UNIVERSIDADE DE SÃO PAULO

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**EVOLUÇÃO TECTÔNICA DA MARGEM ATIVA
NEOPROTEROZÓICA DO ORÓGENO GONDWANA OESTE NA
PROVÍNCIA BORBOREMA (NE-BRASIL)**

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Resumo

O Orógeno Gondwana Oeste (OGO), sugerido nesta Tese, é uma faixa linear que se estende por mais de 5000 km do atual noroeste da África ao Brasil Central. Perspectivas modernas e tradicionais sugerem que este Orógeno resultou do consumo e fechamento do Oceano Goiás-Faruriano que culminou na colisão continente-continente envolvendo os crátons Amazônico/Oeste Africano contra os crátons São Francisco/Congo e Saariano. Nesta Tese foi investigada a evolução crustal de um importante setor do OGO no noroeste da Província Borborema (NE-Brasil), assim como em outras áreas no Togo e Mali, utilizando-se de uma estratégia coerente. A geocronologia U-Pb em zircões de rochas siliciclásticas e ígneas indicam a existência de uma margem convergente de longa duração (de até 350 m.y.) associada ao consumo do Oceano Goiás-Faruriano. Na área investigada, particularmente no NE do Brasil no Domínio Ceará Central, o magmatismo associado a esta convergência pode ser dividido em três principais períodos: i) um período inicial constituído essencialmente por magmatismo de arco juvenil em ca. 880-800 Ma que continua até 650 Ma, como evidenciado indiretamente por zircões detríticos de depósitos sin-orogênicos, ii) um estágio de arco maduro em ca. 660-630 Ma caracterizado por rochas magmáticas híbridas entre crosta e manto e iii) anatexia crustal em 625-618 Ma que continua até 600 Ma. A longa evolução do OGO e o sincronismo do metamorfismo de ultra-alta pressão UAP no NE do Brasil, Togo e Mali indicam que a subducção continental e, portanto, a colisão continental ocorreu simultaneamente por pelo menos 2500 km durante o período Ediacariano (620-610 Ma). Nesta Tese propõem-se que o desenvolvimento da Província Borborema entre 620 a 570 Ma resultou da interação de dois eventos colisionais distintos. A Colisão I ao longo do OGO ocorreu há ca. 620-610 Ma como resultado da colisão entre o Bloco Parnaíba, como parte dos crátons Amazônico/Oeste Africano, contra o embasamento antigo da Província a leste. A zona de sutura associada a esta colisão foi reativada por uma zona transformante dextral (o Lineamento Transbrasiliano), permitindo a aproximação da Província Borborema e sua subsequente colisão contra o cráton São Francisco em ca. 590-580 Ma, marcando a Colisão II ao longo do Orógeno Sergipano. A interação dos esforços de leste relacionados com a Colisão I com os esforços de norte relacionados a endentação cratônica em uma litosfera espessa gerou uma trama extensa de zonas de cisalhamento direcionais que contribuíram para a extrusão da Província na direção nordeste. Nesta Tese, também foi considerado que o sincronismo do metamorfismo UAP (620-610 Ma) por pelo menos 2500 km no OGO registra o primeiro indício de subducção continental na escala do Orógeno Himalaiano e o consequente surgimento de Megamontanhas no registro geológico. A formação destas Megamontanhas ca. 40 m.y. antes da explosão da Vida no período Ediacariano tem a idade ideal para providenciar, por meio da erosão, os sedimentos (nutrientes) que são considerados necessários para a evolução da Vida. Finalmente, a evolução do OGO indica que o supercontinente Gondawa já estava formado desde o Oeste da África ao Brasil Central, dificultando a hipótese da chegada tardia do cráton Amazônico/Oeste Africano e assim a existência de um amplo Oceano Cambriano como previamente proposto.

Abstract

The Neoproterozoic West Gondwana Orogen (WGO), suggested in this Thesis, is a linear belt that extended for more than 5000 km from nowadays northwest Africa to Central Brazil. Traditional views suggest that this orogen resulted from the consumption and closure of the Goiás-Pharusian Ocean that culminated in a continent-continent collision involving mainly the conjoined Amazon and West African cratons against the São Francisco-Congo and Saharan cratons. In this Thesis it has been investigated the crustal evolution of an important sector of the WGO in NE-Brazil at the northwestern portion of the Borborema Province as well as in some areas of Africa in Togo and Mali using a coherent approach. U-Pb Zircon geochronology of siliciclastic and igneous rocks indicates a long-lived convergent tectonics (up to 400 m.y) related to the consumption of the Goiás-Pharusian Ocean. In the studied, particularly in NE- Brazil in the Ceará Central Domain, convergent-related magmatism can be divided into three main periods: i) an early period comprising essentially juvenile arc magmatism at ca. 880-800 Ma and continuing to 650 Ma as evidenced indirectly by detrital zircons from syn-orogenic deposits, ii) a more mature arc period at ca. 660-630 Ma characterized by hybrid mantle-crustal magmatic rocks, and iii) crustal anatexis at 625-618 Ma continuing until ca. 600 Ma. Protracted tectonic evolution in the WGO and synchronicity of UHP metamorphism in NE-Brazil, Togo and Mali indicate that continental subduction, and hence continental collision, occurred simultaneously over at least 2500 km during the Ediacaran period (620-610 Ma). Here, it has been proposed that Borborema Province development from 620 to 570 Ma resulted from two discrete collisional events. Collision I, along the WGO, took place at ca. 620-610 Ma as the result of collision between the Parnaíba Block, as the forefront of the much larger Amazonian-West Africa Craton, and the old basement of the Borborema Province. The suture zone related to this collision was reactivated by a dextral transform zone (the Transbrasiliano Lineament), allowing the Borborema Province to approach and collide against the São Francisco Craton in the south at ca. 590-580 Ma marking collision II along the Sergipano Orogen. The combined stresses related to eastward push from collision I and northward push from the cratonic indentation onto a thickened lithosphere gave rise to an extensive network of strike-slip shear zones across the Province forcing its northeastward extrusion. It has been also considered here that synchronicity of UHP metamorphism (620-610 Ma) over at least 2500 km in WGO records the first Himalayan-scale deep-continental subduction and the consequent appearance of Megamountains in the geological record. The formation of these Megamountains ca. 40 m.y. before the explosion of Life in the Late Ediacaran is perfectly timed to deliver by erosion the sediments (nutrients) that have been deemed necessary for Life evolution. Finally, evolution of the WGO indicate that the Gondwana supercontinent was already assembled from West Africa to Central Brazil, precluding the late arrival of the Amazon-West African craton and hence the existence of a large Cambrian Ocean as proposed earlier.

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1. Introdução

A presente Tese de Doutorado foi realizada no âmbito do Programa de Pós-Graduação em Geoquímica e Geotectônica do Instituto de Geociências da Universidade de São Paulo (IGc-USP). De forma geral, a temática da Tese foi inicialmente discutida com o intuito de entender a evolução crustal da área adjacente ao Lineamento Transbrasiliano no nordeste brasileiro, particularmente no norte do estado do Ceará. Durante a evolução da pesquisa sentiu-se a necessidade de expandir a abrangência do estudo em direção a contraparte africana no Togo, Benin e Mali no sentido de entender e correlacionar as rochas de alto grau (e.g. eclogitos) que ocorrem alinhadas ao longo de uma importante zona de sutura do supercontinente Gondwana.

O **Orógeno Gondwana Oeste** (**fig. 1.1**), sugerido nesta Tese, inclui todas as áreas orogênicas Pan-africanas/Brasilianas que ocorrem ao longo do Lineamento Transbrasiliano-Kandi, que se estende por mais de 5000 km lineares da Argélia ao Brasil Central. Este orógeno resultou do consumo e fechamento do Oceano Goiás-Farusiano (Caby, 2003; Kroener and Cordani, 2003), que culminou na colisão continental envolvendo os crátons Amazônico/Oeste África, São Francisco-Congo e o metacráton Saara. A ocorrência de elementos indicativos de zonas de subducção e colisão continental (e.g. ofiolitos; arcos intra-oceânicos, arcos continentais, metamorfismo de alta e ultra-alta pressão; prismas acrescionários) ao longo das adjacências do Lineamento Transbrasiliano-Kandi levou muitos pesquisadores a interpretar esta estrutura como uma grande zona sutural (e.g. Caby, 1989, Kroener & Cordani, 2003, Cordani et al., 2003a, Cordani et al., 2003b, Ganade de Araujo & Santos, 2008, entre outros), ou mais precisamente, como proposto aqui, uma estrutura desenvolvida dentro da dinâmica colisional Neoproterozóica no Orógeno Gondwana Oeste, semelhante ao sistema de falhas e zonas de cisalhamentos desenvolvidas durante a colisão e escape lateral entre a Índia e Eurásia no Cenozóico.

O Orógeno registra em toda sua extensão um longo período de convergência (> que 400 m.y) condizente com o desenvolvimento de diversos arcos intraoceânicos e continentais que são hoje preservados dentro da zona colisional fóssil profundamente erodida (Pimentel and Fuck, 1992; Caby, 2003; Berger et al., 2011; Ganade de Araujo et al., 2012). As principais assembleias petrotectônicas entre as regiões cratônicas envolvidas na colisão estão representadas por margens passivas, arcos juvenis, arcos continentais tardios e sequências supracrustais sin-orogênicas. Os detritos da erosão das montanhas resultantes da colisão no orógeno estão hoje depositados em bacias molássicas e do tipo *foreland*, e o final da atividade orogênicas é estabelecido em ca. 540-500 Ma com base nas idades dos granitóides pós-colisionais dispostos ao longo do orógeno (Affaton et al., 2000; Caby, 2003; Pimentel et al., 2011; Ganade de Araujo et al., 2012; Castro et al. 2012).

Desde a definição do termo “Pan-Africano” (Kennedy, 1964) para designar eventos termo-tectônicos de idade Neoproterozóica-Cambriana na África, e seu termo equivalente na América do Sul, o “Brasiliano”, inúmeros cinturões orogênicos foram batizados com nomes regionais. A faixa de idade Neoproterozóica que se estende ao longo do importante Lineamento Transbrasiliano-Kandi (Caby, 1989), da Argélia até o Brasil Central, foi segmentada em diversos setores, províncias e blocos tectônicos. Na África, a faixa Dahomey no Togo e Benin (e.g. Affaton et al., 1991), e Hoggar na Argélia e Mali (Caby, 2003) são geralmente agrupadas no Orógeno Trans-Saariano (Caby, 1989; Trompette, 1994). No Brasil, a Província Borborema (Brito Neves et al., 2000) na parte nordeste do país e a Província Tocantins (Pimentel et al., 2000) em sua porção Central são agrupadas nos Orógenos Brasilianos. Nesta Tese, todas as províncias geológicas mencionadas acima foram agrupadas em um só orógeno, batizado aqui de o **Orógeno Gondwana Oeste** (**fig 1.1**). No Brasil, o orógeno abarca o limite oeste da Província Borborema e parte norte da Faixa Brasília.

Há muito tempo, as correlações geológicas entre o nordeste brasileiro e noroeste africano tem sido utilizadas como fonte de informações na reconstrução da paleogeografia do Gondwana Oeste (Almeida & Black, 1968; Torquato & Cordani, 1981; Brito Neves et al., 2001; De Wit et al., 2008a). A correlação entre a Província Borborema e sua expressão na contraparte africana representada pelos domínios Trans-Saarianos, Nigerianos e pelas faixas orogênicas África Central-Oubanguides, no que Tromp alte (1994) designa de Província Nordeste do Brasil/Centro Oeste Africana têm sido alvo correspondente no exercício da correlação (Caby, 1989; Arthaud et al., 2008; Santos et al., 2008a).

Figura 1.1 – Principais crátons, blocos formadores e assembleias lito-tectônicas do Orógeno Godwana Oeste. O Orógeno, que se estende por mais de 5000 km da Argélia ao Brasil central desenvolvido a partir do consumo e fechamento do Oceano Goiás-Farusiano durante o Neoproterozóico entre 950-630 Ma, culminando na colisão continental das principais massas continentais em 620-610 Ma. As flechas negras indicam os nomes regionais dos setores orogênicos do Neoproterozóico, unificados aqui no Orógeno Gondwana Oeste. O retângulo laranja indica a posição da principal área de estudo na Província Borborema. As estrelas em laranja indicam a localização das amostras de rochas de alta-pressão investigadas nesta Tese.

O noroeste da Província Borborema, principal área de estudo desta Tese, configura uma região chave na evolução do Gondwana Oeste, no que De Wit et al. (2008b) denomina de um dos *piercing points* na correlação pré-deriva entre a África e Brasil. Localizada no nordeste do Brasil, a Província Borborema foi originalmente descrita e definida por Almeida et al. (1977,1981) como uma complexa região de domínios tectono-estratigráficos, fortemente afetada pelo(s) evento(s) tectônico(s) de idade Neoproterozóica. Compreendendo uma área de aproximadamente 450.000 km², esta província foi finalmente estruturada, por volta de 0.6-0.58 Ga, pela convergência dos crátons Amazônico-São Luiz-Oeste Africano e São Francisco-Congo, com a participação do bloco Parnaíba, na intitulada colagem Brasileira/Pan-Africana, que culminou na formação da parte oeste do supercontinente Gondwana (Trompette, 1994; Brito Neves & Cordani, 1991).

Por tratar-se de uma região chave dentro da dinâmica orogênica do Neoproterozóico a área de estudo foi amplamente investigada do ponto de vista geocronológico e isotópico para fornecer subsídios ao entendimento e suas respectivas correlações ao longo do Orógeno Gondwana Oeste.

1.1 Objetivos

O objetivo essencial da presente Tese foi entender os eventos tectônicos associados a evolução Neoproterozóica da margem do Orógeno Gondwana Oeste no nordeste do Brasil, bem como estabelecer uma ordem cronológica dos mesmos. A ferramenta escolhida para tal tarefa foi a Geologia Isotópica fortemente aliada com as relações observadas em campo. O sistema U-Th-Pb em zircão e as demais versatilidades petrológicas oferecidas por esta fase mineral foi o ponto mais explorado nesta Tese.

No nordeste do Brasil foram utilizados mapas em andamento ou já confeccionados na escala 1:100000, pelo Serviço Geológico do Brasil. Para a coleta das amostras foram realizadas diversas campanhas de campo no nordeste do Brasil, que contou com o apoio logístico do Serviço Geológico do Brasil. Para coleta de amostras em unidades equivalentes na África (especialmente relacionadas ao metamorfismo de alto grau, *p.e.* eclogitos), foi realizada uma campanha de campo de 20 dias no Togo e Benin. Ainda, nesta Tese algumas amostras provenientes da região do Gourma, no Mali, também foram investigadas. Essas amostras foram coletadas pelo pesquisador Renaud Caby em 1981 e cedidas ao doutorando durante sua estada na França na *Université de Montpellier II*. Vale ressaltar aqui, que na atual conjuntura política essas amostras são de grande valor científico, pois tratam-se de das rochas de ultra-alta pressão mais antigas da Terra (Janh et al., 2001), e a região onde afloram é hoje inacessível devido presença de conflitos armados.

As análises laboratoriais foram em grande parte realizadas no Centro de Pesquisas Geocronológicas (CPGeo) do IGC/USP. Uma outra leva de análises foi realizada na *Research School of Earth Sciences* da *Australian National University* em Camberra, na Austrália.

A forma de abordagem no noroeste da Província Borborema foi dividida em três grandes linhas principais que incluem: i) a sedimentação, ii) o magmatismo e iii) metamorfismo. Esta última linha com particular foco na determinação das idades do pico do metamorfismo de alta/ultra-alta pressão no Brasil e África.

i) *sedimentação*: dentro da linha da sedimentação o objetivo foi entender a proveniência, por meio da datação U-Pb dos zircões detríticos, que compõem as rochas sedimentares (hoje metamorizadas em graus diversos) associadas ao desenvolvimento da margem passiva do Oceano Goiás-Farusiano no nordeste do Brasil bem como bacias marginais relacionadas ao desenvolvimento de arcos Neoproterozóicos, e ainda aquelas associadas embasamento Paleoproterozóico. As principais unidades investigadas no noroeste da Província compõem parte das rochas siliciclásticas dos domínios Médio Coreá e Ceará Central.

ii) *magmatismo*: nesta linha o objetivo essencial foi o de estabelecer o *timing* e fontes das rochas magmáticas relacionadas ao consumo do Oceano Goiás-Farusiano no nordeste do Brasil, bem como o retrabalhamento das mesmas durante a colisão continental que se seguiu. Além dos aspectos geocronológicos U-Pb em zircão das diferentes linhagens magmáticas que compõem o magmatismo foram empregadas técnicas traçadoras de fontes (em rocha total), corriqueiramente utilizadas em estudos de petrologia de granitóides, tais como isótopos de Sr e Nd, combinados com aqueles obtidos diretamente dos zircões, tais como isótopos de Hf e O. A unidade litoestratigráfica investigada nesta linha restringiu-se ao Complexo ígneo-anatético Tamboril-Santa Quitéria localizado no Domínio Ceará Central da Província Borborema.

iii) *metamorfismo*: nesta linha, novamente a geocronologia U-Pb em zircão foi a principal ferramenta utilizada com o objetivo de estabelecer a idade do pico metamórfico de alta pressão e ultra-alta pressão em amostras de eclogitos ao longo do Orógeno Gondwana Oeste no nordeste do Brasil (zona eclogítica Forquilha definida por

Santos et al., 2009), Togo (eclogitos do Monte Lato, Agbessoumondé et al., 2001) e Mali (eclogitos de ultra-alta pressão mais antigos da Terra; Jahn et al., 2001). Ressalta-se que a geocronologia U-Pb em zircão foi aliada a composição química dos mesmos zircões no que se constitui em um moderno ramo da investigação geocronológica chamado de petrocronologia. Nesta linha foram investigadas ainda as condições de P e T aos quais esses eclogitos foram submetidos, por meio da química mineral das fases de pico metamórfico, difusibilidade de Zr em rutilo e Ti em zircão.

1.2 Estrutura da Tese

Com a euforia acadêmica do *publish or perish* (publique ou padeça) (*The Economist*, 19 de outubro de 2013) na busca de melhores índices pessoais e institucionais a publicação de artigos científicos vem sendo incentivada em todos os âmbitos acadêmicos, incluindo as apresentações de teses de doutoramento. Esta nova condição traz vantagens e desvantagens. A vantagem é de gerar teses concisas e objetivas, de certa forma impactantes. A busca por periódicos de alto impacto, em que os artigos são revisados (*peer-reviewed*) por dois a três especialistas, apuram a qualidade da Tese e testam a fidedignidade dos dados analíticos. A desvantagem é de que uma dada Tese pode terminar em um apanhado de artigos desconexos da temática central proposta pela pesquisa. Outra desvantagem é a repetição de itens que devem ser descritos em todos os artigos, como por exemplo o contexto geológico da pesquisa. Contudo, a principal desvantagem surge da pressão, pois dados científicos necessitam de tempo para serem digeridos apropriadamente.

Seguindo esta nova tendência a presente Tese está estruturada sob a forma de seis artigos científicos publicados ou em processo de submissão/revisão em revistas internacionais indexadas. Após aos quatro primeiros capítulos que versam sobre a apresentação da pesquisa, o zircão e suas aplicações, introdução sobre o contexto geológico e metodológico, os resultados são apresentados em seis capítulos que relatam suas implicações dentro do contexto do Domínio Ceará Central, da Província Borborema e do supercontinente Gondwana. Os dois últimos capítulos trazem novas perspectivas aliadas a ideias já tradicionais em relação a evolução Neoproterozóica do Gondwana Oeste no Brasil e África. As opiniões expressadas nestes dois últimos artigos foram fruto do trabalho em colaboração com outros pesquisadores levadas adiante pelo orientador desta Tese.

No capítulo 5, o artigo **“U-Pb detrital zircon provenance of metasedimentary rocks from the Ceará Central and Médio Coreau Domains, Borborema Province, NE-Brazil: Tectonic implications for a long-lived Neoproterozoic active continental margin”** apresenta dados de proveniência sedimentar das bacias de margem passiva e ativa associadas ao desenvolvimento do Orógeno Gondwana Oeste no Domínio Ceará Central da Província Borborema. O mesmo capítulo já encontra-se publicado no periódico internacional *Precambrian Research*.

No capítulo 6, que versa sobre o registro magmático relacionado ao consumo do Oceano Goiás-Farusiano o artigo **“Tracing Neoproterozoic subduction in the Borborema Province (NE- Brazil): clues from U-Pb geochronology and Sr-Nd-Hf-O isotopes on granitoids and migmatites”** traz informações geocronológicas e isotópicas do sistema de subducção relacionado à evolução da margem ativa Neoproterozóica do Orógeno Gondwana Oeste no Domínio Ceará Central. O presente capítulo encontra-se em revisão em no periódico científico internacional indexado *Lithos*.

No capítulo 7 são discutidas as implicações da colisão Ediacarana na Província Borborema por meio do artigo **“Extruding the Borborema Province: a two-stage collision process”**. Neste capítulo é proposto um novo modelo para evolução da Província Borborema com base em duas colisões distintas. No modelo, a interação destas

duas colisões em conjunto com a endentação do Craton São Francisco, no sul da Província, teriam levado ao desenvolvimento do sistema de zonas de cisalhamentos pelas quais a extrusão da Província teria ocorrido. O mesmo modelo interpreta o importante Lineamento Transbrasiliano como um limite de placas transformante durante o Neoproterozóico. O presente capítulo encontra-se publicado no periódico internacional *Terra Nova*.

O *capítulo 8* apresenta informações geocronológicas sobre o metamorfismo colisional ao longo do Orógeno Gondwana Oeste no artigo “**Ediacaran megamountains: evidence for >2500-km-long deep continental subduction in the West Gondwana Orogen**”. Neste capítulo é demonstrado que a colisão continental que procedeu em decorrência do fechamento do Oceano Goiás-Farusiano foi sincrônica ao longo de pelo menos 2.500 km, resultando numa cadeia montanhosa de magnitude equivalente aos Himalaias no período Ediacarano. As implicações desta extensa cadeia de montanhas para o desenvolvimento da Vida na Terra são também discutidas neste capítulo. O presente capítulo encontra-se em revisão em no periódico científico internacional indexado *Nature Communications*.

O *capítulo 9* traz uma ampla revisão do Orógeno Gondwana Oeste ao longo do corredor tectônico Transbrasiliano-Kandi com o artigo “**The significance of the Transbrasiliano-Kandi tectonic corridor for the amalgamation of the West Gondwana**”. Devido a larga extensão do Orógeno, este capítulo foi desenvolvido em parceria com outros colaboradores e a contribuição do doutorando restringiu-se a correlação geológica do nordeste brasileiro e noroeste africano (item 9.4 do mesmo capítulo). Este capítulo foi publicado no periódico *Brazilian Journal of Geology*.

O *capítulo 10* discute a existência de um proposto trato oceânico Cambriano (Oceano Clymene) na América do Sul com o artigo “**Is there a Clymene Ocean in Central South America**”. Neste capítulo é debatido a provável não existência do Oceano Clymene em detrimento da presença do vasto Oceano Goiás-Farusiano e seu fechamento no período Ediacarano. Assim como no capítulo anterior, o mesmo foi elaborado por uma junção de colaboradores e a participação do doutorando restringiu-se ao que tange a história geológica da Faixa Araguaia, Província Borborema e as conexões de ambas entidades com a África (item 10.5.2). Este capítulo encontra-se publicado no periódico internacional *American Journal of Science*.

Por fim, o ultimo capítulo reúne de forma concisa todos os resultados e interpretações apresentados nos capítulos anteriores, constituindo assim, a conclusão da Tese e as avenidas futuras pelo qual a pesquisa científica deverá ser guiada *a posteriori*.

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2. Zircão e sua versatilidade

Como um proeminente geocronômetro, o zircão (ZrSiO_4), exerceu, exerce e provavelmente continuará exercendo um papel singular no entendimento da evolução crustal da Terra. O seu amplo uso na geocronologia, baseado no decaimento de urânio (U) para chumbo (Pb), vem difundido este mineral como um verdadeiro “guardião do tempo” (Harley & Kelly, 2007). As modernas técnicas de análises sobre zircões, em conjunto com outra gama de minerais traços (*e.g.* monazita, badeleíta, titanita, rutilo) revelam idades de eventos de suma importância para a construção de modelos tectônicos sólidos, incluindo eventos de edificação de cadeias montanhosas, magmatismo, idades máximas de sedimentação, e indiretamente parâmetros temporais de construção e dispersão de grandes massas continentais ao longo do tempo (Rubato & Hermann, 2007; Harley et al., 2007; Sherer et al., 2007).

A estrutura do zircão é relativamente aberta e suas vacâncias podem abrigar impurezas de interesse geoquímico da ordem de partes por milhão (ppm) (Harley & Kelly, 2007; Hoskin & Schaltegger, 2003). O zircão pode incorporar muitos elementos externos tais como, P, Sc, Nb, Hf, Ti, Y, U, Th, e Elementos Terras Raras (ETR) via processos de substituição iônica acoplada, principalmente controlada pelos raios iônicos dos cátions Si^{+4} e Zr^{+4} (Hoskin & Schaltegger, 2003).

Além de seu versátil uso como um geocronômetro, recentes avanços nas técnicas de microanálise acabaram ampliando o espectro de problemas geológicos que podem ser abordados com o uso do zircão. Como uma fase que pode acomodar significativas concentrações de elementos traços (sensíveis a temperatura, pressão e processos da época de cristalização), o zircão pode fornecer informações importantes acerca dos processos que operaram na formação de um dado segmento da crosta (Hanchar & van Westrenen, 2007, Harley et al., 2007). Composições isotópicas de oxigênio em zircão trazem informações sobre a relação entre processos de baixa e alta temperatura e auxiliam na investigação da fonte de fundidos magmáticos na crosta (Valley, 2003). A utilização do parâmetro Épsilon Hf obtido diretamente no domínio datado do zircão também é um forte traçador da evolução e fontes de magmas (Hawkesworth & Kemp, 2006). Medidas dos conteúdos de U, Th e He no zircão podem ser utilizados para inferir temperaturas em que as rochas foram expostas a temperaturas próximas a da superfície atual, fornecendo importantes subsídios sobre a história de soerguimento e formação de relevo (Harley & Kelly, 2007). Uma lista das principais características químicas do zircão e suas aplicações pode ser encontrada na **tabela 2.1**.

2.1. Geocronologia U-Th-Pb em zircão

Três séries de decaimento radioativo são conhecidas para o sistema U-Th-Pb, que envolve os seguintes isótopos pais: ^{238}U , ^{235}U e ^{232}Th e seus respectivos isótopos radiogênicos filhos ^{206}Pb , ^{207}Pb e ^{208}Pb (Faure & Mensing, 2005). Cada uma dessas séries envolve uma gama de passos intermediários que geram isótopos intermediários de vida curta. Pelo fato do ultimo estágio do decaimento nessas séries ser muito mais devagar do que os estágios intermediários, o processo de decaimento pode ser matematicamente descrito por uma simples equação, relacionado o número de isótopos pais que restam (*e.g.* ^{238}U) e o número de final de isótopos radiogênicos filhos (*e.g.* $^{207}\text{Pb}^*$) com o tempo:

$$^{206}\text{Pb}^*/^{238}\text{U} = e^{\lambda^{238}t} - 1$$

Tabela 2.1 – Principais características químicas/isotópicas do zircão e suas aplicações, segundo Harley & Kelly (2007).

Propriedade física/química	Substituições	Principais aplicações	Comentários
U e Th	$(U^{+4}, Th^{+4})=Si^{+4}$	Geocronologia U-Pb	O diagrama concórdia pode ser usado na contabilização de isótopos de U e seu conteúdo de Pb radiogênico
He	Decaimento de U e Th	Determinação de taxas de exumação e evolução do relevo usando termocronologia U-Th-He	A cronometria de baixa T e baseada na temperatura de fechamento do zircão para a perda de He
HfO ₂ >3% eq. peso	$Hf^{+4}=Si^{+4}$	Informações sobre residência crustal e crescimento continental; traçador de fontes mantélicas e crustais	O ¹⁷⁶ Lu decai para ¹⁷⁶ Hf. A alta razão de ¹⁷⁶ Hf/ ¹⁷⁷ Hf mudou muito pouco com o tempo e pode ser usada para inferir fontes a partir de um modelo de referência para Terra
Ti>120 ppm	$Ti^{4+}=Si^{4+}$	Termocronologia	Maximizada quando o zircão esta em equilíbrio com o rutilo. Pode fornecer a T de cristalização do zircão ou a T de zircões metamórficos em equilíbrio com rutilo
Y>5000 ppm ETR total>2500 ppm	$(Y^{3+}, ETR^{3+})P^{5+}=Zr^{4+}Si^{4+}$	Reconstrução de histórias magmáticas; traçadores de fontes magmáticas; sintonia entre idades e reações minerais	Requer um grande conhecimento do coeficiente de partição entre os elementos e o zircão
Isótopos de O		Traçador de contribuição sedimentar e crosta na fonte de magmas; investigações sobre reciclagem crustal	O fracionamento de ¹⁸ O para ¹⁶ O ocorre em baixa temperatura. Variações de ¹⁸ O/ ¹⁶ O são utilizadas para discriminação de fontes

onde, e é uma função exponencial, t é o tempo e λ é a constante de decaimento específica para o esquema de decaimento (*p.e.* $\lambda^{238}U = 1,55125 \cdot 10^{-10}$; ²⁰⁶Pb* refere-se ao ²⁰⁶Pb radiogênico formado a partir do ²³⁸U).

No exemplo a seguir, modificado de (Harley & Kelly, 2007), considera-se uma fração de zircões formados durante a cristalização de um magma félsico. Por meio da incorporação de U e Th durante o crescimento magmático desses zircões, três diferentes “cronômetros” isotópicos são iniciados e cada um deles registrará a

história isotópica (ou tempo decorrido) de acordo com sua própria taxa de decaimento (o decaimento do ^{235}U ocorre cerca de sete vezes mais rápido do que o ^{238}U , enquanto o decaimento do ^{232}Th é muito lento).

Em princípio, podem-se obter idades a partir de qualquer um desses isótopos, medindo-se a razão isotópica apropriada e resolvendo a equação acima para o tempo t . Em um sistema ideal as três idades deveriam ser similares com valores concordantes, contudo em zircões reais não podemos assumir que U e Th estão igualmente “fechados” aos processos de recristalização posteriores, que perturbam o sistema isotópico. Igualmente é necessário corrigir o valor de chumbo presente antes de qualquer acúmulo de chumbo radiogênico.

A aplicação dessa sistemática na geocronologia considera os esquemas de decaimento do ^{235}U e ^{238}U em conjunto. Como a taxa moderna da razão $^{235}\text{U}/^{238}\text{U}$ é bem conhecida (1/137,88), não se faz necessária a determinação das razões provenientes dos decaimentos do ^{235}U e ^{238}U separadamente. Desta forma, podem ser usadas mutuamente razões $^{207}\text{Pb}^*/^{235}\text{U}$ e $^{206}\text{Pb}^*/^{238}\text{U}$ desde a formação do zircão investigado [*p.e.* $^{207}\text{Pb}^*/^{235}\text{U} = 137,88 (^{206}\text{Pb}^*/^{238}\text{U})$]. Esta é a base do *diagrama concórdia* desenvolvida por Wheterill (1956). A curva esboçada no diagrama concórdia representa os resultados compatíveis e concordantes das razões $^{207}\text{Pb}^*/^{235}\text{U}$ e $^{206}\text{Pb}^*/^{238}\text{U}$ desde a origem da Terra há 4,6 Ga até o presente.

A curva concórdia é o lugar geométrico onde mutuamente as razões $^{207}\text{Pb}^*/^{235}\text{U}$ e $^{206}\text{Pb}^*/^{238}\text{U}$ são concordantes. No tempo zero, quando o zircão do exemplo é formado, não existe nenhum chumbo radiogênico (Pb^*) nos zircões. Após 1,0 Ga essas razões terão valores 1,677 e 0,167 respectivamente (**curva a na fig. 2.1**). Pela mesma lógica se forem analisados zircões de uma rocha formada há 3,0 Ga esperaríamos que as razões $^{207}\text{Pb}^*/^{235}\text{U}$ e $^{206}\text{Pb}^*/^{238}\text{U}$ tivessem valores de 18,198 e 0,592 respectivamente (**curva b na fig. 2.1**). Essas são ditas idades *concordantes*, pois as duas razões Pb^*/U medidas no zircão correspondem à mesma idade, portanto a posição deste ponto no diagrama concórdia é a medida direta da idade do zircão.

O diagrama também permite conclusões de análises que não caem diretamente sobre a curva concórdia. Para essas idades, em que as razões $^{207}\text{Pb}^*/^{235}\text{U}$ e $^{206}\text{Pb}^*/^{238}\text{U}$ não são concordantes é utilizado o termo *idade discordante*.

No exemplo da **figura 2.2A** as razões Pb^*/U foram geradas pela superposição de um segundo evento geológico, três bilhões de anos depois da formação dos zircões anteriores. Este evento não só forma novos zircões, mas também perturba aqueles já formados, causando perda de chumbo radiogênico previamente acumulado desde o evento que originou estes zircões. A **figura 2.2B** ilustra o efeito no diagrama concórdia caso estes zircões sejam analisados um bilhão de anos depois deste segundo evento, digamos hoje.

No exemplo, todos os zircões terão acumulado chumbo radiogênico (Pb^*) durante este um bilhão de anos transcorridos, desde a ocorrência do segundo evento. Os zircões formados neste segundo evento apresentarão idades concordantes. Nos zircões mais antigos do que o segundo evento, o sistema terá sido “zerado” caso todo chumbo radiogênico acumulado tenha sido perdido durante o segundo evento, e logo estes zircões também forneceriam uma idade de 1,0 Ga. Alternativamente, e mais comumente, a perda de chumbo radiogênico é somente parcial ou confinado a subdomínios nestes zircões.

Figura 2.1 – A. Diagrama Concórdia de Wheterill (1956). Modificado de Harley & Kelly (2007).

Neste caso, uma série de idades podem resultar em um alinhamento de dados, ao longo da curva *discórdia*, que intercepta a curva concórdia em dois pontos: um superior (mais antigo), que fornece à idade de formação dos zircões – 4,0 Ga – e um inferior que fornece uma idade de 1,0 Ga correspondente à idade do segundo evento (**fig. 2.2B**).

Figura 2.2 – A. e B. Diagrama Concórdia e a perda episódica de Pb, neste caso devido ao efeito de um segundo evento que perturba o sistema isotópico há 1000 Ma gerando a curva discórdia – ver texto. Modificado de Harley & Kelly (2007).

2.2. Imageamento de domínios complexos em zircões – catodoluminescência (CL)

Durante as últimas duas décadas, o avanço das técnicas de imageamento elevou consideravelmente a observação das complexidades internas dos zircões. Similarmente, recentes avanços no desenvolvimento da espectrometria de massa de íons secundários (*secondary ion mass spectrometry* – SIMS) e na espectrometria de massa de plasma por ablação a laser (*laser-ablation inductively coupled plasma mass spectrometry* – LA-ICP-MS) tornaram possíveis a determinação de idades U-Pb *in situ* com elevada resolução espacial (spots com $\leq 20 \mu\text{m}$ para técnicas SIMS). Interpretações concretas acerca dos eventos demarcados pelas idades obtidas por essas técnicas requer mais do que a simples utilização de geocronômetros U-Th-Pb, e novamente, modernas técnicas analíticas fornecem a solução. Determinações de elementos traços e terras raras (ETR) e isótopos de oxigênio, realizadas *in situ* em zircões, permite a ligação direta entre interpretações petrogenéticas e a idade do domínio analisado no zircão (Sherer et al., 2007). Adicionalmente, a utilização de isótopos de háfnio (Hf) em zircões datados por U-Pb revela quão grande foi a contribuição de crosta juvenil (*e.g.* diretamente derivada do manto) em relação à quantidade de crosta continental reciclada, fazendo do zircão uma “peça única” – sem a necessidade da análise de outros materiais (*e.g.* rocha total) no estabelecimento da evolução crustal de um dado segmento da crosta (Harrison, et al. 2005).

A observação de imagens pancromáticas de minerais traços alvos de determinações geocronológicas por catodoluminescência (CL), ou alternativamente por imagens obtidas com o auxílio do microscópio eletrônico de varredura (MEV), devem fazer parte da avaliação e interpretação dos resultados analíticos obtidos pelas técnicas SIMS e/ou LA-ICP-MS (Rubatto & Gebauer, 1998; Hoskin, 2000; Silva, 2006). Este procedimento permite o reconhecimento de diferentes domínios com base no padrão textural apresentado pelo zircão, aumentando o controle na investigação geocronológica de diferentes domínios do zircão e consequentemente na interpretação dos dados.

As imagens de catodoluminescência registram com alta resolução a correlação quantitativa entre a intensidade da luminescência e o conteúdo de urânio do zircão (Rubatto & Gebauer, 1998). Texturas magmáticas são caracterizadas por zoneamento oscilatório produzidas pela alternância de halos ricos em urânio (baixa luminescência) e halos pobres em urânio (alta luminescência). Zircões metamórficos são desprovidos de zoneamento oscilatório e são caracterizados por uma textura interna homogênea, levando à destruição da textura ígnea pretérita (Corfu et al., 2003). A **tabela 2.2** de Silva (2006) sintetiza algumas características morfológicas do imageamento por CL e MEV (BSE – *back scattered electron*) e suas respectivas interpretações.

A **figura 2.3** ilustra como as técnicas de imageamento auxiliam na organização e controle dos domínios analisados assim como na interpretação, especialmente quando o mesmo cristal é analisado por mais de uma técnica na busca da integração de diferentes dados de um mesmo zircão ou ainda em um mesmo domínio do zircão. O reconhecimento de fases minerais inclusas no zircão pode indicar as condições P-T de cristalização deste mineral, desta forma permitindo a extração de idades relacionadas com determinadas condições em que o zircão foi formado. A **figura 2.3** (abaixo) mostra zircões metamórficos submetidos a condições metamórficas eclogíticas, com inclusões de onfácita e coesita, em que a datação deste zircão fornecerá a idade do metamorfismo de alta pressão.

Tabela 2.2 – Aspectos morfológicos do zircão e suas respectivas interpretações (modificado de Silva, 2006).

Feições		Interpretações
Morfologia externa do cristal	Idiomorfismo e recristalização periférica	Indica respectivamente preservação das características ígneas ou recristalização metamórfica
Morfologia interna do cristal	Homogeneidade <i>vs.</i> distinção entre núcleo e sobrecrecimentos(s)	Indica evolução simples magmática ou metamórfica
	Zoneamentos oscilatórios por setor ou complexos	Indica precipitação de fundidos (melt-precipitation) do domínio (núcleo e/ou sobrecrecimento)
	Obliteração de texturas magmáticas	Alteração pós magmática, metamorfismo ou metamitização
Forma do núcleo	Idiomórfico	Ígneo (xenocrystal) ou metamórfico (xenocrystal)
	Arredondado	Corrosão magmática, metamórfica ou abrasão (detritico)
	Contorno Irregular	Corrosão magmática e/ou metamórfica
Variações na intensidade de luminescência (CL) e (BSE)	Tonalidades cinza-médio uniforme, tanto em CL quanto em BSE	Origem magmática, Conteúdos de U e Th normais (magmáticos); razões Th/U magmática (0.2-0.8)
	Tonalidades cinza-claro e branco (alta luminescência)	Baixos conteúdos em U (e Th), baixas razões Th/U (>0.1), domínios metamórficos de alto grau
	Tonalidades cinza-escuro e preto (baixa luminescência)	Alto conteúdo de U (>1000 ppm), domínios magmáticos metamitizados

2.3. Datação U-Th-Pb em zircões detriticos

Datação U-Pb de zircões detriticos de rochas metassedimentares siliciclásticas é uma ferramenta poderosa para reconstrução geológica de terrenos polideformados e identificação de núcleos continentais antigos não mais preservados (Froude et al., 1983; Compston & Pidgeon, 1986; Mueller et al., 1992; Condie et al. 2009). A técnica convencional de datação U-Pb por diluição isotópica TIMS (*thermal ionization mass spectrometry*) não é uma metodologia conveniente em estudos de proveniência de zircões detriticos, pelo fato de ser uma técnica não pontual (não *in situ*) e demorada para a aquisição de dados.

Portanto, para estudos de zircões detriticos, métodos como LA-ICP-MS e SHRIMP, são mais adequados, pois permitem a análise pontual (*in situ*) de grande número de zircões em um tempo relativamente curto (50 a 200 grãos em uma sessão de 24 horas). Em estudos de proveniência, a necessidade de datar grande quantidade de zircões é uma premissa estatística para que não se oculte as frações de grãos com menor abundância na amostra (*e.g.* Vermeesch, 2004; 2006). Segundo Vermeesch (2004), um numero ideal de 117 grãos são necessários para que nenhuma fração contendo no mínimo 5 % da população total da amostra seja ocultada em um nível de confiança de 95 %.

Figura 2.3 – Acima: imagens de catodoluminescência (CL) de zircões provenientes da localidade de Jack Hills, mostrando spots analisados para ETR, isótopos de oxigênio e geocronologia U-Th-Pb. (extraído de Cavosie et al., 2006). Abaixo: imagens de catodoluminescência (CL) em zircões submetidos a condições metamórficas eclogíticas de ultra-alta pressão, com inclusões de onfacita e coesita (extraído de Liu et al., 2008).

2.4. Determinação de isótopos de Hf em zircão

Para elementos refratários, tais como Sm, Nd, Lu e Hf, assume-se que a composição destes na Terra é similar àquelas do reservatório uniforme condritico (CHUR – *chondritic uniform reservoir*) definido por meteoritos indiferenciados (Faure & Mensing, 2005). O comportamento do sistema Lu-Hf durante episódios de fusão parcial é similar ao do sistema Sm-Nd, em que o isótopo filho (Hf, Nd) é mais incompatível e, portanto, concentra-se em maior quantidade no fundido magmático do que o isótopo pai (Lu, Sm). Portanto, o sistema Lu-Hf pode ser utilizado da maneira similar ao sistema Sm-Nd para monitorar o grau de heterogeneidade dos reservatórios silicáticos da Terra, no entanto com mais vantagens. As determinações *in situ* Lu-Hf (seguidas daquelas U-Pb) em zircão, além de fornecer informações isotópicas diretamente relacionada ao tempo de formação do domínio analisado no zircão, são aparentemente mais robustas do que aquelas analisadas em rocha total por Sm-Nd, fazendo deste método uma grande ferramenta na análise da evolução crustal de terrenos fortemente metamorfizados (Scherer et al., 2007).

A técnica por LA-MC-ICP-MS (*multi collector*) possibilita a determinação *in situ* da composição isotópica de Hf de diferentes zonas intercrescidas em um zircão (Thirlwall & Walter, 1995).

A **figura 2.4A** ilustra como a composição isotópica de Hf no zircão se comporta na diferenciação da Terra tendo como exemplo um simples cenário, onde um fundido (*melt*) é extraído por fusão parcial do reservatório silicático total da Terra (*Bulk Silicate Earth* – BSE), deixando para trás um manto empobrecido devido a extração do material no **ponto 1**. Devido ao fato do elevado fracionamento de Hf em fundidos em relação ao Lu, o fundido terá uma razão Lu/Hf mais baixa do que o reservatório silicático total da Terra (BSE), enquanto a razão Lu/Hf no manto empobrecido residual terá uma razão mais elevada. Por convenção, a composição isotópica de Hf é expressa em termos da divergência (em partes por 10⁴) em relação aos valores do reservatório condritico uniforme (CHUR), em que cujos valores da razão ¹⁷⁶Hf/¹⁷⁷Hf são assumidos como o mesmo do reservatório silicático total da Terra (BSE). A notação usada para essas divergências, em algum segmento de tempo *t*, é tida como o parâmetro Épsilon Hf(*t*). Na **figura 2.4B**, onde *t* é a idade de cristalização do zircão, o parâmetro Épsilon Hf(*t*) representa a composição isotópica inicial do zircão.

O uso de gráficos que utilizam o parâmetro Épsilon Hf(*t*) na ordenada *versus* a idade de cristalização na abscissa (**fig. 2.4C**) torna-se uma considerável ferramenta na geração de informação sobre a diferenciação da Terra primordial, assim como para um dado segmento da crosta onde pretende-se traçar fontes e processos petrogenéticos associados a evolução de séries magmáticas.

2.5. Combinação dos métodos U-Pb e Lu-Hf em zircão detrítico

O material proveniente da erosão da crosta continental – sedimentos – fornece informações valiosas acerca dos processos de crescimento crustal (Scherer et al., 2007; Harrison, et al., 2005; Bodet & Sharer, 2000), cujo registro já fora erodido. Estudos deste tipo são realizados com o auxílio do sistema Sm-Nd em rocha total, contudo problemas relacionados com a obtenção de idades mistas são muitas vezes interpretados sem o devido cuidado, levando a interpretações errôneas. Métodos como idades K-Ar em micas detríticas e U-Pb são corriqueiramente empregados em estudos de fontes sedimentares. Porém o sistema K-Ar é facilmente desestabilizado durante o metamorfismo e espectros de idades U-Pb em zircão frequentemente geram padrões similares em diferentes domínios crustais.

Figura 2.4 – A. Evolução hipotética da razão $^{176}\text{Hf}/^{177}\text{Hf}$ versus tempo para o (BSE) *Bulk Silicate Earth*, manto empobrecido (DM), para dois reservatórios crustais e para o zircão. **B.** Os mesmos reservatórios representados em termos do parâmetro Épsilon Hf versus tempo. A idade U-Pb no zircão data sua cristalização (3), a idade de Lu-Hf de residência crustal é uma estimativa do tempo decorrido da extração deste domínio crustal do manto empobrecido até a cristalização do zircão hospedado neste domínio. **C.** *Trends* evolutivos hipotéticos para o manto empobrecido (DM) e proto-crosta com valor da razão $^{176}\text{Lu}/^{177}\text{Hf}$ de 0,009 com parâmetro CHUR de Blichert-Toft and Albarede, 1997), (extraído e modificado de Scherer et al., 2007).

Nestes casos, estudos isotópicos combinados pelas técnicas U-Pb e Lu-Hf em zircões detríticos, permitem a distinção entre grãos que tem a mesma idade de cristalização, mas foram formados em domínios crustais com idades de extração mantélica distintas. Estes estudos podem ser realizados usando-se das idades de residência crustal determinadas em grãos individuais. Um valor assumido da razão $^{176}\text{Lu}/^{177}\text{Hf}$ para a crosta continental é usado para delinear a evolução da fonte crustal em tempos passados por meio da razão inicial $^{176}\text{Lu}/^{177}\text{Hf}$

no zircão (**ponto 3 nas figs. 2.4A e B**) até a interseção com a curva do manto empobrecido (**ponto 2 nas figs. 2.4A e B**). Novamente o uso de gráficos que utilizam o parâmetro Épsilon Hf na ordenada *versus* a idade de cristalização na abscissa (**fig. 2.4C**) fornece informações essenciais no estabelecimento de domínios crustais distintos em um dado segmento da crosta continental.

2.6. Isótopos de Oxigênio em zircão

A distinção entre granitóides que possuem assinaturas isotópicas evoluídas derivadas de uma fonte mista entre manto (juvenil) e reciclada (provenientes de rochas metassedimentares) de granitóides provenientes de uma rocha precursora mantélica que se colocou na base da crosta preteritamente é comprometida quando baseada somente em isótopos radiogênicos (e.g. Lu-Hf) (Hawkesworth & Kemp, 2006). Até que ponto idades modelo apontam para as verdadeiras idades de acreção crustal, ou se meramente representam a idade média de residência crustal é ainda um assunto para discussão (Arndt & Goldstein, 1987). O problema é ainda mais grave para zircões detríticos ou herdados, pois dados de geoquímica isotópica de rocha total e observações de campo não são mais disponíveis.

Essa ambiguidade pode ser reduzida adicionando-se aos dados isotópicos radiogênicos os dados isótopos estáveis, cujo o fracionamento é independente do tempo. A razão $^{16}\text{O}/^{18}\text{O}$, expressa como $\delta^{18}\text{O}$ relativo ao SMOW muda somente em processos superficiais de baixa temperatura e assim o $\delta^{18}\text{O}$ de rochas derivadas do manto ($5.7 \pm 0.3\text{‰}$) contrasta com aqueles de rochas que foram submetidas a um ciclo sedimentar ou alteração hidrotermal no assoalho oceânico, que tem valores de $\delta^{18}\text{O}$ elevados (Hawkesworth & Kemp, 2006). Esse processo é registrado nos zircões de granitóides com altos valores de $\delta^{18}\text{O}$, fazendo dos isótopos de oxigênio um traçador para componentes de reciclagem crustal em granitos.

Estudos empíricos estabeleceram que a difusão de oxigênio no zircão é suficiente para manter o $\delta^{18}\text{O}$ original do magma, mesmo durante o metamorfismo e fusão parcial (King et al., 1998; Peck et al., 2003). Os valores de $\delta^{18}\text{O}$ podem ser medidos *in situ* no zircão com excelente precisão ($<0.5\text{‰}$) por sondas iônicas de grande raio com sistema multicoletor (e.g. SIMS) (Valley, 2003).

2.7. Elementos traços e terras raras (ETR) em zircão

Devido à configuração eletrônica similar entre os ETR muitos desses elementos obedecem ao mesmo comportamento geoquímico (Hollinson, 1993). Esses elementos possuem geralmente a mesma valência (e.g. 3^{+}) e ocupam a mesma posição cristalográfica nos minerais que os hospedam. Contudo, o raio iônico destes elementos diminuem sistematicamente em função do seu número atômico desde o La (0,116 nm) até o Lu (0,0977 nm) – em coordenação 8 com oxigênio. É justamente esta diferença entre os raios iônicos, que governam as sutis diferenças no comportamento geoquímico destes elementos e que é amplamente explorada pela petrologia (Hanchar & Westrenen, 2007). Por apresentarem raios iônicos relativamente grandes, os ETR são considerados elementos *incompatíveis*, isto é, eles não substituem facilmente os cátions presentes nos principais minerais formadores de rocha e preferem residir no fundido coexistente. Em contraste, os ETR são geralmente *compatíveis* com relação aos minerais acessórios das rochas, que se formam nos estágios finais da cristalização do magma (Hanchar & Westrenen, 2007; Wark & Miller, 1993). Como resultado, tem-se estabelecido que o balanço dos ETR seja principalmente controlado pelos minerais acessórios ao invés dos minerais formadores de rocha (Wark & Miller, 1993). Devido ao raio dos elementos terras raras leves (ETRL

– La até Gd) ser maior do que os elementos terras raras pesados (ETRP – Tb até Lu), os primeiros são relativamente mais incompatíveis em relação aos últimos.

No zircão os ETR (*p.e.* ETR³⁺) substituem o Zr⁴⁺ (de raio 0,084 nm). Para compensar o desequilíbrio de cargas e assegurar a neutralidade do zircão, o elemento pentavalente P⁵⁺ é requisitado para a substituição do Si⁴⁺, na intitulada substituição da xenotima (*e.g.* Zr⁴⁺ + Si⁴⁺ → ETR³⁺ + P⁵⁺) (Hanchar et al., 2001; Hoskin & Schaltegger, 2003).

Tanto os zircões naturais quanto os sintéticos são enriquecidos em ETRP em relação aos ETRL, sendo esta uma premissa esperada uma vez que os ETRP têm raios iônicos similares ao Zr⁴⁺ (Hanchar & Westrenen, 2007; Hanchar et al., 2001). Muitas pesquisas têm sido realizadas para quantificar os ETR em zircão e como estes elementos se relacionam com o fundido a partir de qual foram cristalizados. Para estes estudos, a relação da partição dos ETR entre mineral e fundido é utilizado o coeficiente de partição de Nernst (D_i) definido por:

$$D_i^{\text{Mineral/Fundido}} = C_i^{\text{Mineral}} / C_i^{\text{Fundido}}$$

Onde D é o coeficiente de partição para um elemento i e C_i^{Mineral} e C_i^{Fundido} são as concentrações em porcentagem de peso do elemento i respectivamente no mineral e no fundido (Beatie et al., 1993). Sabendo-se do coeficiente de partição dos ETR no zircão é possível calcular a composição do fundido a partir do qual o zircão foi cristalizado. Esta informação é importante em casos onde o contexto magmático em que o zircão cresceu foi completamente obliterado. Devido à alta resistência do zircão em relação aos principais minerais formadores de rocha, esses minerais acabam sendo a única testemunha de eventos magmáticos antigos. Essa informação geológica pode ser utilizada em estudos de proveniência de rochas metassedimentares combinados com isótopos traçadores de fontes e processos (*e.g.* Lu-Hf, O) e com a geocronologia U-Th-Pb (Scherer et al., 2007, Cavosie et al., 2008).

2.8. Conexão entre a idade e padrões ETR em rochas de alto grau: *proxies* para P e T

Grande parte dos zircões metamórficos precipitam a partir de uma fase fluida ou fundida saturada em zircão, incorporando U em sua estrutura. O crescimento do zircão ocorre tanto por meio de dissolução sub-sólida quanto reprecipitação diretamente de um fundido ou fluido metamórfico (McClelland & Lapen, 2013). O zircão metamórfico geralmente cresce em resposta ao metamorfismo progressivo ou retrógrado ao longo de trajetórias P-T que vão da fácies anfíbolito a eclogito. Texturas de crescimento *versus* recristalização podem ser reconhecidos em imagens CL. Uma característica importante do zircão é a capacidade de preservar múltiplos domínios que registram eventos metamórficos ao longo da trajetória P-T-t, assim como a história pré-metamórfica do protólito (Rubatto and Hermann 2007).

Mudanças na composição dos elementos traços podem definir se o zircão cresceu sob alta pressão, independentemente de suas inclusões. Zircões da fácies eclogito se formam na presença de granada, que preferencialmente sequestra elementos terras raras pesados (ETRP), e em conjunto com a ausência de plagioclásio conferem ao zircão uma assinatura de ETR característica: um padrão plano caracterizado pelo empobrecimento de ETRP e ausência de anomalia negativa de Eu. Essa assinatura distingue-se dos padrões característicos de zircões de protólito ígneo e das assinaturas de zircões precipitados a partir de fundidos formados na trajetória retrógrada (Rubatto, 2002). A difusibilidade Ti em zircão funciona como um

termômetro capaz de fornecer temperaturas associadas a formação do zircão (Tomkins et al., 2007). A variação de padrões ETR em conjunto com o conteúdo de Ti é ideal na definição de domínios dentro de zircões para amarração destes com as condições de P-T (e.g. Hermann et al., 2001; Mattinson et al., 2006).

