.282399	-13.37	2384.9
RGM53 - 7	1.1	0.282008	0.000027	0.0020000	0.0000068	-27.48	623	0.281985	0.282392	-14.42	2459.3
RGM53 - 8	8.1	0.282012	0.000030	0.0005380	0.0000036	-27.34	623	0.282006	0.282392	-13.68	2412.6
RGM53 - 9	13.1	0.282012	0.000040	0.0009200	0.0000730	-27.34	612	0.282001	0.282399	-14.07	2428.7
RGM53 - 10	12.1	0.282018	0.000048	0.0007490	0.0000180	-27.12	605	0.282009	0.282403	-13.95	2415.1
RGM53 - 11	5.1	0.281955	0.000035	0.0008064	0.0000079	-29.35	610	0.281946	0.282400	-16.09	2552.9
RGM57 - 1	6.1	0.28213	0.000037	0.0003905	0.0000053	-23.16	599	0.282126	0.282407	-9.97	2161.2
RGM57 - 2	5.1	0.282123	0.000035	0.0022910	0.0000270	-23.41	619	0.282096	0.282394	-10.56	2214.0
RGM57 - 3	13.1	0.282131	0.000030	0.0003353	0.0000024	-23.13	604	0.282127	0.282404	-9.80	2154.6
RGM57 - 4	7.1	0.282172	0.000033	0.0013700	0.0000250	-21.68	611	0.282156	0.282400	-8.61	2085.6
RGM57 - 5	12.1	0.282126	0.000021	0.0018590	0.0000590	-23.30	613	0.282105	0.282398	-10.40	2199.3
RGM57 - 6	1.1	0.282131	0.000033	0.0003158	0.0000023	-23.13	605	0.282127	0.282403	-9.77	2153.5
RGM57 - 7	2.1	0.282172	0.000031	0.0003786	0.0000056	-21.68	588	0.282168	0.282414	-8.72	2074.0
RGM57 - 8	3.2	0.28218	0.000032	0.0007930	0.0000190	-21.39	614	0.282171	0.282398	-8.03	2051.3
RGM57 - 9	9.1	0.282178	0.000045	0.0012440	0.0000140	-21.47	605	0.282164	0.282403	-8.48	2072.4

Parâmetros utilizados nos cálculos de isótopos de Lu-Hf.

Constante de decaimento	1,867x 10 ⁻¹¹	(Soderlund <i>et al.</i> , 2004)
¹⁷⁶ Lu/ ¹⁷⁷ Hf CHUR	0,0336	(Bouvier <i>et al.</i> , 2008)
¹⁷⁶ Lu/ ¹⁷⁷ Hf DM	0,0384	(Griffin <i>et al.</i> , 2004)
¹⁷⁶ Lu/ ¹⁷⁷ Hf Bulk Earth	0,015	(Griffin <i>et al.</i> , 2004)
¹⁷⁶ Hf/ ¹⁷⁷ Hf CHUR	0,282785	(Bouvier <i>et al.</i> , 2008)
¹⁷⁶ Hf/ ¹⁷⁷ Hf DM	0,28325	(Griffin <i>et al.</i> , 2004)

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