2.9. Referencias

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3. Contexto Geológico

3.1 O Orógeno Gondwana Oeste

Como explicitado no capítulo introdutório, sugeriu-se aqui chamar de Orógeno Gondwana Oeste todos as áreas orogenéticas Pan-africanas/Brasilianas alinhadas ao longo do Lineamento Transbrasiliano-Kandi. Este orógeno foi resultante do consumo e fechamento do Oceano Goiás-Farusiano (Caby, 2003; Kroner and Cordani, 2003), que culminou na colisão continental envolvendo os crátons Amazônico/Oeste África, São Francisco-Congo e o metacrátón Saara.

O Orógeno registra em toda sua extensão um longo período de convergência (> que 400 m.y) com o desenvolvimento de diversos arcos intraoceânicos e continentais que são hoje observados dentro da zona colisional fóssil profundamente erodida (Pimentel and Fuck, 1992; Caby, 2003; Berger et al., 2011; Ganade de Araujo et al., 2012). As principais assembleias petrotectônicas entre as regiões cratônicas envolvidas na colisão são representadas por margens passivas, arcos juvenis, arcos continentais tardios e sequencias sin-orogênicas. Os detritos da erosão das montanhas resultantes da colisão no orógeno estão hoje documentadas em bacias molássicas e do tipo *foreland*, e o final da atividade orogenética é datado em ca. 540-500 Ma com base nos granitoides pós-colisionais dispostos ao longo do orógeno (Affaton et al., 2000; Caby, 2003; Pimentel et al., 2011; Ganade de Araujo et al., 2012).

Neste capítulo serão apresentados as características geológicas de um setor importante do Orógeno Gondwana Oeste, particularmente no Domínio Ceara Central da Província Borborema, onde se concentraram as investigações desta pesquisa.

3.2. A Província Borborema

A província Borborema Setentrional (**fig. 3.1**) situa-se a norte do Lineamento Patos e é subdividida por Delgado et al. (2003) entre os domínios Médio Coreaú, Ceará Central e Rio Grande do Norte. Alguns autores dividem o Domínio Rio Grande do Norte entre os Domínios Jaguaribeano, Rio Piranhas e Caldas Brandão (**ver fig. 3.1**). Abaixo serão descritos em maior detalhe o Domínios Ceará Central e Médio Coreaú, que abrangem a área física desta pesquisa.

3.2.1. O Domínio Médio Coreaú

Este domínio (DMC) é marcado por um intenso sistema de zonas de cisalhamento SW-NE desenvolvidas em regime compressivo-transpressivo relacionados ao sistema cisalhante Transbrasiliano (Lineamento Transbrasiliano) que separa este do domínio do Ceará Central, a sudoeste. Em termos tectono-estratigráficos DMC compreende o Complexo Granja além da faixa de supracrustais Martinópole-Ubajara tidas com seqüências vulcanossedimentares e metassedimentares marginais ao crátón São Luiz-Oeste Africano.

O Complexo Granja é apresentado como “embasamento” da faixa supracrustal Martinópole-Ubajara e compreende gnaisses para- e ortoderivados, em parte migmatíticos, de fácies anfibolito a granulito, além de ortognaisses de afinidade TTG.

Figura 3.1 – Contexto geológico da área de estudo (a partir de De Wit et al., 2008, Cavalcante, 1999; Delgado et al., 2003; Cavalcante et al., 2003), compilação geocronológica baseada em Osako et al. (2008).

Fetter et al. (1997) reportam idades U-Pb em kondalitos de *ca.* 2,28 Ga. Santos et al. (2008) sugere uma crosta juvenil gerada em ambiente de arco entre 2,3-2,5 Ga, com posterior retrabalhamento crustal no Paleoproterozóico e Neoproterozóico. Idades Ar-Ar e U-Pb em titanita vinculadas ao evento termotectônico do Neoproterozóico estão distribuídas entre 0,575 e 0,554 Ga (Caby et al., 1995).

A faixa supracrustal Martinópolis-Ubajara engloba seqüências de margem passiva, com deposição entre 0,775 e 0,808 Ga e metamorfismo estimado com base em idade U-Pb em titanita em *ca.* 0.650 Ga (Fetter, 1999). O grupo Martinópolis abrange as Formações: 1) Goiabera composta de xistos aluminosos, quartzitos ferríferos e subordinadamente paragneisses; 2) São Joaquim, compostas de quartzitos, metacalcários e metavulcânicas intercaladas; 3) Covão e Santa Terezinha que compreende um conjunto carbonato-psamítico-pelítico de baixo grau metamórfico, com evidências de glaciação (?). A posição estratigráfica interna deste contexto ainda não está perfeitamente estabelecida e, portanto, não deve ser tomada como o verdadeiro empilhamento estratigráfico. Fetter et al., 1997 estabeleceram uma idade neoproterozóica de deposição entre 0,775-0,810 Ga, com base em determinações U-Pb em derrames ácidos intercalados nesta seqüência. O grupo Ubajara correspondente à unidade superior, de ambiente fluvio-marinho, contendo rochas metassedimentares clastopelíticas da Formação Trapiá e Caiçaras, com importante contribuição carbonática pertencentes à Formação Frecherinha, a qual se sobrepõe uma recorrência clastopelítica da Formação Coreá.

3.2.2. O Domínio Ceará Central

O presente conhecimento permite a divisão deste domínio (**figs. 3.1**) entre três unidades cronológicas distintas, sendo estas: (1) Associações/núcleos arqueanos e embasamento paleoproterozóico e; (2) Supracrustais de idade proterozóica indefinida a neoproterozóicas; e (3) Complexos granito-migmatíticos de idade neoproterozóica, bem como uma série de corpos granitóides pós-colisionais a anorogênicos de idade predominantemente cambriana para o primeiro grupo e ordoviciano para o último.

3.2.2.1. Registro arqueano e associações do embasamento paleoproterozóico

A divisão litoestratigráfica entre o registro arqueano e as associações paleoproterozóicas ainda esbarra em certo desentendimento, principalmente pela falta de estudos geocronológicos e isotópicos.

As primeiras denominações para os terrenos arqueanos e das associações paleoproterozóicas deste segmento do Domínio Ceará Central foram designadas por Brito Neves (1975) para o Maciço Tróia-Tauá dividido pela zona de cisalhamento Sabonete-Inharé entre os Blocos Mombaça (gneisses granulíticos do tipo TTG) a sudeste, e Tróia-Pedra Branca (associação do tipo granito *greenstone belt*) a noroeste.

Oliveira e Cavalcante (1993) inserem essas rochas dentro do Complexo Cruzeta (**unidades 7a e 7b na legenda da figura 3.1**), e dividem o mesmo em quatro unidades: 1) Tróia: seqüência metaplutono-vulcanossedimentar; 2) Pedra Branca: ortogneisses cinzentos TTG; 3) Mombaça: gneisses diversos e migmatitos com lentes de metacalcários, anfíbolitos, rochas cálcio-silicáticas e metaultrabásicas; 4) Cedro: metaleucogranitóides tabulares encaixados segundo a superfície tectono-metamórfica regional.

Delgado et al. (2003) referem-se ao Bloco Tróia-Pedra Branca para as rochas deste contexto, e inserem-nas como parte integrante do Complexo Cruzeta, que é dividido entre as unidades Tróia, Pedra Branca, Mombaça e Algodões, incluindo a Suíte Cedro e Madalena.

A unidade Tróia compreende uma associação do tipo *greenstone belt*, composta por metabasaltos, metagabros, metadacitos, metariolitos intercalados com quartzito, grafita xistos, metacalcários, metachert e formações ferríferas bandadas. *Sheets* intrusivos nesta unidade compostos de ortognaisses de composição tonalítica-granodiorítica e leucogranítica da Suíte Cedro forneceram uma idade U-Pb de 2,77 Ga e idades modelo T_{DM} entre 2,92 e 3,04 (Fetter, 1999).

A unidade Pedra Branca que ocorre em associação próxima com a unidade Tróia é composta de ortognaisses de afinidade TTG, com idade U-Pb entre 2,77 e 2,85 (Fetter, 1999). Entretanto, Silva et al. (2002) demonstraram a ocorrência de um evento do Mesoarqueano por meio de análise SHRIMP, em zircões de gnaisses tonalíticos situados a SW de Boa Viagem. A análise revelou um núcleo mais antigo datado de 3,27 Ga e uma borda sobrecrecida datada de 2,08 Ga. Ainda, retrabalhamentos do neoproterozóico (~ 0,57 Ga) foram registrados por Monié et al. (1997) e Fetter (1999) respectivamente, pelos métodos Ar-Ar e U-Pb em monazita.

A unidade Mombaça é formada por ortognaisses diversos, migmatitos e rochas paraderivadas em alto grau metamórfico e distingui-se da unidade Pedra Branca por envolver um maior grau de retrabalhamento (Delgado et al., 2003).

A unidade Algodões (**unidade 6d na legenda da figura 3.1**) é composta por uma associação de rochas metassedimentares, metabásicas, e ortognaisses tonalíticos a granodioríticos, de idade U-Pb entre 2,13 e 2,33 Ga e idades modelo T_{DM} entre 2,24 e 2,44 (Martins, 2000). Estas rochas ocorrem emoldurando o Complexo Cruzeta em seu limite noroeste e oeste (Arthaud, 2007).

A Suíte Madalena (Castro, 2004) é uma associação de quartzo-diorito e diques microdioríticos que cortam o Complexo Cruzeta. Essas rochas mostram fraca deformação e não são migmatizadas. Idades U-Pb em zircão variam em torno de 2,15 e 2,2 Ga (Castro, 2004) para as rochas deste contexto.

Dentro do âmbito do Paleoproterozóico ocorrem ainda no Domínio Ceará Central uma série de associações gnáissicas migmatíticas, carentes de mapeamento geológico e de estudos geocronológicos, designadas como embasamento das supracrustais neoproterozóicas. Estas associações compreendem ortognaisses de composição tonalítica a granodiorítica, geralmente metamorizados em fácies anfibolito de alta temperatura com condições variáveis de migmatização e rara contraparte sedimentar (Cavalcante et al., 2003). Idades U-Pb para essa associação caem entre 2,11 e 2,19 Ga com idades modelo T_{DM} entre 2,42 e 2,48 Ga (Hackspacher et al., 1990; Fetter, 1999; Castro, 2004).

3.2.2.2. Supracrustais de idade proterozóica (neoproterozóica?)

As seqüências supracrustais proterozóicas são representadas pelas rochas metassedimentares do Complexo Ceará (unidades 4a, b, c e d na legenda da figura 1C), o qual abriga as unidades Canindé, Independência, Quixeramobim e Arneiroz (Cavalcante et al., 2003). Essas seqüências ocorrem nas adjacências do Complexo granito-migmatítico Tamboril Santa Quitéria, e apresentam paragênese de moderada a alta temperatura, característica do fácies anfibolito alto a granulito, localmente com relictos de paragênese eclogítica, combinado com um conspícuo bandamento tectônico de baixo ângulo. No Bloco ou Sub Domínio Acaraú (**unidade 6a na legenda da figura 3.1**) as rochas supracrustais do Grupo Novo Oriente (**unidade 4e na legenda da figura 3.1**) representam um segmento de uma margem passiva [com idade máxima de 1,36 Ga,

(Ferreira et al., 2008)] com provável evolução para condições oceânicas francas que foi subsequentemente acrecionada durante a colisão neoproterozóica (Ganade de Araújo et al., 2010a).

De um modo geral, a associação sedimentar original do Complexo Ceará abarca largos tratos psamíticos, psamo-pelíticos e pelíticos, associados ou não a sedimentação química. Ocorrências de anfibolitos representam provavelmente o magmatismo básico sin-sedimentar associado à deposição destes sedimentos (Castro, 2004, Arthaud, 2007). Recentemente, alguns estudos termobarométricos realizados sobre esses anfibolitos granatíferos indicaram registros de condições eclogíticas (Castro, 2004; Garcia, 2006, Santos et al., 2009). Sob uma óptica preliminar, o posicionamento geográfico dessas rochas de alta pressão ocorre emoldurando o Complexo Tamboril Santa Quitéria.

Em termos estruturais, essas seqüências foram fortemente afetadas pelo evento tectono-termal associado à colisão Brasileira-Pan Africana, que materializou nestas rochas um bandamento tectônico de baixo ângulo, que levaram Caby e Arthaud (1986) a interpretar a presença de extensas *nappes* neoproterozóicas com vergência geral para sul, na região a leste do Complexo Tamboril Santa Quitéria. Na região oeste deste complexo, nas proximidades de Sobral, o bandamento tectônico de baixo ângulo evolui e é posteriormente truncado por um sistema de zonas de cisalhamentos transcorrentes dextrais integrante do sistema cisalhante Transbrasiliiano, evidenciando uma mudança no regime tectônico, provavelmente associado ao escape lateral nos estágios finais da colisão continental (Ganade de Araújo & Santos, 2007, Cunha, 2007).

Os primeiros marcos geocronológicos acerca da idade da deposição dos detritos que compõem o Complexo Ceará foram obtidos por Fetter (1999), que obteve uma idade U-Pb de 0,77 Ga derivada de ortognaisses interpretados como derrames ácidos sin-sedimentares, encontrado próximo à localidade de Independência. Castro (2004) reporta uma idade U-Pb semelhante de *ca.* 0,70 Ga para rochas similares na região de Itaitaia. Recentemente, estudos pioneiros de proveniência realizados por Arthaud (2007) na região de Itaitira, assinalaram três populações de zircões detríticos distintas com aglomerados de idades em torno de 0,8 Ga (quatro zircões); 1,0 a 1,2 Ga e diversos zircões em torno de 1,85 Ga. O mesmo autor interpreta as idades em zircões detríticos em torno de 0,8 Ga, em conjunto com uma idade U-Pb de 0,749 Ga obtida em granada anfibolitos, interpretados como magmas básicos sin-sedimentares, como sendo a idade da deposição do Complexo Ceará, dada em um ambiente extensional sobre um embasamento arqueano/paleoproterozóico. Entretanto, o comportamento dos isótopos de Nd (Castro, 2004; Arthaud, 2007) indica que a maior parte da sedimentação foi oriunda da erosão das litologias paleoproterozóicas e provavelmente com subordinada contribuição arqueana. A falta de idades entre 2,0-2,2 Ga e 2,7-2,8 Ga no espectro de zircões detríticos no estudo de Arthaud (2007), característico do embasamento adjacente, aponta que é necessário entender o Complexo Ceará em termos de diversos ciclos deposicionais (*e.g.* sobreposição de bacias) associados respectivamente com o ambiente tectônico vigente para um dado segmento do espaço-tempo (*e.g.* bacias mais antigas paleo a mesoproterozóicas [?] *vs.* bacias brasileiras marginais possivelmente associadas a ambiente de arco). Amaral et al. (2008) reportam uma idade U-Pb/Lu-Hf de *ca.* 1,56 Ga, interpretada como a idade de cristalização de granada anfibolitos inseridos nas rochas metassedimentares do complexo, contudo a validade deste dado na demarcação de um evento termo-tectônico amplo desta idade ainda carece de melhor sustentação. Ainda, a estruturação proposta por Arthaud e Caby (1986), pode justapor por meio de um sistema de *nappes* em uma tectônica *thick-skinned*, onde o embasamento é afetado, conjuntos de rochas metassedimentares mais antigas (*e.g.* meso-paleoproterozóicos) com mais novas (*e.g.* brasileiras), separadas por zonas de cisalhamento de baixo ângulo, cartograficamente indistinguíveis entre si devido à forte impressão termo-tectônica da colisão Brasileira-Pan Africana. Neste sentido, visando o melhor esclarecimento deste contexto estudos isotópicos aplicados à cartografia geológica devem ser realizados com mais frequência.

3.2.2.3. O Complexo granito-migmatítico neoproterozóico Tamboril Santa-Quitéria

O Complexo Tamboril Santa Quitéria (Cavalcante et al., 2003) está alinhado segundo a direção NE-SW, cobrindo uma área de 40.000 km² e representa significativa parte dos complexos granitóides do Domínio Ceará Central. Este complexo consiste de um conjunto litológico que abrange gnaisses metatexiticos e diatexiticos, no sentido de Sawyer (2008) que variam de composição diorítica a granítica. Esparsos corpos de granitóides porfiríticos a equigranulares, mostrando claras relações intrusivas sobre os metatexitos e diatexitos e composição predominantemente granodiorítica a granítica, atestam que diversos pulsos magmáticos foram envolvidos no desenvolvimento deste complexo.

Fetter *et al.* (2003), vem interpretando o complexo como uma bem desenvolvida suíte magmática originada em ambiente de arco continental, com sucessivos episódios magmáticos, ativos durante o neoproterozóico.

Datações U-Pb realizadas por Fetter (1999) geraram idades entre 0,637 e 0,623 Ga para os granitóides deformados do Complexo Tamboril-Santa Quitéria, e ainda idades modelo (T_{DM}) variando entre 0,86 a 1,92 Ga, com Épsilon Nd (600 Ma) variando entre -20 a +4, sugerindo uma fonte mista para a origem dos granitóides. Castro (2004) obteve idades U-Pb similares para estas rochas, variando entre 0,620 a 0,611 Ma. Brito Neves et al. (2003) reportam uma idade de 0,660 Ma para um granitóide inserido neste complexo. Uma idade de 0,795 Ga (Epsilon Nd (800 Ma) = +4,4), foi obtida pelo método Pb-Pb em zircão de ortogneiss tonalítico a granodiorítico da borda leste do Complexo Tamboril-Santa Quitéria, atestando para a contribuição de material juvenil em torno de 0,8 Ga (Ganade de Araujo et al., 2010b).

A descoberta de retroeclogitos por Castro (2004) levou a uma interpretação divergente da discutida por Fetter (2003), em relação à posição da zona de subducção que teria dado origem ao Complexo Tamboril Santa Quitéria. Com base no posicionamento geográfico desses retroeclogitos (a oeste do complexo), Castro (2004), sugere um sentido noroeste para o fechamento oceânico, já Fetter (2003), baseado na posição atual do Batólito e por anomalias gravimétricas positivas (Lesquer et al., 1984) propõem um sentido sudeste para o processo de subducção. Novos dados termobarométricos de Santos et al. (2009), em rochas eclogíticas a oeste do Complexo Tamboril Santa Quitéria reforçam a teoria de uma subducção para leste-sudeste. O panorama atual mostra que estas rochas de alta pressão ocorrem ao redor do Complexo Tamboril Santa Quitéria, sugerindo que o processo colisional pode ter envolvido zonas de alta pressão em ambos os lados do Complexo Tamboril-Santa Quitéria em uma espécie de extrusão tectônica vertical, similares a estruturas de flores positivas. Alternativamente, modelos envolvendo duas ou mais subducções próximas também podem ser empregados a exemplo do modelo proposto por Cabby (2003) na evolução neoproterozóica do escudo Hoggar no centro africano. Este mesmo autor (*com. pess.*) acredita que o Complexo Tamboril Santa Quitéria (independentemente de sua natureza) trata-se de uma unidade alóctone, e, portanto, qualquer esforço atribuído à construção crustal desta região com base nos elementos associados ao arco magmático de Santa Quitéria (*e.g.* subducção a leste ou oeste da presente posição arco, bacias marginais de *forearc*, *foredeep* e/ou *back arc*, prismas acrecionários) deve-se levar em consideração a aloctonia do mesmo.

3.2.2.4. Granitogênese pós-colisional

Adicionalmente, deve ser ressaltado a granitogênese de idade próxima, mas, no entanto, mais jovem que as porções integrantes do Complexo Tamboril Santa Quitéria. O primeiro deles representado pelas suíte Quixadá-Quixeramobim, com idades U-Pb de 0,585 Ga (Almeida, 1999), e granitóide Chaval (no Domínio

Médio Coreau) com idade U-Pb em monazita de 0,591 Ga (Fetter, 1999). O segundo, e mais jovem episódio, é representado por uma série de corpos, no qual se destacam os corpos Mucambo, Meruoca, Barriga, Pagé, Serrote São Paulo e Complexo Anelar Quintas com idades U-Pb variando entre 0,530 a 0,480 Ga (Castro, 2004, Fetter, 1999). A inter-relação do ponto de vista tectônico entre esses episódios magmáticos ainda foi pouco explorada (*e.g.* Ganade de Araujo, 2008, 2011) carecendo ainda de mais dados geocronológicos e do emprego de técnicas traçadoras de fontes e processos, possibilitando uma melhor análise dos processos tectônicos relacionados à geração destes magmas e bem como a história pós-colisional da cadeia.

3.2.2.5. Calhas tardi-brasileiras e o início da sedimentação da Bacia do Parnaíba

Após a colisão neoproterozóica um sistema extensional instalado preferencialmente sobre as zonas de cisalhamento tardias do escape colisional (Sistema Transbrasiliano), representados pelos riftes Jaíbaras, Jaguapari, Cococi, São Julião e Raimundo Nonato sugere um período de colapso orogenético. Este episódio foi acompanhado de magmatismo, já mencionado acima, que compreende corpos intrusivos com idades variando entre 0,53 a 0,48 Ga, assim como vulcanismo representado pelo enxame de diques Coreau e pela Suíte de sills e diques Parapuí. Subseqüentemente a este estágio mecânico, inicia-se a fase termal de subsidência, em que se inicia a sedimentação da bacia intracratônica do Parnaíba (Oliveira & Mohriak, 2003).

O preenchimento dessas calhas extensionais inicia-se com depósitos caracterizados por rápidas variações na espessura e mudança de fácies, associados a espessos pacotes conglomeráticos e conformidades locais e inconformidades de variada extensão (Abreu *et al.*, 1993 Oliveira & Mohriak, 2003). A calha de Jaíbaras apresenta a melhor exposição desta sedimentação e é organizada estratigraficamente da base para o topo nas Formações Massapé, Pacujá e Aprasível.

A Formação Massapé é caracterizada por conglomerados polimíticos clasto-suportados e arenitos de granulação grossa associados à *debris flows* e *mudflows* depositados em leques aluviais, localmente restritos a limites de falhas (Oliveira & Mohriak, 2003). A Formação Pacujá representa a parte distal da Formação Massapé (Gorayeb *et al.*, 1988) e consiste em arenitos, siltitos e folhelhos. A seqüência superior consiste de conglomerados polimíticos que difere daqueles basais por apresentarem fragmentos clásticos de rochas vulcânica e intrusivas (Oliveira & Mohriak, 2003). A atividade magmática crono-correlata ao preenchimento sedimentar desta bacia esta representada por quatro eventos. O enxame de diques Coreau é caracterizado por diques de direção ENE-WNW de composição riolítica-dacítica de textura porfirítica. Os Plutons Mucambo e Meruoca respectivamente com idades de 0,532 Ga (U-Pb) e 0,507 Ga (Rb-Sr) (Fetter, 1999; Sial & Long, 1981) exibem contatos intrusivos e causam auréolas de contato sobre o conteúdo sedimentar da calha Jaíbaras. A suíte Parapuí com idade mínima ordoviciana é composta de andesitos, riolitos, syenitos, basaltos andesíticos e vulcanoclásticas que localmente podem chegar a 350 m de espessura (Oliveira & Mohriak, 2003).

A fase de subsidência termal na Bacia do Parnaíba inicia-se no Siluriano com a deposição das seqüências basais do Grupo Serra Grande, provavelmente a partir de detritos oriundos do desmonte da cadeia neoproterozóica. Este grupo engloba as Formações Ipu, Tianguá e Jaicós onde Góes e Feijó (1994) interpretam que esses pacotes foram depositados em ambientes fluvio-glacial e glacial, passando a transicional marinho e retornando as condições continentais. A área de ocorrência do pacote sedimentar do Grupo Serra Grande delimita a bacia em seus flancos nordeste, leste e sudeste. Apresenta uma notável quebra no relevo regional com característica morfológica de *cuestas*. As maiores espessuras estão situadas a nordeste, particularmente sobre o sistema de zonas de cisalhamento Transbrasiliana, com adelgaçamento em direção a sul e sudeste (Santos & Carvalho, 2004).

A geocronologia dessas bacias é baseada essencialmente em determinações Rb-Sr em rochas vulcânicas, portanto, ainda não se encontra bem definida. Parente et al. (2004) sugere um intervalo entre 0,56-0,53 Ga para a deposição da sequência inferior e 0,53-0,44 Ga para a sequência superior.

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4. Procedimentos Analíticos

4.1. Geoquímica

Os dados concernentes a geoquímica de rocha apresentados aqui foram obtidos independentemente pelo Laboratório SGS-GEOSOL de acordo com o pacote de análises pleiteado pelo Serviço Geológico da Brasil para aquisição de dados geoquímicos no âmbito dos mapeamentos geológicos do território nacional.

Os elementos maiores foram determinados por meio da técnica ICP-AES utilizando o espectrômetro da marca Varian Vista Pro. Os elementos traços e terras raras foram determinados pela técnica ICP-MS utilizando o espectrômetro de massa da marca Perkin-Elmer Sciex ELAN 6000. Análises de padrões de rochas do USGS (BCR-2, BHVO-1 and AGV-1) indicam precisão e acurácia da ordem de 1% para os elementos maiores e de 5% para os elementos traços e terras raras. Os resultados foram tratados com a utilização do programa GCDkit (Janousek et al., 2006) e Petrograph v.beta (Petrelli et al., 2005). Valores de normalização para o manto primitivo de condrito foram retirados de McDonough and Sun (1995) e Sun and McDonough (1989), respectivamente.

4.2. Geocronologia U-Th-Pb em zircão

Nesta Tese foram utilizados duas técnicas distintas para a datação *in situ* pelo método U-Th-Pb em zircão. Para a datação de zircões detriticos optou-se pela técnica LA-ICP-MS (*laser ablation-inductively coupled plasma mass spectrometer*) enquanto que para os zircões ígneos optou-se pela técnica SHRIMP (*sensitive high resolution ion microprobe*). O equipamento utilizado para a datação dos zircões detriticos por LA-ICP-MS foi o Finnigan Neptune acoplado em um laser ArF excimer ($\lambda = 193 \text{ nm}$), instalado no IGc/USP. Em relação aos zircões ígneos foram utilizados os equipamentos da *Australian Scientific Instruments* (ASI) SHRIMP IIe, instalado no IGc/USP e SHRIMP II, instalado na *Research School of Earth Sciences* (RSES) da *Australian National University* (ANU) em Camberra, Austrália. A separação dos zircões investigados seguiram os padrões do laboratório de separação mineral do Centro de Pesquisas Geocronológicas do IGc/USP de acordo com Loios et al. (2009).

4.2.1. Imageamento dos zircões por Catodoluminescência (CL)

Para revelar a textura interna dos cristais de zircão analisados foi utilizados o microscópio eletrônico de varredura Quanta 250 FEG equipado com um espectroscópio de catodoluminescência Mono CL3+ da marca Centaurus, instalado no IGc/USP. Na RSES foi utilizado o microscópio eletrônico de varredura JEOL-6610A acoplado com um sistema de catodoluminescência Robinson. As condições operantes foram 15 kV, 70 μA and a 20 mm para a distancia focal.

4.2.2. O método U-Th-Pb por SHRIMP

Embora os avanços na determinação geocronológica usando separação química e dissolução isotópica (TIMS) tenham trazido excelente progresso, o uso de técnicas pontuais (*in situ*) para análise de minerais mostrando cristais complexos (*e.g.* aqueles contendo sobrecrecimento metamórfico e/ou zonas xenomórficas) tornou-se necessário para melhor extração e interpretação dos dados. No final da década de 70, foi desenvolvido na Austrália, um espectrômetro de massa de alta resolução de íons secundários (*secondary ion mass spectrometer - SIMS*). Este instrumento conhecido como SHRIMP (*sensitive high resolution ion microprobe*) foi utilizado para a determinação de domínios de idade em cristais complexos, já que pequenas porções deste cristal podem ser

analisadas pontualmente. A primeira publicação de idades SHRIMP foi em 1982 (Compson et al., 1982) e somente um ano depois o mesmo instrumento foi usado para a determinação do primeiro mineral terrestre com mais de 4,0 Ga (Froude et al., 1983).

Determinações *in situ* de U-Th-Pb por microsondas de alta resolução são obtidas em superfícies polidas, usualmente com metade da espessura original do cristal, por meio do bombardeamento de um feixe primário de ânions O_2 (Williams, 1998 e referencias citadas neste trabalho). Este bombardeamento de íons (*sputtering*) produzidos pela descarga do oxigênio dentro de um catodo de níquel libera uma fração pequena do zircão gerando uma ampla variedade de íons secundários. Os íons secundários são então duplamente focados, primeiramente por meio de um analisador eletrostático que os filtra segundo seus valores de energia cinética, então esses íons são focados novamente para um setor magnético, que discrimina estes íons com base em suas respectivas massas (Williams, 1998). O setor magnético é similar aquele utilizado nos espectrômetros de ionização termal. Porém, como exibem um raio maior é capaz de operar em uma resolução de massa maior, permitindo a discriminação de isótopos de urânio e chumbo de íons e complexos moleculares de massas similares produzidos na incidência de O_2 no zircão (e.g. ^{96}Zr , ^{94}Zr , ^{16}O versus ^{206}Pb em análises em zircão) (Patchett & Samson, 2005).

Além da análise do zircão e outros minerais traços, determinações pontuais de elementos terras raras e outros elementos-traços proporcionam uma extrema versatilidade da metodologia SHRIMP. Entretanto, é a alta resolução espacial em escala μm , possibilitando a seleção de domínios homogêneos em cristais de zircão com estrutura interna complexa, aliadas à rapidez analítica que fornecem à sistemática SHRIMP sua mais importante vantagem comparativa (Silva, 2006). A **figura 4.1** ilustra o tamanho das cavidades geradas pelas técnicas SHRIMP e LA-ICP-MS.

Figura 4.1 – Ilustração comparativa entre as cavidades geradas pelas técnicas SHRIMP e LA-ICP-MS (modificado de Patchett & Samson, 2005).

Os detalhes acerca da instrumentação e da técnica SHRIMP podem ser encontrados em Williams (1998). Os dados apresentados aqui foram coletados em blocos de cinco *scans* sobre todas as massas de interesse, com o padrão TEMORA (Black et al., 2003) intercalado a cada quatro análises. Para zircões com baixo conteúdo de U, os mesmos procedimentos foram adotados, contudo o número de *scans* utilizado foi de seis, e o tempo de análise em cada canal foi aumentado na busca de uma maior sensibilidade. Todas as análises foram corrigidas para o ^{204}Pb com base nas razões $^{207}\text{Pb}/^{206}\text{Pb}$ medidas de acordo com Williams (1998) e baseado na composição do ^{204}Pb de Stacey and Kramers (1975). Tratamento dos dados e cálculo das idades foram realizados utilizando-se os programas Squid and Isoplot/Ex (Ludwig, 2003).

4.2.3. O método U-Th-Pb por LA-ICP-MS

No começo dos anos 80 foram desenvolvidos espectrômetros de massa quadripolos, utilizando plasma de argônio como fonte de ionização (*inductively coupled plasma mass spectrometers* – ICP-MS). Embora estes instrumentos tenham sido desenvolvidos para a determinação de elementos traços, muitos estudos utilizaram-nos para a obtenção de idades U-Pb. Idades $^{206}\text{Pb}^*/^{238}\text{U}$, $^{207}\text{Pb}^*/^{235}\text{U}$ e $^{208}\text{Pb}^*/^{232}\text{Th}$ podem ser determinadas, assim como idades $^{207}\text{Pb}/^{206}\text{Pb}$, utilizando-se de técnicas de ablação a laser (*laser ablation* – LA-ICP-MS). O apelo óbvio do uso de datações U-Th-Pb pela técnica LA-ICP-MS é a eliminação de procedimentos de separação U-Th-Pb, a não utilização de processos de dissolução ultra-limpas (necessários na técnica TIMS – *thermal ionization mass spectrometer*), a velocidade das análises (menor que 3 minutos por análise) e a possibilidade de análises pontuais *in situ*. Neste sentido esta técnica compartilha diversas similaridades com a técnica de determinação por microsonda iônica U-Th-Pb (SHRIMP). A principal diferença é que o volume do *pit* escavado pelo laser é muito maior do que aquele escavado pela técnica de microsonda iônica, e por isso a técnica de ablação por laser pode ser considerada uma técnica destrutiva (Patchett & Samson, 2005) – **ver figura 4.1.**

As primeiras tentativas para diretamente datar zircões utilizando as técnicas de LA-ICP-MS envolveram o uso de lasers de Nd-YAG para a ablação de zircões, operando com comprimentos de onda de 1064 nm (Fryer et al. 1993). Devido à variação significativa das razões U/Pb esses primeiros estudos por ablação a laser concentraram-se na determinação de idades $^{207}\text{Pb}/^{206}\text{Pb}$, que geravam precisão em torno de 0,5 a 6%. No entanto, o fracionamento elementar diminuiu com a diminuição do comprimento de onda do laser e então pela quadruplicação de lasers de Nd-YAG (266 nm), ou utilizando lasers à base de gás operando na faixa UV (tais como lasers de Ar-F que produzem lasers com comprimento de onda de 193nm), idades U/Pb com maior fidedignidade podem ser obtidas e comparadas com aquelas que utilizam lasers com maiores comprimentos de onda (Patchett & Samson, 2005). Para contrabalancear os efeitos do fracionamento devido à indução do laser, foram analisados padrões conhecidos sob as mesmas condições de calibração e aplicados fatores de correção para os zircões desconhecidos (Patchett & Samson, 2005). A precisão de idades $^{206}\text{Pb}/^{238}\text{U}$ utilizando padrões externos conhecidos em lasers com capacidade quadruplicada é parcialmente dependente da disponibilidade da concentração suficiente de urânio e chumbo, no entanto valores corriqueiramente típicos apresentam concentrações da ordem de vários %. O fracionamento elementar produzido pela ablação é ainda significativo e de qualquer modo correções devem ser aplicadas para a obtenção de idades precisas, as razões Pb/U medidas são menores que as razões reais e o efeito é variável de acordo com o tempo de ablação (Patchett & Samson, 2005).

Nos anos 90 espectrômetros de massa com setores magnéticos e uma gama de coletores Faraday foram acoplados na fonte de plasma. Esses instrumentos multi-coletores (LA-MC-ICP-MS) produzem os mesmos picos planos produzidos pela técnica TIMS e, portanto, são capazes de realizar medidas de razões isotópicas

com maior precisão do que utilizando quadripolos. Existe um considerável interesse na utilização destes novos equipamentos na datação de cristais de zircão e monazita seguindo os mesmos moldes originalmente desenvolvidos para os instrumentos quadripolos.

Os dados U-Pb em zircões detríticos apresentados nesta Tese foram obtidos por meio do espectrômetro da marca Finnigan Neptune acoplado em um laser ArF excimer ($\lambda = 193 \text{ nm}$) da marca Photon, instalado no IGc/USP. Os *mounts* contendo os zircões foram limpos em solução HNO_3 (3%) e depois com água ultralimpa. A ablação foi realizada a uma frequência de 6 Hz e intensidade de 6 mJ gerando uma cavidade (*spot*) de 29 μm no zircão investigado. O material foi carregado pelos gases Ar (0.7 l/min) e He (0.6l/min) em análises de 60 ciclos de 1 segundo. Os zircões investigados foram intercalados com o padrão GJ-1, seguindo a sequência: 2 brancos, 3 padrões, 13 zircões, 2 brancos e 2 padrões. O dado bruto foi reduzido e corrigido para o chumbo comum presente (^{204}Pb), interferências de fundo (*background*) e qualquer possível viés instrumental utilizando-se uma planilha excel desenvolvida pelo CPGeo/USP. As idades foram calculadas por meio do programa Isoplot 3.0 (Ludwig, 2003).

4.3. Isótopos de Sr-Nd

As composições isotópicas Sr-Nd em rocha total foram determinadas pela técnica TIMS (*thermal ionization mass spectrometer*) por meio do espectrômetro VG354 equipado um mono detector Faraday no IGc/USP. As amostras foram digeridas em ácido e os elementos de interesse foram separados em colunas iônicas seguindo os procedimentos descritos em Sato et al. (1995). As razões $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ foram calculadas a partir de análises por Fluorescência de raios-X (Rb e Sr) e ICP-MS (Sm e Nd) realizadas paralelamente.

4.4. Isótopos de Hf em zircão (LA-MC-ICP-MS)

As análises Lu-Hf em zircão também realizadas no espectrômetro Finnigan Neptune acoplado em um laser ArF excimer ($\lambda = 193 \text{ nm}$) Photon, instalado no IGc/USP. O diâmetro da cavidade (*spot*) foi de 39 μm com tempo de ablação de 60 segundos a uma frequência de 7 Hz com o material carregado pelo gás He (0.6l/min) (Sato et al., 2009). As razões $^{176}\text{Hf}/^{177}\text{Hf}$ foram normalizadas para $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$. Os isótopos ^{172}Yb , ^{173}Yb , ^{175}Lu , ^{177}Hf , ^{178}Hf , ^{179}Hf , ^{180}Hf , e $^{176}(\text{Hf}+\text{Yb}+\text{Lu})$ foram simultaneamente medidos. A razão $^{176}\text{Lu}/^{175}\text{Lu}$ de 0.02669 foi usada para o cálculo da razão $^{176}\text{Lu}/^{177}\text{Hf}$. Para a correção do viés de massa nas razões isotópicas Lu-Hf foram empregadas variações obtidas a partir padrão GJ-1 (Sato et al., 2009). A constante de decaimento utilizada para o ^{176}Lu de 1.867×10^{-11} (Söderlund et al., 2004) e os valores do condrito para as razões $^{176}\text{Hf}/^{177}\text{Hf} = 0.282772$ e $^{176}\text{Lu}/^{177}\text{Hf} = 0.0332$ (Blichert-Toft and Albarede, 1997) foram adotados para o cálculo dos valores de ϵ_{Hf} . As idades modelo em duplo estágio foram calculadas usando a razão inicial $^{176}\text{Hf}/^{177}\text{Hf}$ do zircão e a razão $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ para a crosta continental inferior (Griffin et al., 2004).

4.5. Isótopos de O em zircão (SHRIMP)

As composições isotópicas de Oxigênio foram obtidas em seções analíticas posteriores utilizando o SHRIMP-II equipado com uma fonte de Cs na *Research School of Earth Science (RSES)* da *Australian National University (ANU)*. Detalhes da metodologia estão descritas em Ickert et al. (2008). O padrão utilizado foi o TEMORA 2 ($\delta^{18}\text{O} = 8.2\text{‰}$; Black et al., 2004) analisado em conjunto com o padrão FC-1. Os valores isotópicos das análises do FC-1 no SHRIMP-II, normalizados pelo TEMORA 2 forneceram um valor médio de $\delta^{18}\text{O}$ de $5.5 \pm 0.3\text{‰}$

4.6. Química mineral

4.6.1. Zircão e rutilo (LA-ICP-MS)

Elementos traços e terras raras em zircão e rutilo foram analisados por LA-ICP-MS (*laser-ablation inductively coupled plasma mass spectrometry*) na *Research School of Earth Sciences* da *Australian National University* (ANU). As análises foram realizadas no espectrômetro quadrupolo Agilent 7500s conectado a uma célula de ablação 'HelEx' (Eggins et al., 1998) desenvolvida para receber um feixe de ArF Excimer laser (193 nm) pulsado a 5 Hz com energia de saída de 100 mJ. O instrumento foi calibrado para máxima sensibilidade e mínima produção de espécies moleculares mantendo a razão ThO^+/Th^+ em $<0.5\%$. O laser foi operado de forma a gerar uma cavidade de 48 μm . O tempo de análise total foi de 60 s, sendo os primeiros 25 s representantes do background antes da ablação. Vidros sintéticos (NIST 612 para o zircão e NIST 610 para o rutilo) foram utilizados para calibração externa e valores de referencia foram retirados de Pearce et al. (1997). Padrões internos foram SiO_2 (32.45 peso eq. %) para o zircão e TiO_2 (98 peso eq. %) para o rutilo. O vidro natural BCR-2G foi utilizado como um padrão secundário para monitorar acurácia das análises. O tratamento dos dados foi realizado no programa *Iolite* (Paton et al., 2011). Valores para a normalização do condrito foram de Sun and McDonough (1989).

4.6.2. Outros silicatos (ME)

As análises de química mineral para elementos maiores foram realizadas em seções delgadas polidas utilizando uma microsonda eletrônica CAMECA SX100 na *Research School of Earth Sciences* da *Australian National University* (ANU), operando no modo de comprimento de ondas dispersivo. A voltagem e corrente do feixe foram de 15 kV e 20 μA com um feixe focado de 1 μm para piroxênio e granada and 5 μm para micas. O tempo de contagem por elemento foram de 20 s para Na, Mg, Si, Al, K, Ca e Fe e 60 s para Ti, Cr e Mn. K e Na foram sempre analisados primeiramente na rotina analítica. Varias análises em cada fase mineral foram realizadas para obter composições representativas de pares de núcleo e borda também foram investigadas para acessar padrões de zoneamento. Minerais sintéticos e naturais foram utilizados como padrões e que todos foram determinados na camada Ka de emissão de pico. Outros minerais do acervo foram utilizados como padrões secundários.

4.7 Referencias

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5. U-Pb detrital zircon provenance of metasedimentary rocks from the Ceará Central and Médio Coreau Domains, Borborema Province, NE-Brazil: Tectonic implications for a long-lived Neoproterozoic active continental margin

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Abstract

U-Pb geochronological analyses have been acquired on detrital zircon grains collected from 14 samples in the Ceará Central (CCD) and Médio Coreau (MCD) tectonic domains, separated by the Transbrasiliano-Kandi Lineament (TKL) on the Borborema Province. To the West of this lineament, the basement of the MCD has a tectonic affinity with the West African and São Luis Cratons, where Paleoproterozoic rocks predominate. Within this area, three samples of the São Joaquim quartzite yielded Paleoproterozoic and Archean ages only, with the youngest $^{206}\text{Pb}/^{207}\text{Pb}$ ages at *ca.* 1750 Ma. However, the Goiabeira Formation schist, in the same domain, yielded younger $^{207}\text{Pb}/^{235}\text{U}$ ages around 720 Ma, with predominance of zircons within the 750-1100 Ma interval. To the East of the TKL, the Borborema Province correlates well with the Transaharan Belt of West Africa, where the Paleoproterozoic-Archean basement was affected by a strong tectonic imprint of the Neoproterozoic Brasiliano-Pan African orogeny. Three samples from the region between the TKL and the Tamboril-Santa Quitéria granitic-migmatitic Complex (TSQgmC) yielded younger $^{207}\text{Pb}/^{235}\text{U}$ concordant ages (*ca.* 660-700 Ma), which are 20-60 Ma younger than those of the Goiabeira Formation. Further east within the CCD, eastern of the TSQgmC, six samples yielded two distinct detrital patterns, with some samples showing younger concordant ages (*ca.* 900 and 750 Ma), while others demonstrate a strong Paleoproterozoic source component with zircons older than *ca.* 1500 Ma. Finally, one sample of the post-collisional extensional Jaíba Trough, within the main axis of the TKL, yielded a maximum deposition age of *ca.* 540 Ma, with a strong source component ranging from 540 to 640 Ma, derived mainly from the TSQgmC. Evidence of a large ocean basin prior to 800 Ma, along the vicinity of the TKL, is well established in Central Brazil and West Africa. We believe that the strong source component at around 800-700 Ma, and possibly also the older ones at *ca.* 1000 Ma, were shed from magmatic arc systems developed at the margins of the ancient continent which makes up the basement of the Borborema Province. Abrupt cessation of the detrital zircons input at *ca.* 650 Ma, suggests a change in the tectonic regime at this time, marking the onset of collisional tectonics and main metamorphism in the province. This fact is supported by the ages spanning the 640-590 Ma interval found in the metamorphic domains of the studied zircons which are partially synchronous with the development of the main tectonic and magmatic stages of the TSQgmC.

Key words: Borborema Province, Provenance, Detrital Zircon

5.1. Introduction

Detrital zircon age analysis has proven to be a strong tool to unravel tectonic histories and paleogeographical reconstructions of high grade Precambrian terranes, with numerous published studies in the last 20 years. Interpretation of detrital zircon signatures rely on the comparison of the acquired detrital age spectra with the age of surrounding exposed rocks (Condie et al., 2009), however preservation bias of the source rocks must be considered (Hawkesworth et al., 2009). Regardless of interpretation problems, the technique not only provides a useful method in establishing maximum age limits for sedimentary systems, but also valuable information on

whether a particular sedimentary basin evolved in an active or a passive margin setting or yet within an intracontinental environment (Cawood and Nemchin, 2001; DeGraaff-Surpless et al., 2002; Chew et al., 2008; Sun et al., 2009; de Araujo et al., 2010).

The Central Ceará and Médio Coreaú domains of the Borborema Province are located in a pivotal position to understand West Gondwana reconstruction (De Wit et al., 2008). The provenance of the supracrustal rocks of these domains, and their role in the context of the Brasiliano – Pan-African orogenesis is crucial to restore paleogeography of part of Gondwanaland and pre-drift correlations between South America and West Africa.

In this contribution we provide LA-MC-ICP-MS U-Pb detrital zircon geochronology of fourteen samples from metasedimentary rocks of the Ceará Central and Médio Coreaú domains. The main goals of this paper are: 1) compare the provenance of the Médio Coreaú and Ceará Central domains to test the hypothesis of the Transbrasiliano-Kandi Lineament as a major terrane boundary; 2) obtain further information on depositional ages of the different metasedimentary units; 3) provide information about the source areas from which the sediments have been shed and finally; 4) establish paleogeographic reconstructions by comparing ages of the detrital zircons with the ages of the surrounding possible sources, including stable cratonic areas (during the Neoproterozoic orogenesis) and crust-forming events (exposed or not) associated with the formation of the West Gondwana.

5.2. Geological setting

The Ceará Central and Médio Coreaú domains are crustal entities of the northern portion of the Borborema Province (Almeida et al., 1981) which lies in the northeastern portion of the South America Platform (**fig. 5.1**). This province is characterized by magmatic, tectonic, and thermal phenomena spanning from the Archean to the Cambrian-Ordovician period. Final arrangement was accomplished mainly through the Neoproterozoic Brasiliano/Pan-African orogenesis caused by the convergence of major “cratonic” blocks such as the Amazonian-São Luiz-West Africa and the São Francisco-Congo, including the participation of minor blocks during the assembly (ca. 900-530 Ma) of West Gondwana (Brito Neves and Cordani, 1991; Brito Neves et al. 2000; Arthaud et al., 2008).

The northern Borborema Province is limited in the south by the E-W dextral Patos Shear Zone and it is subdivided into four distinct geological domains separated by large NE-SW to ENE-WSW shear zones (Arthaud et al., 2008), including the Ceará Central and Médio Coreaú domains. These domains are separated by the Transbrasiliano Lineament, a major NE-SW trending shear system, which extends into the African territory by the name of Kandi Lineament.

5.2.1. The Médio Coreaú Domain (MCD)

Regional geological framework of the Médio Coreaú Domain was described in detail by Santos et al. (2008a) and Santos et al. (2009b). This domain is primarily composed of: 1) Early Paleoproterozoic gray orthogneiss and migmatite basement from the Granja Complex; 2) Late Paleoproterozoic (?) volcano-sedimentary Saquinho Unit; 3) Neoproterozoic Martinópolis and Ubajara Groups and 4) Post-collisional Chaval and

Figure 5.1 – Geological setting of the Ceará Central and Médio Coreau domains and sample site location (modified from Cavalcante et al., 2003, de Araujo et al., 2010a and De Wit et al., 2008).

Tucunduba (ca. 590 Ma), Meruoca and Mucambo Granitoids (ca. 530 Ma) (**fig. 5.1**).

The Granja Complex is composed of juvenile 2.35-2.27 Ga metatexitic orthogneisses of tonalitic to granodioritic composition and high grade rocks such as kinzigit, charnockite and enderbite (Nogueira Neto et al., 1989; Gaudette et al., 1993; Monié et al., 1997; Fetter, 1999; Santos et al., 2008a). Although Siderian ages (ca. 2.3 Ga) are recorded in the rocks of Granja Complex, Pb-Pb evaporation Rhyacian ages at 2088 ± 24 and U-Pb ID-TIMS at 2176 ± 21 Ma are also reported (Gaudette et al., 1998). The Martinópolis Group is divided into four formations represented from bottom to top by the Goiabeira, São Joaquim, Covão and Santa Terezinha formations (Santos et al., 2008a). The Goiabeira Formation is in tectonic contact with the Granja Complex (Santos et al., 2008b) and it is composed of metapelites, schists and paragneisses (Cavalcante, et al., 2003, Santos et al., 2008a). São Joaquim Formation comprises several quartzite layers with variable mineralogical compositions including kyanite, sillimanite and muscovite. The Covão and Santa Terezinha formations are composed of schists and metadiamictites and metapelites, metacarbonates, with intercalations of metagreywacke, metarhythmites, quartzite and felsic metavolcanic rocks (Santos et al., 2008a). The Ubajara Group is interpreted as a proximal platformal sequence (Santos et al., 2008a) composed of three distinct formations. The Caiçaras Formation is constituted of low-grade fine-grained siliciclastic rocks and the Trapiá-Frecheirinha of metasandstone and metacarbonate while the Coreaú Formation comprises metasandstone and metagreywacke. Fetter et al. (2003) suggest a Neoproterozoic age of deposition at ca. 777 Ma for the Martinópolis Group, based on ID-TIMS U-Pb determinations in zircons from felsic metavolcanic rocks found within São Joaquim and Santa Terezinha Formations. Four granitoid plutons can be recognized in the MDC. The oldest is the deformed Chaval granitoid with an ID-TIMS U-Pb upper intercept age of 591 ± 10 Ma, followed by the deformed Tucunduba granitoid which yielded an ID-TIMS U-Pb upper intercept age of 563 ± 17 Ma (Santos et al., 2008a; Fetter, 1999). The non-deformed Meruoca and Mucambo plutons are younger and comprise alkali to peralkalic granites (Sial et al., 1981). The Mucambo pluton yielded an ID-TIMS U-Pb upper intercept age at 532 ± 7 Ma (Santos et al., 2008a; Fetter, 1999), while the Meruoca pluton yielded a slightly younger weighted mean U-Pb SHRIMP age at 523 ± 10 Ma (Archanjo et al., 2009).

5.2.2. The Ceará Central Domain (CCD)

This domain is composed of (1) Archean remnants of tonalite-trondhjemite-granodiorite (TTG) units of the Cruzeta Complex; (2) juvenile Paleoproterozoic (2.1–2.2 Ga) and high-grade felsic to intermediate orthogneisses and migmatites, including their associated supracrustal rocks of the Algodões Unit (Fetter et al., 2000; Martins et al., 2009); (3) high-grade Early Proterozoic to Neoproterozoic supracrustal rocks partially represented by the units of the Ceará Complex (e.g. Arthaud, 2007; Arthaud et al., 2008) and the Novo Oriente Group (Ganade de Araujo et al., 2010a); (4) granitoids and migmatites of the Tamboril-Santa Quitéria Complex (Fetter et al., 2003; Arthaud et al., 2008); and (5) widespread Neoproterozoic to Ordovician post-collisional to anorogenic granitoids (Fetter, 1999; Castro et al., 2012). The first two associations acted as the basement for the so-called Brasiliano–Pan- African Neoproterozoic orogenesis (**fig. 5.1**).

This basement, particularly the Cruzeta Complex, records an intricate geological history, from the Archean to the Paleoproterozoic period. The Archean remnants (2.85-2.64 Ga). The Ceará Complex is composed of metapelites, metasemipelites and metagreywacke normally showing a prominent schistosity to gneissic fabric; regionally and locally migmatized. Quartzites, marble, calc-silicate rocks and amphibolites also form large

tracts within this complex (Cavalcante et al., 2003; Caby and Arthaud, 1986; Arthaud et al., 2008). Taking into account the degree of partial melting, Cavalcante et al. (2003) divided part of the Ceará Complex into the Independência and Canindé units. The supracrustals rocks preserved from extensive migmatization were grouped in the former, while those that exhibit severe conditions of melting were included in the latter. The age of the Ceará Complex is still a matter of debate. Some authors consider the time of deposition at around 770 Ma (Fetter et al., 2003; Castro, 2004; Arthaud, 2007) based on U-Pb ID-TIMS geochronological data from sheets of granitic gneisses that have been interpreted as alkaline syn-sedimentary rhyolite flows and/or sills. Based on SHRIMP U-Pb detrital zircon studies, Arthaud (2007) argued that the continental fragmentation and passive margin development was around 750 Ma. However, considering that the Early Paleoproterozoic–Archean crust is ubiquitously found as the basement of younger supracrustal rocks in CCD; the lack of older detrital zircons than 1.8 Ga is inconsistent with proposed intracontinental rift setting.

The Neoproterozoic Tamboril-Santa Quitéria granitic-migmatitic Complex (TSQgmC) is a wedge-shaped composite anatetic/igneous association, characterized by a number of magmatic pulses where large volumes of magma intruded in the form of veins, layers, sheets and plutons (Cavalcante et al., 2003; Fetter et al., 2003; Arthaud et al., 2008). The plutonic rocks display syn- to late-magmatic deformation that was in part coeval with the injection of younger and less deformed magma (Arthaud et al., 2008). In general they range from mafic diorite to granite, with predominance of monzogranitic/granitic compositional members (Ganade de Araujo et al., 2012). This plutonic association intruded the supracrustal rocks of Ceará Complex, which are preserved as large resisters, pendants and enclaves of calc-silicate rocks, amphibolites and quartzite, probably derived from infertile portions unable to melt. The granitoids related to the development of this complex range from 650 to 610 Ma (Fetter et al., 2003; Castro, 2004; Santos et al., 2007, de Araujo et al., 2012). However, provenance studies on the adjacent supracrustal rocks (Arthaud, 2007), coupled with a Pb-Pb evaporation zircon age at 795 Ma on granodioritic gneisses from the eastern border of the complex (Ganade de Araujo et al., 2010b) suggest that its development could have started as early as 800 Ma. The Nd isotopic signatures are consistent with variable mixtures between juvenile Neoproterozoic magmas and the surrounding Paleoproterozoic gneisses, indicating that the complex neither represents a juvenile arc nor a suite of crustal melts (Fetter et al., 2003). The tectonic setting of the Tamboril-Santa Quitéria Complex has been interpreted as a Neoproterozoic Andean-type magmatic arc (Fetter et al., 2003), however recent discussions propose a collisional Himalayan setting with an early Andean arc component reworked during the collisional event (Ganade de Araujo et al., 2010b; Costa et al., 2010; de Araujo, 2011; de Araujo et al., 2012).

Magmatic pulses related to the emplacement of granitoids at ca. 580, 530 and 480 are representative of post-collisional manifestations in the Ceará Central Domain (Ganade de Araujo, 2011). The final ca. 480 Ma A-type pulse is marked by small semi-circular stocks which are in some extent temporally close to the deposition of the first strata of the Phanerozoic Parnaíba Basin, developed in the west portion of the collisional Neoproterozoic mountain chain (Castro et al., 2012; Ganade de Araujo, 2011).

5.2.3. The Jaibaras Trough

The Late Neoproterozoic to Early Paleozoic Jaibaras basin is inserted along the main axis of the Transbrasiliano-Kandi Lineament. It consists of a basal fault-scarp-related paraconglomerates package

(Massapê Formation) followed laterally and vertically by fluvial-lacustrine sandstones (Pacujá Formation) interbedded with reddish shales (Oliveira and Mohriak, 2003). The upper Aprazível Formation consists of polymitic conglomerates, and differs from the underlying Massapê Formation by the presence of clasts of plutonic/volcanic rocks from the Meruoca-Mucambo plutons and Parapuí Suite (Oliveira and Mohriak, 2003). The axis of the Jaíbaras Trough is aligned with the main depocenter of the Parnaíba Basin further northwest and it constitutes its precursory stage that was slightly inverted before the initial deposition of the large intracratonic basin.

5.3. Sampling and analytical procedures

5.3.1. Sampling strategy

Taking the Transbrasiliano-Kandi Lineament (TKL) and the Tamboril-Santa Quitéria Complex as regional markers, our sampling approach intended to answer two fundamental questions. The first one is whether the Transbrasiliano-Kandi Lineament demarks a terrane boundary, if so, do the MCD and CCD have distinct provenance signatures or do the detrital ages overlap favoring a correlated development of both domains? The second question is: are there basins related to the Neoproterozoic Tamboril-Santa Quitéria Complex within the adjacent supracrustal rocks in CCD? Bearing these questions in mind, we have collected four samples in the MCD, to the west of the Transbrasiliano-Kandi Lineament, including three quartzites from the São Joaquim Formation (DKE-36, DKE-39, DKE-41) and one schist from the Goiabeira Formation (DKE-30) both from the Martinópolis Group. One reddish sandstone of the Pacujá Formation (DKE-25) from the Jaíbaras Trough, inserted in the main axis of the lineament. Three samples in the zone between the lineament and the Tamboril-Santa Quitéria Complex including two metatextitic paragneiss (DKE-43 and DKE-45) and one quartzite rhythmically interleaved with metapelitic layers (DKE-19). Six samples to east of the Tamboril-Santa Quitéria Complex which include five quartzites (DKE-51, DKE-06, DKE-56, NCEB-427 and NCEB-351) and one metatextitic paragneiss (DKE-53).

5.3.2. Sample preparation

Zircons were separated from the crushed rocks (3-5 kg) using conventional and heavy liquid and magnetic techniques (jaw crusher, disk grinder, Wilfley table, Frantz isodynamic magnetic separator and density separation using bromoform and methylene iodide). To avoid bias introduced during handpicking, no visual morphological or color differentiation was made. Around 250-300 zircons from each sample were mounted in epoxy resin, polished to half of mean grain thickness for further imaging with transmitted light and 43athodoluminescence to unravel its internal complexities. Dodson et al. (1988) states that 59 grains must be analyzed to achieve 95% confidence of finding every population that exists at the 5% level in a given sample, however Vermeesh (2004) taking into account different levels of probability based upon number potential sources areas states that 117 grains should be dated. For each sample about 50 to 80 zircons were dated comprising a total of 836 U-Pb age measurements.

5.3.3. CL imaging and U-Pb geochronology

Cathodo-luminescence (CL) images of zircons were obtained using a Quanta 250 FEG electron microscope equipped with Mono CL3+ cathodoluminescence spectroscope (Centaurus) at the Geochronological Research Center in São Paulo University, Brazil. After that, U–Pb analyses by LA-MC-ICP-MS were carried out using the Finnigan Neptune coupled to an excimer ArF laser ($\lambda=193\text{ nm}$) ablation system also at the Geochronology Research Center of the São Paulo University. The mounts containing zircons were cleaned in a HNO_3 solution (3%) and in ultraclean water bath. The ablation was done with spot size of $29\text{ }\mu\text{m}$, at a frequency of 6 Hz and an intensity of 6mJ. The ablated material was carried by Ar ($\sim 0.7\text{ L/min}$) and He ($\sim 0.6\text{ L/min}$) in analyses of 60 cycles of 1s. Unknowns were bracketed by measurements of the international standard GJ-1, following the sequence 2 blanks, 3 standards, 13 unknowns, 2 blanks, and 2 standards. Raw data were reduced using a home-made spreadsheet and corrections were done for background, instrumental mass bias drift and common Pb. The ages were calculated using ISOPLOT 3.0 (Ludwig, 2003).

5.4. Results

5.4.1 CL images

Selected CL images of dated zircons are given in **figure 5.2**. Zircons display a wide range of morphology and internal textures with predominance of concentric oscillatory zoning, a typical feature of primary magmatic crystals (Corfu *et al.*, 2003). In the majority of samples zircons show rounded grain morphology, suggesting that they have experienced a long distance of transportation. Zircons extracted from sample DKE-25 of the Jaíba Trough are euhedral indicating a near-source region. Xenocrystic cores with magmatic overgrowths are also present in some examples (cf. DKE-30-2.1 and 11.1). Homogeneous strong dark crystals in CL (trace element-rich zircons – cf. DKE-06-35.1) and bright (trace element-poor zircons – cf. DKE-53-29.1) are abundant; however their dark colors in CL images can be suspected of metamictization. Complex zoned zircons and fractured zircons, with domains smaller than the beam diameter were avoided since the results would reflect mixed isotopic patterns. Secondary sub-solidus modifications caused by subsequent thermal events are easily recognizable in some samples. They constitute overgrowths around the rounded pre-existing detrital cores, characterized by low and bright luminescence rims. These overgrowths are predominantly homogeneous and regular concentric around the cores (cf. DKE-45-22.1 and 8.1), but concentrated sometimes at the crystal terminations (cf. DKE-43-42.1 and DKE-53-11.1).

5.4.2. U-Pb ages of detrital zircons

We performed 836 analyzes on detrital zircons and some metamorphic overgrowths. Among them 647 yielded reliable ages with less than 5% discordance ($^{206}\text{Pb}/^{238}\text{U}$ to $^{207}\text{Pb}/^{235}\text{U}$) and 10% discordance ($^{206}\text{Pb}/^{238}\text{U}$ to $^{207}\text{Pb}/^{206}\text{Pb}$). Ages are reported in terms of the $^{207}\text{Pb}/^{206}\text{Pb}$ ratios for grains older than 1.4 Ga, with correlated discordance of $^{206}\text{Pb}/^{238}\text{U}$ to $^{207}\text{Pb}/^{206}\text{Pb}$, or the $^{206}\text{Pb}/^{238}\text{U}$ ratios for zircons younger than 1.4 Ga, with correlated discordance of $^{206}\text{Pb}/^{238}\text{U}$ to $^{207}\text{Pb}/^{235}\text{U}$. Sample descriptions and results (**tables 2 to 15**) are available in the **appendix I** and **supplementary data**.

5.4.2.1. Region to the west of the Transbrasiliiano-Kandi Lineament – Médio Coreá Domain

Samples DKE-36, DKE 39 and DKE-41 (São Joaquim Quartzite – Martinópolis Group). A total of 180 zircons from three samples of São Joaquim Quartzite were analyzed, and among them, 129 yielded reliable Paleoproterozoic and Archean U-Pb ages. The dominant population consists of ages ranging from 2300 Ma to 2050 Ma (59% of total) and from 2500 Ma to 2300 Ma (20% of total) with the youngest concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age at 1745 ± 16 Ma (sample DKE-41) (**fig. 5.3A, B and C**).

Sample DKE-30 (Goiabeira Schist – Martinópolis Group). The youngest $^{206}\text{Pb}/^{238}\text{U}$ age at 705 ± 07 (98% concordant) and the youngest population at *ca.* 720 Ma provides the maximum age of deposition for the Goiabeira schist. Main age clusters are at 705-754 Ma (15% of total), 768-883 Ma (45%) and 930-1139 Ma (16%). A minor population is represented by five Paleoproterozoic grains that spread between 1817 and 2465 Ma and four Archean grains which the oldest concordant $^{207}\text{Pb}/^{206}\text{Pb}$ zircon yielded an age of 2766 ± 13 Ma (**fig. 5.3D**).

5.4.2.2. Region between the Transbrasiliano-Kandi Lineament and the Tamboril-Santa Quitéria Complex – Ceará Central Domain

Sample DKE-25 (Reddish Pacujá sandstone – Jaíba Trough). The youngest zircon found in the Pacujá sandstone within the Jaíba Trough yielded an age of 530 ± 08 Ma (98% concordant) with the youngest population at *ca.* 550 Ma. The Neoproterozoic zircons can be divided into two groups ranging from 550 Ma to 598 Ma (42% of total) and from 600 to 630 (20% of total). Paleoproterozoic zircons were also identified in this sample ranging from 2110 Ma to 2193 Ma (**fig. 5.3E**).

Samples DKE-43 and DKE-45 (metatexitic paragneisses – Canindé Unit of Ceará Complex). The youngest zircon of this sample yielded a $^{206}\text{Pb}/^{238}\text{U}$ age of 648 ± 15 Ma (sample DKE-43) with the young population established at *ca.* 660 Ma. A young low-U homogeneous zircon (#44.1/sample DKE-43) presented a $^{206}\text{Pb}/^{238}\text{U}$ age of 632 ± 20 Ma which is similar to the ages obtained in the metamorphic overgrowths in other zircons from this study, and was therefore interpreted as a metamorphic zircon. Similarly, for the sample DKE-45 the youngest $^{206}\text{Pb}/^{238}\text{U}$ age is at 692 ± 09 Ma, with the youngest population clustering around 705 Ma. Two important Neoproterozoic populations were verified in these samples ranging from *ca.* 700 to 780 Ma (24% of total) and from *ca.* 810 to 920 Ma (40% of total), respectively (**fig. 5.3F and 5.3G**).

Sample DKE-19 (muscovite-sillimanite quartzite – Independência Unit of Ceará Complex). Zircons were extracted from a quartzitic layer of a rhythmically sequence which the youngest one yielded a $^{206}\text{Pb}/^{238}\text{U}$ age at 903 ± 13 Ma, with the youngest population clustering around 940 Ma. The Neoproterozoic zircons (10% of total) spread from 903 to 973 Ma. Mesoproterozoic grains (22 % of total) form a second group comprising the 1027-1196 Ma time interval. Paleoproterozoic zircons (27 % of total) are arranged in two peaks roughly at 1850 and 2000 Ma. Three Archean zircons were dated and the oldest one yielded a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age at 2919 ± 12

Figure 5.2 – Selected CL images and spot placement from the analyzed zircons.

Figure 5.2 – (continued)

Ma. Some metamorphic overgrowths were analyzed, however only one (#32.1) yielded a reliable concordant $^{206}\text{Pb}/^{238}\text{U}$ age at 604 ± 05 Ma (**fig. 5.3H**).

5.4.2.3. Region to the east of the Tamboril-Santa Quitéria Complex – Ceará Central Domain

Sample DKE-51 (muscovite-bearing quartzite – Independência Unit of Ceará Complex). Youngest $^{206}\text{Pb}/^{238}\text{U}$ age at 870 ± 22 Ma and the youngest population at *ca.* 900 Ma constrain the maximum deposition age for this sample. Two younger groups can be observed; one with Neoproterozoic grains ranging from 870 to 967 Ma (20% of total) and a second Late Mesoproterozoic group ranging from 1014 to 1197 Ma (17% of total). Few Mesoproterozoic ages around 1250 Ma and 1650 Ma were also detected. Paleoproterozoic zircons are abundant, comprising 41% of the total grains, with peaks roughly around 1773, 1969, 2076, 2133 and 2444 Ma (**fig. 5.3I**).

Sample NCEB-427 (quartzite – Independência Unit of Ceará Complex). From the zircons analyzed in this sample only two yielded reliable $^{206}\text{Pb}/^{238}\text{U}$ Neoproterozoic ages with the youngest constraining the maximum age of deposition at 963 ± 09 Ma. Mesoproterozoic ages comprise 23% of the total ages and peak mainly at 1066, 1117 and 1324 Ma. Paleoproterozoic zircons dominate this sample constituting 67% of the total zircons, with peaks at 1660, 1833, 2019 and 2122 Ma. (**fig. 5.3J**).

Sample DKE-53 (metatexitic paragneisses - Canindé Unit of Ceará Complex). The youngest analyzed zircon yielded a $^{206}\text{Pb}/^{238}\text{U}$ age at 728 ± 11 Ma with the youngest population clustering at *ca.* 750 Ma. Of the total dated zircons, 28% show Neoproterozoic ages and can be divided into two distinct populations ranging from 728 to 790 Ma and from 860 to 983 Ma. Late Mesoproterozoic zircons are also abundant, spanning from 1012 to 1193 Ma. Mesoproterozoic source component at *ca.* 1250, 1320 and 1550 Ma were also detected. Paleoproterozoic zircons are also rather abundant, comprising 25% of the dated zircons with main peaks at *ca.* 1780 and 1887 Ma with some grains around 2020 and 2200 Ma. Three Archean zircons were analyzed and the oldest one yielded a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age at 3039 ± 21 Ma (**fig. 5.3K**).

Sample NCEB-351 (quartzite – Independência Unit of Ceará Complex). Zircons from this sample spread mainly from the Paleoproterozoic to the Archean. One Mesoproterozoic zircon with a $^{207}\text{Pb}/^{206}\text{Pb}$ age at 1579 ± 12 (102% concordant) constrains the maximum age of deposition. Of the dated zircons, 70% are Paleoproterozoic ranging from 2040 to 2196 Ma. Six Archean zircons (20% of total) ranging from 2658 to 2852 Ma were also dated, with the oldest yielding a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age at 3134 ± 09 Ma (**fig. 5.3L**). Dated metamorphic overgrowths are mainly represented by homogeneous very high-U rims around the detrital cores, however some can be texturally zoned resembling primarily magmatic growth zoned zircons. The $^{206}\text{Pb}/^{238}\text{U}$ ages of these metamorphic domains demonstrate a widespread variation ranging from 645 to 540 Ma.

Sample DKE-06 (feldspathic quartzite – Independência Unit of Ceará Complex). With the exception of the youngest zircon which yielded a Mesoproterozoic $^{207}\text{Pb}/^{206}\text{Pb}$ age at 1579 ± 12 (93% concordant), all zircons from this sample have Paleoproterozoic ages. The main population, comprising 55% of the zircons, spread from 2101 to 2190 Ma. The oldest dated zircon yielded a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age at 2439 ± 22 Ma. Three metamorphic overgrowths yielded reasonable $^{206}\text{Pb}/^{238}\text{U}$ concordant ages at 578, 575 and 540 Ma (**fig. 5.3M**).

Sample DKE-56 (feldspathic quartzite – Independência Unit of Ceará Complex). The youngest zircon yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age at 2031 ± 10 Ma. With the exception of a 2591 ± 91 Ma Archean zircon, all the zircons of this sample yielded Paleoproterozoic ages. Two distinct Paleoproterozoic populations can be identified at 2050-2130 Ma and 2160-2210 Ma time intervals (**fig. 5.3N**).

5.5. Discussion

5.5.1. Metamorphism

Several metamorphic domains were dated in the studied zircons. They are represented mainly by high- and low-U overgrowths around the detrital cores detected in the CL images. The ages of these overgrowths are in agreement with U-Pb zircon and monazite metamorphic ages previously reported (640 to 580 Ma – *e.g.* Fetter, 1999; Castro, 2004; Arthaud, 2007; Amaral et al., 2010). Three distinct groups of metamorphic $^{206}\text{Pb}/^{238}\text{U}$ ages are recorded in the selected overgrowths (**fig. 5.3O**). The older group comprises the interval of *ca.* 650-630 Ma, and it is possible associated with the ages of the eclogitic metamorphism dated by Amaral et al. (2010) in the zone between the Transbrasiliano-Kandi Lineament and the Tamboril-Santa Quitéria Complex. The

second and well defined group ranges from ca. 610 to 590 Ma and is correlated with the widespread regional thermal event that affected the Borborema Province, including part of the migmatization and granitoid emplacement (e.g. Fetter, 1999; Arthaud 2007; Castro, 2004; Neves et al., 2009). The younger group clusters around 540 Ma and is arguably related with the emplacement of post-collisional plutons at 540-520 Ma, or even with isotopic disturbance due to late fluid infiltration. If it proves to be regionally reproducible in the CCD, the metamorphic ages clustering at around 647 Ma in the older group defines the minimum age of deposition of the studied samples in this particular domain.

5.5.2. Detrital zircon provenance patterns

The visual comparison of the dated samples in the probability density distribution (**fig. 5.4A**) and cumulative density distribution (**fig. 5.4B**) diagrams allowed the differentiation of four groups with specific provenance patterns. To assess the heterogeneity of the age distributions we also applied the Kolmogorov-Smirnov (K-S) two-sample test (Berry et al., 2001) (**table 1 in the supplementary data**). This test is a means to mathematically compare two detrital zircon distributions and determine if there is a statistically significant difference between the two distributions (Gehrels et al., 2000). The method is independent of any assumptions about the probability distribution of a sample and allows comparison of both age values (peak locations) and distributions (peak shapes) using the (P) parameter (DeGraff-Surpless et al., 2003). The higher the P value, the more likely it is that the two age distributions were drawn from the same population. To be 95% confident that two populations are not statistically different the P value must exceed 0.05 (DeGraff-Surpless et al., 2003).

Group I

This group is represented by samples that contain dominantly Paleoproterozoic and Archean zircons. The three samples from the São Joaquim Formation in the MCD (DKE-36, DKE-39 and DKE-41) share the same pattern defined by a Statherian youngest population (excepted by sample DKE-36) with a main Rhyacian source component combined with the presence of old Archean grains (>3.0 Ga). The *P* values for this group range from 0.221 to 0.944 (**table 1**), indicating significant similarities in the detrital zircon distribution throughout the São Joaquim Formation.

An associated subgroup in the CCD can be defined by the samples NCEB-351, DKE-06 and DKE-56, with *P* values ranging from 0.363 to 0.414 (**table 1**). These samples have few youngest zircons, but no clustering populations, at the Mesoproterozoic era. The main provenance is defined by Rhyacian zircons with a subordinate Siderian component. With the exception of sample NCEB-351, this subgroup is characterized by the lack of Archean zircons. The *P* values of the sample NCEB-351 varies from 0.038 to 0.126 when compared with samples from the São Joaquim Formation, indicating some similarities in the detrital zircon distribution, but not necessarily sharing the same source areas.

Group II

This group is represented by a set of samples located both in the eastern (DKE-19) and western (DKE-51, DKE-53 and NCEB-427) sides of the Tamboril-Santa Quitéria Complex in CCD. It is characterized by a

heterogeneous distribution including Archean to Neoproterozoic zircons. P values for samples DKE-53, DKE-51 and DKE-19 range from 0.457 to 0.070 suggesting similar detrital zircon distribution, especially for samples DKE-19 and DKE-51 that were sampled on the opposite borders of the complex. Sample NCEB-427 has similar detrital zircon distribution of sample DKE-51 (P value of 0.120), however it shows low P values (>0.05) when compared to the other samples of the group. Except for sample DKE-53, the other ones show the youngest clustering of zircons at the Early Neoproterozoic (Tonian) at *ca.* 900-960 Ma. Some zircons from the Late Mesoproterozoic (Stenian 1000-1200 Ma) are also common, but Ectasian (1200-1400 Ma) zircons are subordinate. There are few zircons at *ca.* 1500 Ma, abundant Statherian (1600-1800 Ma) and extremely abundant Rhyacian (2050-2300 Ma) zircons combined with subordinate occurrence of Siderian (2300-2500 Ma) zircons. Few Archean zircons are present, but they are rarely older than 2800 Ma, unlike the Archean zircons from *Group I* that presented older ages. The sample DKE-53 has the same described pattern, however with younger Neoproterozoic zircons around 730 Ma.

Group III

This group includes the samples DKE-43, DKE-45, DKE-30 which is characterized by the abundance of Cryogenian (660-850 Ma) zircons. The youngest population is around 650 Ma for the sample DKE-43 and *ca.* 700 Ma for the samples DKE-45 and DKE-30. Zircons within the interval of 1000-1300 are rare, with the exception of the sample DKE-30 that contains some Stenian zircons. Although less evident, this group also contain some Paleoproterozoic and few Archean zircons suggesting a rather small contribution of older crystalline rocks. The P value for samples DKE-43 and DKE-45 is 0.122, indicating a good statistical fit for these samples from the Canindé Unit of Ceará Complex.

Group IV

This group was defined solely by the sample DKE-25. It is characterized by a distinctive presence of Ediacaran (542-630 Ma) zircons with the youngest population at *ca.* 550 Ma. An important feature of this group is the lack of any Neoproterozoic grain older than 630 Ma. Some Rhyacian grains are present along with a scarce Archean component.

5.3. Potential source areas

Attributing source areas to detrital zircon provenance patterns is not always a straightforward task and can sometimes be highly speculative.

The record of Archean crust is restricted to certain sites inside the Borborema Province. They can be found in the basement of the Ceará Central (e.g. Fetter, 1999) and Rio Grande do Norte (e.g. Dantas et al., 2004) domains, and also in the Sergipano Belt (e.g. Oliveira et al., 2010). The surrounding cratonic areas represented by the São Francisco and Amazonian-West African cratons also contain numerous Archean inliers and blocks (Teixeira et al., 1989; Tassinari and Macambira, 1999; Barbosa and Sabaté, 2004). Paleoproterozoic crust is extensive found in all domains of the Borborema Province. However, at the present

Figure 5.3 – Concordia (right) and relative age probability (left) diagrams for selected zircons analyzed in this study.

Figure 5.3 – (continued)

Figure 5.3 – (continued)

Figure 5.3 – (continued)

level of knowledge, Siderian crust is abundantly found in the basement of MCD (Santos et al., 2008b) with minor expositions in the CCD (Castro, unpublished). This Siderian event was recognized in the SE portion of the Amazon craton (Vasquez et al., 2008) and also in the West African craton (Lemoine et al., 2006) suggesting that these rocks could form a coherent trend, where great part of it is now hidden by the Paleozoic cover of the Parnaíba Basin. The major crust-forming event in the Borborema Province and within the surrounded cratonic blocks took place during the Rhyacian period in the so called Transamazonian orogenesis (Amazon craton) and Eburnian orogenesis (West African craton).

The samples from group I contain basically Paleoproterozoic and Archean zircons. The quartzites of São Joaquim Formation are dominantly composed of Rhyacian zircons, but due to the widespread area of occurrence of the Rhyacian crust it is impossible to ascribe any particular source region. On the other hand, the same samples demonstrate a combined Siderian source component that can be derived directly from the basement of the Martinópolis Group, particularly of the Granja Complex in the MCD or from further west in the SE Amazonian (to the west of the Araguaia Belt) and West African cratons. This interpretation can also explain the occurrence of Archean zircons (>3.0 Ga) that can be shed from the West African craton, largely from the Archean rocks of Liberia, Sierra Leone and Guinea (e.g. Thiéblemont et al., 2001; Thiéblemont et al., 2004) and central regions of Amazon Craton, particularly from the Carajás-Imataca Block (Tassinari et al., 2000). With the exception of the Late Paleoproterozoic (Statherian) zircons, which could be also shed from the Amazon Craton, the São Joaquim quartzite has similar provenance to those of the Novo Oriente Group (Ganade de Araujo et al., 2010). This group is interpreted as a segment of a passive margin, possibly floored by oceanic crust in its distal portion that flanked the eastern part of the Amazonian-West African craton (present day position); in our opinion this interpretation can be extended for the São Joaquim quartzites in the MCD. Samples NCEB-351, DKE-06 and DKE-56 are also composed of Rhyacian zircons, but with a less evident or even lacking of the Siderian and Archean component, suggesting they do not share the same source area to those of the São Joaquim Formation. The more immature characteristics of these samples, evidenced by their high feldspar content, imply in a proximal source area possibly associated with small depocenters within the Paleoproterozoic-Archean basement of the Borborema Province. Few Late Paleoproterozoic (Statherian) zircons could also been shed from the Orós belt (Sá et al., 2002). The source for the ca. 1.5 Ga zircons is still obscure, but Amaral et al. (2010) reported similar ages for mafic rocks within the basement of the CCD.

The Archean and Paleoproterozoic ages found in the samples that contain Neoproterozoic zircons, especially those from group II, can be derived from the former continent (a possible detached portion of the adjacent cratonic areas – possibly the São Francisco-Congo craton), represented today by the reworked basement of Borborema Province. A provenance from the main cratonic areas is not discarded as well, especially from the São Francisco-Congo craton, but in this case the Sergipano Belt would constitute a significant barrier or drainage divide. The source of the Mesoproterozoic zircons at ca. 1100-1400 Ma in the metasedimentary rocks from the Central Borborema Province is uncertain as there are no obvious local sources for these zircons. Neves et al. (2009) argued for a provenance derived from the Amazon craton for the 1100-1400 Ma zircons, however, the Araguaia Belt or basin, which was inverted only at the end of Neoproterozoic (Moura et al. 2008), would constitute an important obstacle for the zircons from further west and northwest of Amazonian craton, particularly from the Sunsás Province (1.45 to 1.1 Ga) (Santos et al., 2008).

Zircons within the 920-1000 Ma interval are possibly derived from the crystalline rocks related to Cariris Velhos event in the Central portion of the Borborema Province (e.g. Santos et al., 2010). The Cryogenian zircons in the interval of 660-850 Ma are widespread in some samples, especially in those from the group III. Related magmatism to this time interval is still unusual in the Borborema Province (Ganade de Araujo et al., 2010; Neves et al., 2011), however it has a strong component in the detrital zircons from this study and from others in the Borborema Province (e.g. Van Schmus et al., 2003, 2011; Neves et al., 2009). Some authors

Figure 5.4 – A. Relative age probability diagram comparing ages of detrital zircon grains from samples of the Médio Coreáú and Ceará Central Domains. **B.** Cumulative age density distribution diagram comparing ages of detrital zircon grains from samples of the Médio Coreáú (MCD) and Ceará Central Domains (CCD). TSQgmC – Tamboril-Santa Quitéria granitic-migmatitic Complex.

proposed an extensional period for this interval. However, the quasi-continuous spectrum of ages ranging from ca. 900 to ca. 700 Ma are not suggestive of such event, since extensional events are more restricted to a certain duration [*e.g.* ca. 30 m.y. for the Red Sea Rift (Hofmann et al., 1995; Kazmim and Byakov, 2000), ca. 30 m.y. for the Main Ethiopian Rift (Corti et al., 2009)]. In addition, magmatic pulses related to extensional settings are short-lived and less fertile in zircons when compared to those of a convergent magmatic arc setting. The lack of immature terrigenous siliciclastic rocks, or evidence to abrupt facies variations in the lithological record of the CCD also do not fit with the proposed extensional setting. Instead, when recognized, the rocks display a rhythmic intercalation of metagreywacke, metapelites and metapsmites coherent with turbiditic deposits associated or not with marbles. Recently, Ganade de Araujo et al. (2010b) presented a Pb-Pb evaporation age of ca. 795 Ma for a metatextic granodiorite-tonalite in the western portion of the Tamboril-Santa Quitéria Complex and based on preliminary geochemical and Nd isotopic data interpreted this manifestation as pre-collisional subduction-related magmatism in the CCD. The presence of abundant grains in the 900-700 Ma interval reinforces the interpretation of long-lived active continental margin, or subduction systems, since the beginning of the Neoproterozoic time.

The Ediacaran zircons within interval of 630-600 Ma found in the sample from the Jaíbaras Trough are clearly derived from the 640-600 Ma Tamboril Santa Quitéria granitic-migmatitic Complex (Fetter et al., 2003; de Araujo et al., 2012), and also from the post-collisional granitoids ranging from ca. 580 to 540 Ma (Fetter, 1999).

5.5.4. Tectonic implications: A long-lived continental margin?

Evidences from West Transaharan orogen in Africa (Caby et al., 1989; Caby, 2003; Berger et al., 2011) and Central Brazil (Pimentel and Fuck, 1992; Pimentel et al., 2000; Laux et al., 2005) demonstrated that part of the Neoproterozoic growth of western Gondwana occurred firstly during the Late Tonian and Cryogenian, through the development of intraoceanic arcs suggesting the presence of a large ocean separating the São Francisco and Amazonian/West African cratons. These accretionary settings subsequently became collisional orogens that reworked the previous continent (basement) during the Late Cryogenian to Early Ediacaran at ca. 650-620 Ma (Liégeois et al., 1987; Piuze et al., 2003; DellaGiustina et al., 2009; de Araujo et al., 2012). The Borborema Province, specially its crustal domains located in the vicinities of the Transbrasiliano-Kandi shear system, represents the link between the Central Brazil and West Sahara orogens. In this scenario, does the region represented by Borborema Province narrowed into a small ocean or does it remained opened configuring a vast oceanic domain connecting the Goiás (in Central Brazil) and Pharusian (in West Africa) oceans?

Two contrasting views regarding the tectonic reconstruction of the Borborema Province are currently in vogue. In one hand, following the assertion of Van Schmus et al. (2008), some believe that Mesoproterozoic to Early Neoproterozoic break-up of a Paleoproterozoic supercontinent created a large region between Congo/São Francisco and Amazonian/West African cratons, consisting of extensional basins floored by Paleoproterozoic-Archean crust, local basins approaching small oceans, and a larger ocean along the main axis of the Transbrasiliano-Kandi shear system. On the other hand, other authors believe in an essentially intracontinental development, where an autochthonous setting is implied (Neves, 2003; Neves et al. 2009).

In fact, the timing of continental break-up, passive margin development, seafloor spreading, subduction and continental collision related to development of West Gondwana in the Borborema Province is still debatable. The *ca.* 1000-920 Ma Cariris Velho event in the Central Sub-Province (Zona Transversal – see Van Schmus et al., 2011) is regarded as orogenic (Santos et al., 2010); however the lack of any metamorphism of this age has led some authors to interpret this event as an intracontinental extension (*e.g.* Neves, 2003). In the Ceará Central and Médio Coreau domains, some authors favour break-up and rift development at around 770 Ma (Fetter et al., 2003; Castro, 2004; Arthaud, 2007), as proposed by Oliveira and Tarney (1990) for the Canindé Rift within the Sergipano Belt, further southwest.

In the Borborema Province, arc building period is conventionally constrained at *ca.* 640-610 Ma (Fetter et al., 2003; Van Schmus et al., 2011) with initial collisional tectonics at *ca.* 610-590 Ma. Such a short time for arc construction (*ca.* 30 m.y.) is not consistent with a consumption of a large oceanic setting, hosting intraoceanic arcs as evidenced in Central Brazil and in the West Transaharan orogen in Africa. The abundance of detrital zircons younger than the Cariris Velhos event, but older than the main phase of metamorphism and deformation (*ca.* 650 Ma) indicate that this time interval was also an important period of crustal growth in the Borborema Province as noted in the Central Brazil and Northwestern Africa. Although still poorly characterized, a Pb-Pb zircon evaporation age of *ca.* 795 Ma (Ganade de Araujo et al., 2010) in CCD and U-Pb ages at *ca.* 870 and 851 Ma (Neves et al., 2011) in the Central Borborema Province attest for the regional expression of granitoids within this time interval. In our opinion, this period is defined by the pre-collisional arc related systems compatible with a long lived continental margin setting that were overprinted by the collisional tectonics at *ca.* 640-610 Ma (**fig. 5.5**). The sudden cessation of detrital zircons at *ca.* 650 Ma indicates an abrupt change in the tectonic regime arguably caused by the onset of collisional tectonics and inversion of the basins that were fed by arc-related detritus. The absence of Neoproterozoic zircons older than *ca.* 650 Ma in the Jaíba Trough, inserted within the Transbrasiliano-Kandi Shear System, indicate that great part of the provenance was derived from the collisional front associated with the late extensional phases of the orogeny.

The similar provenance of the Goiabeira Formation to the west to the Transbrasiliano-Kandi Lineament to those of the supracrustal rocks to the east of the same lineament indicates that it does not mark the suture itself as previously thought by some authors (*e.g.* Ganade de Araujo and Santos, 2008). This Lineament represents a structure developed during the collisional event, analogous to the Anatolian Fault System in Turkey, or to the shear zones developed due to the Cenozoic India-Eurasia collisional escape tectonics. In this way, the final arrangement of the segmented crustal blocks in the Borborema Province is assigned by the development of strike-slip shear zones of the Transbrasiliano Shear System, which is linked with the ultimate stages of the escape collisional processes, combined with a northeastward extrusion that reflects the final configuration of the province.

Figure 5.5 - Tectonic model for the Neoproterozoic to Cambrian evolution in the Ceará Central and Médio Coreaú Domains of the Borborema Province. **A.** Wide oceanic setting in an early arc stage with inactive precursors arcs (Cariris Velhos) and intraoceanic arcs in Central Brazil and Africa. **B.** Narrowing of the oceanic domain and development of some extensional settings inboard of the eastern (present day position) plate. **C.** Continental collision and development of the Tamboril-Santa Quitéria granitic-migmatitic Complex and strike-slip shear zones of the Transbrasiliano-Kandi Shear System. **D.** Late orogenic stage related with the development of extensional setting (Jaíba Trough) arguably associated with the collapse of the orogen. AC-Amazonian Craton, WAC- West African Craton, SFC-São Francisco Craton.

5.6. Conclusions

Large scale provenance of the metasedimentary rocks from the Ceará Central and Médio Coreaú domains of Borborema Province has been assessed using U-Pb zircon geochronological studies by LA-MC-ICP-MS. Although reconnaissance in nature, several first-order conclusions can be addressed from our study such as:

- i) The rocks from the São Joaquim Formation in the MCD have a particular provenance signature that differs from the supracrustals rocks of the CCD. The detrital zircons from the São Joaquim quartzites were presumably shed from a continental region dominated by Paleoproterozoic and Archean crystalline rocks containing subordinate 1.7-1.8 Ga rocks. The zircon ages are in agreement with a provenance derived from the West African craton and Paleoproterozoic Granja Complex in a passive margin setting. Rocks to the west of the Araguaia belt in the Amazon craton could also have shed detritus to this passive margin system.
- ii) The Goiabeira Formation has similar patterns to the supracrustal rocks of CCD suggesting that the Transbrasiliano-Kandi Lineament does not represent a suture or a terrane boundary, but a large collision-related shear zone.
- iii) Samples from the Ceará Complex in the CCD demonstrate a heterogeneous provenance pattern characterized by deposits exclusively composed by Paleoproterozoic-Archean detritus, probably representative of small basins floored by sialic crust within the orogenic realm, and arc-related deposits with strong Cryogenian source component. In ancient active-margin settings, the sedimentary record contained within forearc strata can provide a more complete history of arc magmatism than the present exposure of the arc itself (DeGraaff-Surpless et al., 2002). Geological relationships along the prolongation of the Transbrasiliano-Kandi Lineament in Africa and Central Brazil, combined with the detrital zircon data presented herein, suggest that the Neoproterozoic paleogeography of this sector of the Gondwana is consistent with a long-lived active margin.
- iv) Zircons with ages in the range of the Tamboril-Santa Quitéria Complex are only found in the molassic basins, represented here by the Jaíbaras Trough, thus indicating that the supracrustals from the MCD e CCD are not related with the development of this complex as previously thought (*e.g.* Fetter *et al.*, 2003).
- v) Metamorphic overgrowths around the detrital zircon cores indicate a pluri-metamorphic evolution with distinct modes at *ca.* 640, 610, 580 and 540 Ma. The ages of metamorphic overgrowths partially overlap the main phases of the Tamboril-Santa Quitéria granitic-migmatitic Complex (*ca.* 640-610 Ma), indicating that this complex developed during the climax of the Neoproterozoic collisional orogenesis in Borborema Province.

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APPENDIX 1 – Sample description and location information

Sample DKE-36 (UTM 331598, 9690964)

This sample from the São Joaquim Formation was collected in the well-known exposition of the São Joaquim quartzite at the Jericoacora beach called Pedra Furada. It consists of a metasiliciclastic sequence of metapsmites with well-preserved regular parallel primary bedding associated with some ferruginous quartzite layers. The sample is a rather pure foliated quartzite composed by quartz (97%) and muscovite (3%) and displays a penetrative lattice preferred orientation of the quartz accompanied by the orientation of the muscovite flakes.

Sample DKE-39 (UTM 314069, 9627149)

This sample from the São Joaquim Formation is a muscovite-bearing quartzite collected near the locality of Jordão along the main road. Petrographically it shows a strong dynamic recrystallization of the quartz grains and well developed lattice preferred orientation that defines a penetrative foliation.

Sample DKE-41 (UTM 314228, 9610804)

This sample from the São Joaquim Formation is a strongly folded muscovite-bearing quartzite collected near the locality of Coreaú along the main road. The sample is composed by quartz (89%), muscovite (8%) and opaque minerals (2%). It displays a penetrative lattice preferred orientation of the quartz grains, accompanied by the orientation of the muscovite flakes.

Sample DKE-30 (UTM 345334, 9616391)

This sample from the Goiabeira Formation is a grayish schist collected along the main road close to the locality of Massapê. It is composed chiefly by quartz (55%), feldspar (20%) with plagioclase predominating over K-feldspar, muscovite (15%) and biotite (5%) and traces of chlorite. The quartz grains are elongated and dynamically recrystallized defining a well-developed lattice preferred orientation and the schistosity is defined by the platy minerals.

Sample DKE-25 (UTM 345334, 9616391)

This sample is a coarse grained reddish brown feldspathic sandstone of the Aprazível Formation from the Jaíba Trough. It demonstrates small cross bedding and ripple marks that are seen on bedding surfaces of fine-grained layers.

Sample DKE-43 (UTM 366196, 9595516)

This sample from the Canindé Unit of the Ceará Complex was collected close to the locality of Caióca and consists in a strongly deformed sillimanite-garnet-biotite metatextitic paragneiss. The leucosomes were avoided during sample preparation; however some very small patches of neosomes were difficult to separate.

Sample DKE-45 (UTM 391833, 9605570)

This sample from the Canindé Unit of the Ceará Complex is also a deformed sillimanite-garnet-biotite metatextitic paragneiss collected at the margin of the Miraima River close to the homonymous locality.

Sample DKE-19 (UTM 369488, 9566121)

This sample from the Independência Unit of the Ceará Complex was collected in a road cut close to the locality of Lisieux. It consists in a sequence of interleaved metapelites with kyanite and metapsamites characterized by muscovite-bearing quartzites with silimanite. Zircons were extracted from the quartzitic layer.

Sample DKE-51 (UTM 464240, 9627273)

Zircons from this sample were extracted from a muscovite-bearing quartzite from the Independência Unit of the Ceará Complex along the eastern border of the Tamboril-Santa Quitéria Complex close of the locality of Lajes.

Sample NCEB-427 (435811, 9513362)

This sample is a muscovite-bearing quartzite from the Independência Unit of the Ceará Complex collected close do the region of Taperuaba town.

Sample DKE-53 (UTM 435809, 9560168)

This sample from the Canindé Unit was collected near the locality of Tejuçuoca and consists in a strongly foliated sillimanite-garnet-biotite metatexitic paragneiss.

Sample NCEB-351 (UTM 410488, 9547439)

This sample is a muscovite-bearing quartzite from the Independência Unit of the Ceará Complex collected close do the region of Taperuaba town.

Sample DKE-06 (UTM 417895, 9491153)

This sample from the Independência Unit of the Ceará Complex was collected close to the locality of Lagoa do Mato and comprises a strongly folded arcosean quartzite with subordinate fibrolite and garnet.

Sample DKE-56 (UTM 502851, 9534867)

This sample from the Independência Unit of the Ceará Complex is a feldspathic-quartzite collected in the Pico Alto locality at the highest elevation of the study area.

6. Tracing Neoproterozoic subduction in the Borborema Province (NE- Brazil): clues from U-Pb geochronology and Sr-Nd-Hf-O isotopes on granitoids and migmatites

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Abstract

The Ceará Central Domain of the Borborema Province is a key tectonic domain within the 5000 km-long West Gondwana Orogen, which extends from Algeria in Africa to Central Brazil. Igneous rocks of the Tamboril-Santa Quitéria Complex, investigated in this study, record a long-lived history of convergent magmatism and crustal anatexis. SHRIMP U-Pb dating and Hf-O isotope analyses of zircons from granitoids and migmatites, coupled with whole-rock Sr-Nd isotopes were used to constrain the evolution of this long-lived continental margin. Magmatism can be divided into three main periods: i) an early period comprising essentially juvenile arc magmatism at ca. 880-800 Ma and continuing to 650 Ma as evidenced indirectly by detrital zircons from syn-orogenic deposits, ii) a more mature arc period at ca. 660-630 Ma characterized by hybrid mantle-crustal magmatic rocks, and iii) crustal anatexis at 625-618 Ma continuing until ca. 600 Ma. Detrital zircons with mantle values of δO^{18} (< 5.7 ‰) in the range of 950 to 650 Ma retrieved from fore-arc deposits indicate that juvenile input persisted throughout whole evolution of the convergent magmatism. Juvenile and mature arc igneous rocks underwent anatexis that gave rise to extensive areas of diatexites within the complex. Anatexis overlap in time with the ages of (ultra)-high pressure (U)HP eclogitic metamorphism dated at 625-615 Ma. In accordance with other continental collision zones, age of UHP/HP metamorphism are interpreted to mark the timing of continental collision and therefore indicate that the anatexis of arc rocks took place during continental subduction in a continent-continent collisional setting. Extensive migmatization continued until ca. 600 Ma and are in part synchronous to the exhumation of the rocks to shallower crustal levels. Thus, the 350 m.y. of magmatic activity in the Ceará Central Domain records the evolution of the West Gondwana margin of the Borborema Province from a juvenile arc setting through a mature arc and continental collision at around 625-600 Ma.

6.1. Introduction

Subduction zones are sites of intensive magmatism and are currently creating >20% of the terrestrial magmatic products (Tatsumi and Eggins, 1995). In these sites, complex compositional variations in the generated magmas arise from interaction between fluids released from the subducting oceanic lithosphere and the overlying mantle wedge, intrinsic heterogeneities from the mantle and magma fractionation (Tatsumi and Kogiso, 2003). Assimilation of crustal material, particularly in Andean-type settings, adds an important component and further variations to the magmas generated in subduction zones (Hildreth and Moorbath, 1988; McMillan et al. 1989).

Subduction of oceanic lithosphere and generation of arcs inevitably precede Himalayan-type collisional orogens. However, in old collisional, deeply eroded terranes, earlier stages of arc magmatism are relatively poorly preserved and have commonly been obliterated by pervasive collisional tectonics. In some extreme cases, earlier arcs can even be subducted along continuous or renewed subduction zones and not be preserved

(Yamamoto et al., 2009). Determining at what stage in the tectonic history of a subduction system a magmatic arc begins to evolve from a juvenile state, dominated by mafic-intermediate magmatism, towards a mature state dominated by felsic granitoid plutonism is critical to understand evolution of arcs and the stages preceding continental collision (Treloar et al., 1996). One important fact to consider is whether these earlier arcs are punctual in time, disconnected from the more mature stage, or are continuously linked to continental arc subduction that precede terminal collision. For example, the Kohistan and Ladakh arcs of northern Pakistan and northwest India represent a Cretaceous early intra-oceanic arc formed during the northward subduction of the Neotethys oceanic lithosphere beneath the Karakoram (e.g. Bard, 1983; Burg et al., 1998; Weinberg and Dunlap, 2000; Schaltegger et al., 2002). This arc was subsequently sutured to the Karakoram Terrane (southern margin of Asia) between 102 Ma and 85–75 Ma (Pettersen, 2010). The incorporated arc then became the Andean-type margin (mature stage) of Eurasia until collision with India at around 50 Ma (Hodges, 2000). Another example is the Mesozoic Sierra Nevada batholith in California, one of the best studied sites for convergent magmatism, where subduction of the Farallon plate beneath North America during the Triassic to early Cretaceous was characterized by early fringing island arcs just off the Paleozoic continental margin. With continued subduction, a mature stage continental arc was developed and a progressively more compressional environment developed as the age of subducting slab continued to young (Busby, 2004; Lee et al., 2007). In this mature arc stage, O-Sr isotopic relations and the variation of $^{147}\text{Sm}/^{144}\text{Nd}$ with ϵ_{Nd} suggests that assimilation of crustal rocks by magmas rising from the mantle and undergoing fractional crystallization could have been the major process responsible for the mixing of crustal- and mantle-derived components (DePaolo, 1981).

Figure 6.1 – Main cratonic blocks and mobile belts of the West Gondwana (modified from De Wit et al., 2008) and the Borborema Province and its main sub-divisions.

In ancient orogenic systems where great part of the petrological history has been obliterated by deformation and/or erosion, zircon can serve as an exceptional crustal growth monitor (Scherer et al., 2007). Coupling of radiogenic and stable isotopes allow measurements of time-stamped of hafnium and oxygen isotopes that can uniquely reveal whether zircon crystallized from a mantle-derived source (juvenile) during crustal generation,

Figure 6.2 – Geological map and structure of the northern portion of the Tamboril-Santa Quitéria Complex and its neighbouring units.

or from magma derived by reworking of pre-existing igneous or sedimentary rocks (Hawkesworth and Kemp, 2006; Scherer et al., 2007).

In this sense, the Lu-Hf system is analogous to the Sm-Nd, and Hf-Nd isotopes form coherent arrays for most mantle-derived rocks (Vervoort et al., 1999). A larger drawback with relying on Hf isotopes from zircons alone to infer episodes of crustal growth concerns the possibility that the zircons crystallized from magmas with mixed source rocks that separated from the mantle at different times (Hawkesworth and Kemp, 2006). The use of oxygen isotopes greatly reduces this ambiguity, because its fractionation is time-independent. The $^{18}\text{O}/^{16}\text{O}$ ratio, expressed as $\delta^{18}\text{O}$ relative to SMOW, is only changed by low temperature and surficial processes, and so the $\delta^{18}\text{O}$ of mantle-derived magmas ($5.7 \pm 0.3\text{‰}$) contrasts with those from rocks that have experienced a sedimentary cycle or hydrothermal alteration on the sea-floor, which have elevated $\delta^{18}\text{O}$ (Hawkesworth and Kemp, 2006). This is reflected in the high $\delta^{18}\text{O}$ of crystallizing zircons and is a fingerprint for a recycled component in granite genesis (Hawkesworth and Kemp, 2006; Hoefs, 2009). Likewise, the Nd-Sr isotopes retrieved from whole-rock analysis also provide a way to make such distinction (DePaolo, 1981; Jacobsen, 1988; DePaolo et al., 1991) and are useful to monitor and evaluate isotopic differences between data acquired from minerals (e.g. zircon) and rocks from the same representative sample.

The Ceará Central Domain of the Northern Borborema Province, NE-Brazil, was part of a long-lived active continental margin of the West Gondwana Orogen that consumed the Goiás-Pharusian Ocean during the Early Neoproterozoic until final collision at Ediacaran times (Fetter et al., 2003; Arthaud et al., 2008; Ganade de Araujo et al., 2012a; Cordani et al., 2013ab; Ganade de Araujo et al., in press). The deep level of exposition, with extensive outcrops of migmatites and exhumed eclogites (Santos et al., 2009), requires the use of isotopic geology to disentangle the evolution of this complex, multi-domain orogenic system. Although timing for arc-building (Andean-type margin) in the Ceará Central Domain is usually attributed to the 650-620 Ma time interval (Fetter et al., 2003; Van Schmus et al., 2008), geochronological evidence from detrital zircons in arc-related basins of the Ceará Complex suggest that arc magmatism could have started as early as 900-800 Ma (Ganade de Araujo et al., 2012a). In addition, several occurrences of Early Neoproterozoic juvenile arc assemblages are described along the length of the orogen in Africa and Central Brazil (e.g. Pimentel and Fuck, 1992; Berger et al., 2011). In some cases, these earlier juvenile arcs subsequently evolved into more mature arc stage preceding final collision that eventually reworked these arcs and precursor basement (continents) during the Late Neoproterozoic period (Liegeois et al., 1987; Caby, 2003; Pimentel et al., 2000).

In this study, we focus in the plutonic rocks of the Tamboril-Santa Quitéria Complex in the Ceará Central Domain, that record a long-lived magmatic system attributed to the subduction of the Goiás-Pharusian Ocean during the Neoproterozoic. Here, we combine U-Pb dating and Hf-O isotope composition of zircons, in addition to whole-rock Sr-Nd isotope compositions from granitoids and migmatite protoliths to unravel the tectonic evolution of this complex and sources (crust vs. mantle) of subduction-related magmas from the Early Neoproterozoic to final collision in the Ediacaran period.

Table 6.1 – Localization and units of the investigated samples from the Tamboril-Santa Quitéria Complex.

Sample	Lithology	Investigated lithology	Unit	UTM
DKE-221	net veined structured granodioritic metatexite	granodioritic paleosome	Lagoa Caiçara	441417/9577893
DKE-200A	mafic tonalitic metatexite	tonalitic paleosome	Lagoa Caiçara	441513/9578525
DKE-269	gray biotite orthogneisses injected by felsic veins	orthogneiss	Lagoa Caiçara	406753/9528308
DKE-231	gray biotite orthogneisses injected by felsic veins	orthogneiss	Lagoa Caiçara	451924/9581494
DKE-277	quartz diorite injected by felsic veins	quartz-diorite	Boi	381531/9560586
DKE-211	porphyritic biotite monzogranite	monzogranite	Santa Quitéria	429040/9601660
DKE-170	granodioritic metatexite with diatexitic portions	granodioritic schollen	Tamboril/Santa Quitéria	408537/9587164
DKE-125A	tonalitic metatexite intruded by felsic granite	tonalitic paleosome	Tamboril	388123/9585000
DKE-125B	tonalitic metatexite intruded by felsic granite	felsic granite	Tamboril	388123/9585000
DKE-273A	biotite diatexite with granodioritic schollen	granodioritic schollen	Tamboril	388830/9524195
DKE-273B	biotite diatexite with granodioritic schollen	diatexite	Tamboril	388830/9524195

6.2. Geological setting: the Ceará Central Domain

Excluding the extensional Mesozoic event that separated South America from Africa, the Borborema Province in northeast Brazil is characterized by magmatic, tectonic, and thermal events spanning the Archean to the Cambrian-Ordovician (Brito Neves et al., 2000). The major cratonic blocks involved in the tectonic events that built the Province include (**fig. 6.1**): 1) the Amazonian-São Luiz-West Africa Craton, including the Parnaíba Block; 2) the São Francisco-Congo Craton, and 3) the Paleoproterozoic-Archean collage forming the basement of the Borborema Province (Brito Neves and Cordani, 1991, Brito Neves et al., 2000; Arthaud et al., 2008; Klein et al., 2008; Ganade de Araujo et al., in press). Its final tectonic arrangement was a result of two Neoproterozoic continental collisions: the first and older along the Ceará Central Domain at ca. 620-615 Ma, as part of the West Gondwana Orogen, followed by the collision at ca. 590-570 Ma of the consolidated Borborema Province against the São Francisco Craton along the Sergipano Orogen in the south (Oliveira et al., 2010; Ganade de Araujo et al., in press).

The Neoproterozoic evolution of West Gondwana Orogen in the Ceará Central Domain results from the development of a convergent margin, related to the consumption of the Goiás-Pharusian Ocean (Cordani et al., 2013a), until the collision between the Parnaíba block (hidden beneath the Phanerozoic Parnaíba basin) and the Paleoproterozoic/Archean basement that extends further east into the Northern Borborema Province (Rio Grande do Norte Domain) (Ganade de Araujo et al., in press).

The Ceará Central domain is composed of several litho-tectonic assemblages that includes: (1) Archean (ca. 2.8-2.7 Ga) remnants of TTG of the Cruzeta Complex; (2) vast tracts of juvenile Paleoproterozoic (ca. 2.2-2.0 Ga) high-grade amphibolites and felsic to intermediate orthogneisses and migmatites (Fetter et al., 2000; Martins et al., 2009); (3) high-grade Neoproterozoic supracrustal rocks represented essentially by the units of Ceará Complex (e.g. Arthaud et al., 2008; Arthaud, 2007, Ganade de Araujo et al., 2012a); (4) large volumes of Neoproterozoic granitoids represented by the Tamboril-Santa Quitéria granitic-migmatitic Complex (Fetter et al., 2003; Arthaud et al., 2008); and (5) widespread Neoproterozoic to Cambrian post-collisional and Ordovician anorogenic granitoids (Castro et al., 2012). The first two associations are considered as the basement for the Neoproterozoic orogeny.

This basement, particularly the Cruzeta Complex, records an intricate geological history, from the Archean to the Paleoproterozoic period. The Archean remnants (2.85–2.64 Ga) are represented by gray gneisses of TTG affinity, locally migmatized and tectonically interleaved with the Paleoproterozoic gneisses (Fetter, 1999). Throughout the Ceará Central Domain there are several small Paleoproterozoic basement inliers preserved within the Neoproterozoic tectonic framework. These rocks are mainly polycyclic intermediate orthogneisses, migmatized not only during the Paleoproterozoic (ca. 2.0 Ga) but also in Late Neoproterozoic collision, with protolith crystallization ages clustering at ca. 2.13–2.15 Ga (Fetter et al., 2000; Castro, 2004).

The Ceará Complex is composed of metamorphosed pelites, semipelites and greywacke, normally showing a prominent schistosity or gneissosity, and are regionally or locally migmatized. Quartzites, marbles, calc-silicate rocks and amphibolites also form large tracts within this complex (Cavalcante et al., 2003; Caby and Arthaud, 1986; Arthaud et al., 2008; Ganade de Araujo et al. 2012a). Taking into account the degree of partial melting, Cavalcante et al. (2003) divided part of the Ceará Complex into the Independência and Canindé units. The supracrustal rocks with only minor migmatization were grouped in the former, while those that exhibit significant melting were included in the latter. Locally in the Ceará Complex, felsic sheets and amphibolites interleaved with metasedimentary rocks are interpreted as former volcanic or sub-volcanic rocks and were dated at ca. 800-750 Ma (Fetter, 1999, Castro, 2004, Arthaud, 2007). U-Pb zircon provenance studies from the Ceará Complex demonstrate a heterogeneous provenance pattern characterized by deposits exclusively composed by Paleoproterozoic-Archean detritus, probably representative of small basins floored by sialic crust within the Neoproterozoic orogenic realm, and orogenic arc-related deposits with strong early to middle Neoproterozoic (900-650 Ma) source component (Arthaud, 2007; Ganade de Araujo et al., 2012a).

In the Ceará Complex, retrogressed eclogites have been described to the east and west of the Tamboril-Santa Quitéria Complex. In the east, in the region of Forquilha, retrogressed eclogites occurs interleaved with high-grade migmatitic metasedimentary rocks (Santos et al., 2009) and protolith crystallization was dated at ca. 1.5 Ga (Amaral et al., 2010). These rocks preserve relics of eclogite facies metamorphism (1.7 GPa, Santos et al., 2009), which may have reached ultra-high pressure (UHP) conditions (Santos et al., 2013) at ca. 615 Ma (Ganade de Araujo, submitted). To the west, in the region of Itaitia retrogressed eclogites were also described by Castro (2004), however peak pressure conditions (1.4 GPa, Castro, 2004) are lower than those estimated for the Forquilha region.

6.2.1. The Tamboril-Santa Quitéria Complex

The Neoproterozoic Tamboril-Santa Quitéria Complex (**fig. 6.2**) is a wedge-shaped composite anatectic/igneous association surrounded by metasedimentary rocks of the Ceará Complex. It is characterized by a number of magmatic pulses where large volumes of magma intruded in the form of dykes, sills, sheets and plutons (Cavalcante et al., 2003; Fetter et al., 2003; Arthaud et al., 2008).

The plutonic rocks display syn- to late-magmatic deformation that was in part coeval with the injection of younger and less deformed magma (Arthaud et al., 2008). In general they range from diorite to granite, with predominance of monzogranitic/granitic rocks (Ganade de Araujo et al., 2012b) of the Santa Quitéria unit in its central part.

Previous age determinations indicate that granitoids of this complex range from 640 to 610 Ma (Fetter et al., 2003; Castro, 2004; Santos et al., 2007; Ganade de Araujo et al., 2012b, Costa et al., 2013). For this time interval, Nd isotopic signatures are consistent with variable mixtures between juvenile Neoproterozoic magmas and older basement, indicating that the granitoids hybrids (Fetter et al., 2003). The tectonic setting of this complex has been interpreted as a Neoproterozoic Andean-type magmatic arc (Fetter et al., 2003), however recent works have proposed an evolution from an arc at ca. 850 to 640 Ma into a collisional Himalayan setting (Ganade de Araujo et al., 2012c, Costa et al., 2013).

In accordance with the nomenclature used by the Geological Survey of Brazil and also those proposed by Castro (2004), in the present study the complex is divided into four different units named Lagoa Caiçara, Boi, Santa Quitéria and Tamboril units. Although majorly composed by distinctive and older rocks, we opted to include the Lagoa Caiçara unit in the complex as it corresponds to the earlier evolution of the long-lived continental active margin proposed here. Investigated samples from these units and their main features are listed in **Table 6.1**.

Figure 6.3 – Field aspects of the studied rocks from the Lagoa Caiçara unit. **A.** Stromatic metatexitic after a 833±6 Ma tonalitic protolith (*sample DKE-221*) with hornblende-bearing leucosomes, interpreted to result from water-fluxed melting. **B.** Stromatic metatexitic after a 650±5 Ma mafic tonalite (*sample DKE-200A*). **C.** 632±5 Ma biotite gneiss with injected leucocratic veins parallel to the gneissic foliation (*sample DKE-269*). **D.** Metatexitic after a 627±5 biotite orthogneiss (*sample DKE-231*).

6.2.1.1. Lagoa Caiçara unit

This unit comprises a heterogeneous meta-igneous association composed predominantly of stromatic metatexites of granodioritic to tonalitic protoliths (**fig. 6.3**). They are in contact with granitic diatexites of the Tamboril unit and the sedimentary-derived metatexites of the Ceará Complex to the east and south. These meta-igneous rocks are also commonly found preserved as blocks, known as schollen or rafts, within the diatexites of the Tamboril unit. Also in the Lagoa Caiçara unit, sheets of biotite-orthogneisses (*c.f.* samples DKE-269 and DKE-231) (**fig. 6.3C and D**) with moderate to small volume of leucosomes cut the more complex deformed migmatitic granodiorite-tonalite. Remnants of sedimentary-derived metatexites, of the Ceará Complex are also present within this unit.

Distinction between the different orthogneisses of Lagoa Caiçara unit is difficult in the field. It seems that this unit comprises multiple intrusions of granitoid rocks. Deformation adds complications and it is challenging in many outcrops to ascribe unambiguously a sample to the broader lithological group. For example, we expect that many late dykes of the Tamboril-Santa Quitéria magmatic event intruded the Lagoa Caiçara unit, and similarly, we have recognized schollen and xenolithic blocks of the Lagoa Caiçara unit in the diatexites of the Tamboril Unit. This distinction although obvious in some places, is less evident when the blocks are of a similar nature to the host magmatic rock. In the present study, geochronological and isotopic data permitted the distinction of three different granitoids protoliths in the Lagoa Caiçara unit: i) ca. 880-830 Ma juvenile tonalitic/granodioritic metatexites with high volume of leucosomes, ii) ca. 650 Ma mafic tonalitic metatexites, and iii) ca. 630 Ma crust-derived orthogneisses with low volume of leucosome.

The regional foliation in this unit is simple and has low to moderate dips (<40 degrees) to northwest and north-northwest (Itapajé structural domain in **figure 6.2**). Along the contact with the diatexites of the Tamboril unit, the stretching lineation has a low rake indicating a strong strike-slip component. They generally plunge gently to ENE and a number of shear sense indicators such as S/C structures suggest a dextral strike-slip movement with a dominant small reverse component. Further south, in the contact of the Lagoa Caiçara unit and the Ceará Complex, the lineation changes to dominantly down-dip, plunging northward and shear sense indicators within the foliations demonstrate a change to top-to north-northeast defining normal movement.

The older (830 Ma and 650 Ma) tonalitic to granodioritic protolith of the metatexites contains biotite (10-20%) and hornblende (5-25%) as the main ferro-magnesian phases. The schollen of this unit found in the Tamboril diatexites have low contents or lack hornblende and are predominantly composed of biotite, plagioclase, K-feldspar and quartz. The neosome of the tonalitic migmatites is composed majorly of plagioclase, quartz and hornblende with no anhydrous peritectic phases, suggesting that melting was due to influx of water rather than hydrate breakdown reactions (Weinberg and Hasalova, submitted). The younger orthogneisses (ca. 630 Ma) have biotite as the main mafic phase accompanied or not of minor muscovite with K-feldspar invariably more abundant than plagioclase.

6.2.1.2. Boi Unit

The Boi unit differs from the Lagoa Caiçara unit by the presence of more homogenous mafic rocks of predominant quartz-diorite to tonalitic/granodioritic composition (**fig. 6.5A**). They are easily recognizable and mapable in the satellite and gamma-ray image due to characteristic low total counts. In the field these rocks may be strongly foliated to rather isotropic. Migmatitic sectors may occur, however the intrusion of felsic melts may generate pseudo-migmatitic patterns. Rocks from this unit are comprised of plagioclase (45-35%), hornblende (10-25%), biotite (15-25%), quartz (15-5%) and K-feldspar (8-3%). They are in part intruded by the Santa Quitéria and Tamboril units. Further south of the study area a U-Pb ID-TIMS zircon age of 637 ± 6.5 Ma was obtained for a juvenile ($\epsilon\text{Nd}(600\text{Ma}) = +3.4$) dioritic migmatitic gneiss (Fetter et al., 2003), possibly associated with the Boi unit.

6.2.1.3. Santa Quitéria Unit

The Santa Quitéria unit forms a large batholith in the central portion of the complex. It is by far the most voluminous magmatic component of the complex and comprises mainly porphyritic K-feldspar monzogranites (**fig. 6.4B**). Composition and strain intensity vary, however toward its central portion, low strain and larger phenocrysts dominate (**fig. 6.2**). The contact of the Santa Quitéria monzogranites with the diatexites of the Tamboril unit is well defined in the satellite and geophysical gamma-ray images (available at the Geological Survey of Brazil), but can be gradual in some places. Locally, close to the town of Iraúçuba, disrupted rafts of the monzogranite can be found within the diatexite indicating that crustal anatexis occurred after the intrusion of this batholith.

One especial feature of this unit is the existence of local disrupted coeval mafic syn-plutonic dikes (**fig. 6.4D**). Geochemical data of these mafic dikes indicate an enriched shoshonitic component derived from mantle sources (Costa et al., 2013; Zincone, 2011). Less common xenoliths of gray orthogneisses, probably derived from the Lagoa Caiçara unit, can also be present within the Santa Quitéria monzogranite. Biotite (20-10%) and hornblende (10-1%) are the main ferro-magnesian mafic phases of Santa Quitéria monzogranites along with plagioclase (40-15%), K-feldspar (35-10%) and quartz (25-15%). Accessories include zircon, titanite, apatite, epidote and opaques. In general, the mafic syn-plutonic dikes are constituted of plagioclase (35-30%), biotite (25-20%), hornblende (20-15%), K-feldspar (15-10%) and quartz (5-2%).

Structurally this unit has a wedge-shaped geometry with foliations in both the NE-SW and E-W trending flanks dipping inwards towards the complex (**fig. 6.2**). In general the regional foliation dips at moderate angles ($35-50^\circ$) to south-southeast in the northern portion of the domain and to north-northwest in its southern portion (Santa Quitéria structural domain in **fig. 6.2**). The stretching lineation within this domain has low angles and plunges predominantly northeast. Shear sense indicators in the monzogranite indicate top-to-east or northeast sense defining a dominantly strike slip motion with both normal and reverse components, broadly the same movement direction as defined in the Itapagé domain. This pattern defines the wedge-shaped geometry that some authors attributed as a product of the necking-down of the Tamboril-Santa Quitéria Complex responsible for its extrusion under a transpressive regime in sort of a positive-flower

structure (Castro, 2004). However, satellite and geophysical image analysis suggest that this geometry could be due to the presence of ENE-WSW late large-scale open upright folds (**fig. 6.2**).

Figure 6.4 – Field aspects of the studied rocks from the Boi and Santa Quitéria units. **A.** 648 ± 5 Ma quartz-diorite of the Boi Unit injected by felsic quartz-feldspathic melt (*Sample DKE-277*). **B.** 638 ± 5 Ma porphyritic monzogranites of the Santa Quitéria unit with mafic enclaves exhibiting crystal-transfer structures (white arrow) (*Sample DKE-211*). **C.** Coeval Santa Quitéria monzogranite with mafic dioritic enclaves showing evidence for transfer of crystals from the granite to the diorite (arrows). **D.** Syn-plutonic dikes of diorites cutting through the Santa Quitéria porphyritic monzogranite.

6.2.1.4. Tamboril Unit

The Tamboril unit represents a gradational unit at the contact between the monzogranite of the Santa Quitéria unit and the gneisses of the Lagoa Caíçara unit, but generally this unit encircles the Santa Quitéria unit. It is dominated by diatexites containing blocks (rafts or schollen) of both Santa Quitéria porphyritic monzogranite and Lagoa Caíçara orthogneisses. Rafts of Santa Quitéria monzogranites dominate close to the contact with the Santa Quitéria unit whereas high-grade metasedimentary and orthogneisses rafts are found close to the contact of the Lagoa Caíçara unit in the vicinity of Itapajé town.

In general the foliation in these diatexites are defined by a well-developed syn-magmatic flow banding usually defined by biotite schlieren (**fig. 6.5D**). Isotropic domains can be found locally. In the south, along the contact with the Lagoa Caíçara unit, foliation in diatexite dips at moderate angles to NNW with an associated

stretching lineation characterized by a strong strike-slip component and shear sense indicators, such as S/C pairs, suggesting a right-lateral movement (top-to-NW). In the north, foliation in the diatexites dips to SSE and E, with a stretching lineation plunging predominantly to SE. Kinematic indicators indicate a top-to-southeast normal displacement, however movement in the opposite direction could also be observed (**fig. 6.2**).

In general, these diatexites lack residual anhydrous peritectic phases, with the exception of rare garnet clusters. Biotite (20-5%) is the main ferro-magnesian phase, but hornblende is present in some samples. In general the rocks tend to have greater concentrations of K-feldspar (45-15%) than plagioclase (25-10%), but in some cases this latter phase can dominate. Previous U-Pb ID-TIMS geochronological data yielded zircon ages for the diatexites of the Tamboril unit at the 620-610 Ma interval (Castro, 2004).

6.3. Analytical Procedures

In order to better understand the temporal evolution and the source of different magmas we carried out in situ U-Pb zircon geochronology coupled with Hf and O isotopes on the same dated zircon domains. Zircon isotopic data were complemented by whole-rock Nd and Sr isotopes to better constrain granite sources for the same representative samples used for zircon investigation.

Zircons were separated from the fresh crushed rocks (3–5 kg) using conventional and heavy liquid and magnetic techniques (jaw crusher, disk grinder, Wilfley table, Frantz isodynamic magnetic separator and density separation using bromoform and methylene iodite). Around 50–80 zircons from each sample were mounted in epoxy resin, polished to half of mean grain thickness for further imaging with transmitted light and cathodo-luminescence to unravel its internal complexities. Cathodo-luminescence (CL) images of zircons were obtained using a Quanta 250 FEG electron microscope equipped with Mono CL3+ cathodoluminescence spectroscope (Centaurus) at the Geochronological Research Center in São Paulo University, Brazil.

U-Pb analyses were done using SHRIMP IIe at the Geochronological Research Centre (CPGeo) at the São Paulo University. The data have been reduced in a manner similar to that described by (Williams 1998, and references therein), using the SQUID Excel Macro of Ludwig (2001). Uncertainties given for individual U-Pb analyses (ratios and ages) are at the 1σ level, however uncertainties in the calculated weighted mean ages are reported as 95% confidence limits and include the uncertainties in the standard calibrations where appropriate. For the age calculations, corrections for common Pb were made using the measured ^{204}Pb and the relevant common Pb compositions from the Stacey and Kramers (1975) model. Concordia plots, regressions and any weighted mean age calculations were carried out using Isoplot/Ex 3.0 (Ludwig, 2003) and where relevant include the error in the standard calibration. U-Pb geochronological results are presented in **Table S1 of supplementary data**.

Lu-Hf analyses were also carried out at the Geochronological Research Centre (CPGeo) at the São Paulo University on a Neptune laser-ablation multi-collector inductively coupled plasma mass spectrometer equipped with a Photon laser system. The laser spot used was 39 μm in diameter with an ablation time of 60 seconds, repetition rate of 7 Hz, and He used as the carrier gas (Sato *et al.*, 2009). $^{176}\text{Hf}/^{177}\text{Hf}$ ratios wer

Figure 6.5 – Field aspects of the studied rocks from Tamboril unit. **A.** Composite outcrop of patchy metatexite after a 882 ± 7 Ma granodioritic orthogneiss (schollen) embedded in a 618 ± 5 Ma granitic diatexite of Tamboril unit within Lagoa Caiçara unit (*Sample DKE-273A and B*). **B.** Raft of a 663 ± 7 Ma granodioritic orthogneiss embedded in a granitic host close to the contact between Santa Quitéria and Tamboril units (*Sample DKE-170*). **C.** Folded stromatic metatexite tonalite to diorite (Boi unit) injected by crustal granitic veins of Tamboril unit (*Sample DKE-125*). **D.** Characteristic flow banding defined by schlieren diatexite of the Tamboril unit. **E.** Characteristic schollen diatexite of the Tamboril unit. **F.** Hornblende-bearing leucosomes in diatexite of Tamboril unit.

normalized to $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$. Zircon Hf isotopic data are presented in table 3. The isotopes ^{172}Yb , ^{173}Yb , ^{175}Lu , ^{177}Hf , ^{178}Hf , ^{179}Hf , ^{180}Hf , and $^{176}(\text{Hf}+\text{Yb}+\text{Lu})$ were simultaneously collected. $^{176}\text{Lu}/^{175}\text{Lu}$ ratio of 0.02669 was used to calculate $^{176}\text{Lu}/^{177}\text{Hf}$. Mass bias corrections of Lu-Hf isotopic ratios were done applying

the variations of GJ1 standard. A decay constant for ^{176}Lu of 1.867×10^{-11} (Söderlund *et al.*, 2004), the present-day chondritic ratios of $^{176}\text{Hf}/^{177}\text{Hf} = 0.282772$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0332$ (Blichert-Toft and Albarede, 1997) were adopted to calculate ϵHf values. A two-stage continental model (T_{DM}) was calculated using the initial $^{176}\text{Hf}/^{177}\text{Hf}$ of zircon and the $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ ratio for the lower continental crust (Griffin *et al.*, 2004). Zircon Lu-Hf isotopic results are presented in **Table S2 of supplementary data**.

Oxygen isotopic compositions were obtained in three separate analytical sessions using the SHRIMP-II equipped with a Cs-gun at the Research School of Earth Science (RSES) in The Australian National University as described by Ickert *et al.* (2008). TEMORA 2 zircon ($d^{18}\text{O} = 8.2\text{‰}$; Black *et al.*, 2004) was analyzed along with FC1 zircon. The results are presented in **Table S3 of supplementary data** and plotted on **figure 6.10A**. No corrections for IMF/gain drift or EISIE were necessary. Oxygen isotope analyses of FC1 on SHRIMP II, normalized to TEMORA 2, yield a mean $d^{18}\text{O}$ value of $5.5 \pm 0.3\text{‰}$.

Nd-Sr isotopic compositions were determined thermal ionization mass spectrometry (TIMS) in a VG354 spectrometer equipped with a single Faraday detector at the Geochronological Research Centre (CPGeo) at the São Paulo University. The same powders used for whole-rock elemental analyses were taken into solution by acid digestion, and the elements of interest were separated in ion-exchange columns following the procedures described in Sato *et al.* (1995). No spikes were added; $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios were calculated from whole-rock analyses obtained by XRF (Rb and Sr) and ICP-MS (Sm and Nd). Nd-Sr isotopic results are presented in **Table S4 of supplementary data**.

Major and trace elements, were analyzed at the SGS GEOSOL laboratories according to the package used by the Geological Survey of Brazil. Major element oxides were determined using a Varian Vista Pro ICP-AES. Trace elements were determined using a Perkin-Elmer Sciex ELAN 6000 ICP-MS. Analyses of USGS rock standards (BCR-2, BHVO-1 and AGV-1) indicate precision and accuracy better than 1% for major elements and 5% for trace elements and REE. Whole rock geochemical results are presented in **Table S5 of supplementary data**.

6.4. Results

Isotopic results for the granitoids and migmatites of the Tamboril-Santa Quitéria Complex are available in the **supplementary data** related to this article. Zircon U-Pb, Lu-Hf and oxygen isotopic measurements were all carried out on the same textural domain in each zircon, which permitted us to link age and isotopic parameters directly. A summary of the isotopic data acquired herein is provided in **table 6.2**.

6.4.1. Zircon SHRIMP U-Pb ages, zircon Hf-O and whole-rock Nd-Sr isotopes

6.4.1.1. Lagoa Caiçara unit

As described earlier, it is difficult to distinguish the igneous rocks of this unit based solely on their field characteristics. The isotopic results summarized in **table 6.2** define three groups of igneous rocks based on the age of the protoliths and their sources, which revealed how subduction-related magmas developed through time.

Table 6.2 – Summary of the main isotopic features of the investigated samples.

Sample	Lithology	Unit	Age (Ma)	inheritance (Ma)	eHf(t)zircon	(source)	$\delta^{18}\text{O}_{\text{zircon}}$ (‰)	(source)	eNd(t)	(source)	$(87\text{Sr}/86\text{Sr})$	(source)
DKE-221	granodioritic metatexite (protolith)	Lagoa Caiçara	833 ±6.1	no	(+0.5 to +19.3)	mantle	(5.09 to 6.24)	mantle	(+4.98)	mantle	(0.7025)	mantle
DKE-200A	mafic tonalitic metatexite (protolith)	Lagoa Caiçara	654.6±4.7	no	(-3.6 to +1.5)	mantle/crustal	(6.73 to 8.19)	mantle	(-5.45)	crustal	(0.7105)	crustal
DKE-269	gray biotite orthogneiss	Lagoa Caiçara	632±5.1	820-761	(-1.4 to +5.4)	mantle/crustal	(8.69 to 10.82)	crustal	(-10.75)	crustal	(0.7108)	crustal
DKE-231	gray biotite orthogneiss	Lagoa Caiçara	627±4.9	no	(-18.7 to -6.1)	n.a.	n.a.	n.a.	(-9.65)	crustal	(0.7143)	crustal
DKE-277	quartz diorite	Boi	648±4.1	no	(-6.6 to -0.8)	crustal	(5.48-6.25)	mantle	(-5.87)	crustal	(0.7056)	crustal/mantle
DKE-211	porphyritic monzogranite	Santa Quitéria	637.8±4.8	no	(-12.2 to -2.9)	crustal	(7.06 to 8.57)	crustal	(-4.25)	crustal	(0.7107)	crustal
DKE-170	granodioritic schollen	Tamboril	663±6.6	no	(+3.7 to +13.2)	mantle	(5.94-9.06)	crustal	(+1.80)	mantle	(0.7028)	mantle
DKE-125A	tonalitic metatexite (protolith)	Tamboril	646±4.5	no	n.a	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
DKE-125B	tonalitic metatexite (schollen)	Tamboril	625.9±4.6	no	n.a	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
DKE-273A	granodioritic metatexite (schollen)	Tamboril	892±7.5	no	(-3.6 to +1.5)	mantle/crustal	(5.20 to 6.44)	mantle	(+3.84)	mantle	(0.7020)	mantle
DKE-273B	diatexite matrix	Tamboril	618±4.1	728-879	(-1.4 to +5.4)	mantle/crustal	(6.41 to 9.10)	mantle	(-3.55)	crustal	(0.7079)	crustal

Figure 6.6 – Cathodoluminescence images from zircons selected for U-Pb geochronology and Hf-O isotopic investigation.

Sample DKE-221

This sample is a hornblende-biotite stromatic metatexite of tonalitic composition (**fig. 6.3A**). Zircons were extracted from the paleosome (or the protolith), avoiding contamination with the neosome, and are euhedral, translucent and colorless. In general they range in size from 80 to 200 μm and have a length to width ratios ranging from 2:1 to 4:1. Cathodoluminescence images reveal a well-developed oscillatory zoning typical of magmatic zircons (**fig. 6.6**). Some zircons have low-U, thin metamorphic rims, too small for SHRIMP analysis. Analyzed zircons have U contents between 52-256 ppm and Th/U ratios ranging from 0.50 to 0.78. Fourteen analyses were done in the zircons and a calculated Concordia age using all analyzed zircons yielded an age of 833 ± 6.1 Ma (1σ) (**table 2**), interpreted as the crystallization age of the tonalitic protolith (**fig. 6.7**).

Zircons have a significant variation of $^{176}\text{Hf}/^{177}\text{Hf}$ as a function of $^{206}\text{Pb}/^{238}\text{U}$ ages with values ranging from 0.282261 to 0.282800 for ages between 880 and 795 Ma. Despite such variation all analyzed zircons yielded consistently positive ϵHf_0 varying from +0.5 to +19.3 indicating that the tonalitic protolith was derived from mantle or juvenile sources at *ca.* 830 Ma (**fig. 6.10A**). Oxygen isotopes further support the mantle origin

indicated by the Hf isotopes in the zircons. The $\delta^{18}\text{O}$ values of 5.09 to 6.24‰ are in agreement with values of mantle zircon ($5.7 \pm 0.3\text{‰}$, Hawkesworth and Kemp, 2006) (**fig. 6.10B**). Whole rock Sr-Nd isotopes also support a juvenile origin for the tonalitic protolith, with low initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7025 and positive $\epsilon\text{Nd}_{(t)}$ value of +4.98 at the time of crystallization at 833 Ma (**fig. 6.11**).

Sample DKE-200A

This mafic tonalitic metatexite was collected in quarry close to the Itapajé town and differs from the previous sample, not only in terms of age and source, but also by higher content of hornblende (**fig. 6.3B**). Zircons were extracted from the paleosome, avoiding contamination with the neosome. In general they are subhedral to euhedral, translucent and colorless, with dimensions ranging from 60 to 150 μm . They have complex zoned patterns (c.f. zircon #7.1 – **fig. 6.6**) to well-developed oscillatory zoning. Most of the grains have a pronounced metamorphic overgrowth possibly due to anatexis of the protolith, not dated in this study. Th/U ratios of the dated zircon spots range from 0.55 to 0.85. A Concordia age defined by nine concordant zircons yielded an age of 650.6 ± 5.1 Ma (1σ) (**table 6.2**), much younger than the previous sample and interpreted as the crystallization age of the igneous protolith (**fig. 6.7**). $^{176}\text{Hf}/^{177}\text{Hf}$ ratios from the analysed zircons vary from 0.282226 to 0.282428 with $\epsilon\text{Hf}_{(t)}$ varying from -3.6 to + 1.5. The $\delta^{18}\text{O}$ values for the same zircons in the same CL zones range from 6.73 to 8.19‰ and combined with whole-rock initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7105 and negative $\epsilon\text{Nd}_{(t)}$ value of -5.45 suggest that this grantoid was predominantly sourced from crustal material, in contrast to the previous sample.

Sample DKE-231

This orthogneiss differs from the surrounding migmatitic gneiss found in the same unit by the incipient anatexis (e.g. small leucosome volume) and by the absence of hornblende and a more granitic composition (s.l.) than previous samples (**fig. 6.3C**).

Investigated zircons are colourless and mostly euhedral ranging in size from 80 to 200 μm . They have prominent high-U rim related to late thermal events (c.f. zircon #6.1 – **fig. 6.6**). Analysed magmatic zircons have Th/U ratios varying from 0.27 to 0.60 and define a twelve-point concordia age of 627 ± 4.9 Ma (1σ) that reflects crystallization of the protolith to the orthogneiss (**fig. 6.7**). One zircon with a $^{206}\text{Pb}/^{238}\text{U}$ age of 691 ± 18 Ma represents an outlier and is likely inherited. No oxygen analysis was carried out for this sample. $^{176}\text{Hf}/^{177}\text{Hf}$ ratios for the analyzed zircons in spots along the same CL zone range from 0.281848 to 0.282207 with correspondent $\epsilon\text{Hf}_{(t)}$ varying from -18.7 to -6.1, and together with a high initial whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7143 and negative $\epsilon\text{Nd}_{(t)}$ value of -9.65, suggests this magma was essentially sourced from older crustal rocks.

Sample DKE-269

This migmatitic orthogneiss is compositionally similar to the previous one and was found in the same geological context. Zircons from the protolith are euhedral to subhedral with sizes ranging from 50 to 150

μm . Most zircons have a well-developed rim surrounding inherited cores (c.f. zircons #3.1 and #8.1 – **fig. 6.7**). In general U/Th ratios vary from 0.17 to 1.52 (0.17-0.58 for inherited cores). A concordia age of 632 ± 5.1 Ma (1σ) was defined by eleven concordant points and reflects the age of crystallization of the protoliths. Three inherited zircons with $^{206}\text{Pb}/^{238}\text{U}$ ages of 823 ± 23 , 796 ± 19 and 761 ± 19 Ma suggest that Early Neoproterozoic protoliths, such as the 800 Ma, juvenile tonalite of sample DKE-221, were involved in the genesis of the protolith. $^{176}\text{Hf}/^{177}\text{Hf}$ ratios from zircons with $^{206}\text{Pb}/^{238}\text{U}$ ages in the range of 600 to 658 Ma vary from 0.282323 to 0.282523 with ϵHf_t of -1.4 to +5.4, pointing to a juvenile component in the genesis of the precursor magmas. One inherited core yielded a highly radiogenic $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282685 with correspondent ϵHf_t of +14.5, further supporting the suggestion that juvenile sources were involved in the genesis of the protolith of this orthogneiss. However, $\delta^{18}\text{O}$ values range from 8.69 to 10.82‰. This contrasts with expectations from magmas generated by juvenile sources and suggests either crustal material contributed to the formation of the precursor magmas or external, isotopically evolved water was present during melting of the source (**see discussion in section 6.5.1.3**). High initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7108 and strong negative ϵNd_t value of -10.75 also support the participation of older crustal material in the genesis of the magma.

6.4.1.2. Boi Unit

Sample DKE-277

Zircons from this mafic tonalite are subhedral with ovoid shapes ranging in size from 40 to 100 μm . In general, they have a well-developed igneous oscillatory zoning surrounded by a thin metamorphic overgrowth not accessible by the SHRIMP analyses (c.f. zircons 6.1 and 4.1 – **fig. 6.6**). The dated igneous zircons have U/Th ratios of (0.56-0.97) and yielded a twelve-point concordia age of 648 ± 4.1 Ma (1σ) that reflects the age of crystallization of tonalite (**fig. 6.7**). $^{176}\text{Hf}/^{177}\text{Hf}$ ratios from these zircons have a narrow variation between 0.282201 and 0.282348 which corresponds to ϵHf_t values between -6.6 to -0.8. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7056 and negative ϵNd_t value of -5.87 indicate that both mantle and older crust were involved in the magma genesis, however $\delta^{18}\text{O}$ values for the dated zircons range from 5.48-6.25‰, which fall within the proposed range for mantle zircons (5.7 ± 0.3 ‰, according Hawkesworth and Kemp, 2006).

6.4.1.3. Santa Quitéria unit

Sample DKE-211

This sample of porphyritic monzogranite from the core of the batholith is representative of the most voluminous igneous unit found within the complex. Zircons from this sample (**fig. 6.4B**) are euhedral (80-200 μm) and display nicely developed oscillatory zoning (**fig. 6.6**) with U/Th ratios ranging from 0.45 to 1.06. Eleven spot analyses yielded a concordia age of 637.8 ± 4.8 Ma (**fig. 6.7**), which reflects the age of the crystallization of the monzogranite. This age is slightly younger than the mafic sample DKE-277 from the Boi unit. $^{176}\text{Hf}/^{177}\text{Hf}$ ratios from the analyzed zircons range from 0.282028 to 0.282314 corresponding to a ϵHf_t between -12.2 to -2.9, indicating participation of crustal material in the genesis of the monzogranitic magma, as also suggested by the high $\delta^{18}\text{O}$ values of 7.06 to 8.57. Despite of evident interaction with mafic magmas of

the Boi unit, high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7107 and negative ϵNd_0 value of -4.25 also point to involvement of crustal sources in the genesis of this monzogranite.

Figure 6.7 – U-Pb Whetheril Concordia diagrams for the investigated samples. Green ellipses indicate protolith crystallization; yellow ellipses indicate inheritance; pink ellipses indicate melt precipitated zircons; white ellipses were excluded from the dataset.

6.4.1.4. Tamboril Unit

This unit is dominated by granitic diatexites that often contain rafts (schollen) from older igneous rocks of the complex. It represents a gradational unit at the contact between the monzogranite of the Santa Quitéria unit and the gneisses of the Lagoa Caiçara unit and Ceará Complex in the north. The isotopic results does confirm field observations with samples with characteristics similar to those of the Santa Quitéria (DKE-170, DKE-125A) and samples of older juvenile material similar to the Lagoa Caiçara (DKE-273A).

Sample DKE-170

Zircons from this metatextitic granodiorite raft (**fig. 6.5B**) extracted from the contact between the Santa Quitéria unit with the diatexites of the Tamboril unit are mostly euhedral (80-150 μm) and characterized by a prominent oscillatory zoning surrounded by a thin high-U metamorphic overgrowth (c.f. zircons 7.1 and 9.1 –

fig. 6.6). Th/U ratios for the dated zircons vary significantly from 0.11 to 1. Seven concordant analyses fall in a group yielding a concordia age of 663 ± 6.6 Ma (1σ) (**fig. 6.7**). $^{176}\text{Hf}/^{177}\text{Hf}$ ratios from the analyzed zircons are slightly radiogenic with values ranging from 0.282471 and 0.282741, with correspondent $\epsilon\text{Hf}_{(t)}$ of +3.7 to +13.2, indicating involvement of juvenile sources in the genesis of the magmas. Low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7028 and positive $\epsilon\text{Nd}_{(t)}$ value of +1.80 also lend support to partial melting of depleted mantle sources. However, the $\delta^{18}\text{O}$ values (5.94-9.06‰) for the dated igneous zircons fall outside the field of mantle zircons and suggest that crustal contaminants or isotopically evolved water interaction during crystallization could contribute to the observed higher $\delta^{18}\text{O}$ values.

Sample DKE-125

We collected two samples in this outcrop. Sample DKE-125A is a mafic stromatic metatextitic diorite raft embedded in the granitic diatextite of the Tamboril unit. Sample DKE-125B represents the host granite diatextite (**fig. 6.5C**). Field evidence does not support derivation of the diatextite from the partial melting of the diorite because the leucosomes in the diorite have different composition of the host diatextite evidenced by abundant plagioclase. Zircons from the metatextitic diorite are euhedral to subhedral (60-200 μm) and have well-defined igneous oscillatory zoning with Th/U ratios ranging from 0.50 to 0.80. Twelve zircons form a group in the concordia line yielding a mean age of 646 ± 4.5 Ma (1σ) for the dioritic protolith crystallization (**fig. 6.7**). Zircons from the host granitic diatextite are also euhedral to subhedral and have well-defined igneous oscillatory zoning with Th/U ratios from 0.13 to 0.84. A concordia age of 625.9 ± 4.6 Ma (1σ) defined by eleven concordant analyses reflects the age of the crystallization of this diatextite (**fig. 6.7**). No zircon Hf-O isotopes or whole-rock Sr-Nd analyses were performed for either of these samples. These results suggest that mafic intrusive rocks of an age similar to that of the Boi unit were involved in an anatectic event that occurred only 20 m.y. after their crystallization.

Sample DKE-273

The composite sample DKE-273 is divided into a schollen of granodioritic composition (sample DKE-273A) and the host diatexite of the Tamboril unit (sample DKE-273B) (**fig. 6.5A**). Different from sample DKE125, field evidence such as continuity between the host diatexite and leucosomes in the schollen, as well as textural similarity supports partial melting of the granodioritic schollen as one of the sources of the diatexite. Zircons from the granodioritic schollen are euhedral, transparent, colorless to light yellow. Most of them are equant to short prismatic. Crystals range in length from 80 to 200 μm . Most zircons are oscillatory zoned and interpreted as the result of magmatic growth (c.f. zircons 16.1 and 15.1 – **fig. 6.6**), but newly developed rims around magmatic cores also with a characteristic oscillatory zoning are interpreted as melt-precipitated zircons from the partial melting event (c.f. zircons 5.2 and 7.1 – **fig. 6.6**). A third type of zircon is characterized by homogenous domains that crosscut the two types described above (c.f. zircon 12.2 – **Fig. 6.6**). Two clusters of crystallization ages were obtained from zircons in the schollen. The older, with a calculated Concordia age of 892 ± 7.5 Ma is considered to be the protolith age, and was obtained from both old cores (c.f. zircons 5.2 and 7.1 – **fig. 6.6**) and from zircons with prominent oscillatory zoning but lacking overgrowths (c.f. zircons 16.1 and 15.1 – **fig. 6.6**). The younger cluster with a calculated concordia age of 620 ± 5.1 Ma (**fig. 6.7**) is interpreted as the age of anatexis and was obtained from magmatic overgrowths (melt-precipitated) around older cores (c.f. zircons 5.2 and 7.1 – **fig. 6.7**). We note that this age is similar within error to the age of the anatectic matrix of the previous sample DKE125B. Th/U ratios in this sample vary systematically with younger zircons showing lower ratios (0.07-0.22) while the older zircons demonstrate higher values (0.22-0.67).

The analyzed zircons for sample DKE-273A zircons have initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratio with values ranging from 0.282104 to 0.282249 for ages between 904 to 846 Ma and correspondent ϵHf_t varying from -3.6 to +1.5. The highest $(^{176}\text{Hf}/^{177}\text{Hf})_i$ ratio of 0.282348 occurs in an inherited zircon with a $^{206}\text{Pb}/^{238}\text{U}$ age of 959 Ma, corresponding to the maximum ϵHf_t value of +6.2. Two grains with well-defined younger melt-precipitated rims were analyzed with $^{206}\text{Pb}/^{238}\text{U}$ ages of 638 and 622 Ma and correspondent ϵHf_t of -0.9 and +0.1, respectively. Although the ϵHf_t for the zircons of the granodioritic protolith yielded mostly neutral values hampering the possibility of evaluation between the distinction of juvenile vs. crustal material, time-resolved oxygen isotopes on the same zircons were more conclusive. The $\delta^{18}\text{O}$ values for the older zircons (830-959 Ma) of 5.20 to 6.44‰ fall mostly within the range of mantle zircon ($5.7 \pm 0.3\text{‰}$), indicating addition of juvenile mantle-derived material in the referred time. Conversely, $\delta^{18}\text{O}$ values of 7.69 to 8.17‰ for the melt-precipitated rims (643-581 Ma) are significantly higher than the mantle zircon, indicating involvement with crustal material or addition of water during the melting event. The granodioritic schollen also have also low initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7020 and positive ϵNd_t of +3.84 at ($t=892$ Ma), pointing to derivation of juvenile mantle-derived sources.

Sample DKE-273B, representative of the diatexite matrix yielded younger ages and several inherited zircons from the melted protolith. Zircons from this sample are also euhedral, transparent, colorless, with crystals ranging in length from 80 to 200 μm (**fig. 6.6**). The calculated Concordia age at 618 ± 4.1 Ma (**fig. 6.7**) was acquired from newly formed zircons from the melt (c.f. zircons 3.1 and 9.1 – **fig. 6.6**) or from melt-precipitated overgrowths around older magmatic cores (c.f. zircons 2.2 and 7.1 – **fig. 6.6**). This age is

equivalent to that obtained from the melt-precipitated overgrowths found in the zircons from the schollen in the sample DKE-273A and also from the diatexite sample DKE125B, and represents more precisely the time of the anatexis. Ages from older cores (c.f. zircons 2.1 and 4.1 – **fig. 6.6**) scatter between 728 and 879 Ma and do not define a precise age in the concordia diagram, suggesting an inherited nature from the precursor source material prior the melting event. In general, variations between the initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratio and the $^{206}\text{Pb}/^{238}\text{U}$ ages for the melt-precipitated zircons in diatexite of the sample 273B are significantly higher than the zircons extracted from the schollen, with values ranging from 0.282152 to 0.282687 for ages between 637 to 607 Ma and correspondent ϵHf_t varying from -1.4 to + 5.4 (**fig. 6.10A**). Two older cores, inherited from the schollen were also analyzed and yielded ϵHf_t of -0.5 and +14.5, suggesting some incorporation of juvenile material from the schollen protolith, as expected from field observations. The $\delta^{18}\text{O}$ values for the melt-precipitated rims and newly formed zircons of 6.41 to 9.10 ‰ are also higher than the mantle zircon, indicating addition of water during the melting event and or contamination with crustal material (**fig. 6.10B**). As also expected, the older cores inherited from the schollen have mantle signatures with zircons values ranging from 4.64 to 5.53‰ (**fig. 6.10B**). This diatexite have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7079 and negative ϵNd_t of -3.55 at $t=618$ Ma, suggesting that crustal material was also involved in the genesis of the diatexites (**fig. 6.11**).

6.4.2. Zircon SHRIMP O isotopes in detrital zircons

Forty-one analyses of Neoproterozoic zircons (939-648 Ma) extracted from two samples of metatextitic paragneisses (samples DKE-43 and 45) of the Ceará Complex (close to Miráima town) were also performed to evaluate the changes in mantle and crustal involvement with time. $^{206}\text{Pb}/^{238}\text{U}$ ages of the same analyzed zircons were previously acquired by Ganade de Araujo et al. (2012a) and the ages of the paragneisses are younger than 650 Ma and their anatexis was estimated to be at 640-600 Ma.

According to these authors zircons were shed from a long-lived arc system (the Tamboril-Santa Quitéria Complex) and deposited in a forearc basin. In general both samples have a significant variation between low and high $\delta^{18}\text{O}$ values, however, lower mantle-like values ($\delta^{18}\text{O} < 6.0\text{‰}$) are consistently more abundant in the sample DKE-43. The $\delta^{18}\text{O}$ values for this sample range from 3.64 to 8.11‰ with 78% of the total analyzed zircons ($n=22$) exhibiting values of $\delta^{18}\text{O} < 6.0\text{‰}$ throughout the range of 949 to 648 Ma (**fig. 6.10B**). For the sample DKE-45 $\delta^{18}\text{O}$ values varies from 5.09 to 7.73‰ with 36% of the analyzed zircons ($n=19$) showing values $< 6.0\text{‰}$ for a narrower range of time between 711 and 932 Ma (**fig. 6.10B**). $^{206}\text{Pb}/^{238}\text{U}$ ages and $\delta^{18}\text{O}$ values indicates that mantle derived sources persisted throughout time since the beginning of the Neoproterozoic arc magmatism, however the presence of zircons with high $\delta^{18}\text{O}$ values ($> 6.0\text{‰}$) from 869 to 662 Ma also suggests that the sources (magmas) of these zircons also have interacted with crustal materials.

Figure 6.8 – **A.** K_2O versus SiO_2 diagram of Peccerillo and Taylor (1976), showing that granitoids are high-K calc-alkaline to shoshonitic in nature. **B.** A/NK vs. ASI diagram modified from Shand (1947). **C.** Rb versus $Ta+Yb$ tectonic discrimination diagram of Pearce et al. (1984). **D.** Th/Hf versus Ta/Hf discrimination diagram between continental active margins and within plate volcanic zones of Shandl and Gorton (2002).

6.4.3. Major and trace elements

Geochemical results does not allow any particular discrimination among the major units of the Tamboril-Santa Quitéria Complex, instead, granitoids show similar trace and REE patterns mostly characteristic of convergent plate margins.

6.4.3.1. Lagoa Caiçara unit

Geochemically, the non-melted portions of from the older group of gneisses and migmatites (ca. 830 Ma) have SiO_2 ranging from 65.3 to 68.2 wt.%. The K_2O contents range between 2.1 and 5.9 wt.% with an average of 3.9 wt.% with the samples plotting mostly in the high-K calc-alkaline field in the K_2O versus SiO_2 classification diagram of Peccerillo and Taylor (1976) (**fig. 6.8A**). Their Al_2O_3 contents ranges from 13.8 to 19.6 wt.% yielding a metaluminous to subordinately weak peraluminous signatures ($ASI=0.73-1.08$) (**fig. 6.8B**). The geochronological data presented herein identified not only Early Neoproterozoic migmatitic orthogneisses, but also orthogneisses, which protoliths have crystalized at ca. 650 Ma. These ca. 650 Ma orthogneisses have SiO_2 ranging from 55.5 to 62.2 wt.% with an average of 57.7 wt.% and similar K_2O (1.77-4.86 wt.%) contents of the older gneisses.

In the primitive mantle-normalized spidergram (**fig. 6.9**), the samples from the both groups (ca. 800 and ca. 650 Ma) show characteristic negative anomalies of Th, Nb, La, P and Ti. In the case of P and Ti this is attributed to a residue of apatite and ilmenite in the parental magma. These rocks have similar REE contents when compared with typical I-type granites. All samples of the older group (ca. 800 Ma) exhibit high REE contents, relatively enrichment of LREE ((La/Yb)_N ratios of 4.3 to 44.5 with an average of 10.2), flat HREE patterns ((Tb/Yb)_N ratios of 1.0 to 2.4) and strong to weakly negative Eu anomalies (Eu/Eu* ratios of 0.56 to 0.99) (**fig. 6.9**). Samples from both groups plot within the VAG field in the tectonic discriminant diagram of Pearce et al. (1982) and in the active-margin granites of Shandl and Gorton (2002) (**figs. 6.8C and D**).

6.4.3.2. Boi unit

In general, samples from this unit have SiO₂ ranging from 67.0 to 69.1 wt.% with an average of 67.8 wt.%. They have rather high K₂O (2.4-7.4 wt.%) and low MgO (0.63-1.0 wt.%) contents with samples plotting mostly in the high-K calc-alkaline field in the K₂O versus SiO₂ diagram (**fig. 6.8A**). Their Al₂O₃ contents are between 14.9 and 15.7 wt.% giving the rock a weak peraluminous signature (ASI=0.99–1.06) (**fig. 6.8B**). They have low Ba (479-868 ppm) and Sr (152-329 ppm) contents, and characteristic negative anomalies of Nb and Ti and positive anomalies of U, K and Ce in the primitive mantle-normalized spidergram (**fig. 6.9**). Normally, the samples show a relatively enrichment of light rare earth elements (LREEs)((La/Yb)_N ratios of 8.2 to 91.5 with an average of 11.8)), and a predominant strong negative Eu anomalies (Eu/Eu* ratios ≈ 0.63).

6.4.4.3. Santa Quitéria unit

Geochemically, the samples of Santa Quitéria unit have SiO₂ contents in between 58.7 and 75.4 wt.%, with an average of 61.1 wt.%. K₂O contents range between 1.8 and 7.4 wt.% with an average of 3.1 wt.% with the samples plotting mostly in the high-K calc-alkaline and shoshonitic fields in the K₂O versus SiO₂ classification diagram of Peccerillo and Taylor (1976) (**fig. 6.8A**). The samples demonstrate overall patterns of decreasing Mg, Fe, Ca, Ti, Al and P with increasing SiO₂. Their Al₂O₃ contents are in between 13.3 and 17.4 wt.% indicating a metaluminous to weak peraluminous character (ASI=0.73–1.07) (**fig. 6.8B**). The samples display an enriched LILE pattern, defining a downward sloping profile in the primordial mantle normalized spidergram, combined with positive anomalies of K, Pb and Nd and negative Nb, Th, P and Ti anomalies (**fig. 6.9**). In spite of the significant variance of Ba and Sr, the former appears especially abundant, with average values of 727 and 223 ppm, respectively. Generally the analysed samples exhibit high REE contents, relatively enrichment of light rare earth elements (LREEs)((La/Yb)_N ratios of 3.3 to 67.8 with an average of 14.9), flat HREE patterns ((Tb/Yb)_N ratios of 0.7 to 3.1) and predominant negative Eu anomalies (Eu/Eu* ratios ≈ 0.83).

6.4.4.4. Tamboril unit

In general terms, samples of the Tamboril diatexite are geochemically similar to that of the Santa Quitéria unit. They have SiO₂ ranging from 62.4 to 68.3 wt.% with an average of 64.3 wt.%. K₂O contents range

Figure 6.9 – REE and spider diagrams for granitoid rocks of the protolith of granitoids and migmatites from Tamboril-Santa Quitéria Complex. Primitive mantle and Chondrite normalized values from McDonough and Sun (1995) and Sun and McDonough (1989), respectively.

between 1.6 and 6.5 wt.% with an average of 3.9 wt.% with the samples plotting mostly in the high-K calc-alkaline and shoshonitic fields in the K_2O versus SiO_2 diagram (**fig. 6.8A**). Their Al_2O_3 contents are in between 13.6 and 16.9 wt.% that gives a metaluminous to subordinately weak peraluminous characteristic ($ASI=0.84-1.04$) (**fig. 6.8B**). In the primitive mantle-normalized spidergram samples show characteristic negative anomalies of Nb, P and Ti that should be attributed in part to the residue of apatite and ilmenite in the parental magma. The samples exhibit high REE contents, relatively enrichment of light rare earth elements (LREEs) ($(La/Yb)_N$ ratios of 2.9 to 85.1 with an average of 20.4), flat HREE patterns ($(Tb/Yb)_N$ ratios of 0.7 to 4.7) and predominant negative Eu anomalies (Eu/Eu^* ratios ≈ 0.63) (**fig. 6.9**).

6.5. Discussion

6.5.1. Magmatic Evolution

Geochemistry, U–Pb zircon ages, time-resolved zircon Hf–O isotopic determinations and whole-rock Sr–Nd isotopes of the Tamboril-Santa Quitéria Complex obtained in this study provide important constraints on the magmatic and tectonic evolution of the Ceará Central Domain. Trace element concentrations of the investigated samples display a typical spectrum of arc-related igneous rocks, the so-called “arc-signature”, characterized by the enrichment of highly mobile large ion lithophile elements (LILE) relative to high field strength elements (HFSE) (McMillan et al., 1989). However, it is the isotopic composition that characterizes better the source of the investigated granitoids. Essentially, magmatism can be divided into three main periods with their particular characteristics: i) an early period comprising essentially juvenile arc magmatism at ca. 880–800 Ma, ii) a more mature arc period at ca. 660–630 Ma characterized by hybrid mantle-crustal components, and iii) crustal anatexis at 625–618 Ma continuing until ca. 600 Ma. In the following discussion we will avoid the unit nomenclature based on mapping, and divide the investigated samples according to their age and isotopic signatures.

6.5.1.1. Early 880–800 Ma juvenile arc-related magmatism

Samples 273A and DKE-221 of granodioritic/tonalitic, yielded the oldest zircon crystallization ages at 892 ± 7.5 and 833 ± 6.1 Ma, respectively. These samples have predominantly positive $\epsilon Hf(t)$ (–3.6 to +19.3) and positive $\epsilon Nd(t)$ (+4.98 to +3.84) combined with low initial $^{87}Sr/^{86}Sr$ (<0.7025), suggesting derivation from a depleted mantle (juvenile) source. Nonetheless, detrital zircons from forearc deposits of the Ceará Complex suggests that magmatism was continuously active from at least ca. 900 to ca. 650 Ma (Ganade de Araujo, 2012a) (**fig. 6.12**). The $d^{18}O$ values retrieved from the same detrital zircons previously dated by these authors (samples DKE-43 and DKE-45 of the Ceará Complex), indicate that the juvenile input persisted throughout great part of the convergent magmatism ascribed to the consumption of the Goiás-Pharusian Ocean (**fig. 6.10B**).

Geochemistry of these 880–800 Ma tonalitic to granodioritic rocks suggests that this juvenile signature was acquired in a arc-related setting rather than during rifting. In the Ceará Central Domain, some authors favour break-up and rift development at around 770–750 Ma (Fetter et al., 2003; Castro, 2004; Arthaud, 2007; Brito Neves et al. 2013), however the lack of characteristic features of rift settings such as concomitant immature

terrigenous sedimentation, abrupt tectonically-controlled facies variations and abundant bimodal volcanism, does not support this idea.

Figure 6.10 – A. Variations of $\delta^{18}\text{O}$ values with age for zircons from protolith of granitoids and migmatites from Tamboril-Santa Quitéria Complex. **B.** Schematic diagram for Lu-Hf isotopic evolution vs. U-Pb age for zircons from protolith of of granitoids and migmatites from Tamboril-Santa Quitéria Complex.

Instead, such extensional event may be related to an extensional subduction setting and development of diachronous back-arc basins to the east of the Lagoa Caiçara unit. On the other hand, a U-Pb ID-TIMS age of ca. 770 (Fetter et al. 2003) retrieved from volcanic rocks found associated with passive margin deposits of the Martinópolis Group in the Médio Coreáú Domain (west of the Transbrasiliano Lineament in **fig. 6.2**) suggest that extension and passive margin development was concurrent with subduction and arc development in the Ceará Central Domain.

Figure 6.11 – A. Relationship between zircon $\epsilon\text{Hf}(t)$ and whole-rock $\epsilon\text{Nd}(t)$ for protolith of granitoids and migmatites from Tamboril-Santa Quitéria Complex. Mantle and crust arrays are from Vervoort et al. (1999). **B.** $\epsilon\text{Nd}(t)$ vs. $\epsilon\text{Sr}(t)$ diagram protolith of granitoids and migmatites from Tamboril-Santa Quitéria Complex. The line separating materials derived from upper (high positive ϵSr) to lower crust (low positive ϵSr) have been proposed by DePaolo and Wasserburg (1979).

Evidence from the West Gondwana Orogen in Africa (Caby, 1989, 2003; Berger et al., 2011; Dostal et al. 1994) and Central Brazil (Pimentel and Fuck, 1992; Pimentel et al., 2000; Laux et al., 2005) demonstrates that part of the Neoproterozoic growth of western Gondwana occurred firstly during the Late Tonian and Cryogenian (950-750 Ma), through the development of intraoceanic juvenile arcs, suggesting the presence of a large ocean separating the São Francisco and Amazonian/West African and Saharan cratons. In Hoggar, within the Silet region (Algeria), diorite-tonalite and monzogranite plutons from the Iskel magmatic arc yielded U-Pb zircon ages at *ca.* 868 and 839 Ma (Caby et al., 1982). Occurrence of slices of pre-Pan-African

basement directly overlain by shelf sediments and capped by arc volcanic rocks in several localities suggests that the Iskel magmatic arc was built on attenuated continental crust adjacent to possible slices of oceanic lithosphere (Lapierre et al., 1986; Caby, 2003). Further south, in the Gourma region (Mali) the Tilemsi-Amalaoulaou intraoceanic arc assemblages (Dostal et al., 1994) were dated within the 790-710 Ma time interval (Caby et al., 1989; Berger et al., 2011). The Tilemsi arc is considered the upper crust supra-structure equivalent of the Amalaoulaou complex (Berger et al., 2011). Although precise geochronological data is lacking for the Kabyé massif in the Dahomeyan belt (Togo), geochemical and field characteristics suggest that this massif could in fact represent the roots of a continental arc (Duclaux et al., 2006).

In Africa the active continental margin (Andean-type) are located east of the oceanic terranes (Caby, 2003; Berger et al., 2011). This stage of ocean-continent subduction was dated at 696 ± 5 Ma within the Kindal Terrane and at 716 ± 6 Ma in the Idras des Iforas region in Mali (Caby and Andréopoulos-Renaud, 1987; Bruguier et al., 2008), indicating that it was partially coeval with the ocean-ocean subduction stage active further west. In Central Brazil, the Neoproterozoic Goiás magmatic arc in the Brasília Belt is composed of juvenile orthogneisses ranging from ca. 920 to 780 Ma (Pimentel and Fuck, 1992; Laux et al., 2005; Matteini et al., 2010). Younger ages at ca. 670-630 Ma were also reported (Laux et al., 2005) and may represent a second stage of the pre-collisional magmatism in Central Brazil, also with hybrid mantle-crustal isotopic signatures.

In other words, the juvenile nature of these rocks in the CCD and geological relationships along the orogen in Africa and Central Brazil suggests that the large Goiás-Pharusian Ocean were connected and did not narrow into a small ocean in the Borborema Province as suggested by some authors (Castaing et al. 1994; Neves et al., 2003; Brito Neves et al., 2013).

6.5.1.2. Mature Andean-type arc magmatism: ca. 660-630 Ma

The granitoids of the younger magmatism marked by the Santa Quitéria and Boi units (samples DKE-211 and DKE-277) together with the gneissic granitoids found in the Lagoa Caíçara unit (samples DKE-269 and DKE-231) and granitoid schollen (samples DKE-170 and DKE-125A) found within the diatexites of the Tamboril unit range in age between 663 and 627 Ma. These rocks have negative to positive ϵHf_t (-18.7 to +13.2) and ϵNd_t (-10.75 to +1.80) combined with moderate to high initial $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7056-0.7143). Isotopic results for the granitoids within this 30 m.y. span of magmatism indicate sources ranging from mantle to continental (**table 6.2**), which characterizes a mature arc stage.

After the juvenile granitoids of the 890-800 Ma arc, the oldest granitoid (663 ± 6.6 Ma, sample DKE-170) within the Tamboril-Santa Quitéria Complex occurs as a raft inserted in the Tamburil unit close to the contact with the Santa Quitéria unit. The ϵHf_t , ϵNd_t and initial $^{87}\text{Sr}/^{86}\text{Sr}$ indicate that this granitoid was derived from the partial melting of depleted mantle sources (**figs. 6.10 and 6.11**). However, high zircon $\delta^{18}\text{O}$ values (5.94-9.06‰) suggests that these juvenile magmas would have also interacted with (meta)sedimentary rocks that contributed to increased $\delta^{18}\text{O}$ values. Sample DKE-170 contrasts with sample DKE-200A, the next oldest rock in this group, however. Sample DKE-200A is a mafic tonalite dated at 650.6 ± 5.1 Ma. Its isotopic composition indicates that old continental rocks were its main source (initial

$^{87}\text{Sr}/^{86}\text{Sr}=0.7105$; $\epsilon\text{Hf}_i=-5.45$ and $\delta^{18}\text{O}$ 6.73-8.19, **table 6.2**). The difference between the two samples is taken to indicate that contrasting sources (crust and mantle) were mobilized in this period. This is in fact confirmed by consideration of the remaining samples in this group.

The high-K to shoshonitic mafic dioritic and tonalitic rocks of the Boi Unit are the oldest (648 ± 4.1 Ma, sample DKE-277) coherent and mappable magmatic rocks identified within the complex. Whilst ϵHf_i for sample DKE-277 is negative (-6.6 to -0.8) and suggestive of crust participation, juvenile ϵNd_i signatures (Fetter et al., 2003), attest for mixing between rocks with mantle and crustal signatures. Mantle involvement is further supported by the zircon $\delta^{18}\text{O}$ values (5.48-6.25‰). The diorite gneiss of sample DKE-125A although lacking isotopic data, has a similar zircon U-Pb age of 646 ± 4.5 Ma and is correlated with the Boi unit magmatism.

The youngest magmatic intrusive pulses in the Tamboril-Santa Quitéria Complex are represented by the 632 ± 5.1 and 627 ± 4.9 Ma biotite granitic magmatism found in the Lagoa Caiçara unit. Nd-Sr isotopic data for these rocks are coherent with a crustal origin as also suggested by high zircon $\delta^{18}\text{O}$ values (6.73-10.82‰). Inherited zircons with ages at 823 ± 23 , 796 ± 19 and 761 ± 19 Ma indicate that Early Neoproterozoic juvenile protoliths from the Lagoa Caiçara unit were also important sources for this granitic magmatism, and may contributed for the partially positive ϵHf_i in the sample DKE-269.

As discussed above, one of the main features of the Santa Quitéria monzogranitic magmatism is the close association with syn-plutonic mafic dikes of enriched mantle affinity, likely connected with the Boi unit magmatism. This mantle input is geochemically enriched and predominantly shoshonitic in nature (Costa et al., 2013; Zincon, 2011). Available geochronological data for the high-K to shoshonitic porphyritic granites of the Santa Quitéria unit allow us to bracket its formation to within the 640-635 Ma time interval (Fetter et al. 2003 and our data). Negative ϵHf_i (-12.2 to -2.9) and ϵNd_i (-4.25) values together high initial $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7107) and high zircon $\delta^{18}\text{O}$ values (7.06-8.57‰) indicate that crustal also were involved in the formation of the Santa Quitéria monzogranites. The Boi and Santa Quitéria units are part of the same magmatic system and illustrate well the interaction of crust-mantle sources commonly described in mature arcs (DePaolo, 1981).

The enriched signatures observed in the Santa Quitéria-type granitoids could be explained by partial melting of a modified metassomatic mantle combined with significant crustal contamination, rather than an asthenosphere input. In many arcs described worldwide magmas have enriched geochemical features, which are consistent with a derivation from mantle sources modified by metasomatic fluids. These fluids can be derived from subducted incompatible element-rich sediments (Tatsumi et al., 1986), or from slab melts (Martin et al., 2005). The relative roles of crustal contamination and mantle source enrichment (e.g. through the contribution of subducted terrigenous sediments or slab fluids) are often debated in arc petrogenesis (e.g. Fourcade et al., 1994), but difficult to quantify. The expected modifications in the underlying mantle would arise from the long-lasting interaction of subduction derived melts since the ca. 850 Ma, initiated by the Lagoa Caiçara juvenile magmatism.

Along the West Gondwana Orogen, other late Andean-type arcs have also been identified in the time bracket between 650 and 600 Ma. As mentioned above, in Hoggar (Mali) such arc magmatism is related with the

consumption of the Goiás-Pharusian Ocean by east-southeast directed subduction (Caby et al., 1981) and the formation of the large Adrar des Iforas continental arc batholith at around 630 Ma (Liegiois et al., 1987). In the Dahomey section of the orogen in Togo and Benin, arc-type Neoproterozoic granitoid rocks dated at ca. 650-630 Ma (Kalsbeek et al., 2012) are also related with an east-dipping subduction zone evolved during consumption of the Goiás-Pharusian Ocean. Finally, in the central Brazil branch of the orogen, final magmatic pulses of the Goiás magmatic arc at ca. 630 Ma (Pimentel et al., 1999; Laux et al., 2005) could be correlated with this mature arc setting that predates final collision in the West Gondwana Orogen.

6.5.1.3. Reworking of arc rocks: the 620-610 Ma crustal anatexis event

Samples of neosomes resulting from crustal anatexis define the youngest group of rocks within the complex at 625-610 Ma, generally grouped in the Tamboril unit. In the field the Tamboril magmatism results from the remelting of the surrounding protoliths (**fig. 6.5**), mainly orthogneisses of intermediate compositions of the Lagoa Caiçara and Santa Quitéria units and minor metasedimentary rocks of the Ceará Complex.

The resulting magmatism in the Tamboril unit dated here at 625-618 Ma (Samples DKE-273 and DKE-125) consists of neosomes and their isotopic composition reflects the variations of their source from juvenile to hybrids with some crustal input.

The schollen diatexite of sample DKE-273 suggests that it was derived from the partial melting of the juvenile rocks of the Lagoa Caiçara unit, which is found as rafts within the diatexites. Intermediate granitoids, such as those of the Lagoa Caiçara unit have no muscovite and small amounts of biotite or hornblende (10-25 %), precluding generation of large melt fractions by dehydration melting (Sawyer, 2008). The large melt fraction and the lack of residual anhydrous phases in these migmatites, such as garnet, sillimanite, orthopyroxene or cordierite, suggest melting by influx of water close to the solidus temperature promoting water-saturated melting of quartz+plagioclase-K-feldspar (Kenah and Hollister, 1983; Sawyer 1998; Sawyer, 2008). In support of this interpretation is the existence of peritectic hornblende in leucosomes in some sections of the Lagoa Caiçara unit (see sample DKE-221). Gardien et al. (2000) have demonstrated that the stability of residual hornblende formed from biotite breakdown requires addition of external water.

Close investigation of our analyses of diatexitic granite and schollen of samples DKE-273A and 273B suggest however that other sources were also involved in the generation of melts surrounding the schollen. Although melting of the tonalitic/granodioritic paleosome and generation of the diatexite melt is evident in the field, zircon $eHf(t)$ values diverge from the whole-rock $eNd(t)$, suggesting Hf-Nd isotope decoupling in the diatexites (**fig. 6.11A**). The behavior of the Lu-Hf system during melting is analogous to that of the Sm-Nd system, with the daughter element Hf and Nd fractionating into the melt to a higher degree than the parent element Lu and Sm (Scherer et al., 2007). Because Hf and Nd fractionate more strongly into melts than Lu and Sm, the melt will have lower Lu/Hf and Sm/Nd values than the protolith and over time the isotopic compositions of the melt and protolith will diverge into lower and higher $^{176}Hf/^{177}Hf$ and $^{143}Nd/^{144}Nd$ values, respectively. The $(^{176}Hf/^{177}Hf)_i$ values for the zircons of the diatexite of the Lagoa Caiçara unit are higher or equal to the $(^{176}Hf/^{177}Hf)_i$ of the source juvenile material (schollen), indicating that radiogenic ^{176}Hf remained constant or slightly increased during the melting event. We interpret this feature as a direct consequence of the isotopic

inheritance of the juvenile source zircons to the melt-precipitated zircons. We believe that the Hf budget in the melt is being controlled mainly by the zircons with high $(^{176}\text{Hf}/^{177}\text{Hf})_i$ derived from the juvenile protoliths and that rapid melting by addition of water would preclude radiogenic ^{176}Hf to homogenize with other possible sources and thus reflect the direct isotopic composition of the protolith (**fig. 6.12**). The decoupling of zircon Hf versus whole-rock Nd isotopes in the Lagoa Caiçara diatexites is due to the retention of radiogenic Hf during partial melting of juvenile arc-derived zircons, similarly suggested Wu et al. (2006) for the reworking of juvenile crust in South China. Since the bulk ^{143}Nd is available from a variety of minerals and sources, rather than zircon which is the main container of Hf in crustal rocks (Hoskin and Schaltegger, 2003) we believe that during partial melting the whole-rock Sm-Nd system was readily equilibrated with the new melt, and reflects the addition of other external, old continental sources that contributed to the lower and less radiogenic $\epsilon\text{Nd}_{(t)}$ value. Hf provided by zircons from the external contaminants were minor compared with the Hf provided by the juvenile protolith, and this may reflect: i) low zircon fertility of the crustal contaminants; ii) low magmatic resorbing of these zircons in the melt; or iii) a bias introduced by our low resolution sampling. This external sources are also observed in the field as preserved schollen of metasedimentary rocks, granites and older Paleoproterozoic (2.1 Ga) orthogneisses from the basement.

The rapid addition of water during melting could explain the conservation of the protolith Hf isotopic signature of the melt-precipitated zircons as well as their high $\delta^{18}\text{O}$ values (**fig. 6.12**). The origins of the fluids in geological processes are always intriguing and difficult to address. The time of diatexite formation is in agreement with the time of continental collision in Ceará Central Domain (see discussion below) and thus fluids associated with subducted material and underlying metasomatized mantle wedge are not possible sources. Instead, fluids released by prograde collisional metamorphic dehydration-type reactions of the adjacent rocks are suitable candidates, as proposed White et al. (2005) at a smaller scale by for the diatexites of Broken Hill, Australia.

6.5.1.4. Bracketing collision time

The fundamental question that arises when addressing temporal relationship of magmatic lineages of a given orogenic system, using the prefixes pre-, syn- and post-collisional is: *when did the collisional stage start?* Initial collision, starting at the first contact of the continental blocks, evolves into crustal thickening (due to plate overriding) followed much later by thinning due to gravitational adjustments in response to delamination of crustal root (Leech et al., 2001). Each of these tectonic stages can be fingerprinted by a related tectono-thermal and magmatic manifestation preserved within the final orogenic record.

Retrogressed eclogitic rocks found between the western border of the Santa Quitéria Complex and the Transbrasiliano Lineament (Santos et al. 2009) are an essential piece of the collisional story of de orogen. Santos et al. (2013) reported the find of coesite inclusions within garnet, suggesting UHP (>2.7 GPa) metamorphic conditions at depths major than 90 km. It is well known from recent collisional orogens, as well as in some fossil collisional zones, that eclogite facies metamorphism, including UHP rocks, is one the best markers of the onset of the collisional process (e.g. Liou et al. 2004; Liu et al., 2008; de Sigoyer et al., 2000; Leech et al., 2005, Gilotti, 2013). Petrochronology for the (U)HP metamorphism in the Forquilha eclogitic zone in CCD and along the West Gondwana Orogen in Togo and Mali indicate that the timing of continental

collision was nearly synchronous for at least 2500 km along the orogen around 615-610 Ma (Ganade de Araujo et al., submitted).

Figure 6.12 – Schematic illustration of the water-fluxed melting of the juvenile protoliths. Hf budget in the melt is mainly controlled by high initial $(^{176}Hf/^{177}Hf)_i$ of the juvenile arc-related protoliths, thus yielding a melt with high initial $(^{176}Hf/^{177}Hf)_i$. Nd budget is controlled by the mixing of juvenile protoliths and crustal contaminants yielding neutral ϵ_{Nd} signatures in the melt. Water-fluxed melting of the juvenile protoliths are indicated by the lack of anhydrous peritectic phases in the melt, as well as by high $\delta^{18}O$ signatures.

Given the marked change in the nature of magmatism, from primary arc magma intrusion down to 625 Ma, to secondary magmatic rocks derived from the remelting of these primary magmatic rocks, at around 620-615 Ma, we postulate that this marks a change from early magmatism related to plate convergence associated to the consumption of the Goiás-Pharusian Ocean along the West Gondwana Orogen to crustal recycling due to collision.

The India-Asia collision is our type locality for large-scale continental collision. There collision started ca. 55 Ma (Klootwijk et al., 1992), ultimately creating the Himalaya and Tibet. The most obvious metamorphism occurred during partial melting ca. 20 Ma, but rare relict metamorphic minerals, textures, and isotope ages as old as 35-55 Ma attest to earlier Himalayan metamorphism (e.g., see Hodges, 2000; de Sigoyer et al., 2000). In the Himalayas ages for the coesite-bearing UHP eclogites are 45-55 Ma (de Sigoyer et al., 2000; Kaneko et al., 2003; Donaldson et al., 2013) while geochronology on the partially melted rocks indicates that melting was

almost simultaneous at 18-22 Ma (Harrison et al., 1998), constraining a gap between 37-23 m.y. since the beginning of the collision and the main period of melting. Thus, widespread melting of mid-crustal levels is thought to have started at around 30 Ma, with more voluminous magmatism at around 20 Ma, and with melt present today ~15 km below the surface underneath the Tibetan Plateau (Harrison et al., 2006). Thermal modeling of crustal thickening suggests that the concomitant thickening of the layer enriched in heat producing elements leads to crustal heating generating crustal anatexis some tens of millions of years after crustal thickening in accordance to observations (England et al. 1998). This seems not to be the case in the Ceará Central Domain.

If our interpretation is correct, the swap from arc magmatism to crustal anatexis occurred concurrently with collision suggesting one of several possibilities: a) the CCD is a deeper section of the collisional belt than the one presently exposed in the Himalayas, b) the thermal evolution of the exposed section of the CCD was different from that of the Himalayan front upon collision, with the CCD remaining hotter, c) influx of water related to the arrival of continental sediments into the subduction zone bringing water to the arc. It should be stated that although melting was synchronous to the onset of collision in CCD, younger leucosomes containing anhydrous peritetic garnet and sillimanite derived from the partial melting of the metasedimentary rocks of the Ceará Complex were dated at 610-600 Ma (Castro, 2004; Arthaud, 2007) and possibly younger at 580 Ma (Fetter, 1999).

6.6. From a juvenile to mature arc setting and terminal collision.

Our new data indicate that subduction initiation of the Goiás-Pharusian Ocean may have been active as early as 890 Ma in the Ceará Central Domain and, as suggested by detrital zircon studies from supracrustal rocks (Ganade de Araujo et al., 2012a), may have been continuous until terminal collision at ca. 620-615 Ma. However, the continuity of detrital zircon spectra contrasts with the apparent long pause between ca. 800 Ma and 660 Ma recorded by the magmatic rocks of the CCD alone. This apparent gap could be due to erosion of the earlier arc granitoids or by insufficient geochronological data (**fig. 6.13**).

In any case, the igneous samples investigated here record two main arc-building stages. The first, early to middle Neoproterozoic stage I, comprising mainly juvenile tonalites and granodiorites from the Lagoa Caíçara unit, followed by a second stage (stage II) that comprises abundant diorites, tonalites and mainly high-K monzogranites with mixed mantle-crustal signatures from the Santa Quitéria and Boi units and younger orthogneisses found in the Lagoa Caíçara unit.

The juvenile nature of the stage I arc granitoids suggest an initial emplacement outboard of the leading edge of the continental margin of the Paleoproterozoic-Archean basement of the Borborema Province to the east, at ca. 890 Ma possibly in an oceanic environment. This scenario is similar to that described in the earlier stages of Mesozoic convergent margin of Baja California, Mexico (Busby, 2004). In this area, the subducting Farallon plate at that time was old and cold at the trench and therefore the subduction zone was in retreat and the arc was thus emplaced in an extensional setting, generating intra-arc to backarc basins.

Figure 6.13 – A. Comparison between zircon ages acquired from the granitoids of the Tamboril-Santa Quitéria in this study versus detrital zircons from the back arc and fore arc basins from Ceará Complex (data from the detrital zircons from Ganade de Araujo, 2012a). **B.** Summary of magmatic ages of the granitoid rocks of the Tamboril-Santa Quitéria Complex.

Similarly, if the oceanic plate of the Pharusian-Goiás Ocean was old and cold at the time of subduction in the Ceará Central Domain an extensional setting would have developed between stage I arc and the former continental margin explaining for example the sediments deposited in the rear area of the arc in a possible back-arc setting between the juvenile Lagoa Caiçara unit and the Paleoproterozoic/Archean basement to east. However, provenance studies through detrital zircon investigation in these marine sediments of the Ceará Complex have both arc and continental signatures (Ganade de Araujo et al., 2012a) suggesting that stage I arc magmatism was not far off the continental margin (**fig. 6.14A and B**). Furthermore, some authors have proposed that the bimodal alkaline (high-Nb) and mafic magmatism associated with these sediments between

840-750 Ma is related to extension (Castro, 2004; Arthaud, 2007; Arthaud et al. 2008). Imprecise upper intercept ID-TIMS U-Pb zircon ages at ca. 840 Ma from alkaline rhyolites with high-Nb content close to Itataia town (Castro, 2004) and ID-TIMS U-Pb zircon ages of 772 Ma from felsic gneissic sheets found further south close to Independência town may constrain the period of extension.

Development of a back-arc basin during stage I arc magmatism and extension of the continental crust to the east of the Lagoa Caiçara unit, yielding space that was filled with progradational back-arc deposits that record arc growth above sea level to the west (**fig. 6.14B**). No clear evidence is available to say if the back-arc basin developed into an incipient oceanic crust, however the mafic rocks close to Pentecoste town could be candidates and should be studied in detail.

The mature arc stage II magmatism is comprised of several pulses of granitoids and overprints magmatism related to stage I arc at ca. 660-630 Ma (Fetter et al., 2003; Castro, 2004; Ganade de Araujo et al., 2012b) (**fig. 6.14C**). These magmatic rocks are geochemically enriched when compared with the intermediate granitoids of stage I arc (**fig. 6.8A**). Likewise, contrasting to the stage I arc granitoids, isotopic signatures of stage II rocks show variable mixtures between juvenile and crustal material.

We postulate that after the last pulse of arc magmatism at ca. 627 Ma (sample DKE231), initial continent-continent collision in Ceará Central Domain is marked by the first contact between the stretched passive margin of Paleoproterozoic-Archean basement to the east (the Northern Borborema basement) and the Paleoproterozoic basement to the west (the Parnaíba + Granja Complex). Continental subduction is evidenced by the (U)HP eclogitic metamorphism in the Forquilha HP domain, which may have initiated as early as ca. 624 Ma reaching peak P conditions at ca. 615 Ma (Ganade de Araujo, submitted) (**fig. 6.14D**). At this stage remelting of the arc assemblages took place in the Tamboril-Santa Quitéria Complex.

The period following continental subduction at ca. 615 Ma is related to exhumation of the (U)HP eclogites, especially those found at the Forquilha (U)HP domain. The emplacement of the (U)HP rocks into shallower crustal levels was probably facilitated by extensional tectonics and buoyancy-aided exhumation (**fig. 6.14E**).

6.7. Conclusions

The Ceará Central Domain of the Borborema Province is a Neoproterozoic orogenic area (Brito Neves et al. 2000), part of the 5000 km-long West Gondwana Orogen (Ganade de Araujo, in press), which extends from Algeria in Africa to Central Brazil. Our results allowed determination of three stages of magmatism reflecting three distinct tectonic environments: i) an early period of essentially juvenile arc magmatism at ca. 880-800 Ma, ii) a second, mature arc period between 660-630 Ma, characterized by hybrid mantle-crustal components, and iii) remelting of the arc-related igneous rocks during continental collision, evidenced by abundant extensive migmatization dated to between 625 and 600 Ma. These ages overlap with those of (U)HP eclogitic metamorphism at 624-615 Ma suggesting that migmatization occurred during continental subduction in a continent-continent collisional setting.

A. ca.880 Ma

B. ca.800-750 Ma

C. ca.660-630 Ma

D. ca.620-615 Ma

E. ca.615-600 Ma

Figure 6.14 – Sketch tectonic model for Neoproterozoic tectonic evolution for the continental convergent margin of Ceará Central Domain. **A.** Early subduction stage in an extensional setting, due to old oceanic lithosphere subduction and juvenile magmatism accretion on a stretched continental margin. **B.** Continuous subduction with development of extensional back-arc basins with associated magmatism and both arc- and continental-derived detritus. **C.** Compressive arc-setting and development of the Santa Quitéria arc. **D.** Terminal collision with subduction of stretched continental crust to the west of the Santa Quitéria arc and subduction of the stretched continental crust (e.g. back-arc basin) to the east of the Santa Quitéria arc. Collisional metamorphism on both sides of the arc are evidenced by (U)HP-eclogite rocks of Forquilha (Santos et al., 2009; Santos et al., 2013; Ganade de Araujo, submitted) and Itatiaia HP eclogites (Castro, 2004). **E.** Post-collision extension and exhumation of the (U)HP and HP rocks.

The apparent gap between the two periods of arc magmatism could be explained by incomplete exposure and erosion. Evidence for continuous magmatism comes from abundant detrital zircons in the fore- and back-arc basins with ages in the range of 900 to 650 Ma. Oxygen isotopes from detrital zircons in the fore-arc indicate that juvenile input persisted throughout the entire evolution of convergent magmatism. Igneous rocks of the Tamboril-Santa Quitéria Complex record a long-lived history of convergent magmatism lasting up to 350 m.y.

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6.8. References

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7. Extruding the Borborema Province (NE-Brazil): a two-stage Neoproterozoic collision process

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Abstract

We propose that Borborema Province development from 620 to 570 Ma resulted from two discrete collisional events. Collision I, along the West Gondwana Orogen on the west side of the Province, took place at ca. 620-610 Ma as the result of collision between the Parnaíba Block, as the forefront of the much larger Amazonian-West Africa Craton, and the old basement of the Borborema Province. The suture zone related to this collision was reactivated by a dextral transform zone (the Transbrasiliano Lineament), allowing the Borborema Province to approach and collide against the São Francisco Craton in the south at ca. 590-580 Ma marking collision II along the Sergipano Orogen. The combined stresses related to eastward push from collision I and northward push from the cratonic indentation onto a thickened lithosphere gave rise to an extensive network of strike-slip shear zones across the Province forcing its northeastward extrusion.

7.1. Introduction

Lateral escape of continental blocks occurs in many active collisional areas (e.g., Sengör et al., 1985; Tapponier et al., 1982). Extrusion tectonics has been suggested for the Borborema Province, in northeast Brazil (e.g., Brito Neves et al., 2000; Alkmim et al., 2001; Bueno et al., 2009), but the relation to its large-scale tectonic evolution remains unclear. Here we present the hypothesis that the evolution of the Province during the Neoproterozoic (620-570 Ma) results from interference between two collisions.

The Province is bound to the west by the Parnaíba Block, which is inferred from geophysics and considered here as the forefront of the much larger Amazonian-West African Craton (**fig. 7.1**). This boundary is adjacent to a set of dextral high-T shear zones collectively known as the Transbrasiliano Lineament (equivalent to the Kandi Lineament in Africa). To the south the Province is bound by the São Francisco Craton.

It is composed of large areas of Archean/Paleoproterozoic gneissic/migmatitic rocks, as well as restricted Mesoproterozoic/Early Neoproterozoic rocks of the Cariris Velhos Belt, which make up the basement of metamorphosed supracrustal rocks, Neoproterozoic to Cambrian intrusions, and its well-known network of transcurrent shear zones (Caby, 1989; Vauchez et al., 1995; Brito Neves et al., 2000; Weinberg et al., 2004; Neves et al., 2012; Archanjo et al., 2013). In our view, by the end of the Neoproterozoic, the Province constituted a coherent block of stable continental crust of pre-Neoproterozoic age, with small intracontinental basins in its central portion (Neves, 2003), which was highly remobilized during and after the following collisional events described here.

In the Province, a number of crustal-scale shear zones branch out of the major Transbrasiliano shear system (**fig. 7.2**). Continuity between high-strain zones, similar P-T conditions of deformation and kinematic coherence suggest that this network represents a single system (Vauchez et al., 1995).

Figure 7.1 – Position of cratons, blocks, Brasiliano Neoproterozoic orogens and Neoproterozoic to Cambrian fold and thrust belts in Brazil (modified from Alkmim et al., 2001). Cratons in the inset: AM: Amazonian, WA: West Africa, SF: São Francisco, C: Congo, S: Saharan “metacraton”.

The E-W trending Patos and Pernambuco dextral shear zones divide the province into Northern, Central and Southern sub-provinces (Brito Neves et al., 2000; Neves, 2003). Dextral strike-slip shear zones trending NE-SW characterize the Northern sub-province. The Central sub-province is characterized by a network of conjugate shear zones, comprising a set of E-W trending dextral shear zones and a sinistral set trending NE-SW (Neves et al., 2012). The Southern sub-province is limited by the São Francisco Craton in the south and is characterized by south-verging thrusting with a small dextral component (Oliveira et al., 2010). Geochronological (U-Pb) and thermochronological (Ar-Ar) data (table 7.1 and figure 7.2) indicate that deformation along these shear zones peaked with associated magmatism from 590 to 560 Ma and extended into lower T conditions from 550 to 500 Ma in the Central sub-province (e.g., Monié et al., 1997; Corsini et al., 1998; Guimarães et al., 2004; Neves et al., 2008; Hollanda et al., 2010; Neves et al., 2012; Archanjo et al., 2013).

Regional deformation and metamorphism were synchronous throughout the Northern and Central sub-provinces, starting before ca. 630 Ma, but are younger in the Southern sub-province, starting at ca. 610-570 Ma (Arthaud et al., 2008; Oliveira et al., 2010; Amaral et al., 2010; Neves et al., 2012). High-P/high-T regional metamorphism prevails in the northern part, where eclogites have been found between the Transbrasiliano Lineament and the Santa Quitéria continental arc (Santos et al., 2009).

Low-P/high-T metamorphism characterizes the central part, dominated by migmatites and gneisses (Neves et al., 2012), and lower P/T conditions (amphibolite to greenschist facies) are dominant in the Southern sub-province (Oliveira et al., 2010).

Table 7.1 – Summary of the main U-Pb zircon and Ar-Ar ages of the Borborema Province.

Northern Borborema Province		Mineral	Technique	Reference
<i>Magmatic ages</i>				
876±6	Lagoa Caiçara Complex granodiorite	zircon	SHRIMP	Ganade de Araujo et al. (2012a)
831±7	Lagoa Caiçara Complex tonalite	zircon	SHRIMP	Ganade de Araujo et al. (2012a)
655±5	Tamboril Santa Quitéria Complex granodiorite	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
652±5	Tamboril Santa Quitéria Complex granodiorite	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
648±4	Tamboril Santa Quitéria Complex diorite	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
646±5	Tamboril Santa Quitéria Complex diorite	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
642±15	Novo Oriente granites	zircon	Pb evaporation	Ganade de Araujo et al. (2012c)
639±3	Novo Oriente granites	zircon	Pb evaporation	Ganade de Araujo et al. (2012c)
638±3	Novo Oriente granites	zircon	LA-ICP-MS	Ganade de Araujo et al. (2010)
637±7	Tamboril-Santa Quitéria Complex diorite	zircon	ID-TIMS	Fetter (1999)
637±4	Tamboril Santa Quitéria Complex monzogranite	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
629±22*	Lagoa Caiçara Complex diatexite	zircon	ID-TIMS	Castro (2004)
628±4	Lagoa Caiçara Complex two-mica orthogneiss	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
627±4	Lagoa Caiçara Complex two-mica orthogneiss	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
625±4.6	Tamboril Santa Quitéria Complex diatexite	zircon	SHRIMP	Ganade de Araujo et al. (unpublished)
623±6	gnaissic granodiorite	zircon	ID-TIMS	Castro (2004)
619±6*	Tamboril Santa Quitéria Complex diatexite	zircon	ID-TIMS	Castro (2004)
618±5*	Tamboril Santa Quitéria Complex diatexite	zircon	SHRIMP	Ganade de Araujo et al. (2012a)
618±5*	Tamboril Santa Quitéria Complex diatexite	zircon	SHRIMP	Ganade de Araujo et al. (2012a)
618±3*	Tamboril Santa Quitéria Complex diatexite	zircon	ID-TIMS	Castro (2004)
611±3*	Tamboril Santa Quitéria Complex diatexite	zircon	ID-TIMS	Castro (2004)
597±6	Tororó diorite	zircon	SHRIMP	Archanjo et al. (2013)
595±3	Tororó gabbro norite	zircon	SHRIMP	Archanjo et al. (2013)
591±4	Tororó granite	zircon	SHRIMP	Archanjo et al. (2013)
591±10	Chaval granite	zircon	ID-TIMS	Fetter (1999)
587±5	Quixeramobim monzonite	zircon	ID-TIMS	Nogueira (2004)
585±5	Quixeramobim monzonite	zircon	ID-TIMS	Nogueira (2004)
577±5	Acarí porphyritic granite	zircon	SHRIMP	Archanjo et al. (2013)
572±4	Acarí leucogranite	zircon	SHRIMP	Archanjo et al. (2013)
571±3	Pereiro granite	zircon	ID-TIMS	Magini (2001)
563±17	Tucunduba monzonite	zircon	ID-TIMS	Fetter (1999)
560±60	Padre cosme granite	zircon	ID-TIMS	Magini (2001)
532±7	Mucambo syenite	zircon	ID-TIMS	Fetter (1999)
523±5	Meruoca monzosyenite	zircon	U-Pb SHRIMP	Archanjo et al. (2009)
522±5	Barriga granite	titanite	ID-TIMS	Fetter (1999)
495±14	Taparuaba granite	zircon	U-Pb SHRIMP	Castro et al. (2012)
467±7	Taparuaba granite	zircon	ID-TIMS	Castro et al. (2012)
460±15	Pajé granite	zircon	ID-TIMS	Teixeira (2005)
<i>Metamorphic ages</i>				
650±3	Calc-silicate rock	zircon	LA-ICP-MS	Amaral et al. (2010)
644±3	Aluminous paragneiss	Monazite	U-Pb EPMA	Castro (2004)
614±4	Mafic retro-eclogite	zircon	LA-ICP-MS	Amaral (2010)
629±10	Aluminous paragneiss	Monazite	U-Pb EPMA	Castro (2004)
626±19	Aluminous paragneiss	Monazite	U-Pb EPMA	Castro (2004)
617±1.7	Aluminous paragneiss	zircon	ID-TIMS	Castro (2004)
614±2	Aluminous paragneiss	zircon	ID-TIMS	Castro (2004)
612±5.5	Aluminous paragneiss	zircon	SHRIMP	Arthaud (2007)
612±3.4	Mafic granulite	zircon	LA-ICP-MS	Amaral et al. (2012)
607±6.7	Leucosome	zircon	ID-TIMS	Arthaud (2007)
607±1.6	Leucosome	zircon	ID-TIMS	Arthaud (2007)
604±3.8	Aluminous paragneiss	zircon	ID-TIMS	Castro (2004)
603±3	Diatexite	zircon	ID-TIMS	Castro (2004)
603±1.5	Aluminous paragneiss	zircon	ID-TIMS	Castro (2004)
601±9	Aluminous paragneiss	Monazite	U-Pb EPMA	Castro (2004)
589±10	Mafic granulite	zircon	LA-ICP-MS	Amaral et al. (2012)
586±1.6	Aluminous granitoid	zircon	ID-TIMS	Fetter (1999)
Central Borborema Province		Mineral	Technique	Reference
<i>Magmatic ages</i>				
618±5	Curral de Cima tonalite	zircon	SHRIMP	Ferreira et al. (2011)
616±5	Timbaúba pluton	zircon	SHRIMP	Guimarães et al. (2011)
616±4	Caruarú orthogneiss	zircon	LA-ICP-MS	Neves et al. (2012)
606±8	Jupi two mica gneiss	zircon	LA-ICP-MS	Neves et al. (2008)
592±7	Bom Jardim granite	zircon	ID-TIMS	Guimarães et al. (2004)
592±5	Esperança pluton	zircon	ID-TIMS	Archanjo and Fetter (2004)
591±5	Caruaru Arco Verde batolith	zircon	Pb evaporation	Neves et al. (2004)
591±5	Texeira pluton	zircon	SHRIMP	Archanjo et al. (2008)
588±12	Caruaru Arco Verde batolith	zircon	ID-TIMS	Guimarães et al. (2004)

Table 7.1 – (continued)

587±8	Cachoeirinha pluton	zircon	LA-ICP-MS	Neves et al. (2008)
587±5	Caruaru Arco Verde batolith	zircon	Pb evaporation	Neves et al. (2004)
586±21	Pajeú Complex	zircon	ID-TIMS	Van Schumus et al. (1995)
586±2	Pancas pluton	zircon	LA-ICP-MS	Neves et al. (2012)
581±3	Alagoinhas pluton	zircon	ID-TIMS	Mariano et al. (2009)
581±2	Campina Grande Complex	zircon	ID-TIMS	Guimarães et al. (2004)
577±4	Lourenço monzodiorite	zircon	SHRIMP	Ferreira et al. (2011)
576±3	Serra Redonda pluton	zircon	SHRIMP	Archanjo et al. (2008)
575±14	Serra Branca Complex	zircon	ID-TIMS	Guimarães et al. (2004)
573±4	Cabanas pluton	zircon	LA-ICP-MS	Neves et al. (2008)
570±24	Queimadas pluton	zircon	ID-TIMS	Guimarães et al. (2004)
564±5	Mamanguape pluton	zircon	LA-ICP-MS	Ferreira et al. (2011)
548±4	Sucuru dike	zircon	SHRIMP	Hollanda et al. (2010)
543±7	Pereiro pluton	zircon	LA-ICP-MS	Guimarães et al. (2004)
542±5	Uruçu gabbro	zircon	SHRIMP	Hollanda et al. (2010)
538±23	Serra do Velho Zuza pluton	zircon	LA-ICP-MS	Guimarães et al. (2004)
537±6	Monteiro dike	zircon	SHRIMP	Hollanda et al. (2010)
534±4	Sumé pluton	zircon	SHRIMP	Hollanda et al. (2010)
533±4	Santa Catarina pluton	zircon	SHRIMP	Hollanda et al. (2010)
Metamorphic ages				
626±15	leucosome of migmatitic paragneiss	zircon	LA-ICP-MS	Neves et al. (2006)
625±24	zircons from banded orthogneiss	zircon	LA-ICP-MS	Neves et al. (2006)
632±17	Alcantil orthogneiss	zircon	LA-ICP-MS	Neves et al. (2012)
623±6	zircon overgrowth in a paragneiss	zircon	LA-ICP-MS	Neves et al. (2009)
612±54	metamorphic zircons in orthogneiss	zircon	LA-ICP-MS	Neves et al. (2006)
600±22	metamorphic zircons in orthogneiss	zircon	LA-ICP-MS	Neves et al. (2006)
Southern Borborema Province				
Magmatic ages				
628±12	Pre-collision Camará tonalite	zircon	SHRIMP	Bueno et al. (2009)
625±2	Coronel Sá pre-collisional granodiorite	zircon	ID-TIMS	Long et al. (2005)
584±10*	Angico syn-collisional granite	titanite	ID-TIMS	Bueno et al. (2009)
571±9*	Pedra Furada syn-collisional granite	titanite	ID-TIMS	Bueno et al. (2009)
Metamorphic ages				
573±1	Macururé garnet-mica schist	WR-garnet	Sm-Nd isochron	Oliveira et al. (2010)
*magmatic and metamorphic ages (melting ages)				
Northern Borborema Province				
Colling ages				
641±2	Orthogneiss	biotite	Ar-Ar	Castro (2004)
611±3	Amphibolite	amphibole	Ar-Ar	Castro (2004)
601±4	Granja kinzigite	biotite	Ar-Ar	Monic et al. (1997)
601±2	Amphibolite	amphibole	Ar-Ar	Castro (2004)
599±8	Orthogneiss	biotite	Ar-Ar	Castro (2004)
599±2	Amphibolite	amphibole	Ar-Ar	Castro (2004)
598±2	Amphibolite	amphibole	Ar-Ar	Castro (2004)
597±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
595±2	Amphibolite	amphibole	Ar-Ar	Castro (2004)
595±1	Orthogneiss	biotite	Ar-Ar	Castro (2004)
594±2	Amphibolite	amphibole	Ar-Ar	Castro (2004)
594±2	Orthogneiss	biotite	Ar-Ar	Castro (2004)
592±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
588±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
584±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
583±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
582±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
582±1	Aluminous paragneiss	muscovite	Ar-Ar	Castro (2004)
578±1	Aluminous paragneiss	muscovite	Ar-Ar	Castro (2004)
576±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
574±6	Granja granulite	amphibole	Ar-Ar	Monic et al. (1997)
573±1	Aluminous paragneiss	muscovite	Ar-Ar	Castro (2004)
573±1	Aluminous paragneiss	muscovite	Ar-Ar	Castro (2004)
572±6	Mombaca granulite	amphibole	Ar-Ar	Monic et al. (1997)
568±5	Mombaca granulite	biotite	Ar-Ar	Monic et al. (1997)
563±5	Granja granulite	biotite	Ar-Ar	Monic et al. (1997)
562±1	Aluminous paragneiss	muscovite	Ar-Ar	Castro (2004)
561±3	Protomylonitic granite	biotite	Ar-Ar	Corsini et al. (1998)
557±1	Aluminous paragneiss	muscovite	Ar-Ar	Castro (2004)
556±1	Aluminous paragneiss	biotite	Ar-Ar	Castro (2004)
555±5	Pre-migmatitic tonalite	amphibole	Ar-Ar	Monic et al. (1997)

Table 7.1 – (continued)

552±7	Chaval granite	muscovite	Ar-Ar	Monie et al. (1997)
549±5	Archean basement	amphibole	Ar-Ar	Monie et al. (1997)
549±3	Mylonitic tonalite	biotite	Ar-Ar	Corsini et al. (1998)
544±3	Mafic layer	amphibole	Ar-Ar	Corsini et al. (1998)
542±3	Biotite-rich dyke	amphibole	Ar-Ar	Corsini et al. (1998)
540±3	Mylonitic granite	amphibole	Ar-Ar	Corsini et al. (1998)
536±5	Oros schist	phlogopite	Ar-Ar	Monie et al. (1997)
534±5	Paragneiss (Ceara Central)	muscovite	Ar-Ar	Monie et al. (1997)
534±3	Mafic boudin	amphibole	Ar-Ar	Corsini et al. (1998)
529±3	Mafic boudin	biotite	Ar-Ar	Corsini et al. (1998)
525±5	Paragneiss (Ceara Central)	biotite	Ar-Ar	Monie et al. (1997)
525±5	Archean basement	muscovite	Ar-Ar	Monie et al. (1997)
525±2	Pegmatite	biotite	Ar-Ar	Araujo et al. (2005)
524±5	Proterozoic cover	muscovite	Ar-Ar	Monie et al. (1997)
524±3	Metapelite	biotite	Ar-Ar	Corsini et al. (1998)
524±3	Biotite-rich dyke	biotite	Ar-Ar	Corsini et al. (1998)
521±3	Mafic layer	biotite	Ar-Ar	Corsini et al. (1998)
520±3	Muscovite-quartz mylonite	muscovite	Ar-Ar	Araujo et al. (2005)
511±3	Metapelite	biotite	Ar-Ar	Corsini et al. (1998)
510±3	Mylonitic granite	biotite	Ar-Ar	Corsini et al. (1998)
509±3	Quartzite	muscovite	Ar-Ar	Corsini et al. (1998)
506±2	Hydrothermal muscovite	muscovite	Ar-Ar	Araujo et al. (2005)
505±3	Mylonitic orthogneiss	muscovite	Ar-Ar	Corsini et al. (1998)
505±2	Mylonitic schist	muscovite	Ar-Ar	Araujo et al. (2005)
502±5	Aluminous granitoid	biotite	Ar-Ar	Corsini et al. (1998)
502±3	Granite vein	muscovite	Ar-Ar	Corsini et al. (1998)
502±3	Mylonitic orthogneiss	biotite	Ar-Ar	Corsini et al. (1998)
501±3	Serido metapelite	biotite	Ar-Ar	Corsini et al. (1998)
500±3	Equador schist	biotite	Ar-Ar	Corsini et al. (1998)
500±3	Sheared granite	muscovite	Ar-Ar	Corsini et al. (1998)
500±2	Mylonitic schist	biotite	Ar-Ar	Araujo et al. (2005)
496±3	Equador schist	biotite	Ar-Ar	Corsini et al. (1998)
491±3	Sheared granite	biotite	Ar-Ar	Corsini et al. (1998)
Central Borborema Province				
cooling ages				
509±5	Santa Cruz do Capibaribe pluton	biotite	Ar-Ar	Hollanda et al. (2010)
510±5	Coxixola mylonites	muscovite	Ar-Ar	Hollanda et al. (2010)
511±2	Coxixola mylonites	muscovite	Ar-Ar	Hollanda et al. (2010)
519±5	Alcantil orthogneiss	biotite	Ar-Ar	Hollanda et al. (2010)
520±5	Coxixola mylonites	muscovite	Ar-Ar	Hollanda et al. (2010)
521±5	Jupi orthogneiss	biotite	Ar-Ar	Hollanda et al. (2010)
529±5	Alcantil orthogneiss	amphibole	Ar-Ar	Hollanda et al. (2010)
530±2	Prata mafic stock	biotite	Ar-Ar	Neves et al. (2012)
535±5	Metagranodiorite	biotite	Ar-Ar	Neves et al. (2012)
547±2	Prata mafic stock	amphibole	Ar-Ar	Neves et al. (2012)
547±4	Coxixola mylonites	muscovite	Ar-Ar	Neves et al. (2012)
548±2	Coxixola mylonites	muscovite	Ar-Ar	Neves et al. (2012)
552±5	Cachoeirinha pluton	biotite	Ar-Ar	Neves et al. (2012)
Southern Borborema Province				
cooling ages				
544±10	Leucogranite Major Isidoro	biotite	Ar-Ar	Brito et al. (2008)
557±10	Leucosome Major Isidoro	muscovite	Ar-Ar	Brito et al. (2008)
551±10	Kinzigitic gneiss Rio Couripe	biotite	Ar-Ar	Brito et al. (2008)
579±10	Biotite gneiss Rio Couripe	biotite	Ar-Ar	Brito et al. (2008)
566±10	Leucosome Rio Couripe	biotite	Ar-Ar	Brito et al. (2008)

In this work, we use existing data on the temporal and spatial tectonothermal and magmatic evolution of the Province to propose a new integrated tectonic model for its evolution between 620 and 550 Ma resulting from two collisional events.

7.2. The West Gondwana Orogen: the 620-600 Ma Collision I

The Ceará Central Domain of the Northern sub-province is part of a large collisional belt (denominated here as the West Gondwana Orogen) that extends from Hoggar in Africa to Central Brazil (e.g., Caby, 1989, Trompette, 1994, Cordani et al., 2013a). Due to continued post-collisional convergence and escape, this collisional belt was subsequently reactivated by dextral shear zones (Caby, 1989, Castaing et al., 1994). In the Ceará Central Domain, relics of retrogressed eclogitic rocks (800°C; 17 kbar) have been dated at ca. 615 Ma (Amaral, 2010), although older ages at ca. 650 Ma from related calc-silicate rocks have also been putatively attributed to high-P conditions. These rocks are roughly aligned with other HP and UHP rocks in Africa also dated at ca. 620-610 Ma (Bernard Griffiths et al., 1991; Affaton et al., 2000; Jahn et al., 2001) defining a large Himalayan-scale collisional orogen.

Along this orogen, and pre-dating collision, there are older, 950-630 Ma magmatic and sedimentary assemblages, interpreted as remnants of intraoceanic or continental arcs (e.g., Pimentel and Fuck, 1992; Caby, 2003; Duclaux et al., 2006; Berger et al., 2011; Ganade de Araujo et al., 2012a,b), related to the Pharusian-Goias Ocean, consumed before collision (Kröner and Cordani, 2003; Cordani et al., 2013b). This view of the large Pharusian-Goias Ocean challenges the previous proposed Cambrian Clymene Ocean, where a younger suture is inferred along the Paraguay-Araguaia fold and thrusts-belts (Trindade et al., 2006; Thover et al., 2012; Cordani et al. 2013b).

The Tamboril-Santa Quitéria Complex in the Ceará Central Domain is a large igneous unit representing arc granitoids (Fetter et al., 2003) emplaced slightly before high-P metamorphism at ca. 650-630 Ma. These arc rocks, together with supracrustal rocks, were reworked and partially melted soon after the inferred collision at ca. 620-600 Ma (Arthaud, 2007; Ganade de Araujo et al., 2012a).

Throughout the Northern and Central sub-provinces a gently-dipping foliation associated with thrusting developed at the time of collision I (e.g. Caby and Arthaud, 1986; Neves et al., 2012). In the Central sub-province, where detailed studies have been carried out, this foliation is associated with peak metamorphic conditions of 640-750°C and 6-8 kbar, at ca. 620 Ma (Neves et al., 2012).

We argue that collision of the Borborema Province against the Parnaíba block, representing part of the Amazonian-West-African Craton, is marked in Ceará Central Domain by the end of arc-related magmatism at ca. 630 Ma and high-P metamorphism at ca. 620-615 Ma, and was followed by thickening of the crust with the development of the thrust-related foliation (Caby and Arthaud, 1986), and heating, reaching peak thermal conditions and widespread anatexis at ca. 620-600 Ma.

7.3. The Sergipano Orogen: the 590-570 Ma collision II

The Sergipano Orogen in the Southern sub-province is composed of supracrustal rocks related to the development of the Sergipano Basin that may have initiated before 900 Ma (Brito Neves et al., 2000; Oliveira et al., 2010). Rifting at ca. 700 Ma is well-documented in one of its domains (the Canindé Domain) and could have lasted until ca. 640 Ma (Oliveira et al., 2010). Like in the Ceará Central Domain, early convergent magmatism has arc-related signatures dated at ca. 630-625 Ma (Long et al., 2005; Bueno et al., 2009), generated during consumption of a restricted Sergipano Ocean between the São Francisco Craton and the Pernambuco-Alagoas Massif (PEAL) (Oliveira et al., 2010).

Figure 7.2 – Temporal and spatial distribution of granitoid rocks in the Borborema Province. The figure depicts a systematic younging of arc-related pre-collisional magmatism from north to south across the Province, as well as in the timing of metamorphism. Arc magmatism ends at ca. 630 Ma in the NW-section of the province but starts at ca. 630 Ma in the south (Fetter et al., 2003; Oliveira et al., 2010). Whilst the most voluminous magmatism is centred around 580 Ma across the entire Province, the timing of collisional magmatism in each region varies: 620 Ma in the NW section of the Province, along the trend of the Transbrasiliano Lineament and 590-570 Ma in the south, along the Sergipano Orogen, contemporaneous with peak magmatism in the Province. Ar-Ar cooling ages also show the same systematic decreasing age pattern from the site of collision I in the Ceará Central Domain to the site of collision II in the Sergipano Belt,

but suggest final cooling was established only at *ca.* 500 Ma, which is interpreted to indicate the end of collisional deformation. Voluminous magmatism during development of the shear zones at *ca.* 590-560 Ma has dominant lithospheric mantle affinities (Neves et al., 2000; Guimarães et al., 2004) possibly related to delamination of the orogenic crustal root after thickening promoted by collision I (Ganade de Araujo, 2011). Ar-Ar ages also indicate slow cooling rate with continuous heat supply until the Cambrian (Monié et al., 1997; Corsini et al., 1998; Hollanda et al., 2010). References for ages are listed in table 1.

Three deformation events were recognized in the Sergipano Orogen (Bueno et al., 2009; Oliveira et al., 2010): south-verging D₁ nappes and thrusts zones, which thrust the supracrustal rocks over the São Francisco Craton; reactivation of D₁ structures during D₂ transpression associated with significant vertical movements and emplacement of most granitoids; and brittle to ductile-brittle D₃ structures associated with exhumation in response to continued compression. Peak amphibolite facies metamorphism occurred during D₂ at *ca.* 570 Ma (garnet/whole-rock Sm-Nd isochron, Oliveira et al., 2010) similar to the U-Pb titanite ages of syn-D₂ leucogranites, such as the 584 ± 10 Ma Angico and the 571 ± 9 Ma Pedra Furada leucogranites (Bueno et al., 2009).

Combining the nature of deformation and timing of leucogranites, Bueno et al. (2009) interpreted that peak metamorphism resulted from collision between the São Francisco Craton and the Pernambuco-Alagoas Massif (PEAL), thrusting the supracrustal rocks onto the craton during D₂. The climax of this collision and associated granitic magma production can be reasonably bracketed to between 590–570 Ma (Bueno et al., 2009; Oliveira et al., 2010), some time after peak temperatures in collision I.

7.4. Extrusion Tectonics (*ca.* 590-570 Ma)

Figure 7.3 summarises the evolution of the two orogenies starting with collision I closing the Goiás-Pharusian Ocean, marking the beginning of the West Gondwana Orogen, followed by closure of the Sergipano Ocean and collision of the São Francisco Craton with the Borborema Province, producing collision II at *ca.* 590 Ma, marking the beginning of the Sergipano Orogen. Interaction between the two collisions, led to the extrusion of the Province between 590 and 560 Ma along the network of strike slip shear zones accompanied by intrusion of syn-kinematic granitoids. The *ca.* 30 m.y. time gap between collision I and II (*ca.* 620 and 590 Ma, respectively) is also reflected in the time gap between peak temperatures (620-600 Ma and 590-570 Ma, respectively).

Structures developed during the interaction of the two orogenies change systematically from the SE to NW across the Province (**fig. 7.4**). In the Southern sub-province, south-verging thrusting of the Sergipano Belt with a small dextral component indicates dominant N-S shortening and crustal thickening (Bueno et al. 2009; Oliveira et al., 2010).

In contrast, the Central sub-province has an older (*ca.* 620 Ma) penetrative foliation, possibly related to far-reaching stresses from collision I, overprinted by the conjugate sets of E-W dextral and NE-SW sinistral, subvertical, mylonitic belts that characterize its main deformation phase (Neves et al., 2012). Here, the 600-590 Ma time interval corresponds to transition from the contractional event to transcurrent regime, which reached slightly lower P-T conditions at 690-730 °C and 4-6 kbar (Neves et al., 2012) between 590 to 570 Ma, as constrained by ages of syn-transcurrent granitoids (e.g., Guimarães et al., 2004; Neves et al., 2008; Neves et al., 2012). This late deformational event defines a pure shear deformation with maximum shortening oriented

NW-SE and strain taken up by the conjugate pair of transcurrent shear zones with first-order dextral E-W shear zones associated to second order sinistral NE-SW shear zones (**fig. 7.4**).

The Northern sub-province also has its early low-angle contractional foliation overprinted by regional NE-SW trending dextral shear zones. Movement in these shear zones, dated by different methods bracket movement to between 590 and 570 Ma (Fetter, 1999; Archanjo et al., 2013; Souza et al., 2006), extending into low-T conditions until the Cambrian period (530-500 Ma) (Corsini et al., 1998; Araújo et al., 2005; Hollanda et al., 2010). Unlike the Central sub-province, the lack of well-developed sinistral conjugate sets of shear zones in the Northern sub-province suggests a dominant simple shear transcurrent movement at 590-570 Ma, dominated by a NNE-directed block extrusion, and characterized by an inferred maximum shortening strain axis trending approximately E-W (**fig. 7.4**).

Accordingly, coeval ca. 590-570 Ma regional structures indicate a large scale anticlockwise rotation of the maximum shortening axis from N-S in the southeast to E-W in the northwest of the Province. This is accompanied by a change from thrusting with a small dextral component in the south and southeast, to pure shear expressed by a conjugate set of transcurrent shear zones in the centre, to dextral transcurrent movement and NNE block extrusion in the northwest (**fig. 7.4**). It is important to notice that deformation was not simple rigid block rotation along faults, but consisted also of internal ductile block deformation through widespread deformation. Regional-scale rotation of strain axes is interpreted to be a result of superposition between a continued eastward push of the conjoined Parnaíba Block and the Amazonian-West African Craton, and the stresses generated by the northward push of the São Francisco Craton in the south. In this case, axes rotation reflects the relative impact of each collision: collision I dominating deformation in the northwest imposing NNE to NE escape, and collision II dominating in the southeast, causing south-verging thrusting. Deformation in the inter-collisional Central sub-province reflects interaction between the two collisions leading to its conjugate transcurrent system.

Whilst our focus has been restricted to South America, the conclusions can be expanded to Africa. The São Francisco Craton was part of the much wider São Francisco-Congo Craton, and collision II with the Borborema Province was a result of the closure of the Sergipano-Oubanguides Ocean (e.g. Van Schmus et al., 2008).

Likewise, collision I was part of a broader collisional belt, the West Gondwana Orogen, including the Dahomey and Hoggar in Africa, where magmatism, eclogitization, thrusting and later dextral reactivation follow the same overall timing as in the Ceará Central Domain (Caby, 1989; Bernard Griffiths et al., 1991; Caby, 2003). Moreover, the large scale shear zone system that resulted from the interaction between the two collisions can be correlated to large lineaments in the Benino-Nigerian Province and the E-W trending shear zones of Cameroon (Caby, 1989; Trompette, 1994; Van Schmus et al. 2008).

7.5. Transbrasiliano-Kandi Strike-Slip Belt (TKSSB): A Neoproterozoic Transform Plate Boundary?

The dextral TKSSB formed as a result of relative movement obliquity during the West Gondwana Orogen, collision I (**fig. 7.3**).

Figure 7.3 – Simplified Neoproterozoic tectonic evolution of the Borborema Province and adjoining regions. **A.** Position of main tectonic components of the region in the pre-collision I stage (ca. 800-650 Ma), based on Caby (1989), Pimentel and Fuck (1992), Brito Neves et al. (2000), Berger et al. (2011), Ganade de Araujo et al. (2012a,b), including the Cariris Velhos extensional event (Neves, 2003). The Parnaíba Block is inferred from geophysical evidence (de Castro et al., 2003) and is separated from the Amazonian-West-African Craton by the Gurupi and Araguaia volcano-sedimentary belts (Klein et al., 2005; Moura et al., 2008). Opening of the Sergipano Basin (>800 Ma) and continued rifting (ca. 700-640 Ma) separating the PEAL from the rest of the São Francisco-Congo Craton (Oliveira et al., 2010). **B.** Collision I (ca. 620-610 Ma) in the west, leading to the West Gondwana Orogen marked by HP and UHP metamorphism and anatexis of continental crust (Bernard-Griffiths et al., 1991; Agbassoumonde et al., 2001; Jahn et al., 2008; Fetter et al., 2003; Santos et al., 2009) and arc magmatism at the Sergipano Orogen due to initiation of subduction (Oliveira et al., 2010). **C.** Collision II (ca. 590-570 Ma) resulting from the closure of the Sergipano-Oubanguides Ocean and leading to thrusting of sedimentary rocks onto the craton, and development of inboard orogenic basins in the Borborema Province (e.g., Van Schmus et al., 2003), and syn-collisional magmatism (Bueno et al., 2009; Oliveira et al., 2010). Inversion of the Gurupi and Araguaia basins (Klein et al., 2005; Moura et al., 2008). **D.** Final craton indentation and northeastward extrusion stage (ca. 580-550 Ma) with development of major shear zones (Neves et al., 2012, Archanjo et al., 2013) emanating from the main Transbrasiliano-Kandi Strike-Slip Belt. White arrows: direction of mass escape. Dashed line: shore line.

Figure 7.4 – Extrusion of Borborema Province. **A.** Simple squeezing model that requires sinistral movement of the Transbrasiliano shear zone after collision II. **B.** Squeezing and anticlockwise internal block rotation due to ductile deformation of the Borborema Province allowing northeast escape and dextral movement on the Transbrasiliano shear zone. **C.** Estimated orientations of 2-D strain axes for the different domains illustrating their counter-clockwise rotation from southeast to northwest. Borborema Province scale block rotation (black thick arrow) and domain-scale rotations (solid black arrows) illustrating east and northeast mass escape. Straight red gray arrows: relative movement direction at ca. 590-570 Ma. Straight green arrows: mass escape direction at ca. 590-570 Ma.

This strike-slip belt reactivated the previous sutures (e.g. Caby, 1989; Castaing et al., 1994), and the onset of dextral movement along it started possibly as early as 615 Ma soon after collision I. In our view, it may have functioned as a transform plate boundary allowing approximation of the Borborema Province to the São Francisco Craton, and leading to closure of the Sergipano Ocean and collision II (**fig. 7.3**).

Extrusion of the Borborema Province via a simple squeezing model after indentation of the São Francisco Craton requires a change from dextral to sinistral shear movement in the TKSSB (**fig. 7.4A**). This movement switch at ca. 590-580 Ma has not been documented, and therefore we suggest instead an anticlockwise block rotation of the Borborema Province in relation to the São Francisco Craton, associated with internal ductile deformation of its tectonic domains so as to maintain dextral shear sense along the TKSSB (**fig. 7.4B**). It is

this rotation resulting from collision interference and cratonic indentation that triggered the inferred northeastward extrusion of Borborema Province.

7.6. Conclusion

We present the first attempt to integrate the tectonothermal and magmatic history of the Borborema Province within the 650-550 Ma timespan. Granitoid and migmatite ages indicate that its Neoproterozoic evolution started with an early collision associated with the closure of the large Goiás-Pharusian Ocean at 620-600 Ma and generalized crustal thickening, marked by the development of eclogites and high-T thrusting foliation defining the West Gondwana Orogen. The site of this collision was subsequently reactivated by a set of dextral shear zones, forming the Transbrasiliano-Kandi Strike Slip Belt, which acted as a transform plate boundary, allowing the closure of the restricted Sergipano Ocean and collision against São Francisco Craton at ca. 590 Ma. Interaction between the two collisions between 590 and 570 Ma and continuous cratonic indentation led to the province-wide switch to transcurrent deformation and block escape generally to the NE, associated with province-wide magmatism and regional rotation of the maximum shortening axis.

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8. Ediacaran megamountains: evidence for >2500-km-long deep continental subduction in the West Gondwana Orogen

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Abstract

The 5000-km-long, deeply eroded West Gondwana Orogen (WGO) is a major continental collision zone that exposes numerous occurrences of deeply subducted rocks (i.e. eclogites) along its strike. The position of these eclogites marks the suture zone between colliding cratons and the age of metamorphism constrains the transition from subduction-dominated tectonics to continental collision and mountain building. Here we investigate the metamorphic conditions and age of ultrahigh-pressure (UHP) eclogites from three key localities within the WGO in Mali, Togo and NE-Brazil by coupled U-Th-Pb and rare earth element zircon analyses. Protracted tectonic evolution in the WGO and synchronicity of UHP metamorphism indicate that continental subduction occurred simultaneously over at least 2500 km during the Ediacaran period (620-610 Ma). We consider this to be the first record of modern, Himalayan-scale deep-continental subduction and the consequent appearance of megamountains in the geological record. The formation of these megamountains c. 40 m.y. before the explosion of Life in the Late Ediacaran is perfectly timed to deliver by erosion the sediments (nutrients) that have been deemed necessary for Life evolution.

8.1. Introduction

In rare cases, the buoyant continental crust is subducted to ultrahigh-pressure (UHP) conditions (i.e. depth > 90 km) and exhumed back to the surface. It has been recently recognised that exhumed UHP rocks derive from subducted continental margins, where the cold and dense slab of oceanic crust pulls the attached continental crust to mantle pressures (Beltrando et al., 2010, Hacker et al., 2013, Gilotti, 2013). These UHP rocks are exhumed during the collision that immediately follows subduction of the margin – within 2-30 m.y. – at rates of cm to mm/year (e.g. Rubatto and Hermann 2001, Kylander Clarke et al., 2012). Deep subduction of felsic continental crust leads to collisional zones with deep roots that consist of material with a density lower than mantle rocks (Austrheim, 1991, Gilotti, 2013). This density contrast produces an isostatic rebound that gives rise to significant topographic relief (Fischer, 2002; Burov and Toussaint, 2007). Indeed, most UHP rocks are exposed within large mountain belts like the present Alps and the Himalayas or the now-eroded Paleozoic orogenies (Gilotti, 2013). Therefore, the finding of rocks that underwent relatively low temperatures and HP to UHP conditions is a tracer not only for paleo subduction zones (sutures), but also for large mountain edifices.

UHP crustal rocks appear relatively late in the Earth's history. The oldest coesite-bearing eclogite is c. 620 Ma old and is found in Mali (Caby, 1994; Jahn et al., 2001), which is part of the West Gondwana Orogen. It has been argued that the first appearance of such low-temperature – high-pressure rocks during the Neoproterozoic marks the onset of modern plate tectonics dominated by the subduction of very cold slabs

(Stern, 2005; Brown, 2008). During the Neoproterozoic Era and particularly the Ediacaran Period (635-542 Ma), Earth and Life underwent significant changes with the rise of the metazoans (Waggoner, 2003) and increase in atmospheric pO_2 to the presently known levels (Canfield et al., 2007). It has been argued that major rises in atmospheric pO_2 throughout the Earth history (the largest in absolute amount of oxygen added occurred at c. 600-570 Ma) are linked to massive supply of sediments from mountain belts that were produced during the amalgamation of supercontinents (Campbell and Allen, 2008). The existence of such mountain belts in the Ediacarian period was postulated on the chronology of sediments and rare metamorphic rocks (Squire et al., 2008, Campbell and Squire 2010). However, key factors for the model such as the exact timing and spatial distribution of the change from subduction- to collision-dominated tectonics, and the size of the mountain belt have not been constrained.

In this article, we investigate the timing and metamorphic conditions of UHP rocks from three key localities distributed along the West Gondwana Orogen and compile the magmatic and sedimentary record of this vast collision zone. We argue that the simultaneous age of subduction of the continental margin marks the onset for the formation of the first large mountain chain (megamountains) similar in size to the Himalayas and we discuss potential consequences for the evolution of Life on Earth.

8.2. The West Gondwana Orogen (WGO)

The Neoproterozoic West Gondwana Orogen (WGO) (**fig. 8.1**) is a linear belt that extended for more than 5000 km from nowadays northeast Africa to Central Brazil (Caby, 1989, Cordani et al., 2013). Modern views suggest that this orogen resulted from the consumption and closure of the Goiás-Pharusian Ocean (Caby, 2003; Cordani et al., 2013) that culminated in a continent-continent collision involving mainly the conjoined Amazon and West African cratons against the São Francisco-Congo and Saharan cratons (**fig. 8.1A**). The rocks involved in this orogen record long-lived, accretionary convergent tectonics since the Early Neoproterozoic with development of several intra-oceanic and continental arcs that are now preserved within a deeply eroded, fossil collisional zone (Pimentel and Fuck, 1992; Caby, 2003; Berger et al., 2011; Ganade de Araujo et al. 2012). The orogen has a protracted tectonic history (up to 400 m.y., fig. 8.1) from the inception of oceanic arc-related juvenile magmatism at ca. 850 Ma to terminal continental collision at c. 615-610 Ma, to Molasse deposition and post-collisional granites at 550-500 Ma. The orogen can be subdivided into four different sectors (**fig. 8.1C**) that all have key features that are observed in modern collision zones such as the Alps and the Himalayas. Ophiolites represent remnants of oceanic crust that once separated the continental blocks and they can be found as elongated units within the suture zone, as exemplified by the Bou Azzer ophiolite in Morocco (El Hadi et al., 2010). The dominant rock types between the cratonic blocks are passive margin deposits, early (juvenile) and late arc-related rock assemblages and syn-orogenic supracrustal sequences. The erosion of mountains that resulted from the collision is documented in molasses and foreland sedimentary deposits and the termination of orogenic activity is dated by the age of post-collisional granitoids at ca. 540 Ma (Caby, 2003; Ganade de Araujo et al., 2012).

8.3. Deep subduction in the West Gondwana Orogen

One main characteristic of the WGO is the existence of several mafic-ultramafic massifs containing relics of UHP/HP metamorphism [fig. 8.1, Agbossoumondé et al., 2001; Jahn et al., 2001; Santos et al., 2009).

Figure 8.1 – **A.** Main cratons and tectonic blocks involved in the formation of the collisional West Gondwana Orogen. **B.** Spatial distribution of HP and UHP rocks and main lithotectonic assemblages in the 5000-km-long collisional orogen. **C.** Temporal distribution of the main lithotectonic units along each sector of the orogen together with main global events. Sr-isotope values for sea water after Veizer (1989) and pO₂ after Canfield et al. (2007). Timing of pre-collision geological events are mostly based on Pimentel (1992), Caby (1989), Caby (2003), Ganade de Araujo et al. (2012).

In order to constrain the geochronological and petrological record of HP and UHP rocks across the WGO we investigated eclogitic samples from Mali (S-506), Togo (DKE-350) and NE-Brazil (DKE-107) using a coherent approach. Eclogites from Mali and Togo are interlayered with UHP/HP quartzites from passive margin deposits (Jahn et al., 2001; Agbossoumondé et al., 2001). They preserve a fresh peak paragenesis of garnet, omphacite, phengite, coesite/quartz and accessory rutile (Fig. 2). Thermobarometric calculations for the peak eclogitic conditions were performed using a set of pressure sensitive net-transfer equilibria coupled with Fe-Mg exchange thermometry for the assemblage omphacite-garnet-phengite (Ravna and Terry, 2004) combined

with Zr-in-rutile (Tomkins et al., 2007) and Ti-in-zircon (Watson et al., 2006) thermometers (see details in Supplementary **Figure 8.1** and **Tables S1** and **S2**).

Figure 8.2 – Photomicrographs (polarized light) of the investigated samples. Eclogites from Mali and Togo exhibit phase equilibria among garnet (grt), omphacite (omph), phengite (phe) and rutile (rt). Retrogressed eclogites from NE-Brazil have abundant amphibole (amph), garnet and simplectic clinopyroxene (cpx) – plagioclase (pl) resulting from the breakdown of former omphacite. In this sample rutile is often rimmed by titanite (ttn).

To calculate the peak-pressure condition we selected phases with petrographic evidence of equilibrium among omphacite with the highest jadeite content, garnet with highest grossular and pyrope contents, and phengite with the highest Si content, as proposed by Ravna and Terry (2004). Uncertainties are presented as stated by the authors.

Peak P-T condition in the Gourma region (Mali) was retrieved from an omphacite-garnet-phengite quartzite (sample S-514), which occurs in contact with the eclogites. Thermobarometric calculations using omphacite-garnet-phengite yielded a P-T range of 3.0–3.3 GPa (± 0.3 GPa) and 690–705°C ($\pm 65^\circ\text{C}$) (fig. 8.3). Metamorphic zircons in the UHP Mali eclogite have 6–26 ppm Ti, corresponding to Ti-in-zircon temperatures of 700–830 °C with a mean temperature of 720 ± 20 °C in concordance with a mean Zr-in-rutile temperature of $740 \pm 35^\circ\text{C}$. These metamorphic conditions are consistent with the rare occurrence of coesite-bearing eclogites and testify UHP conditions for this area, indicating subduction of the Gourma passive margin to depths of > 100 km (Caby, 1994).

Further south, in the Dahomey segment (Togo), UHP and HP eclogitic and granulitic rocks are exposed in the Latto region (Agbossoumondé et al., 2001). For the Togo eclogite, calculations indicate a P-T condition in the range of 2.8–3.0 GPa (± 0.3 GPa) and 620–700° ($\pm 65^\circ\text{C}$) (**fig. 8.3**). Ti-in-zircon temperatures range from 640 to 760°C with a mean temperature of 680 ± 20 °C, which is in agreement with the mean temperature of Zr-in-rutile of 720 ± 30 °C. Although no coesite relics have been found so far in Togo, calculated P-T conditions are well within the coesite stability field, suggesting UHP metamorphism.

Eclogites from NE-Brazil are found as *boudins* within high-T partially melted metasedimentary rocks (Santos et al., 2009). As a consequence, they are intensely retrogressed and composed of garnet, Ca-rich pyroxene, hornblende and symplectite of Na-augite and plagioclase. From the composition of relic phases, a minimum P-T conditions for the decompression stage of 1.7 GPa and 770 °C have been established (Santos et al., 2009), however recent finds of coesite inclusion in garnet from the same area examined here suggests UHP metamorphism (Santos et al. 2013). In the investigated sample DKE 107, zircon displays striking similarities in age and composition to the zircon retrieved from the UHP rocks. Ti-in-zircon temperatures yield a mean temperature of 700 ± 15 °C for type-1 and 690 ± 15 °C for type-2 zircons (see below). The mean Zr-in-rutile temperature of 790 ± 35 °C are slightly higher than those recorded by Ti-in-zircon, potentially recording increasing temperature during decompression from the eclogite to the granulitic stage.

In Hoggar, the ca. 620 Ma Tideridjaouime-Tileouine HP metamorphic belt (550-600°C, 1.4-1.8 GPa) represents a slab of subducted continental material exposed along the edge of the IOGU terrane, which has been interpreted as a microcontinent (Caby and Monie, 2003). In the southern LATEA terrane, the Tin Begane eclogite (790°C, 1.5 GPa) is preserved as *boudins* within the thrust zone of an allochthonous supracrustal sequence (Bougrara, 1999). The Azroun N'Fad eclogite represents the deepest unit within LATEA terrane owing to the highest pyrope content of relict garnet (up to 44%), with a decompression path from 1.5 to 1.1 GPa between 800 and 700°C (Zetounou et al., 2004).

8.4. Timing of deep continental subduction

The UHP samples investigated contain zircons that were imaged by cathodoluminescence to show internal zoning, dated by the U-Pb method with a Sensitive High Resolution Ion Microprobe (SHRIMP) and characterised for their trace element composition by Laser Ablation Inductively Coupled Plasma Mass Spectrometer (LA-ICP-MS) (**analytical methods in Appendix**).

The zircon crystals from the Mali and Togo eclogite are composed of relict cores and metamorphic rims and minor rounded homogeneous zircons. The cores show regular zoning (**fig. 8.4**) and high Th/U (0.5-0.4) and rare earth element (REE) contents. The zircon REE pattern is enriched in heavy-REE and shows a marked negative Eu-anomaly (**fig. 8.4**). All these features indicate a magmatic origin (Rubatto, 2002). The zircon cores from the Mali eclogite are severely discordant and yield spurious individual ages (Supplementary **table S3**), which suggest partial resetting of the U-Pb system. An upper discordia intercept points to a protolith age of > 1.0 Ga. For the Togo eclogite the zircon cores yield a protolith age of 703.2 ± 8.1 Ma.

The zircon rims in the Mali and Togo eclogite show a distinct internal texture, chemistry and age. They have weak or no zoning, low REE contents with flat HREE in the Mali sample and steep to flat HREE in the Togo

Figure 8.3 – Pressure-temperature diagram comparing P-T paths (dashed curves are inferred) for the HP-UHP terranes from Mali, Togo and NE-Brazil. Peak metamorphic conditions are discussed in the text and data are available in the electronic appendix. No geothermobarometric information is available for the retrograde P-T path in Mali, but petrographic (Caby, 1994) and geochronological (Jahn et al., 2001) evidence suggests rapid exhumation without passing through the granulite field. Geothermobarometric data from Lato terrane (Agbossoumondé et al., 2001) indicate a retrograde path through granulite facies (1.0-0.8 GPa and 700-750 °C) and later amphibolite facies (0.6-0.4 GPa and 500-600°C). Minimum calculated P-T condition from NE-Brazil for the retrogression (1.7 GPa and 770°C) was followed by a granulitic stage (1.4 GPa and 870°C) and then an amphibolitic stage at (0.5-0.75 GPa and 530-700°C) (Santos et al., 2009).

sample. In contrast to the cores, the zircon rims commonly have lower Th/U (mostly 0.01-0.09) and no significant Eu-anomaly. Depletion of HREE together with the lack of negative Eu-anomaly indicates that these zircons have grown in the presence of garnet and in the absence of plagioclase (Rubatto, 2002). This mineral assemblage is diagnostic of eclogite-facies conditions. For the Togo zircons, the decreasing HREE+Y content and increasing Eu/Eu* (**fig. 8.3 inset**) is consistent with zircon rim growth over increasing garnet modal abundance, as expected during prograde metamorphism. For the Mali samples 15 analyses yield a concordia age of 611.3 ± 6.1 Ma, and for the Togo samples the rims are dated at 608.7 ± 6.1 Ma (**fig. 8.4**). Since individual zircon domains that grew during prograde to peak metamorphism in the Togo sample show no age difference within the analytical resolution (c. 12 Ma) a rapid tectonic burial rate can be inferred for the Togo eclogite.

Zircons from the NE-Brazil eclogite have rare magmatic cores and two distinct metamorphic domains. The texturally older metamorphic domain is U-poor, is commonly rounded and displays no internal structure (type 1 in **fig. 8.2**). The texturally younger, U-rich domain shows weak or no zoning (type 2). Type-1 zircon is HREE enriched and middle-REE depleted when compared with the type-2 zircon, which has a flat HREE pattern with no negative Eu-anomaly, similar to the eclogite-facies zircon rims from Togo and Mali. The

Figure 8.4 – Left: Wetherill Concordia plots of the U-Pb zircon data. Ellipse colours reflect zircon growth domains as defined by zoning and REE composition. Grey ellipse represents the Concordia age. **Right:** Internal structure of the zircons revealed by cathodoluminescence and rare earth element (REE) composition (chondrite-normalised patterns) for each zircon type.

significant change in REE pattern is attributed to MREE-rich titanite breaking down to rutile and increasing garnet growth during burial to (U)HP conditions. The two chemically different zircon domains yield ages that are within error, but support their textural relationships: the inner, U-poor rims is 624.7 ± 7.0 Ma old ($n=5$, 2-sigma) while the outer rim is dated at 616.0 ± 6.1 Ma.

8.5. The West Gondwana megamountains and implications for the Ediacaran Earth

The P-T determinations for the Mali and the Togo sample provide evidence for subduction of a continental margin to UHP conditions. The metamorphic zircon domains in the HP/UHP rocks from Mali, Togo and NE-Brazil have a trace element composition indicating growth at high-pressure conditions (flat HREE pattern and no negative Eu-anomaly). In the Togo and NE-Brazil samples, zircon growth also occurred during prograde metamorphism and burial. The geochronological data (forced to 1% to account for external errors) from the Mali (611 ± 6 Ma), Togo (609 ± 6 Ma) and NE-Brazil (616 ± 6 Ma) eclogites indicates contemporaneous subduction in these localities of the WGO. Although precise geochronological control is missing in the eclogites of Hoggar, the protracted geological evolution of this sector (**fig. 8.1**) is similar to that of other sectors where timing for eclogite metamorphism is better constrained. Overall this record indicates that subduction, and the continental collision that followed, occurred simultaneously along at least 2500-km of the WGO (**fig. 8.1**).

This is a unique scenario in the Ediacaran Earth comparable in size to the modern Himalayan continental collision. While small UHP units can initially be exhumed by tectonic forces, continental collision that follows UHP subduction of continental margins give rise to significant relief (Fischer, 2002; Burov and Toussaint, 2007). Relief is the primarily regulator of erosion rate. This dynamic is best documented in present mountain belts. In the Alps, the raise of the mountains is related to the continental collision, immediately following the subduction of the European continental margin at c. 35 Ma (Rubatto and Hermann, 2001; Beltrando et al. 2010). In the Himalayas, UHP metamorphism is dated at c. 50 Ma (Leech et al. 2005) providing evidence that a high relief can be sustained for at least 50 m.y. postdating subduction of the continental margin.

The high mountain relief produced as a result of the Himalayan-scale continental subduction and collision along the WGO in the Late Ediacaran (615-610 Ma) has the appropriate time to shed sediments vital to the rise of the metazoans at ca. 600-575 Ma. We propose that a high-altitude mountain range existed in the WGO in the period 610-560 Ma and was, therefore, most susceptible to surface erosion. In the Himalayan collisional orogen, 4-5 km thick deltaic sediments extending over 3000 km into the Indian Ocean were deposited over the last 7 m.y., in the late-stage exhumation of the orogen (Yin, 2006). The sink for the detritus related to the dissection of the WGO collisional chain is preserved within basins that records the transition from a rifted passive margin to fully developed foreland basins (**fig. 8.1**), usually to the west of the orogen. The best examples are the Gourma and Taoudeni basins (Bertrand-Sarfati et al., 1995) in Algeria and Mali, the upper portion of the Volta Basin (Carney et al., 2010) in Togo and Ghana and the Paraguay Basin in Central Brazil where Ediacaran biota have been described (Warren et al., 2012). The Ediacaran sedimentary rocks in South America are also masked by younger sedimentary cover, however evidence from the Parecis basin in the

southwestern portion of the Amazon craton suggest that the extent of the Ediacaran forelands can be larger than previously thought (Vasconcelos et al., in press).

Based on the oldest known UHP rocks in Mali (Jahn et. al., 2001), it has been proposed that this time marks the onset of modern plate tectonics (Stern, 2005; Brown, 2008). Here we show that deep and cold subduction of continental crust occurred simultaneously over a distance of at least 2500 km and thus is a key characteristic of the West Gondwana Orogen. The appearance of this first megamountains in the geological record at ca. 610 Ma is followed by remarkable biological, climatic and geochemical turnovers, including the radiation of life in the Late Ediacaran to Early Cambrian periods (ca. 600-575 to 510 Ma).

A number of authors have recognized that nutrients such as P, Ca, K, Fe, Mg, Zn and Mo from continental sediments are essential to sustain complex life forms (Brasier, 1990). Ediacaran and Cambrian life radiation was also aided by increased oxygen in the atmosphere and oceans (Canfield et al., 2007). Environmental conditions suitable for life radiation was strongly influenced by high and rapidly increasing rates of sediment accumulation and subsidence associated with amalgamation of the Gondwana supercontinent (Brasier and Lindsay, 2001). The systematic increase of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the seawater during Ediacaran period reflects the addition of soluble Sr derived from erosion of old continents into oceans (Veizer, 1989). Erosion of mountains that were produced during the amalgamation of supercontinents has been linked to steep rises in atmospheric O_2 (Campbell and Allen, 2008) as it provided nutrients for algae and cyanobacteria responsible for photosynthetic production of O_2 . Increased sedimentation and burial of organic carbon and pyrite (Campbell and Squire et al. 2010) prevents their reaction with oxygen in the environment, leading to further increases in atmospheric oxygen.

Previous geochronology of zircons from sediments has suggested that the erosion of the Transgondwana supermountains that formed at the collision of East and West Gondwana (Squire et al. 2006) is responsible for the supply of nutrients necessary for life radiation. However, the timing is problematic as this poly-phased orogen was completely assembled only at ca. 520-500 Ma (Collins and Pisarevsky, 2005), and therefore post-dates the Ediacaran life radiation. These Ediacaran organisms first appeared in interglacial and post-glacial environments in deep water between 600 and 560 million years ago and later colonized environments around the world in a variety of facies, in both shallow and deep water showing little geographic endemism (Waggoner, 2003). Our rationale is also consistent with the view that the ancestries of the Cambrian explosion extend some 50-70 million years before its expression in the fossil record (Meert and Lieberman, 2008), thus placing the West Gondwana Orogen as one of the potential environmental triggers for Life sustainability. We report the precise timing and significant extent of the oldest known UHP rocks in geological history, which makes the West Gondwana Orogen one of the events that could have changed the way evolution proceeded on Earth.

8.6. Appendix: Analytical methods

Mineral chemistry. Chemical analyses of major elements in minerals (**Supplementary Table S1**) were performed on polished thin sections with a CAMECA SX100 electron microprobe of RSES, ANU, operating in the wavelength-dispersive mode. Acceleration voltage and beam current were 15 kV and 20 nA with a

focused beam diameter of 1µm for pyroxene and garnet and 5µm for micas. Counting times per element were 20 s for Na, Mg, Si, Al, K, Ca and Fe and 60 s for Ti, Cr and Mn. K and Na were always analysed first in the routine. Several analyses on each phase were performed to obtain representative compositions and core-rim analyses were done to assess zoning patterns. Standards were natural and synthetic minerals, which were all determined on the K α emission peak. In-house mineral standards were used as secondary standards.

Supplementary Figure 1 – Calculated P-T conditions for the Mali (left: sample S-514) and Togo (right: sample DKE-350) using the geothermobarometry of Ravna and Terry (2004). Temperatures of garnet-phengite exchange reaction of and Zr-in-rutile accordingly Green and Hellman (1982) and Tomkins et al. (2007), respectively.

Imaging of internal zoning. Zircons were separated from the crushed rocks using conventional heavy liquid and magnetic techniques. Grains were mounted in epoxy resin and polished down to expose the near-equatorial section. Imaging of grain sections was carried out at the RSES in Canberra and in the Geochronological Research Centre (CPGeo) at the University of São Paulo (USP). Cathodoluminescence (CL) investigation at RSES employed a JEOL-6610A scanning electron microscope (SEM) supplied with a Robinson detector for cathodoluminescence. Operating conditions for the SEM were 15 kV, 70 µA and a 20 mm working distance. In Sao Paulo CL images were acquired using a Quanta 250 FEG SEM equipped with a Centaurus Mono CL3+ detector for cathodoluminescence.

Trace elements. Trace elements of zircon and rutile (**Supplementary Tables S2 and S3**) were analysed by laser-ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at the Research School of Earth Sciences (ANU). The instrument includes an ANU ‘HelEx’ laser ablation cell built to receive a pulsed 193 nm wavelength ArF Excimer laser with 100 mJ output energy at a repetition rate of 5 Hz and coupled to an Agilent 7500s quadrupole ICPMS. The instrument was tuned for maximum sensitivity and minimum production of molecular species, maintaining ThO⁺/Th⁺ at <0.5%. The laser was operated in drilling mode

with spot sizes of 47 μm for rutile and 28 μm for zircon. Total analysis time was ~ 60 s, the first ~ 25 s of which was background acquisition prior to ablation. Synthetic glasses (NIST 612 for zircon and NIST 610 for rutile) were used for external calibration. Internal standards were SiO_2 of 32.45 weight % for zircon and TiO_2 of 98 weight % for rutile. The BCR-2G natural glass was used as a secondary standard to monitor accuracy. Data evaluation was done with the software package *Iolite v.2.5*. Chondrite values for normalization are from Sun and McDonough (1989).

SHRIMP. Zircon was analysed for U, Th and Pb in the epoxy mount using the SHRIMP-II at the ANU in Canberra and the SHRIMP-IIe at USP in São Paulo. For zircon, instrumental conditions and data acquisition were generally as described previously (Williams, 1998). The data were collected in sets of six scans throughout the masses and a reference zircon (TEMORA 2) was analysed each fourth analysis. The analyses (Supplementary Table S4) were corrected for common Pb on the basis of the measured $^{207}\text{Pb}/^{206}\text{Pb}$ ratios as described previously (Williams, 1998). The common Pb composition was assumed to be that predicted by the model in (Stacey and Kramer, 1975). U-Pb data were collected over five analytical sessions using the same standard, with the different sessions having calibration errors between 1.23% and 2.16% (2 sigma), which was propagated to single analyses. Data evaluation and age calculation were done using the software Squid and Isoplot/Ex (Ludwig, 2003), respectively. Average $^{206}\text{Pb}/^{238}\text{U}$ ages are quoted at the 2-sigma confidence level and forced to at least 1% to account for external errors.

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9. The significance of the Transbrasiliano- Kandi tectonic corridor for the amalgamation of West Gondwana

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Abstract

The assembly of West Gondwana was completed by the end of the Precambrian, when the Amazonian, West African, São Francisco-Congo, Kalahari and Rio de la Plata cratons, as well as the Saharan metacraton and the Parnaíba, Parapanema and Luiz Alves cratonic fragments were united by means of the Brasiliano-Pan African orogeny, a geotectonic process that was active from the late Neoproterozoic to the early Paleozoic, related to the closure of a large oceanic domain, the Goiás- Pharusian Ocean. Several accretionary complexes and possible microcontinents were trapped within the Brasiliano-Pan African mobile belts, and they have been accommodated within a few hundred kilometers of the Transbrasiliano-Kandi tectonic corridor. The supercontinent was already formed at about 600 Ma, as indicated by the existence of a very large Ediacaran epicontinental sea covering large areas of west-central Brazil and southern Uruguay along the margins of the Amazonian and Rio de la Plata cratons, demonstrating the connection of both cratonic units at that time and making the idea of a collisional suture closing a supposed Clymene Ocean unsustainable. In the Cambrian, a major plate reorganization occurred, being responsible for the initiation of subduction of the oceanic lithosphere along an open and unconfined Pacific Ocean. The resulting Pampean Orogeny correlates nicely in time with the Saldania, Ross, and Tasmanian belts along the southern Gondwana margin. Simultaneously, extensional-type post-tectonic episodes occurred repeatedly along the Transbrasiliano-Kandi tectonic corridor.

9.1. Introduction

Almost all studies on the formation of Gondwana suggest that the supercontinent was formed by the amalgamation of a few building blocks of different sizes, in a series of continental collisions. Most of these blocks originated from the breakup of Rodinia, covering the entire timeframe of the Neoproterozoic (Li et al., 2008).

Several large-scale models were put forward for the assembly of Gondwana. The simplest one describes the final amalgamation of two large continental masses, West Gondwana (made of South America and Africa) and East Gondwana (made of Antarctica, Australia, India and Madagascar), forming the Mozambique belt (Kröner, 1980; McWilliams, 1981; Shackleton, 1996). With the progress of geological knowledge in recent years, especially in the fields of paleomagnetism and geochronology, the mechanisms of Gondwana assembly are now more precisely constrained, especially concerning the timing of the successive collisions between continental building blocks. Different models for the assembly of Gondwana were suggested, such as those by Meert (2003), Cordani et al. (2003), Yoshida et al. (2003), Collins and Pisarevski (2005), Trindade et al. (2006), among others.

The schematic map of **figure 9.1** shows the nomenclature used in this article for supracontinental building blocks: (1) the Amazon-West African block is formed by the Amazonian and West African cratons, as well as the small São Luis craton and a possible microcontinent covered by the Phanerozoic Parnaíba basin; (2) the Central African block includes Congo-São Francisco, Rio de la Plata and Kalahari cratons, plus the Paranapanema block concealed beneath the Paraná basin and most of northern Africa, named the Saharan metacraton; (3) the Indo- Arabian block includes the Indian shield, Madagascar, Sri Lanka, and the eastern basement of the Arabian-Nubian shield; and (4) the Australian-Antarctic block includes East Antarctica and Australia, excluding the Tasman orogen. Sizes and relative positions are only indicative, and **figure 9.1** should not be considered as a palinspastic reconstruction.

The main objective of this paper was to review the process of amalgamation of West Gondwana due to the convergence of the Amazon-West African and the Central African blocks, related to the closure of a large oceanic domain, the Goiás-Pharusian Ocean. From our point of view, this was responsible for the Brasiliano-Pan African orogeny, a geotectonic process that was active in the late Neoproterozoic. We will also review the alternative scenario proposed by Trindade et al. (2006), as well as Tohver et al. (2012), arguing that the final assembly of Gondwana would have occurred in the Cambrian, as a result of the closure of a different ocean, called Clymene. This topic, which is relevant to the age of amalgamation of West Gondwana, will be briefly discussed in the appropriate chapter.

Figure 9.1 – Crustal building blocks for the amalgamation of Gondwana, after the closing of the Goiás-Pharusian and Mozambique oceans. Location of the Iapetus Ocean between SW Gondwana, Laurentia and Baltica, and location of the Proto-Pacific Ocean before the onset of the subduction of the Pacific Plate.

9.2. Geotectonic Setting of West Gondwana

The Amazon-West African block and the Central African block (**fig. 9.1**) are formed by several juxtaposed continental nuclei, which are residual fragments from the disruption of Rodinia, in a process that started around 900 Ma, and finished only at ca. 570 Ma, when Laurentia was finally separated from the Amazon-West African block, with the formation of the Iapetus Ocean (Meert 2003, Li et al., 2008, among others). However, the disruption of Rodinia is not free of controversy. For example, based on a great deal of geochronological and tectonic evidence, Cordani et al. (2003) and Kroener & Cordani (2003) suggested that the Central African block may have never been part of Rodinia. Tohver et al. (2006) reached a similar conclusion based on their review regarding the available paleomagnetic data from Africa and South America. The position of the Goiás-Pharusian Ocean (Kröner & Cordani 2003), where intraoceanic island arcs were formed at about 850 – 900 Ma, is indicated in **figure 9.2**, adapted from Cordani et al. (2013). In many paleomagnetic reconstructions, such as those by Meert (2003) or Cordani et al. (2003), this ocean was named “Brasiliano”, or sometimes “Adamastor”. A more restricted Adamastor Ocean (Hartnady et al., 1985), located along the western- southwestern boundary of the Central African block, is illustrated in **figure 9.2**. Its formation and disappearance are related to initial rifting, followed by dispersion, and later the reassembly of two important cratonic nuclei, Kalahari and Rio de la Plata, and some smaller cratonic fragments, such as Paranapanema and Luís Alves, against the larger Congo-São Francisco craton. Although data from the early arc assemblages in the Adamastor Ocean yielded juvenile signatures for ca. 800 Ma granitoids (Tupinambá et al., 2012), its extension seems to be much more restricted than the ones of Goiás-Pharusian Ocean due to the confined nature of the Araçuaí Orogen. As a consequence of the subduction of oceanic lithosphere related to the closure of the Goiás-Pharusian Ocean, several accretionary complexes and possible microcontinents were trapped within mobile belts formed during the Neoproterozoic collisional events. The tectonic process was extremely complex, leading to the formation of several sutures. The mobile belts were the result of the Brasiliano-Pan African orogeny, and are now exposed in very large areas of West Africa and South America. They may be classified into two types of orogenic units, showing different ages, tectonic environments and evolution:

1. An older component (dated at 950 – 650 Ma) made of magmatic and sedimentary assemblages, many of which have mantle-derived intraoceanic features, constituting accretionary-type orogenic belts. They essentially comprise plutonic-volcanic magmatic associations, which are exposed at upper-middle crustal levels, such as the Iskel, Tilemsi, Amalaoulaou, Kabyé and Goiás magmatic arcs, trapped between the Amazon- West African and the Central African blocks (Dostal et al., 1994, Caby 2003, Laux et al., 2005, among others). The tectonic evolution is coeval with that of the Arabian-Nubian Shield, whose intraoceanic magmatic arcs are exposed between the Central African and Indo-Arabian blocks;
2. A younger component (dated at ca. 700 – 520 Ma) formed by the collage of orogenic belts located along the cratonic margins, comprising reworked basement plus collisional fold-and-thrust and metamorphic belts. They consist of metasedimentary and metavolcanic rocks, which were intruded by large amounts of granitoid rocks, exposed at deep to shallow crustal level, and were tectonically affected by a protracted Brasiliano- Pan African orogeny. These tectonic units may include oceanic assemblages, such as ophiolites, accretionary prisms and

island-arc magmatic suites, and in some places HP, UHP and UHT metamorphic assemblages. Examples are the Trans-Saharan, Dahomeyan, Brasília and Paraguay belts (**fig. 9.3**).

In addition, the Gondwana Supercontinent was subjected to widespread Ediacaran-Cambrian tectono-thermal reactivation in almost all regions that were previously affected by the Neoproterozoic accretionary, collisional or intracontinental orogenies, be it within the mobile belts of that age or at the marginal parts of cratonic areas. This tectono- thermal overprint is also detected over very large areas, such as the Saharan metacraton in Africa and its counterpart in South America, within the Borborema Province.

Figure 9.2 – Major tectonic elements related to West Gondwana at about 800 – 900 Ma ago, prior to the final amalgamation. Cratons: AM = Amazonian; CO = Congo; KA = Kalahari; LAU – Laurentia; RP = Rio de La Plata; SF = São Francisco; SM = Sahara metacraton; WA = West African. Smaller cratonic fragments: BO = Borborema; GO = Goiás Central Massif; LA = Luiz Alves; PA = Paranapanema; PB = Parnaíba; PP = Pampia. Intra-oceanic magmatic arcs: A = Amalaoulou; G = Goiás; I = Iskel; K = Kabyé; T = Tilemsi. Adapted from Cordani *et al.* (2013).

9.3. Closure of the Goiás-Pharusian Ocean

The Goiás-Pharusian Ocean occupies a very large area and includes many intraoceanic magmatic arcs, whose tectonic evolution started as early as *ca.* 900 Ma. Approximately 300 Ma later, this ocean closed due to successive continental collisions, which sutured the West African Craton against the Saharan metacraton (Abdelsalam *et al.*, 2002) in the north, and the Amazonian against the São Francisco Craton in the south.

The several Brasiliano-Pan African orogenic belts, which were created in this process, are aligned along a very long corridor in South America and Africa that is dominated by a megashear zone, which is one of the major

tectonic elements in the world. Schobbenhaus (1975) coined the name “Transbrasiliano lineament” in his compilation of the tectonic map of Brazil, showing that this structure crosses a large part of the continent, from northeast Brazil down to Paraguay and Argentina. Caby (1989), Trompette (1994), Fairhead and Maus (2003), Santos *et al.* (2008), among many others, have shown that it extends to Africa, where it crosses the western part of the continent, from Togo to Algeria, along the Hoggar 4050'-Kandi shear system (**fig. 9.3**). The megashear is formed by a series of ductile shear zones, which occur in very large areas. It probably reaches the bottom of the lithosphere, and the shear zone motion must have started shortly after the closure of the Goiás-Pharusian Ocean, taking advantage of the several weak lithospheric zones formed during continental collisions. All collisional sutures related to the Brasiliano-Pan African belts are accommodated within a few hundred kilometers of the lineament, in a region that will be referred to as the Transbrasiliano-Kandi tectonic corridor.

The coherence of the lineament is clearly marked by the strong linear magnetic anomalies obtained from the CHAMP satellite survey and reported by Fairhead & Maus (2003). Within the Trans-Saharan and Dahomey belts of West Africa, a string of positive gravimetric anomalies, locally associated with linear magnetic anomalies, is observed near the margin of the West African craton, associated with a series of mafic and ultramafic massifs. Their tectonic significance may be attributed to the rise of mantle diapirs, which indicate the position of Neoproterozoic suture zones (Trompette 1994). In South America, the lineament is clearly visible in the aeromagnetic mosaic of central and northeast Brazil, forming a series of low amplitude magnetic anomalies, which can be traced across the country from NE to SW. Elongated gravimetric and aeromagnetic anomalies along the main trend of the lineament have also been observed in Brazil, such as a strong anomaly observed within the Parnaíba basin, associated with the main depocenter of the Paleozoic sedimentary sequences (Nunes 1993).

9.4. The Borborema Province and the Trans-Saharan belt

The Saharan metacraton, named by Abdelsalam *et al.* (2002), is not well defined. It is characterized as a large portion of cratonized continental crust of the pre-Neoproterozoic age dominated by medium to high-grade gneissic and migmatitic terrains, which were highly remobilized during the Pan-African orogeny. These authors interpreted the evolution of this tectonic unit as an initially coherent cratonic mass that was subjected to a widespread extensional tectonic regime, which caused, possibly during the early Neoproterozoic, pervasive rifting and the formation of narrow oceanic basins. These basins closed during the late Neoproterozoic, forming a collage of continental blocks. The aforementioned tectonic evolution seems to be similar to the one described for the basement of the central part of the Borborema Province of Brazil, where the Archean to Paleoproterozoic sialic basement (Brito Neves *et al.*, 2000, Van Schmus *et al.*, 2008, among others) underwent widespread Neoproterozoic rejuvenation and pervasive granite magmatism.

In **figure 9.4**, the dividing line between the Trans-Saharan belt and the Saharan metacraton established by Abdelsalam *et al.* (2002) is the Raghane shear zone, located in the eastern part of the Tuareg Shield (Liégeois *et al.*, 1994; 2000), where the Barghot and Aouzegueur terranes have been thrust from west to east across a rigid cratonic block.

Figure 9.3 – Outline of the Transbrasiliano-Kandi mega-shear zone in a late Paleozoic pre-drift reconstruction of South America and Africa, with the relative position of cratons, cratonic fragments and late-Proterozoic- Cambrian mobile belts. Adapted from Cordani et al. (2013).

The Raghane shear zone extends to the south, cutting through the Air Massif and disappearing beneath undeformed Phanerozoic rocks. Although we acknowledge the need for better control an additional evidence, we propose that the dividing line between the Dahomeyan belt and the Saharan metacraton, as a continuation of the Raghane shear zone, could be represented by the important lineament at the eastern limit of the “Nigerian schist belt”, which marks the boundary between the western and eastern Nigeria terranes, as depicted by Arthaud et al. (2008). Also in **figure 9.4**, along the eastern side of the West African craton, there are two typical marginal sequences deposited at the boundary of the cratonic region and facing an eastern ocean. These are the Gourma and Volta basins, in which there are several kilometres thick sedimentary sequences accumulated. The deposits are mainly flyschoid material made up of siltstones, shales and greywackes, with some intercalation of carbonatic rocks with stromatolites, which indicate the late Neoproterozoic age (Trompette 1994).

Early Neoproterozoic (*ca.* 870 – 700 Ma) oceanic terranes have been identified in many parts of the Trans-Saharan belt (**fig. 9.4**), from the Hoggar to the Dahomeyan segments (Caby 1989, 2003; Berger et al., 2011).

In Hoggar, in the Silet region (Algeria), diorite-tonalite and monzogranite plutons from the Iskel island arc yielded U-Pb zircon ages of *ca.* 868 and 839 Ma. The occurrence of slices of pre Pan-African basement directly overlain by shelf sediments and capped by volcanic arc rocks in several localities suggests that the Iskel magmatic arc was built on attenuated continental crust, adjacent to possible slices of oceanic lithosphere (Lapierre *et al.*, 1986, Caby 2003). Further to the south, in the Gourma region (Mali), the Tilemsi-Amalaoulaou intraoceanic arc assemblages (Dostal *et al.* 1994) were dated at *ca.* 790 – 710 Ma, and the Tilemsi arc is considered as the upper crust superstructure, equivalent to the Amalaoulaou complex (Berger *et al.* 2011). Although there are not enough precise geochronological data for the Kabyé massif in the Dahomeyan belt (Togo), geochemical and field characteristics suggest that this massif may be the continuation of the Iskel-Tilemsi-Amalaoulaou intraoceanic arc system (Duclaux *et al.*, 2006). The late Neoproterozoic Andean-type active continental margin, which produced extensive arc plutonism, is located east of the oceanic terranes. This stage of ocean-continent subduction was dated at 696 ± 5 Ma in the Kindal terrane and 716 ± 6 Ma in the Adrar des Iforas region, in Mali (Bruguier *et al.*, 2008), indicating that it was partially coeval with the ocean-ocean subduction active further west.

To the south, the Dahomey belt is characterized by a complex thrust stack and suture, representing the convergence and subsequent collision between the Benino-Nigerian province, part of the Saharan metacraton, and the West-African craton. This belt comprises a series of metasiliciclastic rocks (quartzites and schists) from the Atakora and Kante units, but it also contains high-grade metamorphic rocks (up to eclogite facies) with mafic and ultramafic protoliths (Agbossoumondé *et al.*, 2001). In Benin, possible arc-type granitoids related to the consumption of the Pharusian Ocean were dated at *ca.* 660 – 650 Ma (Kalsbeek *et al.*, 2013).

The subsequent closure of the Pharusian oceanic domain, by means of a continent-continent collision (Himalayan-type orogen), is constrained by the presence of UHP and HP rocks in the Trans-Saharan orogenic belt (**fig. 9.4**). In the Gourma region, coesite-bearing eclogites (up to 25 kbar) and blueschists have been identified (Caby 1994, Jahn *et al.*, 2001). Geochronological studies on these rocks indicated the age of eclogitization at *ca.* 620 Ma. Eclogites (*ca.* 19 kbar) and HP granulites have also been described in Togo (Attoh 1998, Agbossoumondé *et al.* 2001). Geochronological ages of these rocks are scarce. However, a single Pb-Pb zircon age of 612 ± 1 Ma obtained from a HP granulite (Affaton *et al.* 2000) suggests that collision was already going on at that time in this sector of the Trans-Saharan orogenic belt.

As shown in **figure 9.3**, the central part of the Borborema tectonic province in Brazil is very probably the counterpart of the Saharan metacraton of northern Africa. The north-western part of this province has been correlated with the Trans-Saharan domains of West Africa for many years (Torquato & Cordani 1981, among many others). In particular, the region close to the Transbrasiliano lineament in northeast Brazil, represented by the Médio Coreáú and Ceará Central domains (**figs. 9.4 and 9.5**), is considered to be the counterpart of the Dahomeyan rocks in Togo and Benin along the area of the Kandi lineament (Caby 1989, Arthaud *et al.*, 2008). In that region, evidence of pre-collisional magmatic assemblages (intraoceanic and Andean-type settings) is still scarce. However, an important period of crustal growth at 900-700 Ma can be inferred from the record of detrital zircon grains from the regional supracrustal rocks (Ganade de Araújo *et al.*, 2012); the presence of abundant grains spanning the entire interval of about 200 Ma allows us to infer the presence of a

long-lived active continental margin, where subduction-related magmatism had been ongoing since the beginning of the Neoproterozoic.

In the Ceará Central domain, east of the Transbrasiliano lineament (TB, from now on), the Tamboril-Santa Quitéria Complex, a large area formed by different types of 640 – 610 Ma old granitoid rocks and migmatites, was described as a continental magmatic arc by Fetter et al. (2003). Van Schmus et al. (2011) indicate the age of collision at *ca.* 610 – 590 Ma. However, Amaral et al. (2010) reported ages of *ca.* 650 – 630 Ma for some high-pressure metamorphic rocks also located east of the lineament, in the Forquilha Eclogite Zone (**fig. 9.5**). These ages have been interpreted as being related to the eclogite formation, and suggest that the continental collision may have taken place earlier. More recently, however, Amaral et al. (2012) reported ages between *ca.* 613 and 590 Ma for the metamorphism of granulite facies in mafic granulites of the nearby Cariré area, and therefore the precise age of continental collision remains controversial. After isostatic uplift, cooling, and denudation, the mobility and tectonic activity along the megashear continued for a long time.

Considering the continuity of major faults, the similarity of regional lithostratigraphic trends and the westward polarity of structural features, the correlation between the Trans-Saharan belt and the northwestern part of the Borborema Province is highly probable. However, as pointed out by Santos et al. (2008), the eclogites and related rocks of the Hoggar, as well as the HP metamafic rocks of the Dahomeyides, were located west of the Kandi-Hoggar 4050' lineament, but UHP or HP metamorphic rocks were not identified in the Médio Coreaú domain. On the other hand, the known high-grade metamorphic rocks occurring in the Ceará Central domain are located to the east of TB.

9.5. The Brasília Belt, the Goiás Magmatic Arc and the Paraguay Belt

The Brasília Belt, in central Brazil, presents unequivocal evidence indicating the closure of the long-lived (900-630 Ma) Goiás Ocean at *ca.* 630 Ma. The belt is one of the largest and better preserved Neoproterozoic orogenic belts in Brazil (Pimentel et al., 2000), comprising: (i) a thick Neoproterozoic metasedimentary pile, including the Paranoá, Canastra, Araxá, Ibiá, Vazante, and Bambuí groups, mostly overlying Paleoproterozoic and occasionally Archean basement at the western margin of the São Francisco Craton; (ii) the Goiás Massif, a microcontinent (or allochthonous sialic terrain) composed of the Archean Crixás-Goiás granite-greenstones and associated Proterozoic formations; and (iii) the large juvenile Neoproterozoic Goiás Magmatic Arc in the west (**fig. 9.5**).

The several low to medium-grade supracrustal rock units of the Neoproterozoic metasedimentary pile show tectonic vergence to the east, towards the cratonic area (Dardenne, 2000) and metamorphic grade increases westward. Recent zircon provenance data suggest that some of these units (e.g., the Ibiá, Araxá and part of the Serra da Mesa groups) were deposited and deformed within a short interval between *ca.* 650 and 630 Ma (Pimentel et al., 2011). Moreover, a Neoproterozoic ophiolitic mélange has been identified in the Araxá Group as a representative of the oceanic crust (Strieder and Nilson 1992).

Figure 9.4 – Geological correlations between northeastern South America and north-western Africa, in a late Paleozoic pre-drift reconstruction.

In the western part of the Brasília belt, a large area formed of Neoproterozoic juvenile crust records the closure of a large oceanic domain between the Amazonian and São Francisco paleocontinents from at least *ca.* 900 to 600 Ma (Pimentel & Fuck 1992). This is known as the Goiás Magmatic Arc, which represents one of the most important tectonic components of the Brasília Belt (**fig. 9.5**). It is divided into the Arenópolis arc to the south, and the Mara Rosa arc to the north, comprising: (i) juvenile island-arc (*ca.* 900-800 Ma; e.g., Mara Rosa and Arenópolis sequences), volcanic-sedimentary sequences associated with mantle-derived tonalite-granodiorite-granite orthogneisses; and (ii) younger (*ca.* 650-630 Ma) continental arc-type volcano-sedimentary sequences intruded by a series of tonalite-granodiorite plutonic complexes (Junges et al., 2002, and references therein). The juvenile signature of the older metavolcanic and metaplutonic rocks with tholeiitic to calc-alkaline signature is demonstrated by their low initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios, positive ϵNd values and Sm-Nd TDM model ages mostly between 0.8 and 1.1 Ga (Laux et al., 2005, and references therein). Tonalite and granodiorite of the younger arc association have older Nd TDM model ages and slightly negative ϵNd values, suggesting a continental arc setting for this magmatic event. For the amalgamation of Gondwana, the intraoceanic magmatic arcs were fused together by a series of soft collisions, and granitoid magmatism persisted during the tectono-magmatic episodes of the Brasiliano orogeny (Pimentel et al., 2000). The predominant calc-alkaline composition of these magmatic rocks indicates the action of continued subduction-related active margin processes.

It is not an easy task to identify the main sutures related to the closure of the Goiás Ocean, and the available geophysical evidence is just starting to reveal some important discontinuities. Deep crustal and lithospheric studies using seismic tomography (Assumpção et al., 2004, Feng et al., 2007), as well as deep seismic refraction and teleseismic receiver function investigations (Soares et al., 2006, Ventura et al., 2011) were conducted over the surface outcrops of the Goiás Magmatic Arc, where a large positive Bouguer anomaly occurs along the TB. The hot and dense lithospheric mantle underlying the magmatic arc, where the crust is only 36-38 km thick, compensates this Bouguer anomaly. To the east, crustal thickness increases to up to 43 km below the marginal Brasília belt and the western part of the São Francisco craton. Westwards, seismological data show an abrupt 16 km step in the Moho discontinuity in the passage from the Goiás arc to the Araguaia belt and the Amazonian craton. Such structure is considered as the result of the duplication of the lower mafic crust of the Amazonian paleoplate during late Neoproterozoic subduction below the Goiás magmatic arc (Ventura et al., 2011). In this context, the Serra Azul Archean metamorphic rocks are interpreted as an obducted sliver of the Amazonian craton basement (Soares and Fuck 2011). The geochronological evidence that is available so far indicates that the main subduction event ended at *ca.* 630-600 Ma, and that the main regional metamorphic peak occurred at *ca.* 650-630 Ma, as recorded by the granulites of the Anápolis-Itaçu Complex indicated in **figure 9.5** (Della Giustina et al., 2009), as well as by several other rock units of the Brasília Belt (Baldwin and Brown 2008, and references therein). In this granulite complex, UHT sapphirine-bearing rocks yielded U-Pb metamorphic ages of *ca.* 650 Ma, which are roughly coeval with the emplacement of mafic-ultramafic complexes, thus representing the metamorphic core of the Brasília orogen (Piazana et al., 2003).

The Paraguay Belt (location in **figure 9.3**) is a typical fold-and-thrust belt located along the southeastern margin of the Amazonian Craton (Alvarenga et al., 2000, Campanha et al., 2011). It is affected by tectonic deformation, which is almost imperceptible at the border of the craton, but increases up to tight isoclinal folds towards its inner areas. It comprises the older Cuiabá Group, including glacial sediments of the Puga

Formation, and the younger Corumbá Group, comprising carbonates and pelites with Ediacaran-age fossils (Boggiani et al., 2010). The Cuiabá Group was affected by low-grade metamorphism, up to the biotite zone of the greenschist facies. Within the Cuiabá Group, some “cap carbonates” of the Araras Group, directly overlying diamictites of the Puga Formation, were dated at *ca.* 630 Ma (Babinski et al., 2006). In addition, in Planalto da Serra, Mato Grosso, DeMin *et al.* (2013) reported $^{40}\text{Ar}/^{39}\text{Ar}$ dating on phlogopite, as well as additional Rb-Sr and Sm-Nd ages, for high-K ultramafic rocks, plugs and dykes affecting an area about 30 km long, intrusive into the low-grade metasedimentary rocks of the Cuiabá Group. Their ages, close to 600 Ma, represent the minimum age for the deposition and deformation of this unit. Later, during Cambrian, the Cuiabá Group was thrust over the Corumbá Group in a thin skin deformation, basically westerly directed. Moreover, rocks of the upper part of the Corumbá Group, located farther to the west, were virtually undeformed and unconformably deposited on the sialic basement of the Amazonian Craton. A final regional deformational phase of very low intensity, extensional in character and related to a few intrusions of granite bodies, took place during the Cambrian, or even later.

The sedimentary environment of the Corumbá Group is generally attributed to a restricted marine shelf within an epicontinental sea overlying the southeastern margin of the Amazonian Craton, and the resulting deposits correlate with the sedimentary filling of the nearby Tucavaca aulacogen, in Bolivia. Some other sedimentary sequences that are similar in age and tectonic setting to the Corumbá Group have been recently attributed to the Ediacaran and seem to represent very extensive marine transgressions, in a general context of epicontinental seas. For example, recent geochronological data by Pimentel et al. (2011) for the Bambuí Group in the western part of the São Francisco Craton suggest Ediacaran (*ca.* 600 Ma or younger) depositional ages for this foreland sequence. Moreover, Gaucher et al. (2003, 2008, 2009) and Poiré and Gaucher (2010) demonstrated the existence of a very important close correlation of the Corumbá Group with the Arroyo del Soldado Group, in Uruguay, which practically have the same stratigraphy and the same fossiliferous content of the Ediacaran age. They would therefore belong to the same continental shelf, along the margins of the Amazonian and Rio de la Plata cratons, and this reasoning is a powerful paleogeographic indicator for a connection of these two tectonic units in the Ediacaran.

The Amazonian-Rio de la Plata link in the Ediacaran is the main argument to deny the existence of oceanic lithosphere in central South America, as proposed by Trindade et al. (2006) and Tohver et al. (2012). In the latter, the suture resulting from the collision between the

Amazonian and São Francisco-Congo cratons and the closure of a supposed Clymene Ocean is crossing the entire South American continent. These authors reviewed the tectonic history of the Pampean, Paraguay, and Araguaia belts along the margins of the Amazonian and Rio de la Plata cratons, and tried to demonstrate that these belts were tectonically active from the late Ediacaran to the late Cambrian, as the final stages of Gondwana formation.

In addition to the already mentioned close correlation between the Corumbá and the Arroyo del Soldado groups, located along the same Ediacaran continental shelf, which precludes the existence of a wide ocean, a few other arguments against the idea of a Clymene Ocean in central South America were presented and discussed with the pertinent details in Cordani et al. (2013). Some of them are briefly summarized here:

1. The most important evidence for the hypothesis of the Cambrian Clymene Ocean, the Puga paleopole, is located at low latitude, not far from the present pole, and therefore could be related to a younger remagnetization.
2. The assembly of West Gondwana was completed by *ca.* 600 Ma. After this, there is no geological evidence of an oceanic lithosphere (e.g., ophiolites, magmatic arcs etc.) in central South America.
3. The tentative correlation between the Pampean and Paraguay belts cannot be accepted, because their tectonic significance is totally different. There is no similarity in lithology, metamorphism, or structural trends.
4. The Araguaia Belt started as a Neoproterozoic in- traplate aulacogenic-type basin, formed over an ancient sialic basement, which may have extended into a premature oceanic stage with limited width, with the possibility of once having been connected to the main Goiás-Pharusian Ocean.

9.6. Ediacaran/Cambrian Tectonic Evolution in Southern West Gondwana

The Goiás-Pharusian Ocean closed at the end of the Neoproterozoic, and from then on, West Gondwana became a single continental mass (Cordani et al., 2013). After the probable uplift following the Brasiliano-Pan African orogeny, orogenic collapse and extension took place not only within the Transbrasiliano tectonic corridor, but also in adjacent areas. Simultaneously, a major spreading center was developing between West Gondwana and Laurentia, which led to plate reorganization, responsible for the initiation of convergence along the Pacific margin of Gondwana.

Cawood (2005) suggested that the subduction of the Pacific oceanic lithosphere occurred at the Gondwana margin at *ca.* 570 Ma. The name “Terra Australis Orogen” was proposed for a very large tectonic province located along the southern margin of Gondwana, forming an open and unconfined Pacific Ocean and comprising several accretionary orogens. The Pampean orogeny (Ramos 1988, Rapela et al., 1998) is the South American representative in Terra Australis. According to Ramos (1988), the Eastern Pampean ranges, in which high-grade metamorphic rocks are recognized, were formed as a result of normal subduction of oceanic lithosphere, followed by continent-continent collision between the Rio de la Plata Craton and the Pampia microcontinent. Geochronological data indicate a Cambrian age for the entire tectonic development of the Pampean orogen, which correlates in time with the Saldania, Ross, and Tasmanian companion belts of Terra Australis (Cawood 2005, Schwartz et al., 2008).

Escayola et al., (2011), dealing with the study of the Pampean orogen, summarized the available lithological, stratigraphic, and structural knowledge of the Puncoviscana formation in northern Argentina and presented conclusive evidence for its syntectonic character as an accretionary complex. They showed that the orogenic process had already started in the Ediacaran, around 560 Ma, being tectonically active during most of the Cambrian, until at least 520 Ma.

Figure 9.5 – Main tectonic elements within the Borborema and Tocantins provinces and the Parnaíba Basin, in South America, in the vicinity of the Transbrasiliano Lineament.

As already mentioned, this period was marked by the subduction of the Pacific oceanic lithosphere, which produced tectonic compression and regional metamorphism, associated with voluminous granite magmatism of orogenic and subduction-related nature (Ramos 1988, Rapela et al., 1998). Given the time involved, 560-520 Ma, the Pampean orogenic system is much younger than the Neoproterozoic collisional belts responsible for closing the Goiás-Pharusian Ocean.

While subduction was going on at the southern part of Gondwana, extensional post-tectonic episodes occurred along the Transbrasiliano-Kandi megashear, and the overall extensional tectonic scenario in central Brazil clearly contrasts with that of the coeval compressional scenario of the Pampean orogen, in Argentina. Extensional tectonic reactivation occurred repeatedly at specific intervals along the megashear. The Kandi-Hoggar 4050 lineament in Africa cuts through the region of the Pan-African Trans-Saharan orogen and is covered, in some parts, by relatively young and shallow cratonic covers (**fig. 9.4**). In South America, TB crosses the entire Borborema and Tocantins tectonic provinces, and it also cuts through the basement of three large and relatively thick cratonic basins, the Parnaíba, to the north, and the Paraná and Chaco-Paraná, to the south (**fig. 9.5**). As indicated by Cordani et al. (2013), when TB leaves the Parnaíba Basin, immediately to the southwest of it, the linear structures of the megashear truncate the north-south structural trends of the Araguaia Belt of the Neoproterozoic age.

A series of small extensional cratonic sedimentary basins were formed along the TB in graben troughs, such as the Jaibaras, Monte do Carmo, Água Bonita, and Piranhas basins (Brito Neves et al., 1984). They are early Paleozoic, formed by brittle reactivation processes that affected older shear zones of the lineament. The Jaibaras rift is located at the northwestern corner of the Borborema Province (Oliveira and Mohriak 2003, Aguiar et al., 2011), and is described as an extensional structure forming a graben, which continues to the southwest, beneath the sedimentary rocks of the Parnaíba basin. Brito Neves et al. (1984) showed that it represents a precursor intracratonic rift for the thermal subsidence that started in the Silurian with the deposition of the Serra Grande formation. The main depocenters for this formation and for the younger sedimentary sequences, which continue into the Carboniferous, are located along the TB. Further south, within the Tocantins Province in central Brazil, TB maintains a northeast-south-west trend and affects parts of the Goiás Magmatic Arc. The Monte do Carmo rift, with a similar tectonic evolution to the Jaibaras rift, as well as the Água Bonita and Piranhas grabens, filled with Paleozoic sediments, are also located along the same structural trend.

Towards the southwest, TB disappears beneath the Paraná Basin. From geophysical evidence produced by Mantovani and Brito Neves (2005), and as previously suggested by Cordani et al. (1984), the megashear separates the supracrustal rocks of the Paraguay Belt to the west from the Paranapanema cratonic fragment to the east. Finally, continuing into Paraguay and Argentina, the TB is present within the basement of the Chaco-Paraná basin, where it has affected the tectonic evolution of sedimentary systems, as shown by the prominent depocenters of the Pilar and Las Breñas basins, where a total thickness of several kilometers is found (Wiens 1985).

Concomitantly to extensional tectonics along the Transbrasiliano tectonic corridor, mafic magma underplating and anatexis of the continental crust may have been responsible for the onset of postorogenic

bimodal magmatism, active from the Ediacaran to the Ordovician (580-450 Ma), accompanied by the intrusion of practically undeformed K-rich, A-type granite bodies. The related rifting of the lithosphere may well have been produced by transtensional stresses, as a distant reflection of the more or less coeval compression produced at the southern border of Gondwana, during the subduction of the Pacific lithosphere.

A few examples of the Ediacaran to Cambro-Ordovician magmatism along the TB are given, as follows:

- In the state of Ceará, the emplacement of the Mucambo (530 Ma) and Meruoca (523 Ma) granites intrusive into the sedimentary rocks of the Jaibaras rift (Archanjo et al., 2009), as well as the Quintas Ring Complex (495 Ma), located to the east of the TB (Castro et al., 2012).
- In the region of the Goiás Magmatic Arc and vicinities, the emplacement of a number of intrusive complexes comprising small gabbro-diorite bodies associated with large K-rich granite plutons and A-type granitic intrusions, such as the Serra Negra (508 Ma), Iporá (490 Ma), and Serra do Impertinente (485 Ma) (Pimentel et al., 1996).
- Within the area of the Paraguay belt in Mato Grosso do Sul, in the vicinity of the TB megashear, the emplacement of the São Vicente (521 Ma), Coxim (542 Ma), Rio Negro (549 Ma), Sonora (549 Ma) and Taboco (546 Ma), intruding deformed metasedimentary rocks of the Cuiabá Group (Ferreira et al., 2008, McGee et al., 2012).

9.7. Conclusion

In conclusion, the collage of West Gondwana was largely completed by the end of the Precambrian, when the Amazonian, West African, São Francisco-Congo, Kalahari and Rio de la Plata cratons, the Saharan metacraton and the Parnaíba, Paranapanema and Luiz Alves cratonic fragments were united and tectonically stabilized. The geological evidence available so far indicates that the Neoproterozoic Goiás-Pharusian Ocean was already closed at about 600 Ma, and the resulting sutures are located within or close to the Transbrasiliano-Kandi tectonic corridor.

Later, in the Ediacaran and continuing at least during early Paleozoic, following the final stages of the Brasiliano orogeny, extension was predominant over West Gondwana and this structural regime may have been a result of two main factors: (1) the orogenic collapse of the folded belts produced by the Brasiliano orogeny; and (2) a distant tectonic reflection of the compressional Pampean orogeny, which was in action at the south-western margin of Gondwana. As a corollary, we argue that the idea of a collisional suture in central South America, closing a supposed Cambrian Clymene Ocean in the Cambrian, is not sustainable.

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10. Was there an Ediacaran Clymene Ocean in Central South America?

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Abstract

Previous studies have proposed that a major suture resulted from the collision between the Amazonian and São Francisco-Congo cratons during the Cambrian, following the closure of a supposed Clymene Ocean. The proposal tentatively located this ocean along the Araguaia and Paraguay belts at the eastern margin of the Amazonian Craton, and its southern extension reached the Pampean belt in Argentina. In the present study we will argue that the existence of Ediacaran-Cambrian oceanic lithosphere in central South America is highly unlikely. West Gondwana was assembled during the convergence between the Amazonian, West African, São Francisco-Congo and Rio de La Plata cratons as well as the Saharan Metacraton, leading to the closure of the Goiás-Pharusian Ocean during the Neoproterozoic. Final closure and continental collision resulted in the development of the Transbrasiliano-Kandi mega-shear zone that cuts through several mobile belts, but leaves the cratonic areas totally untouched. Consistent results of radiometric dating along the Transbrasiliano (TB) mega-shear in South America and of metamorphic rocks of the Brasília Belt have indicated that the Neoproterozoic collision finished at *ca.* 620 Ma. After isostatic uplift, cooling, and denudation, between 590 and 500 Ma, emplacement of undeformed K-rich postorogenic granites represented the main tectonic event. At this time or afterwards, a series of small extensional sedimentary basins formed in graben troughs, most of which are within the TB tectonic corridor. They all were of extensional character, contrasting clearly with the convergent tectonics occurring within the coeval Pampean Orogen in Argentina. The main arguments showing that an Ediacaran to Cambrian oceanic closure in central Brazil is untenable include: (i) the assembly of West Gondwana was completed by *ca.* 600 Ma, when the convergence between the Amazonian, São Francisco and Rio de La Plata cratons had already ended. After this, there is no geological evidence of an oceanic lithosphere (for example, ophiolites, magmatic arcs, *et cetera*), ruling out the possible existence of an Ediacaran or Cambrian Clymene Ocean in Central Brazil; (ii) the Gurupi and Araguaia belts in Brazil, as well as the Bassaride and Rokelide belts in West Africa, are regarded as aulacogenic-type systems formed within an intraplate tectonic setting. Their tectonic history precedes the collision between the Amazonian and São Francisco-Congo cratons, as demonstrated by the linear structures of the Transbrasiliano megashear which truncate the N-S structural trends of the Araguaia Belt; (iii) there is a close correlation between the Corumbá Group of the Paraguay Belt in Brazil and the Arroyo del Soldado Group in Uruguay. These sedimentary sequences belonged to the same Ediacaran continental shelf and this is a powerful indicator for an Ediacaran connection between the Amazonian and Rio de La Plata cratons, which precludes the existence of a wide ocean (for example, the Clymene) between them. On the other hand, the tentative correlation between the Sierras de Córdoba and the Paraguay Belt cannot be accepted, because these are far apart and there is no similarity in lithology, metamorphism, or structural trends; (iv) the Puga paleopole is the most important evidence for the hypothesis of the Cambrian Clymene Ocean, however the age of about 600 Ma for this paleopole, taken on the basis of Sr and C isotopes, is loosely constrained. In addition this is located at low latitude, not far from the present pole, and therefore could be related to a younger remagnetization; (v) the Pampean Orogen is made up of medium- to high-grade metamorphic rocks constrained between 560 Ma and 520 Ma and therefore was tectonically active during most of the Cambrian. However, at this time, an oceanic lithosphere is not evident in the vicinity of the Paraguay belt, and in central Brazil extensional rather than convergent tectonic processes have been observed.

10.1. Introduction

Almost all publications regarding the formation of Gondwana stress that it was formed by the amalgamation of several building blocks of different sizes, most of them originating from the breakup of Rodinia. They assembled through a series of continental collisions that covered the entire Neoproterozoic (see, for example, Li et al., (2008) and references therein). West Gondwana, the largest building block of the supercontinent, included the Amazonian and West African cratons, the Congo-São Francisco, Rio de La Plata, and Kalahari cratons, as well as a few smaller continental fragments, such as the Paranapanema, the Goiás Massif, and the Luiz Alves (see **figure 10.1**). It also included a large region of northern Africa, named the Saharan Metacraton.

West Gondwana was assembled during the convergence between the Saharan Metacraton and the Amazonian, West African, and Congo-São Francisco cratons. This event led to the closure of the Goiás-Pharusian Ocean and to the development of several orogens during the Neoproterozoic Era. In central Brazil the Brasília Belt occurs along the western margin of the São Francisco Craton. Further to the west, along the eastern/southeastern margins of the Amazonian Craton, they are known as the Paraguay and Araguaia belts. According to most existing tectonic, stratigraphic, and geochronological evidences, West Gondwana was already assembled by ca. 650 Ma to 600 Ma.

As an alternative scenario, Trindade et al. (2006) postulated that the final assembly of Gondwana occurred in the Cambrian, after the closure of a large oceanic basin, named the Clymene Ocean. This event was a result of the convergence and collision between the Amazonian and Congo-São Francisco continents. Their rationale for this model was based on paleomagnetic data for sedimentary rocks of the Araras Formation, which were obtained by Trindade et al. (2003) from the Paraguay Belt of Central Brazil. These rocks were believed to have been deposited at ca. 600 Ma. However, its paleopole plotted quite far from the other Gondwana poles for that time. This observation was understood to imply that the Amazonian Craton and the rest of Gondwana were far apart during Ediacaran to Cambrian times. Although the Clymene Ocean was not properly described in their work, the concept was readily accepted by researchers working in southern South America, such as Li et al. (2008), Pisarevski et al. (2008), Cordani et al. (2009), Tohver et al. (2010), and Ramos et al. (2010).

The hypothesis of a Cambrian ocean seems appealing in terms of its elegance and simplicity, and has been seriously considered. For instance, Cordani et al. (2009) indicated two alternative timeframes for the assembly of West Gondwana. In the first timeframe, the Amazonian and West African cratons joined the São Francisco-Congo Craton before the Ediacaran. In the second timeframe, the model put forward by Trindade et al. (2006) was followed, and the assembly of West Gondwana was completed only after the Ediacaran.

Supporting the idea of the Cambrian Clymene Ocean and looking for a suitable location for the suture resulting from its closure, Tohver et al. (2012) proposed that the suture zone crossed the entire South American continent (as shown in **figure 10.2**). They presented new geochronological data indicating Ediacaran to Cambrian magmatism at the Sierras de la

Figure 10.1- Major tectonic elements related to West Gondwana, prior to the final amalgamation. Major cratons: AM = Amazonian; CO = Congo; KA = Kalahari; RP = Rio de La Plata; SF = São Francisco; SM = Sahara Metacraton; WA = West African. Smaller cratonic fragments: AA = Arequipa-Antofalla; BO = Borborema; GO = Goiás Central Massif; LA = Luiz Alves; PA = Paranapanema; PB = Parnaíba; PP = Pampia. Intra-oceanic magmatic arcs: A = Amalaoulaou; G = Goiás; I = Iskel; K = Kabyé; T = Tilemsi.

Ventana of Argentina. They also suggested a correlation with the Pampean belt, although the shallow level of magma emplacement in the Sierra de la Ventana contrasted with the deeply exhumed high-grade rocks of the Pampean Orogen. Moreover, they reviewed the tectonic history of the Pampean, Paraguayan, and Araguaia belts along the margins of the Amazonian and Rio de La Plata cratons. They tried to demonstrate that these three belts were tectonically active from the late Ediacaran to the late Cambrian times along the Clymene suture zone, marking the closure of the ocean and the final stages of formation of Gondwana.

In this paper, we will present the available evidence indicating that the South American Platform was already in place in the Neoproterozoic. We will argue that the existence of an area with an oceanic lithosphere in its central region during the period from Ediacaran to early Cambrian is highly unlikely. As a corollary, we will suggest that the Clymene Ocean and its Cambrian closure, as put forward by Trindade et al. (2006) and Tohver et al. (2012), are untenable on the grounds of the available evidence.

10.2. Closure of the Goiás-Pharusian Ocean

Figure 10.1 presents a likely model for the relative positions of the main cratonic elements around 750 Ma to 700 Ma, prior to closure of the Goiás-Pharusian Ocean and the final collisions that led to West Gondwana. This figure shows the approximate location of the oceanic realms, where the intraoceanic island arcs were

already in place; outlines of the cratonic masses and fragments, together their marginal basins that would become Neoproterozoic mobile belts; some of the accretionary units within the intervening oceans; smaller microcontinents (e.g., Arequipa-Antofalla and Pampia); and smaller ancient blocks located in the Borborema and Tocantins Provinces. **Figure 10.1** also illustrates the position of the confined Adamastor Ocean, located in Central Gondwana, whose tectonic history ended only in the Cambrian.

The name “Goiás-Pharusian Ocean” was suggested by Kröner and Cordani (2003), although many paleomagnetic reconstructions, such as those by Meert (2003) and Cordani et al. (2003), have named this ocean “Brasiliano” or “Adamastor”. This ocean occupied a very large area and included intraoceanic magmatic arcs, whose tectonic evolution started as early as ca. 900 Ma and covered the entire Neoproterozoic. As mentioned above, this ocean closed ca. 650 Ma to 600 Ma, after a series of successive continental collisions that gave rise to the many mobile belts of the Brasiliano-Pan African orogenic cycle. In most recent paleomagnetic reconstructions, such as these by Rapalini (2006) or Tohver et al. (2010), all workers agree on a consolidated West Gondwana by Middle Cambrian.

The closure of the Pharusian Ocean, in consequence of the convergence between the West African Craton and the Saharan Metacraton, was characterized by Himalayan-type continental collisions, with the development of UHP and HP rock associations in the Trans-Saharan orogenic belt (see **figure 10.3**). Blue-schists and eclogites have been identified in the Gourma region and in Togo, and geochronological studies have estimated the age of metamorphism at ca. 620 Ma (Caby, 1994; Trompette, 1994; Jahn et al., 2008; Attou, 1998; Agbossoumoundé et al., 2001; Affaton et al., 2000). Oceanic terranes, relicts of the Pharusian Ocean, have been identified in many regions, from the Hoggar to the Dahomeyan segments (Caby, 1989, 2003; Berger et al., 2011; Dostal et al., 1994; Duclaux et al., 2006). These terranes, which have been dated within the 900 Ma to 700 Ma timeframe, correspond to the Iskel island arc in the Hoggar, the Tilemsi-Amalaoulaou intraoceanic arc assemblages in the Gourma region, and the Kabyé massif in the Dahomeyan Belt of Togo.

Abdelsalam et al. (2002) set the boundary between the Trans-Saharan Belt and the Saharan Metacraton along the eastern margin of the Tuareg Shield (Liégeois et al., 2000). However, their Saharan Metacraton was not well-defined. It was characterized as a large portion of pre-Neoproterozoic cratonized continental crust dominated by medium- to high-grade gneissic and migmatitic terrains, which were highly remobilized during the Neoproterozoic. The NE portion of the Borborema Province of Brazil—where large areas with Paleoproterozoic or older crust have been found, may correlate with the Saharan Metacraton. Moreover, a correlation between the northwest part of the Borborema Province and the Trans-Saharan belt is highly likely, given the similarity of the regional lithostratigraphic trends (Caby, 1989; Arthaud et al., 2008), the continuity of the major faults, and the correlation between the granitoid rocks of Dassa and Savá in Benin (Cordani and others, 1993) with rocks of the Tamboril-Santa Quitéria Complex in the Ceará State of Brazil, a large area formed by different types of granitic rocks and migmatites (Fetter et al., 2003). Ceará HP retroeclogites that are located close to the eastern side of this complex (see **fig. 10.5**), dated at ca. 650 Ma by Amaral et al. (2010), are considered to be representative of the Neoproterozoic suture.

Similar to the Pharusian Ocean, the Goiás Ocean was consumed as a consequence of the convergence between the Amazonian and São Francisco Cratons. As one of the largest, most complete, and most preserved

Neoproterozoic orogenic belts in Brazil (Pimentel and others, 2000), the Brasília Belt presents compelling evidence for closure of the Goiás Ocean at ca. 650 Ma to 630 Ma, as recorded by granulitic rocks of the Anápolis-Itaçu Complex (A in **fig. 10.3**) and several other rock units (Baldwin and Brown, 2008). This belt comprises a thick Meso-Neoproterozoic sedimentary pile in the east, a microcontinent composed of Archean rock units and associated Paleoproterozoic formations (the Goiás Massif), and a large magmatic arc in the west, the so-called Goiás Magmatic Arc (GMA in **figure 10.3**). Due to its tectonic significance and areal magnitude, the GMA represents the most important tectonic element of the Brasília Belt, and is formed by Neoproterozoic juvenile crust that records the closure of the Goiás Ocean from ca. 900 Ma to 600 Ma (Pimentel and Fuck 1992).

Figure 10.2 – Outline of the Transbrasiliano-Kandi mega-shear zone in a pre-drift reconstruction of South America and Africa. The suggested position of the Cambrian suture proposed by Tohver et al. (2012) is indicated. Phanerozoic covers are omitted.

The GMA is divided into the Arenópolis arc to the south and the Mara Rosa arc to the north, which are separated from each other by the Goiás Massif. The arc includes: (i) juvenile island arcs (ca. 900–800 Ma) with volcanic-sedimentary sequences that are spatially associated with tonalitic-granodioritic-granitic orthogneisses with a mantle signature; (ii) younger (ca. 650–630 Ma), island arc-type volcano-sedimentary sequences and

associated tonalite-granodiorite plutonic complexes; and (iii) late- to post-orogenic (<600 Ma) granites that are associated with gabbro-diorite bodies (Junges et al., 2002, and references therein).

The Brasiliano-Pan-African belts with Neoproterozoic sutures are aligned along a very long area (>6000 Km) of South America and Africa that is dominated by one of the most important tectonic elements of the world. This megashear zone is called the Trans-Brasiliano lineament in Brazil, continues as the Kandi-Hoggar 4°50' lineament into Africa (Caby, 1989), and will be referred to in this work as the “Transbrasiliano-Kandi”. As shown in **figures 10.2 to 10.5**, the megashear zone cuts through many of the Brasiliano-Pan African belts, but leaves the cratonic areas untouched. It may be the largest coherent shear zone on Earth (Attoh and Brown, 2008). Its coherence has been well-characterized by the very long linear magnetic anomalies obtained from the CHAMP satellite survey and reported by Fairhead and Maus (2003). This lineament probably reaches the bottom of the lithosphere and is formed of a series of ductile shear zones, comprised of many parallel sets of faults, which may cover very broad areas. The shear zone motion must have started shortly after the closure of the Goiás-Pharusian Ocean. Tectonic reactivations have repeatedly occurred along the shear zone, where low-intensity seismic activity continues through the present day.

The 650 Ma to 600 Ma timeframe for the orogenic tectono-magmatic episodes in northwest Africa and central South America indicate that the convergence of the continental blocks leading to the closing of the Goiás-Pharusian Ocean was finished at that time along the corridor of the Transbrasiliano-Kandi megashear.

10.3. Ediacaran and Cambrian subduction of the oceanic lithosphere in southern South America

In the previous chapter, we showed that the Goiás-Pharusian Ocean closed at the end of the Neoproterozoic, at which time West Gondwana was in place as a single continental mass. At about the same time, the Mozambique Ocean was also closing, with termination of the convergence between West Gondwana and the different components of East Gondwana. A major center was developing and spreading between West Gondwana and Laurentia, which led to a need for major plate reorganization. This reorganization was associated with the initiation of convergence along the Pacific margin of Gondwana.

Cawood (2005) suggested that the subduction of the Pacific oceanic lithosphere occurred at the Gondwana margin at ca. 570 Ma, more or less simultaneously with the separation of Laurentia and the opening of the Iapetus Ocean. The name “Terra Australis Orogen” was proposed for a very large tectonic province that was located at the southern Gondwana margin, along an open and unconfined Pacific Ocean and comprising a collection of accretionary orogens. These orogens were formed by the stacking of magmatic arc complexes, which were formed in successive subduction zones by the tectonic processes of “soft collision” and accretion, accompanied by the extensive production of felsic volcanic and granitoid magmas. Knowledge about such major tectonic processes is necessary for understanding the tectonic development of southern South America within the 600 Ma to 500 Ma timeframe and for assessing the possible presence of the suggested Clymene Ocean. The initiation of the Pacific subduction and the development of the Pampean Orogen are the key elements needed to analyze the plate reorganization that occurred just after the amalgamation of Gondwana.

Figure 10.4 shows a schematic tectonic outline for the southwestern part of South America. Describing the general tectonic setting of the Sierras Pampeanas from a mobilistic perspective, Ramos (1988) pointed to differences between the Puncoviscana Formation in the north and the Eastern Pampean ranges, including the Sierra Norte and the Sierra de Cordoba, in the southeast. Covering a very large region of northern Argentina and southern Bolivia, the Puncoviscana Formation consisted of turbidites, which have been interpreted as a deep-water “flysch-type” sequence and as part of a passive margin, whose basin was linked to a stable craton to the east but was probably open to the west to a proto-Pacific Ocean. The Eastern Pampean ranges, in which high-grade metamorphic rocks have been found, were thought to be formed as the result of normal subduction of oceanic lithosphere, followed by a continent-continent collision between the same Pampean terrane (Pampia in **figure 4**) and the Rio de La Plata Craton.

Rapela et al. (1998), working on the polymetamorphic basement of the Sierras de Cordoba, presented a comprehensive report on what they called the “Pampean Orogeny,” in which the low-grade meta-sedimentary rocks of the Puncoviscana Formation and the medium- to high-grade metamorphic rocks of the Sierras Pampeanas were attributed to the same tectonic episode. Their geochronological data indicated a clear Cambrian age for the entire tectonic development. Detrital zircons from the low-grade sediments established a maximum age of deposition at 560 Ma to 550 Ma. An age of 530 Ma was obtained for the emplacement of the calc-alkaline granites, as a result of northeast-directed subduction. These authors followed the same interpretation offered by Ramos (1988) and proposed that the high-grade metamorphism characteristic of the Pampean Orogeny, dated shortly after emplacement of the granites at ca. 525 Ma, was produced by a continental collision of the western side of Gondwana (the Rio de La Plata Craton in **figure 10.4**) with an exotic terrane, retaining for it the name of “Pampean Terrane”.

Schwartz et al. (2008), studying the Sierra Norte, performed an additional geochronological program and extended the time interval for the calc-alkaline magmatism from 555 Ma to 525 Ma. These authors discussed the implications of the tectono-magmatic evolution of the Pampean Orogen, correlating it with the Saldania, Ross, and Tasmanian companion belts of the Terra Australis. In particular, they noted that the plutonism in all of them was roughly coeval from Ediacaran to early Paleozoic and included very similar calc-alkaline and strongly peraluminous suites.

Escayola et al. (2011), dealing mainly with the Puncoviscana Formation in the northern part of Argentina, proposed a new tectonic interpretation. They summarized the available lithological, stratigraphic, and structural knowledge on that formation and presented conclusive evidence for the syntectonic character of it as an accretionary complex. They also highlighted the almost north-south strike of the belt along the proto-Andean active margin of West Gondwana, where they located an oceanic opening called the “Puncoviscana tract” (see **figure 10.4**). This opening is in approximately the same position as that of the oceanic basins envisioned by Ramos (1988) and Rapela et al. (1998).

Recalling that Escayola et al. (2007) had identified ophiolite remnants within the high-grade metamorphics of the Sierra de Cordoba, and considering the presence of some Early Cambrian felsic and mafic arc volcanic rocks in the older parts of the Puncoviscana Formation, as well as the arc-like composition of the associated turbidites, Escayola et al. (2011) inferred that sedimentation occurred adjacent to an approximately coeval arc

terrane. In their model, this process occurred in a basin that formed on the accretionary wedge, associated with a west-facing Pampean arc built upon the proto-Andean margin of West Gondwana. They proposed that their Puncoviscana tract was closed by a collision between Gondwana and the exotic Arequipa-Antofalla microcontinent. They indicated that this tectonic episode could be related to the separation of Laurentia and the spreading of the Iapetus Ocean.

The above discussion demonstrates the controversy surrounding the issue of the tectonic significance of the Pampean Orogeny. Regardless of whether the Puncoviscana Formation represents an active or passive margin, whether the tectonic process is collisional or non-collisional, or whether the Puncoviscan Ocean is open or restricted, the timing of the orogenic processes is clearly well-constrained between 560 Ma and 520 Ma. Therefore, the Pampean orogenic system is much younger than the Neoproterozoic belts responsible for closing the Goiás-Pharusian Ocean. The Pampean system had already started in the Ediacaran and was tectonically active during most of the Cambrian. Given that the Amazonian, São Francisco-Congo, and Rio de La Plata Cratons were already together after the Brasiliano Orogeny, it can be concluded that instead of representing the final amalgamation of Gondwana, the Pampean Orogeny marks the oldest evidence of subduction of the Pacific oceanic plate under the southwestern margin of Gondwana.

10.4. Extensional-type post-tectonic episodes along the Transbrasiliano Lineament in South America

The Kandi-Hoggar 4°50' lineament in Africa cuts through the mobile belts of the Trans-Saharan orogen and is covered, in some parts, by relatively young and shallow cratonic covers. Its South American counterpart, the Transbrasiliano lineament (hereafter, TB), cuts through parts of the Borborema Province and Tocantins tectonic provinces, as well as the basement of three large and relatively thick cratonic basins, the Parnaíba basin (**figure 10.3**) and the Paraná and Chaco-Paraná basins to the south. The TB is clearly visible in the aeromagnetic mosaic of the central and eastern parts of the Brazilian territory, crossing the country from northeast to southwest. Comprised of a series of low-amplitude magnetic anomalies, it starts at the northwestern tip of the Borborema Province, continues through the eastern part of the Parnaíba Basin to the central part of Goiás, travels to the northern part of the Paraná basin, and goes to the southwest, to Paraguay and Argentina, where it seems to end near the city of Cordoba.

Consistent results of radiometric dating along the TB have indicated that the Neoproterozoic collisional tectono-magmatic events finished at ca. 620 Ma. However, after isostatic uplift, cooling, and denudation, the mobility and tectonic activity along the megashear continued for a long time. Within the period between 590 and 500 Ma, emplacement of virtually undeformed K-rich postorogenic granites represented the main geological/tectonic event, which was sometimes associated with gabbros and diorites. This fact can be verified all along the TB. At the north, in Ceará State, a few granitic plutons emplaced at 530 Ma to 500 Ma have been identified that are representative of late to postcollisional processes in the Ceará Central Domain (Fetter et al., 2003; Castro et al., 2012). In Mato Grosso do Sul State, several undeformed granitic plutons with ages between 550 Ma and 520 Ma (Ferreira et al., 2008, and references therein) have been encountered along the TB, which cut through the supracrustal rocks of the Paraguay Belt, as will be described later.

Figure 10.3 – Geological correlations between north-eastern South America and north-western Africa, in a pre-drift reconstruction. The location of specific tectonic features mentioned in the text are indicated: HP and UHP metamorphic units of Neoproterozoic age; extensional early Paleozoic basins; and basement mantled domes within the Araguaia Belt.

In Goiás State, deep crustal and lithospheric seismic tomography (Assumpção et al., 2004; Feng et al., 2007) and deep seismic refraction and teleseismic receiver function investigations (Soares et al., 2006; Ventura et al., 2011) have been conducted at the location of a large positive Bouguer anomaly, corresponding to the surface exposure of the GMA. In this region, using group-velocity tomography and lithospheric s-velocity studies, Feng et al., (2007) demonstrated that there was a region of thinner lithosphere along the shear system. This finding may suggest a process of delamination of the crustal root of the collisional orogen, which gave way to an asthenospheric uplift, with the heating and formation of posttectonic high-K and A-type granitic intrusions. Especially within the area of the Goiás Magmatic Arc, these granitic plutons, whose radiometric ages were between 560 Ma and 520 Ma, were virtually undeformed, indicating that they were related to extensional-type tectonics along the megashear (Pimentel et al., 1996). One point yet to be clarified is the presence of small exposures of medium- to high-grade rocks (migmatites, enderbites, and mafic-ultramafic bodies), dated between 560 Ma to 520 Ma, along the TB in Goiás and Tocantins (e.g., Lima et al., 2008; Motta-Araujo et

al., 2003). The tectonic significance of these local occurrences is poorly understood, but they might represent exposures of lower crustal sections related to the emplacement of mafic magmas and subsequent crustal melting during Late Neoproterozoic to Cambrian times.

At the time of the intrusion of these granites or afterwards, a series of small extensional intracratonic sedimentary basins formed in graben troughs, such as the Jaíbaras, Monte do Carmo, Água Bonita, and Piranhas basins, along the TB in Ceará, Tocantins, Goiás, and Mato Grosso States (Brito-Neves et al., 1984). Most of them were rift basins formed between the Cambrian and the Silurian by brittle reactivation processes that affected older shear zones of the TB. They were mainly formed by fault-scarp-related paraconglomerates at the base, followed laterally and vertically by fluvial-lacustrine sandstones interbedded with shales.

The Jaíbaras rift, located at the northwest corner of the Borborema Province (Ja in figure 3), is the best representative of these extensional structures (Oliveira and Mohriak, 2003; Aguiar et al., 2011). Its age is well-controlled by the Mucambo (530 Ma) and Meruoca (510 Ma) granitic plutons. The graben trough continues to the southwest, beneath the sedimentary rocks of the Parnaíba basin, where it represents a precursor intracratonic rift for the thermal subsidence of the cratonic basin, which started in the Silurian with the deposition of the Serra Grande Formation. The main depocenters for this formation and younger sedimentary sequences until the Carboniferous were located along the TB (Brito-Neves and others, 1984), demonstrating that the successive extensional tectonic reactivations occurred in the Paleozoic. Brito-Neves et al. (1984) suggested the presence of a cratonic nucleus, herein named as the “Parnaíba block” (see **figures 10.2 and 10.3**), within the basement of the Parnaíba Basin, to the west of the TB trend.

Within the Tocantins Province in central Brazil, the TB maintains a northeast-southwest trend and is located over the Goiás Magmatic Arc. Immediately to the southwest of the Parnaíba Basin, the Monte do Carmo rift (Mo in **figure 10.3**) seems to have had a very similar tectonic evolution to the Jaíbaras rift. The Água Bonita graben, also located along the TB (Ag in figure 3), is filled with Paleozoic sediments (Brito Neves et al., 1984). The Piranhas basin, filled with Cambrian and Ordovician sediments, is located right next to the border of the Paraná cratonic basin.

To the southwest, the TB disappears below the northeastern corner of the Paraná Basin. As suggested by Cordani et al. (1984) and as inferred by Mantovani and Brito-Neves (2005) from geophysical evidence, within the basement of this basin, the megashear separates the supracrustal rocks of the Paraguay Belt to the west from a cratonic fragment (the Paranapanema block) to the east (**figures 10.4 and 10.5**). Along the western boundary of the Paraná Basin, not far from the influence of the TB megashear, Ferreira et al. (2008) dated a few practically undeformed granitic plutons, which yielded ages between 550 Ma and 520 Ma and will be discussed later. These plutons are filling spaces related to extensional features. They intrude deformed metasediments belonging to the older lithostratigraphic system of the Paraguay Belt. Continuing into Paraguay and Argentina, the TB is present within the basement of the Chaco-Paraná cratonic basin. Here, the TB has affected the tectonic evolution of the sedimentary systems, as shown by the more prominent depocenters of the linear basins of Pilar in Paraguay (Wiens, 1985) and Las Breñas in Argentina (LB in **figure 10.4**), where a total thickness of more than 6000 meters can be found.

Figure 10.4 – Geotectonic interpretation of the south-eastern part of South America. It includes cratonic units (Amazonian, Rio de La Plata, Paranapanema and Luiz Alves), allochthonous terranes (Arequipa-Antofalla, Famatina, Cuyania South America during the Paleozoic and the tectonic units of the Pampean orogeny: the eastern Pampean ranges and the Puncoviscana Tract. Phanerozoic covers are omitted. Tectonic features related to the Transbrasiliano Lineament.

In summary, the Goiás-Pharusian Ocean was already closed at 600 Ma. Therefore, later events, most of which located within or near the TB, were of an extensional character. This overall extensional tectonic scenario in central Brazil clearly contrasts with that of the coeval Pampean Orogen in Argentina. In this latter area, the period between ca. 560 Ma and 520 Ma was marked by the subduction of the Pacific oceanic lithosphere,

including tectonic compression and regional high-to medium-grade metamorphism, associated with voluminous granite magmatism of an orogenic and subduction-related nature (Ramos 1988; Rapela et al., 1998).

Figure 10.5 – Late Neoproterozoic geotectonic features of eastern South America. The location of the proposed suture resulting from the closure of a supposed Ediacaran/Cambrian Clymene Ocean is indicated.

10.5. Was there an Ediacaran Clymene Ocean in Central South America?

In this chapter, we present some arguments showing that a Cambrian oceanic closure, as suggested by Trindade et al. (2006) and Tohver et al. (2012) and depicted in **figure 10.5**, is untenable on the grounds of the available evidence.

10.5.1. Age of the Amazonian-São Francisco-Congo collision along the region of the Transbrasiliano Megashield

As we showed in the previous chapters, the amalgamation of Gondwana involved a long process of plate convergence. In South America, it started at ca. 900 Ma, during the earliest magmatic phases of the juvenile, intraoceanic, and accretionary Goiás Magmatic Arc. Evidence suggests that the main collisional episodes of the Brasiliano-Pan African orogens occurred at the end of the Neoproterozoic. Several recent articles with robust geochronological control (mainly by means of U-Pb zircon ages) demonstrated that the tectonic

evolution of the region dominated by the TB was developed roughly between 650 Ma and 600 Ma. Considering the possible high-grade metamorphic events that could be related to Himalayan-type collisions as tracers for the main episodes related to the convergence between the Amazonian and São Francisco-Congo Cratons, we find that these events are restricted to the oldest phase of the process at ca. 650 Ma (Baldwin and Brown, 2008; Della Giustina et al., 2009). Younger tectono-magmatic episodes have been registered until ca. 600 Ma. However, within the Ediacaran to Cambrian time-period, the few magmatic episodes observed in the area have been related to either posttectonic or anorogenic granitic plutons associated with an extensional tectonic regime.

In summary, the assembly of West Gondwana was completed by 600 Ma, when the convergence between the Amazon-West African and the Central African blocks had already been terminated. Therefore, there is no geological evidence of a possible Ediacaran Clymene Ocean in Central Brazil.

10.5.2. The Bassarides, Rokelides, Araguaia, AND Gurupi belts

Figure 10.3 shows a possible Neoproterozoic tectonic scenario related to the process of plate interaction and convergence between the Amazonian and West African Cratons. This figure shows the position of the Brasiliano-Pan African orogenic belts, such as the Bassarides, Rokelides, Gurupi, and Araguaia, formed by the interplay and consequent collisions between the mentioned cratonic elements. It also accounts for the São Luis cratonic fragment and the possible Parnaíba Block microcontinent, which is concealed below the sediments of the Parnaíba basin and was envisioned by Brito-Neves et al. (1984). The shape of this cratonic fragment is based on the geophysical interpretation of Nunes (1993) and is taken from Klein and Moura (2008).

Villeneuve (2008) extensively reviewed the orogenic belts on the western side of the West African Craton. The oldest is the Bassaride belt, which is cut to the north by the Paleozoic Mauritanide belt and to the south by the Rokelide belt, although parts of it are incorporated within the two younger belts. It is comprised of three lithostratigraphic units, the oldest of which is a volcano-sedimentary sequence formed within a rift-related basin. According to Villeneuve (2008), a few rhyolites stratigraphically below the oldest sequence of the unit yielded ages between 1050 Ma and 1000 Ma. A range of ages between 700 Ma and 650 Ma were obtained for the volcanic and plutonic rocks of the Niokolo-Koba Group, affected by compression and crustal thickening around 660 Ma. These were followed by the deposition of the flyschoid sedimentation of the Mali and Batapa Groups, whose final metamorphism has been dated at ca. 555 Ma (Villeneuve and Dallmeyer, 1987; Dallmeyer, 1989). A molassic phase consisting of reddish sandstones concluded the sedimentary history of the belt.

According to the same author (Villeneuve, 2008), the younger Rokelide Belt consists of the high-grade basement gneisses of the Kasila thrust belt in the west, extensively mobilized during the Pan-African tectonothermal event, and the Rokelide trough to the east, filled with the low-grade and weakly deformed metasediments of the Rokel River Group and interpreted as a foreland system. Rocks of the Kasila thrust belt (Ka in **figure 10.3**) present Archean ages, similar to those encountered in the nearby Kénéba-Man domain of the West African Craton (Hurley et al., 1971; Williams, 1988). The Rokel River Group (Ro in figure 3) comprises glaciogenic deposits in the lower part and clastic rocks associated with different volcanic types at the

top. Delor et al. (2002) reported U-Pb zircon ages of ca. 570 Ma from the high-grade rocks, interpreted to be the timing of the granulitization. K-Ar and Ar-Ar ages on hornblendes were obtained by Dallmayer (1989), who reported that the range (580–550 Ma) constrained the entire tectonic evolution of the belt to a pre-Ediacaran time.

Since the article by Hurley et al. (1967), the São Luis Craton (**figures 10.3 and 10.5**) has been considered to be a small cratonic fragment that was separated from the much larger West African Craton and remained in South America when the Atlantic Ocean was formed in Mesozoic time. The Neoproterozoic Gurupi belt, which occurs on the western side of the craton, displays structural transport and metamorphic polarity towards the cratonic area. Klein and Moura (2008) analyzed the tectonic development of the belt, suggesting a direct continuation of the Rokelide Belt. The predominant rocks exposed in the outcropping area were basement gneisses with Paleoproterozoic ages of ca. 2200 Ma. These rocks, affected by the Brasiliano-Pan African orogeny, included granitoids that were correlative of similar Tromaí calc-alkaline bodies belonging to the São Luis Craton. They formed the continental basement over which a Neoproterozoic belt was developed, starting with a rifting phase at ca. 730 Ma (Klein et al., 2005). At the southeastern part of the belt, the medium-grade metasedimentary rocks of the Marajupema Formation were found, thrust over the basement rocks. Klein and Moura (2008) suggested that the Neoproterozoic tectonic event was due to a collision between the São Luis Craton with the concealed Parnaíba cratonic block, located below the sedimentary rocks of the Parnaíba Basin.

In pre-drift Brazil-Africa reconstructions, the Araguaia Belt in Brazil has been routinely correlated with the Rokelides; both belts exhibit similar north-northwestern trends (see **figure 10.3**) and virtually the same age for the tectonothermal event responsible for their final tectonic configuration (Trompette, 1994; Moura et al., 2008). However, these belts have opposite structural vergence. The Araguaia Belt shows tectonic transport to the west, in the direction of the Amazonian Craton, whereas the Rokelide Belt is transported easterly, against the West African Craton. The Araguaia Belt (Alvarenga et al., 2000) comprises two main tectonic units: the low-grade metasediments and associated mafic and ultramafic bodies of the Tocantins Group to the west (To in **figure 10.3**), and the medium- to high-grade gneisses of the Estrondo Group to the east (Es in **figure 10.3**), thrust over the former tectonic unit. The Tocantins Group exhibits north-south structural trends that cut through the west-northwest-east-southeast trends of the Archean and Paleoproterozoic rocks of the Carajás domain of the Amazonian Craton. Its mafic-ultramafic rocks are included within tectonic slices and are interpreted to be remnants of ophiolitic complexes. One of these, the Quatipuru ophiolite, was dated at ca. 750 Ma by Paixão et al. (2008) with the Sm-Nd method. Moura et al. (2008) reported one very precise Pb evaporation zircon age of 817 ± 5 Ma for an intrusion of metagabbro. A rather imprecise Pb-Pb evaporation age of ca. 1000 Ma for the Serra da Estrela alkaline syenitic gneiss has been interpreted as indicating a major event of crustal rifting and formation of the Araguaia basin (Alvarenga et al., 2000).

Within the Estrondo Group, several basement inliers occupying mantled gneissic dome-like structures have been recognized (Hasui et al., 1984; Herz et al., 1989), yielding Archean and Paleoproterozoic ages. Four of them were located in **figure 10.3**. This finding led Moura and Gaudette (1999) to suggest that they could be extensions of the Amazonian Craton. Some zircon Pb evaporation ages ranging from 650 Ma to 550 Ma were obtained for a few syntectonic granitoid rocks, constraining the timing of structural development within the

belt (Alves et al., 2006). Radiometric ages between 560 Ma and 530 Ma obtained by the K-Ar method in a few micas and amphiboles (Moura et al., 2008) indicated that the region of the Araguaia belt was already below ca. 300 °C in the Cambrian. Although a more robust geochronological control is needed for the Araguaia belt, from the perspective of regional cooling, these K-Ar mica ages alone preclude the onset of major regional thermo-tectonic episodes, such as a major continental collision, in Cambrian times.

The Bassarides, Rokelides, Gurupi, and Araguaia belts have been regarded as originating within an intraplate-like tectonic setting (Brito-Neves and Cordani, 1991; Villeneuve and Cornée, 1994; Alvarenga et al., 2000; Villeneuve, 2008, among many others). In this work they will be considered as members of a few activated aulacogens, located between the West African and Amazonian Cratons. Although some of these Neoproterozoic belts may have given birth to a poorly evolved oceanic rift where the ocean floor was formed in restricted areas, such as the Serra do Quatipuru within the Araguaia Belt (Kotschoubey et al., 2005; Paixão et al., 2002), the lack of arc-related magmatic rocks along all them strongly indicates the possibility of an intracontinental tectonic evolution. Finally, the close correlation in age between the basement rocks encountered in the orogenic belts and their respective cratonic areas (see **figure 10.3**) reinforces this reasoning. For the Gurupi belt, the basement granitoid gneisses are a direct continuation of the Tronái granites of Paleoproterozoic age of the São Luis Craton. For the Rokelides, the allochthonous Kasila high-grade gneisses show Archean ages of the same order as those encountered in the adjacent Kénéba-Man domain of the West African Craton. For the Araguaia belt, the granitoid rocks occupying the mantled gneissic domes, basement of the medium-grade rocks of the Estrondo Group, display either Archean or Paleoproterozoic ages, similar to the ages obtained from the Carajás domain of the Amazonian Craton.

The possibility of the existence of a concealed Parnaíba Block is strengthened when several U-Pb dates from detrital zircons ages of Neoproterozoic formations of the Gurupi and Araguaia belts are considered. Based on results obtained from a quartzite belonging to the Estrondo Group in the southern part of the Araguaia belt, Moura et al. (2008) demonstrated a derivation from predominant Neoproterozoic to Mesoproterozoic sources, with ages between 1200 Ma to 800 Ma. Possible source rocks were not observed within the neighboring regions of the Amazonian Craton. These authors postulated a possible eastern provenance, from sources now below the sediments of the Parnaíba basin. Similarly, the presence of detrital zircon crystals as young as 1100 Ma from the Marajupema Formation (Klein and Moura, 2008) was taken to indicate the presence of source rocks from the south because provinces from the Mesoproterozoic age were not found on the São Luis Craton, within the southern part of the West African Craton, or from the eastern part of the Amazonian Craton.

In conclusion, we consider that the Bassarides, Rokelides, Gurupi, and Araguaia are intracontinental belts formed by the tectonic evolution of aulacogenic-type systems. Consequently, their tectonic history would precede the collision between the Amazonian and São Francisco-Congo Cratons along the Transbrasiliano corridor. This conclusion is supported by the evidence that the linear structures of the TB megashear truncate the north-south structural trends of the Araguaia Belt (see **figure 10.3**). The origin of the stresses related to the basin inversion within these belts is poorly understood. As a tentative speculation, we attribute these events, at least partially, to distant plate adjustments accompanying the successive stages of the closing of the Goiás-Pharusian Ocean along the Transbrasiliano-Kandi tectonic corridor.

10.5.3. The Corumbá - Arroyo del Soldado Epicontinental Sea

The Paraguay belt is a typical thrust-and-fold belt located along the southeastern margin of the Amazonian Craton (Alvarenga et al., 2000; Campanha et al., 2011; see location in **figures 10.4 and 10.5**). It comprises the older Cuiabá Group, which includes the glacial sediments of the Puga Formation, and the younger Corumbá Group, which comprises carbonates and pelites with Ediacaran-age fossils (Boggiani et al., 2010). According to Babinski et al. (2006), some “cap carbonates” of the Araras Group, which directly overlie diamictites of the Puga Formation, were dated at ca. 630 Ma (by Pb-Pb whole-rock isochron method). In general, the sedimentary environment indicates a restricted marine shelf with shallow marine waters. This situation could be related to an epicontinental sea overlying a continental basement that corresponds to the extension of the Amazonian Craton. The sediments of the Corumbá Group are mainly marine and correlate with the sedimentary filling of the nearby Tucavaca aulacogen in Bolivia (see **figures 10.4 and 10.5**). Moreover, Gaucher et al. (2003) demonstrated the existence of a close correlation between the Corumbá Group of the Paraguay Belt and the Arroyo del Soldado Group in Uruguay (Co and AS in **figure 10.4**). They observed that these sedimentary sequences, which belonged to the same continental shelf that comprised virtually the same succession with the same Ediacaran age, deepened to the east along the eastern edge of the Rio de La Plata and Amazonian Cratons (e.g. Gaucher et al., 2003, 2008 and 2009). This finding is a powerful paleogeographic indicator for the Ediacaran connection of both cratonic units, which precludes the existence of a wide ocean (e.g., the Clymene) between them.

The rocks of the Paraguay belt are affected by low-grade metamorphism up to the biotite zone of the greenschist facies. They are also affected by tectonic deformation, which is almost imperceptible on the border of the craton but increases up to tight isoclinal folds towards its inner areas (Alvarenga et al., 2000). A detailed regional study by Campanha et al. (2011) in the southern part of the region demonstrated the presence of different structural domains. In particular, rocks of the Cuiabá Group (named Agachi schists by these authors) were affected by deformation and low-grade metamorphism while the sedimentary rocks of the upper part of the Corumbá Group (Tamengo and Guaicurus formations of Ediacaran age) were virtually undeformed and unconformably deposited over the sialic basement of the Rio Apa Block (**figure 10.4**). De Min et al. (2012) provided an important piece of evidence regarding the tectonic evolution of the area. In the northern sector of the Paraguay belt, they studied a series of K-rich, undeformed ultramafic bodies and associated carbonatites that intruded the low-grade metamorphic rocks belonging to the Puga Formation of the Cuiabá Group. The intrusions were plugs and dykes affecting an area about 30 km long, not far from Planalto da Serra, Mato Grosso. A few of these rocks were dated by Ar-Ar, Rb-Sr, and Sm-Nd methods, which revealed high-quality radiometric ages of ca. 600 Ma, placing a lower limit for the deformation of the Cuiabá Group. Moreover, the extensional tectonics of Planalto da Serra could be coeval with the one that produced the Tucavaca aulacogen. We believe that such late Neoproterozoic regional tectonics may indicate the starting point for the separation of Laurentia from West Gondwana.

A final regional deformational phase of very low intensity, extensional in character and related to a few intrusions of granitic bodies, took place during the Cambrian or even later. Within the area of the Paraguay belt in the vicinity of the TB megashear, Ferreira et al. (2008) and McGee et al. (2012) dated some

undeformed granitic bodies, such as the São Vicente (521 ± 8 Ma), Coxim (542 ± 4 Ma), Rio Negro (549 ± 4 Ma), Sonora (549 ± 5 Ma) and Taboco (546 ± 4 Ma), which intruded the deformed supracrustal rocks of the Cuiabá Group. These plutons were coeval with felsic tuffs interlayered with sedimentary rocks of the much younger Tamengo Formation, recently dated at 543 ± 4 Ma (Boggiani et al., 2010). In the Cambrian, the Pampean Orogeny was in action at the southern margin of Gondwana, but no evidence for an oceanic lithosphere has been found in the vicinity of the Paraguay belt. Moreover, the possible suture envisioned by Tohver et al. (2012) along the edge of the main thrust front of the Paraguay Belt (see **figure 10.5**) cannot be defended, because the sedimentary sequences are autochthonous marginal basins located at the border of the Amazonian Craton.

10.5.4 - Significance of the Puga Paleomagnetic Pole

Several large-scale models based on paleomagnetic measurements have been proposed for the assembly of Gondwana. The oldest such model showed the final collision of two large continental masses, West Gondwana (formed of South America and Africa) and East Gondwana (formed of Antarctica, Australia, India and Madagascar), during the period from late Neoproterozoic to early Paleozoic along the Mozambique belt (see, for example, original works by Kroener, 1980 or McWilliams, 1981). In recent years, the mechanism for Gondwana assembly has become more precisely constrained, due to a better understanding of the tectonic evolution and timing of the successive collisions between continental building blocks. However, the Ediacarian interval is notorious for its difficulty in paleomagnetic interpretation, and different models for the assembly of Gondwana have been suggested, such as those by Meert, (2003); Cordani et al. (2003); Yoshida et al. (2003); Collins and Pisarevski, (2005); Pisarevski et al. (2008); Meert and Lieberman, (2008), among many others.

Paleomagnetic measurements were obtained by Trindade et al. (2003) for dolomites of the Puga cap carbonate of the Araras Formation within the Paraguay Belt (location in **figure 10.5**), whose age has been supposed to be very late Neoproterozoic or early Ediacaran. Because this apparent pole plotted quite far from the other poles of Gondwana of the same age, the authors proposed that the Amazonian Craton and the rest of Gondwana were not united, and a large ocean existed between them during the Ediacarian. In their figure 10, they named this ocean as “Clymene”.

The Puga paleopole is the most important and, perhaps, the sole evidence to which the hypothesis of the Cambrian Clymene Ocean is dependent. This finding is considered to be of high quality because of internal coherence. However, Trindade et al. (2006) recognized that the age of 630 to 600 Ma for this paleopole, taken on the basis of Sr and C isotopes, is loosely constrained. Moreover, as emphasized by Pisarevski et al. (2008), this Puga paleopole is located at high latitude, not far from the present-day pole; thus, it could be related to a recent remagnetization.

At present, no other reliable paleomagnetic poles are available for the Amazonian Craton for the Neoproterozoic–early Paleozoic time interval. In our view, there is currently no evidence to show definitively that the Amazonian Craton was far away from the São Francisco-Congo Craton at the time when the Iapetus Ocean started to form. New paleomagnetic measurements from Neoproterozoic to Cambrian rocks, located within the Amazonian Craton, are needed. Specifically for the Paraguay belt, it will be very important to

obtain significant paleomagnetic results from the undeformed magmatic rocks of Planalto da Serra, which exhibit robust radiometric ages close to 600 Ma.

10.5.5. The Pampean Ocean and its northern continuation

Figure 10.4 approximately follows the model of Escayola et al. (2011), in which their Ediacaran oceanic opening (called the Puncoviscana tract) runs along the proto-Andean active margin of West Gondwana, with a north-south trend. The Pampean Orogen, whose medium- to high-grade metamorphic rocks would be coeval with the less metamorphic meta-sediments of the Puncoviscana Formation, is constrained in time between 560 Ma and 520 Ma. Therefore, it was tectonically active during most of the Cambrian. The Amazonian, São Francisco-Congo, and Rio de La Plata Cratons were already together after the collisions related to the Brasiliano-Pan African orogeny; thus, extensional processes, instead of convergence, are observed along the possible sutures, close to the TB megashear zone. In our view, the active margin represented by the Pampean Orogen, with its northern continuation, is the first evidence of the subduction of the Pacific oceanic plate under the southwestern margin of Gondwana. The closure of the Puncoviscana tract may have been the result of an interaction between the Rio de La Plata Craton and either the Pampean terrane of Rapela and others (1998), the Pampia terrane of Ramos et al. (2010), or the Arequipa-Antofalla microcontinent of Escayola et al. (2011). In any case, the subducted oceanic lithosphere would have disappeared below the Rio de La Plata Craton. The Pampean Orogeny, being coeval with the comparable Phanerozoic orogenic systems of the Southern Hemisphere, would be one of the oldest evidences for the onset of the Terra Australis Orogen, as proposed by Cawood (2005). Finally, the Puncoviscana Formation is younger than 570 Ma, as demonstrated by the age of the detrital zircons (Escayola et al., 2011). Thus, at the time of the Pampean orogeny, the Puncoviscana Ocean could not have reached central South America because the collisional sutures related to the Brasiliano orogenic cycle were already closed by this time.

Trindade et al. (2006) did not provide conclusive arguments to prove the existence of an oceanic domain north of the Pampean ranges in Ediacaran to Cambrian times. Tohver et al. (2012) tried to find a possible model to justify the existence of a Clymene Ocean of that age and suggested the location of a Clymene suture as represented in **figure 10.5**. First, they envisioned an Ediacaran link between the rocks of the Sierras de Cordoba with those within the basement of the Paleozoic sedimentary rocks of the Sierras Australes (see **figures 10.4 and 10.5**). Second, they suggested a correlation of these systems with the Paraguay Belt. Although we have some difficulties in correlating the collisional rocks of the Sierras de Cordoba with the granitic rocks of the Sierra de La Ventana, we could consider an Ediacaran to Cambrian link between these regions, maintaining an unconfined position for them, while surrounding the Rio de La Plata Craton. This composite belt, extending to the northern Sierras Pampeanas as in **figure 10.4**, would be typically accretionary and related to the subduction of the proto-Andean Pacific oceanic lithosphere. However, we cannot accept the tentative correlation between the Sierras de Cordoba and the Paraguay Belt, for several reasons. First, the localities are very far apart (>500 km). Second, there is not any similarity of lithology, metamorphism, or structural trends. Third, and most importantly, there was no oceanic connection in the Ediacaran between the Pampean and Paraguay belts, as was mentioned above when referring to the Corumbá-Arroyo del Soldado epicontinental shelf.

10.6. Conclusions

From the discussion presented in the previous chapters, the tectonic history of West Gondwana can be summarized as follows:

1. The orogenic belts observed along the margins of the West African, Amazonian, São Francisco-Congo, and Kalahari Cratons and the Saharan Metacraton were formed during the Brasiliano-Pan African orogenic cycle through the closing of the Goiás-Pharusian Ocean, the only large domain with an oceanic lithosphere occurring among them. After a long-lived convergence, starting ca. 900 Ma, the main collisional episodes of Himalayan type that formed West Gondwana were dated roughly ca. 650 Ma and 620 Ma, when Laurentia and Baltica were still attached to the supercontinent.
2. In Ediacaran time, Gondwana was already formed as a coherent continental block with a continental crust. Laurentia and Baltica started to separate from it, as the final stage of Rodinia fragmentation, ca. 600 Ma or a little later, with the onset of a major spreading center and the formation of the Iapetus Ocean. This development was concomitant with the tectono-magmatic evolution in the Brasiliano-Pan African belts, especially along the region of the Transbrasiliano-Kandi megashear.
3. At the beginning of the Paleozoic when major plate reorganization was necessary, the subduction of the oceanic lithosphere started along the Pacific margin of the supercontinent. The accretionary Pampean Orogen in Argentina was developed in the Cambrian between 550 Ma and 520 Ma. It showed tectonic shortening and regional high- to medium-grade metamorphism, associated with voluminous subduction-related granite magmatism.

In conclusion, the Ediacaran to Cambrian timeframe (see **itens 2 and 3** above) is critical for the assessment of the possible existence of a Clymene Ocean in South America. At that time, southern Gondwana was surrounded by marginal accretionary orogens. Early Paleozoic subduction and consequent accretionary tectonics were also occurring, which provided the kinematic framework for distant plate adjustments, such as local collisions and minor tectonic adjustments in the interior of the supercontinent. Most of these changes were of an extensional character and were along the Transbrasiliano-Kandi megashear, which was established on the continental crust. Therefore, from the Ediacaran to early Cambrian time in the central region of South America, there is no geological evidence of any area with an oceanic lithosphere, ruling out the possible existence of a Clymene Ocean in that age.

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11. Conclusões

Os dados produzidos nesta Tese, em conjunto com aqueles mais relevantes já existentes na literatura geológica, referenciados ao longo dos capítulos anteriores, permitiram apresentar algumas conclusões tectônicas em escalas que vão do nível regional a supra-continental, envolvendo o Domínio Ceará Central, a Província Borborema e o supercontinente Gondwana. Grande parte dessas conclusões referem-se aos produtos (*p.e.* sedimentação, magmatismo e metamorfismo) derivados do consumo e fechamento do Oceano Goiás-Farusiano (OGF).

Dentro do Domínio Ceará Central, o estudo de proveniência de zircões detríticos nas rochas metassedimentares do Complexo Ceará, permitiu o reconhecimento de bacias sin-orogênicas relacionadas com o sistema de subducção do OGF. Além disso, as idades obtidas nesses zircões permitiram indiretamente reconhecer um importante e longo período de magmatismo entre 900-650 Ma, interpretado como indicativo do magmatismo de arco resultante do consumo da litosfera oceânica do OGF. No Domínio Médio Coreaú, a oeste do Domínio Ceará Central, as sequências quartzíticas de margem passiva da Formação São Joaquim revelaram somente a presença de zircões antigos, derivados de crosta Arqueana e Paleoproterozóica, provavelmente derivadas do craton Oeste-Africano. Os zircões detríticos da Formação Goiabeira, no Domínio Médio Coreaú, forneceram idades similares às rochas metassedimentares do Complexo Ceará, o que poderia indicar um contexto similar para a deposição dos sedimentos associados a estas unidades estratigráficas em um mesmo ambiente orogênico.

Com relação ao magmatismo relacionado ao consumo do OGF as investigações isotópicas e geocronológicas do Complexo Tamboril-Santa Quitéria, no Domínio Ceará Central, permitiram o reconhecimento de três estágios magmáticos principais. O estágio inicial condiz com a colocação de granitóides juvenis da unidade Lagoa Caiçara em um arco magmático instalado na borda oeste do continente que constitui hoje o embasamento antigo da Província Borborema Norte, durante 890-800 Ma. Esse estágio foi seguido por um aparente período de ausência de magmatismo no Domínio Ceará Central. Contudo, como colocado acima, evidências indiretas para a continuidade do magmatismo de arco entre 800-650 Ma são observadas nas idades dos zircões detríticos de bacias sin-orogênicas (bacias de *fore-arc* e *back-arc*) a leste e a oeste do Complexo Tamboril-Santa Quitéria. Isótopos de $^{18}\text{O}/^{16}\text{O}$ nesses zircões detríticos indicaram que fontes juvenis e crustais contribuíram para o magmatismo neste intervalo de tempo. O estágio de arco juvenil é seguido por um estágio de arco maturo com assinaturas híbridas entre crosta e manto, representados pelas unidades Boi e Santa Quitéria além de ortognaisses da unidade Lagoa Caiçara, entre 660-630 Ma. O último estágio é representado pelo magmatismo essencialmente crustal, que gerou os diatexitos da unidade Tamboril entre 625-618 Ma. Este magmatismo é temporalmente associado ao início da colisão continental, evidenciado pela idade do metamorfismo eclogítico entre 624-616 Ma.

O metamorfismo associado a subducção de longa duração descrito acima foi obliterado pelo metamorfismo associado a colisão continental subsequente. Sobrecrecimentos metamórficos datados entre 650-630 Ma nos zircões detríticos do Complexo Ceará poderiam estar relacionados com o metamorfismo de subducção, contudo, o erro das análises individuais não permitem conclusões sólidas. Idades precisas para o metamorfismo colisional foram obtidas em eclogitos da Zona eclogítica de Forquilha, a oeste do Complexo

Tamboril-Santa Quitéria. Idades U-Pb de domínios metamórficos distintos em combinação com a química dos zircões analisados permitiram datar o início do metamorfismo colisional em $624 \pm 7,0$ Ma, com pico em $616 \pm 6,1$ Ma. Essas idades estão em conformidade com idades de outros eclogitos que afloram no ao longo OGO no Togo ($608,7 \pm 5,8$ Ma) na região de Lato e Mali ($611,3 \pm 3,6$ Ma) no sul do Hoggar, na região de Gourma.

No âmbito da Província Borborema, os novos dados apresentados aqui, aliados aos dados geocronológicos já existentes, permitiram a construção de um modelo de evolução tectônica entre 620-550 Ma envolvendo a hipótese de interferência entre duas colisões distintas ao longo da borda oeste e sul da Província. O modelo propõe que a interação dos esforços gerados por estes dois eventos colisionais foi responsável pela geração do feixe de zonas de cisalhamento que caracterizam a Província, e por sua subsequente extrusão em direção nordeste. No modelo, o importante Lineamento Transbrasiliano é interpretado como uma zona de transferência que operou como um limite de placa transformante permitindo a aproximação e colisão da Província com o cráton São Francisco.

A história de convergência tectônica e geração dos elementos relacionados a esta convergência (*p.e.* sedimentação, magmatismo e metamorfismo) associada ao consumo e fechamento do OGF é similar ao longo do Lineamento Transbrasiliano-Kandi que se estende por mais de 5000 km da Algéria ao Brasil Central. Este corredor tectônico abrange diversos blocos, províncias e áreas orogênicas Neoproterozóicas que possuem diferentes nomes regionais. Nesta Tese, todas essas regiões alinhadas ao longo deste corredor tectônico foram agrupadas em um só orógeno: o Orógeno Gondwana Oeste (OGO). O sincronismo do metamorfismo eclogítico de ultra-alta pressão ao longo do OGO sugere que a colisão continental entre as massas continentais envolvidas na construção desse Orógeno foi simultânea por mais de 2500 km entre 610-615 Ma. Este seria o primeiro registro de uma colisão continental de magnitude similar ao Orógeno Himalaiano (≈ 2500 km) envolvendo subducção continental profunda em pressões superiores a 2,7 GPa (>90 km de profundidade). A edificação e erosão dessas Megamontanhas, com elevações superiores a 8000 km (em analogia com o Orógeno Himalaiano) podem ter contribuído para a elevada taxa de sedimentação na transição entre os períodos Ediacariano-Cambriano, e consequentemente para o aumento da pO_2 na atmosfera e oceanos e dos nutrientes (sedimentos) bio-disponíveis necessários para a radiação e manutenção da Vida.

Por fim, a colisão simultânea entre as massas continentais (crátons) responsáveis para a formação deste extenso orógeno sugerem que o supercontinente Gondwana já estava consolidado ao longo do OGO no final do período Ediacariano. Esta perspectiva contrasta com outras que sugerem a presença de um amplo oceano Cambriano na América do Sul. A ocorrência de bacias marinhas Cambrianas (e.g. Faixa Paraguai) nas proximidades do OGO é um fato inegável, contudo a deposição das mesmas devem ter ocorrido, em um sentido amplo, como *forelands* associados ao desmonte OGO. A deformação destas sequências ocorreu em consequência da compressão final ao longo do OGO no Cambriano.

A geração de novos dados estimula a pesquisa científica e abre espaço para novas interpretações. Muitas lacunas ainda estão por serem preenchidas e a reprodutibilidade das interpretações geradas nesta Tese precisam ser testadas em suas diversas escalas. Dentro do Domínio Ceará Central, a estrutura e evolução tectônica associada a exumação das rochas eclogíticas devem ser investigadas em detalhe. O posicionamento

temporal das diversas gerações de leucossomas relacionados ao estágio colisional ainda estão por serem definidas e carecem de melhor investigação. A validade do modelo proposto para Província Borborema com base na interferência entre duas colisões ainda carece de melhores dados acerca do período em que as diversas zonas de cisalhamentos foram operantes. Por fim, dentro do contexto do OGO, o timing dos eclogitos expostos no extremo norte na região do Hoggar ainda esbarra em certa deficiência de dados, bem como a extensão do mesmo a sul do Brasil Central. Essas, entre outras, fazem parte de uma gama de questões que ainda necessitam de uma resposta mais sólida.

ANEXO I

(análises U-Th-Pb LA-ICP-MS em zircões detriticos)

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-36	2.1	C	PMhll	10.1457	0.1546	0.4399	0.0046	2.2731	0.0238	0.1661	0.0033	0.4728	1.2199	2.350	0.021	2.448	0.014	2.517	0.034	96	93
DKE-36	3.1	R	PMoz	5.2082	0.0713	0.3152	0.0023	3.1729	0.0228	0.1197	0.0023	0.3406	0.7438	1.766	0.011	1.854	0.012	1.952	0.033	95	90
DKE-36	4.1	C	PMoz	7.4244	0.1009	0.3821	0.0027	2.6174	0.0184	0.1408	0.0027	0.3481	0.6591	2.086	0.013	2.164	0.012	2.233	0.032	96	93
DKE-36	5.1	C	PMoz	5.9942	0.0809	0.3359	0.0023	2.9767	0.0206	0.1290	0.0024	0.3195	0.5338	1.867	0.011	1.975	0.012	2.081	0.032	95	90
DKE-36	6.1	R	PMoz	6.7296	0.0903	0.3605	0.0025	2.7742	0.0192	0.1341	0.0025	0.3316	0.4958	1.984	0.012	2.077	0.012	2.149	0.032	96	92
DKE-36	7.1	C	PMoz	7.0678	0.0956	0.3698	0.0027	2.7044	0.0194	0.1387	0.0026	0.2808	0.3851	2.028	0.012	2.120	0.012	2.207	0.032	96	92
DKE-36	8.1	C	PMoz	6.2597	0.0830	0.3515	0.0024	2.8451	0.0195	0.1290	0.0024	2.3510	2.9045	1.942	0.011	2.013	0.012	2.081	0.032	96	93
DKE-36	9.1	C	PMhml	21.7037	0.2899	0.5987	0.0043	1.6702	0.0121	0.2604	0.0048	0.6774	0.7130	3.025	0.017	3.171	0.013	3.246	0.028	95	93
DKE-36	14.1	C	PMoz	8.4129	0.0681	0.4032	0.0026	2.4804	0.0158	0.1497	0.0016	0.4044	0.2300	2.184	0.012	2.277	0.007	2.338	0.018	96	93
DKE-36	16.1	C	PMhll	8.0697	0.0619	0.3883	0.0024	2.5753	0.0156	0.1515	0.0015	0.5023	0.3087	2.115	0.011	2.239	0.007	2.359	0.017	94	90
DKE-36	20.1	C	PMoz	9.3820	0.0713	0.4163	0.0026	2.4021	0.0149	0.1637	0.0017	0.2165	0.1621	2.244	0.012	2.376	0.007	2.492	0.018	94	90
DKE-36	23.1	C	PMhhl	9.6530	0.1441	0.4441	0.0020	2.2517	0.0099	0.1571	0.0022	0.8549	0.4019	2.369	0.009	2.402	0.014	2.421	0.024	99	98
DKE-36	24.1	C	PMhhl	14.7124	0.2143	0.5383	0.0023	1.8575	0.0079	0.1993	0.0027	0.4601	0.2255	2.777	0.010	2.797	0.014	2.826	0.022	99	98
DKE-36	25.1N	C	PMoz	7.2543	0.1073	0.3909	0.0017	2.5581	0.0112	0.1355	0.0018	0.2343	0.1200	2.127	0.008	2.143	0.013	2.167	0.023	99	98
DKE-36	16.1	C	PMhll	7.9536	0.1159	0.4101	0.0015	2.4382	0.0087	0.1411	0.0019	0.5768	0.3572	2.216	0.007	2.226	0.013	2.236	0.023	100	99
DKE-36	29.1	C	PMoz	9.0180	0.1311	0.4287	0.0015	2.3325	0.0080	0.1544	0.0021	0.2105	0.1390	2.300	0.007	2.340	0.013	2.391	0.023	98	96
DKE-36	30.1	C	PMoz	7.4594	0.1090	0.3928	0.0013	2.5459	0.0082	0.1390	0.0019	0.2402	0.1683	2.136	0.006	2.168	0.013	2.211	0.023	99	97
DKE-36	31.1	C	PMhhl	8.7154	0.1315	0.4232	0.0018	2.3628	0.0103	0.1497	0.0021	0.1115	0.0833	2.275	0.008	2.309	0.014	2.338	0.024	99	97
DKE-36	32.1	R	PMoz	7.2310	0.1058	0.3880	0.0013	2.5771	0.0085	0.1369	0.0019	0.1984	0.1584	2.114	0.006	2.140	0.013	2.184	0.023	99	97
DKE-36	33.1	C	PMhml	6.5555	0.0955	0.3709	0.0011	2.6959	0.0081	0.1291	0.0018	0.1505	0.1292	2.034	0.005	2.053	0.013	2.083	0.023	99	98
DKE-36	34.1	C	PMoz	9.3857	0.1368	0.4384	0.0014	2.2812	0.0074	0.1578	0.0022	0.1432	0.1328	2.343	0.006	2.376	0.013	2.429	0.023	99	96
DKE-36	35.1	C	PMoz	9.2921	0.1508	0.4374	0.0037	2.2863	0.0192	0.1564	0.0025	0.2191	0.0573	2.339	0.016	2.367	0.015	2.414	0.028	99	97
DKE-36	36.1	C	PMoz	7.2500	0.1187	0.3869	0.0033	2.5847	0.0219	0.1358	0.0022	0.2351	0.0604	2.108	0.015	2.143	0.015	2.171	0.028	98	97
DKE-36	37.1	R	PMoz	22.8947	0.3713	0.6391	0.0055	1.5646	0.0135	0.2630	0.0043	0.1344	0.0339	3.186	0.022	3.222	0.016	3.262	0.025	99	98
DKE-36	38.1	C	PMhml	7.3934	0.1197	0.3869	0.0033	2.5846	0.0218	0.1400	0.0023	0.2854	0.0705	2.108	0.015	2.160	0.014	2.223	0.028	98	95
DKE-36	39.1	C	PMhll	9.2165	0.1483	0.4329	0.0036	2.3097	0.0194	0.1562	0.0025	0.2687	0.0652	2.319	0.016	2.360	0.015	2.412	0.027	98	96
DKE-36	40.1	C	PMcz	9.0033	0.1458	0.4359	0.0037	2.2941	0.0194	0.1506	0.0024	0.3116	0.0743	2.332	0.017	2.338	0.015	2.348	0.028	100	99
DKE-36	41.1	C	PMhll	6.0208	0.0970	0.3492	0.0029	2.8637	0.0241	0.1265	0.0020	0.3098	0.0727	1.931	0.014	1.979	0.014	2.049	0.028	98	94
DKE-36	42.1	R	PMhml	9.0721	0.1458	0.4365	0.0037	2.2911	0.0196	0.1531	0.0025	0.1120	0.0258	2.335	0.017	2.345	0.015	2.377	0.027	100	98
DKE-36	43.1	C	PMoz	7.1881	0.1160	0.3846	0.0032	2.6002	0.0218	0.1358	0.0022	0.1505	0.0341	2.098	0.015	2.135	0.014	2.171	0.027	98	97
DKE-36	44.1	C	PMoz	12.1906	0.1950	0.4779	0.0040	2.0923	0.0176	0.1872	0.0030	0.3059	0.0681	2.518	0.018	2.619	0.015	2.721	0.027	96	93
DKE-36	45.1	C	PMoz	6.5003	0.1042	0.3666	0.0031	2.7278	0.0231	0.1299	0.0021	0.2249	0.0492	2.013	0.015	2.046	0.014	2.093	0.027	98	96
DKE-36	46.1	C	PMcz	16.1074	0.2573	0.5387	0.0046	1.8563	0.0157	0.2188	0.0035	0.1080	0.0233	2.778	0.019	2.883	0.015	2.978	0.025	96	93
DKE-36	47.1	R	PMoz	13.6501	0.2192	0.5076	0.0044	1.9701	0.0172	0.1970	0.0031	0.1542	0.0327	2.646	0.019	2.726	0.015	2.806	0.026	97	94
DKE-36	48.1	C	PMhhl	6.9477	0.1377	0.3753	0.0048	2.6648	0.0340	0.1346	0.0003	0.1869	0.0404	2.054	0.022	2.105	0.017	2.155	0.004	98	95
DKE-36	49.1	C	PMhml	6.5867	0.1294	0.3619	0.0045	2.7630	0.0347	0.1335	0.0002	0.3624	0.0781	1.991	0.022	2.058	0.017	2.141	0.002	97	93
DKE-36	50.1	R	PMoz	6.2243	0.1220	0.3538	0.0044	2.8266	0.0354	0.1289	0.0002	0.2192	0.0472	1.953	0.021	2.008	0.017	2.081	0.002	97	94
DKE-36	51.1	R	PMoz	6.3765	0.1250	0.3583	0.0045	2.7909	0.0350	0.1301	0.0002	0.2393	0.0515	1.974	0.021	2.029	0.017	2.096	0.002	97	94
DKE-36	52.1	C	PMhml	7.2847	0.1427	0.3860	0.0048	2.5908	0.0325	0.1376	0.0002	0.6832	0.1470	2.104	0.023	2.147	0.017	2.193	0.002	98	96
DKE-36	53.1	C	PMhll	8.6721	0.1689	0.4117	0.0051	2.4292	0.0303	0.1540	0.0001	0.2310	0.0497	2.222	0.023	2.304	0.018	2.387	0.001	96	93
DKE-36	54.1	C	PMhml	5.8504	0.1143	0.3436	0.0043	2.9101	0.0363	0.1246	0.0002	0.1714	0.0369	1.904	0.021	1.954	0.017	2.021	0.002	97	94
DKE-36	56.1	C	PMoz	12.0933	0.2357	0.4751	0.0060	2.1048	0.0264	0.1859	0.0002	0.4854	0.1043	2.506	0.026	2.612	0.018	2.709	0.002	96	92
DKE-36	57.1	C	PMhml	11.0054	0.2145	0.4527	0.0056	2.2092	0.0275	0.1759	0.0002	0.3351	0.0720	2.407	0.025	2.524	0.018	2.615	0.002	95	92
DKE-36	58.1	C	PMhml	6.8167	0.1323	0.3764	0.0047	2.6569	0.0330	0.1318	0.0002	0.1568	0.0337	2.059	0.022	2.088	0.017	2.119	0.002	99	97
DKE-36	59.1	C	PMcz	7.0999	0.1382	0.3902	0.0049	2.5627	0.0321	0.1331	0.0003	0.3060	0.0657	2.124	0.023	2.124	0.017	2.136	0.004	100	99
DKE-36	62.1	R	PMoz	6.9130	0.0867	0.3752	0.0057	2.6649	0.0406	0.1354	0.0016	0.1337	0.0190	2.054	0.027	2.100	0.011	2.165	0.020	98	95
DKE-36	64.1	R	PMoz	6.3559	0.0803	0.3630	0.0055	2.7551	0.0419	0.1280	0.0015	0.3522	0.0489	1.996	0.026	2.026	0.011	2.068	0.020	99	97
DKE-36	66.1	C	PMoz	6.7597	0.0843	0.3651	0.0056	2.7393	0.0417	0.1361	0.0016	0.1903	0.0260	2.006	0.026	2.080	0.011	2.174	0.020	96	92
DKE-36	68.1	C	PMoz																		

Sample	Spot #	Spot pos.	Interp. Cl.	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-39	1.1	C	PMoz	7.8151	0.0576	0.4057	0.0024	2.4650	0.0146	0.1408	0.0006	0.4847	0.1794	2.195	0.011	2.210	0.007	2.232	0.008	99	98
DKE-39	2.1	C	PMoz	9.5278	0.0711	0.4373	0.0028	2.2870	0.0145	0.1600	0.0008	0.3445	0.1309	2.338	0.012	2.390	0.007	2.453	0.009	98	95
DKE-39	3.1	C	PMoz	7.4750	0.0563	0.3873	0.0023	2.5822	0.0152	0.1399	0.0007	0.3078	0.1198	2.110	0.011	2.170	0.007	2.222	0.008	97	95
DKE-39	4.1	C	PMhml	6.4733	0.0734	0.3716	0.0046	2.6908	0.0335	0.1275	0.0006	0.1833	0.0689	2.037	0.022	2.042	0.010	2.061	0.008	100	99
DKE-39	6.1	C	PMhml	9.8507	0.1132	0.4577	0.0057	2.1847	0.0271	0.1572	0.0009	0.3086	0.1077	2.430	0.025	2.421	0.011	2.422	0.010	100	100
DKE-39	7.1	C	PMhhl	6.6072	0.0877	0.3696	0.0052	2.7057	0.0380	0.1306	0.0012	0.3786	0.1278	2.027	0.024	2.060	0.012	2.104	0.016	98	96
DKE-39	8.1	C	PMhhl	9.8078	0.1209	0.4511	0.0060	2.2170	0.0296	0.1604	0.0012	0.2290	0.0749	2.400	0.027	2.417	0.011	2.456	0.013	99	98
DKE-39	9.1	C	PMhhl	5.7417	0.0734	0.3351	0.0046	2.9843	0.0408	0.1258	0.0010	0.3417	0.1077	1.863	0.022	1.938	0.011	2.038	0.014	96	91
DKE-39	12.1	C	PMoz	7.4342	0.0838	0.3965	0.0049	2.5223	0.0314	0.1367	0.0007	0.2905	0.0834	2.153	0.023	2.165	0.010	2.183	0.009	99	99
DKE-39	13.1	C	PMoz	7.3505	0.0859	0.4014	0.0050	2.4912	0.0311	0.1335	0.0007	0.2302	0.0642	2.176	0.023	2.155	0.010	2.140	0.009	101	102
DKE-39	14.1	C	PMoz	6.7924	0.0790	0.3711	0.0047	2.6946	0.0342	0.1337	0.0008	0.0777	0.0214	2.035	0.022	2.085	0.010	2.144	0.010	98	95
DKE-39	15.1	C	PMhml	9.1477	0.1024	0.4170	0.0052	2.3983	0.0298	0.1591	0.0008	0.3910	0.1030	2.247	0.024	2.353	0.010	2.443	0.008	95	92
DKE-39	17.1	C	PMoz	7.2822	0.0941	0.3971	0.0041	2.5182	0.0257	0.1342	0.0017	0.4190	0.3716	2.156	0.019	2.147	0.011	2.151	0.022	100	100
DKE-39	18.1	C	PMoz	6.2535	0.0814	0.3473	0.0034	2.8795	0.0283	0.1321	0.0017	0.4461	0.4051	1.922	0.016	2.012	0.011	2.123	0.022	96	90
DKE-39	19.1	C	PMoz	7.0608	0.0878	0.3821	0.0036	2.6169	0.0248	0.1345	0.0016	0.5728	0.5330	2.086	0.017	2.119	0.011	2.154	0.020	98	97
DKE-39	20.1	C	PMoz	6.7407	0.0836	0.3685	0.0035	2.7137	0.0255	0.1351	0.0016	1.7326	1.6537	2.022	0.016	2.078	0.011	2.162	0.020	97	94
DKE-39	21.1	R	PMoz	6.8463	0.0846	0.3665	0.0034	2.7283	0.0254	0.1359	0.0016	0.7831	0.7672	2.013	0.016	2.092	0.011	2.172	0.020	96	93
DKE-39	24.1	C	PMoz	8.3184	0.1053	0.4116	0.0040	2.4294	0.0234	0.1487	0.0017	0.5111	0.5437	2.222	0.018	2.266	0.011	2.327	0.020	98	95
DKE-39	26.1	C	PMoz	9.4631	0.1172	0.4268	0.0040	2.3428	0.0222	0.1641	0.0019	0.4860	0.5484	2.291	0.018	2.384	0.011	2.496	0.020	96	92
DKE-39	30.1	C	PMcz	7.3867	0.0685	0.3788	0.0042	2.6398	0.0291	0.1436	0.0008	1.2884	0.4423	2.071	0.019	2.159	0.008	2.267	0.010	96	91
DKE-39	33.1	C	PMoz	8.6041	0.0851	0.4064	0.0046	2.4604	0.0281	0.1585	0.0008	0.6917	0.2618	2.199	0.021	2.297	0.009	2.436	0.009	96	90
DKE-39	34.1	C	PMoz	4.7094	0.0465	0.3067	0.0034	3.2607	0.0362	0.1118	0.0007	0.9415	0.3690	1.724	0.017	1.769	0.008	1.831	0.012	97	94
DKE-39	36.1	C	PMoz	7.5884	0.0680	0.3924	0.0043	2.5484	0.0278	0.1416	0.0007	0.3128	0.1313	2.134	0.020	2.183	0.008	2.242	0.009	98	95
DKE-39	37.1	C	PMoz	8.3263	0.0828	0.4214	0.0048	2.3728	0.0270	0.1436	0.0008	0.3539	0.1552	2.267	0.022	2.267	0.009	2.267	0.009	100	100
DKE-39	38.1	C	PMoz	12.8425	0.1158	0.5057	0.0055	1.9777	0.0217	0.1887	0.0011	0.7399	0.3377	2.638	0.024	2.668	0.008	2.734	0.009	99	96
DKE-39	40.1	C	PMoz	7.7245	0.0698	0.3835	0.0043	2.6078	0.0291	0.1473	0.0009	0.3797	0.1891	2.092	0.020	2.199	0.008	2.310	0.010	95	91
Zircons not included on related diagrams																					
DKE-39	11.1	R	PMoz	5.4929	0.0391	0.3145	0.0017	3.1794	0.0175	0.1287	0.0005	0.3038	0.1484	1.763	0.008	1.899	0.006	2.077	0.007	93	85
DKE-39	5.1	C	PMoz	6.2390	0.0692	0.3442	0.0042	2.9055	0.0357	0.1333	0.0006	0.2142	0.0776	1.907	0.020	2.010	0.010	2.138	0.008	95	89
DKE-39	10.1	C	PMoz	4.4744	0.0533	0.2612	0.0033	3.8282	0.0487	0.1245	0.0006	0.1416	0.0434	1.496	0.017	1.726	0.010	2.021	0.009	87	74
DKE-39	16.1	C	PMoz	6.8901	0.0816	0.3509	0.0044	2.8497	0.0360	0.1426	0.0008	0.3688	0.0946	1.939	0.021	2.097	0.010	2.255	0.010	92	86
DKE-39	22.1	C	PMoz	3.7665	0.0495	0.2280	0.0022	4.3857	0.0428	0.1213	0.0015	0.4538	0.4566	1.324	0.012	1.586	0.010	1.975	0.021	84	67
DKE-39	23.1	C	PMoz	5.8866	0.0767	0.3174	0.0033	3.1503	0.0331	0.1361	0.0016	0.4639	0.4779	1.777	0.016	1.959	0.011	2.174	0.020	91	82
DKE-39	25.1	C	PMoz	8.5016	0.1049	0.3928	0.0037	2.5456	0.0238	0.1587	0.0018	0.4579	0.5015	2.136	0.017	2.286	0.011	2.438	0.019	93	88
DKE-39	27.1	C	PMoz	4.6024	0.0591	0.2916	0.0029	3.4296	0.0339	0.1150	0.0014	0.7229	0.8414	1.649	0.014	1.750	0.011	1.882	0.021	94	88
DKE-39	28.1	C	PMoz	4.5745	0.0593	0.2647	0.0027	3.7784	0.0378	0.1265	0.0015	0.3824	0.4595	1.514	0.014	1.745	0.011	2.048	0.020	87	74
DKE-39	29.1	C	PMoz	4.4473	0.0568	0.2530	0.0025	3.9529	0.0387	0.1280	0.0015	0.4653	0.5778	1.454	0.013	1.721	0.011	2.069	0.020	84	70
DKE-39	31.1	C	PMoz	5.4699	0.0531	0.2664	0.0031	3.7542	0.0438	0.1483	0.0008	0.5110	0.1812	1.522	0.016	1.896	0.008	2.323	0.009	80	66
DKE-39	32.1	C	PMoz	2.8931	0.0564	0.1937	0.0032	5.1633	0.0854	0.1119	0.0017	0.5233	0.1939	1.141	0.017	1.380	0.015	1.832	0.027	83	62
DKE-39	35.1	C	PMoz	8.9778	0.0818	0.4053	0.0047	2.4673	0.0283	0.1639	0.0010	0.4405	0.1791	2.193	0.021	2.336	0.008	2.494	0.011	94	88
DKE-39	39.1	C	PMhhl	6.1995	0.0956	0.3358	0.0040	2.9784	0.0358	0.1358	0.0020	0.6318	0.3014	1.866	0.019	2.004	0.013	2.170	0.025	93	86
DKE-39	42.1	C	PMoz	4.1320	0.0393	0.2779	0.0031	3.5988	0.0396	0.1092	0.0007	0.5416	0.2961	1.581	0.015	1.661	0.008	1.790	0.011	95	88
DKE-39	41.1	C	PMoz	2.5789	0.0277	0.1621	0.0018	6.1706	0.0696	0.1169	0.0007	0.2135	0.1113	0.968	0.010	1.295	0.008	1.910	0.010	75	51

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-41	1.1	C	PMoz	18.6487	0.2434	0.5628	0.0022	1.7770	0.0070	0.2428	0.0026	0.1674	0.0381	2.878	0.009	3.024	0.013	3.140	0.016	95	92
DKE-41	2.1	C	PMoz	6.1464	0.0797	0.3489	0.0013	2.8659	0.0105	0.1288	0.0014	0.2398	0.0547	1.929	0.006	1.997	0.011	2.079	0.018	97	93
DKE-41	3.1	C	PMoz	8.4775	0.1116	0.4078	0.0016	2.4520	0.0096	0.1509	0.0016	0.1914	0.0438	2.205	0.007	2.284	0.012	2.351	0.018	97	94
DKE-41	5.1	C	PMoz	6.9916	0.0903	0.3719	0.0014	2.6888	0.0100	0.1375	0.0015	0.2274	0.0522	2.038	0.007	2.110	0.011	2.192	0.018	97	93
DKE-41	6.1	C	PMoz	7.1826	0.0939	0.3843	0.0014	2.6024	0.0097	0.1357	0.0015	0.2798	0.0644	2.096	0.007	2.134	0.012	2.169	0.018	98	97
DKE-41	8.1	C	PMoz	4.5499	0.0601	0.2999	0.0012	3.3346	0.0133	0.1096	0.0012	0.2562	0.0593	1.691	0.006	1.740	0.011	1.795	0.019	97	94
DKE-41	9.1	C	PMcz	7.5060	0.1004	0.3832	0.0016	2.6098	0.0112	0.1414	0.0015	0.8764	0.2030	2.091	0.008	2.174	0.012	2.241	0.019	96	93
DKE-41	10.1	C	PMoz	7.5039	0.0982	0.3851	0.0014	2.5966	0.0098	0.1409	0.0015	0.3214	0.0746	2.100	0.007	2.173	0.012	2.235	0.018	97	94
DKE-41	11.1	C	PMoz	7.1755	0.0933	0.3787	0.0014	2.6403	0.0095	0.1370	0.0015	0.2077	0.0483	2.070	0.006	2.133	0.012	2.186	0.018	97	95
DKE-41	13.1	C	PMcz	5.6926	0.0749	0.3312	0.0012	3.0189	0.0113	0.1239	0.0013	0.2590	0.0605	1.844	0.006	1.930	0.011	2.012	0.018	96	92
DKE-41	14.1	C	PMoz	7.5535	0.2225	0.3993	0.0084	2.5044	0.0528	0.1395	0.0015	0.6519	0.0492	2.166	0.039	2.179	0.026	2.217	0.019	99	98
DKE-41	18.1	C	PMoz	7.3781	0.2022	0.3931	0.0076	2.5437	0.0492	0.1364	0.0014	0.2248	0.0169	2.137	0.035	2.158	0.024	2.179	0.018	99	98
DKE-41	20.1	C	PMoz	7.6578	0.2088	0.3999	0.0077	2.5003	0.0484	0.1388	0.0015	0.4020	0.0303	2.169	0.036	2.192	0.024	2.209	0.018	99	98
DKE-41	21.1	C	PMoz	17.8265	0.4846	0.5948	0.0115	1.6812	0.0324	0.2188	0.0023	0.0722	0.0056	3.009	0.046	2.980	0.026	2.978	0.017	101	101
DKE-41	22.1	C	PMoz	7.5259	0.2048	0.3956	0.0076	2.5277	0.0487	0.1392	0.0015	0.4047	0.0305	2.149	0.035	2.176	0.024	2.213	0.018	99	97
DKE-41	24.1	C	PMoz	7.5982	0.2061	0.3948	0.0076	2.5328	0.0487	0.1399	0.0015	0.3370	0.0254	2.145	0.035	2.185	0.024	2.221	0.018	98	97
DKE-41	25.1	C	PMoz	7.3809	0.1996	0.3985	0.0077	2.5096	0.0483	0.1353	0.0014	0.1601	0.0121	2.162	0.035	2.159	0.024	2.164	0.018	100	100
DKE-41	26.1	C	PMoz	6.0556	0.1637	0.3503	0.0067	2.8549	0.0550	0.1257	0.0013	0.0804	0.0062	1.936	0.032	1.984	0.023	2.037	0.018	98	95
DKE-41	27.1	C	PMoz	7.0496	0.0779	0.3781	0.0052	2.6451	0.0366	0.1360	0.0015	0.2904	0.0382	2.067	0.024	2.118	0.010	2.173	0.018	98	95
DKE-41	28.1	C	PMoz	9.2314	0.0956	0.4211	0.0056	2.3750	0.0316	0.1596	0.0016	0.2610	0.0338	2.265	0.025	2.361	0.009	2.449	0.017	96	93
DKE-41	29.1	C	PMoz	4.4728	0.0532	0.3026	0.0041	3.3046	0.0453	0.1078	0.0012	0.2818	0.0365	1.704	0.020	1.726	0.010	1.766	0.020	99	97
DKE-41	30.1	C	PMoz	7.2451	0.0770	0.3875	0.0052	2.5804	0.0349	0.1354	0.0014	0.4126	0.0529	2.111	0.024	2.142	0.009	2.165	0.018	99	98
DKE-41	31.1	C	PMoz	7.4064	0.0774	0.3916	0.0053	2.5539	0.0344	0.1379	0.0014	0.2138	0.0273	2.130	0.024	2.162	0.009	2.197	0.018	99	97
DKE-41	32.1	C	PMoz	12.4215	0.1737	0.5019	0.0080	1.9923	0.0318	0.1839	0.0019	0.2441	0.0309	2.622	0.034	2.637	0.013	2.690	0.017	99	97
DKE-41	33.1	C	PMoz	7.9459	0.0883	0.4100	0.0057	2.4391	0.0338	0.1409	0.0016	0.3722	0.0470	2.215	0.026	2.225	0.010	2.234	0.019	100	99
DKE-41	34.1	C	PMoz	5.8746	0.0606	0.3444	0.0046	2.9039	0.0390	0.1236	0.0012	0.1140	0.0142	1.908	0.022	1.957	0.009	2.007	0.017	97	95
DKE-41	35.1	C	PMoz	7.3845	0.0786	0.3871	0.0054	2.5830	0.0357	0.1377	0.0014	0.2957	0.0367	2.110	0.025	2.159	0.009	2.195	0.018	98	96
DKE-41	36.1	C	PMoz	8.4913	0.0889	0.4128	0.0056	2.4228	0.0327	0.1492	0.0015	0.3860	0.0476	2.227	0.025	2.285	0.009	2.332	0.017	97	96
DKE-41	38.1	C	PMoz	6.7709	0.0700	0.3802	0.0051	2.6300	0.0355	0.1302	0.0013	0.3966	0.0482	2.077	0.024	2.082	0.009	2.098	0.018	100	99
DKE-41	39.1	C	PMoz	7.3857	0.0770	0.3915	0.0053	2.5540	0.0344	0.1372	0.0014	0.3539	0.0428	2.130	0.024	2.159	0.009	2.188	0.017	99	97
DKE-41	41.1	C	PMoz	6.1136	0.1153	0.3483	0.0056	2.8707	0.0462	0.1275	0.0009	0.2160	0.0635	1.927	0.027	1.992	0.016	2.062	0.012	97	93
DKE-41	42.1	C	PMoz	5.9591	0.1119	0.3501	0.0057	2.8566	0.0462	0.1238	0.0009	0.2345	0.0685	1.935	0.027	1.970	0.016	2.010	0.012	98	96
DKE-41	43.1	C	PMoz	12.1116	0.2276	0.4916	0.0079	2.0340	0.0329	0.1792	0.0013	0.4169	0.1209	2.578	0.034	2.613	0.017	2.647	0.012	99	97
DKE-41	44.1	C	PMoz	5.9885	0.1121	0.3411	0.0055	2.9317	0.0473	0.1272	0.0009	0.2562	0.0738	1.892	0.026	1.974	0.016	2.058	0.012	96	92
DKE-41	46.1	C	PMoz	6.3695	0.1563	0.3546	0.0075	2.8198	0.0599	0.1322	0.0011	0.1219	0.0352	1.957	0.036	2.028	0.021	2.125	0.014	96	92
DKE-41	48.1	C	PMoz	6.3401	0.1183	0.3605	0.0058	2.7737	0.0444	0.1277	0.0009	0.3632	0.1018	1.985	0.027	2.024	0.016	2.064	0.012	98	96
DKE-41	49.1	C	PMoz	7.5276	0.1401	0.3990	0.0064	2.5062	0.0403	0.1369	0.0010	0.4069	0.1132	2.164	0.029	2.176	0.017	2.184	0.012	99	99
DKE-41	50.1	C	PMhml	7.2043	0.1339	0.3863	0.0062	2.5885	0.0415	0.1358	0.0010	0.4712	0.1303	2.106	0.029	2.137	0.016	2.171	0.012	99	97
DKE-41	51.1	R	PMoz	16.9115	0.3130	0.5831	0.0094	1.7151	0.0276	0.2115	0.0015	0.3727	0.1024	2.961	0.038	2.930	0.018	2.923	0.011	101	101
DKE-41	52.1	C	PMoz	7.4960	0.1384	0.3910	0.0062	2.5576	0.0408	0.1398	0.0010	0.3741	0.1021	2.127	0.029	2.173	0.016	2.220	0.012	98	96
DKE-41	53.1	C	PMhill	9.8783	0.0948	0.4428	0.0065	2.2584	0.0330	0.1620	0.0014	0.1314	0.0373	2.363	0.029	2.423	0.009	2.474	0.014	98	96
DKE-41	54.1	C	PMoz	7.5753	0.0754	0.4090	0.0061	2.4447	0.0363	0.1356	0.0012	0.5051	0.1469	2.211	0.028	2.182	0.009	2.168	0.015	101	102
DKE-41	55.1	C	PMoz	7.2958	0.0760	0.3826	0.0057	2.6136	0.0389	0.1369	0.0012	0.3053	0.0909	2.088	0.027	2.148	0.009	2.184	0.015	97	96
DKE-41	56.1	C	PMoz	7.1576	0.0699	0.3866	0.0057	2.5863	0.0382	0.1352	0.0012	0.7114	0.2171	2.107	0.027	2.131	0.009	2.163	0.015	99	97
DKE-41	57.1	C	PMoz	9.9500	0.0954	0.4507	0.0066	2.2187	0.0325	0.1618	0.0014	0.4603	0.1440	2.398	0.029	2.430	0.009	2.472	0.015	99	97
DKE-41	58.1	C	PMoz	20.5761	0.1970	0.6166	0.0090	1.6218	0.0238	0.2444	0.0021	0.5167	0.1658	3.096	0.036	3.119	0.009	3.150	0.013	99	98
DKE-41	59.1	C	PMoz	7.7585	0.0745	0.3947	0.0058	2.5336	0.0372	0.1442	0.0012	0.2342	0.0769	2.145	0.027	2.203	0.009	2.274	0.014	97	94
DKE-41	60.1	C	PMoz	5.9932	0.0626	0.3416	0.0051	2.9278	0.0441	0.1276	0.0011	0.2337	0.0791	1.894	0.025	1.975	0.009	2.063	0.015	96	92
DKE-41	62.1	C	PMoz	4.5270	0.0455	0.3117	0.0046	3.2085	0.0475	0.1066	0.0009	0.3214	0.1152	1.749	0.023	1.736	0.008	1.745	0.016	101	100
DKE-41	63.1	C	PMoz	7.1284	0.0749	0.3800	0.0057	2.6314	0.0395	0.1372	0.0012	0.2892	0.1067	2.076	0.027	2.128	0.009	2.188	0.015	98	95
DKE-41	64.1	C	PMoz	7.2273	0.0706	0.3843	0.0057	2.6021	0.0384	0.1369	0.0012	0.2210	0.0841	2.096	0.026	2.140	0.009	2.185	0.015	98	96
DKE-41	65.1	C	PMoz	7.0716	0.0711	0.3688	0.0055	2.7113	0.0407	0.1387	0.0012	0.3570	0.1401	2.024	0.026	2.120	0.009	2.207	0.014	95	92
Zircon not used on the related diagrams																					
DKE-41	4.1	C	PMoz	4.1952	0.0613	0.1962	0.0018	5.0961	0.0460	0.1582	0.0017	0.2448	0.0561	1.155	0.010	1.673	0.012	2.433	0.019	69	47
DKE-41	7.1	C	PMoz	5.6565	0.0738	0.3241	0.0013	3.0855	0.0122	0.1264	0.0013	0.4205	0.0969	1.810	0.006	1.925	0.011	2.047	0.018	94	88
DKE-41	1																				

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-30	1.1	C	PMoz	1.7891	0.0350	0.1749	0.0037	5.7174	0.1224	0.0742	0.0007	0.6221	0.5947	1.039	0.021	1.041	0.013	1.049	0.019	100	99
DKE-30	2.1	C	PMzc	1.2550	0.0255	0.1366	0.0031	7.3194	0.1680	0.0679	0.0007	1.0473	0.9417	0.826	0.018	0.826	0.011	0.862	0.021	100	96
DKE-30	3.1	C	PMoz	6.7753	0.1266	0.3879	0.0082	2.5779	0.0548	0.1279	0.0010	0.4908	0.4164	2.113	0.038	2.083	0.016	2.067	0.014	101	102
DKE-30	4.1	C	PMcz	10.6234	0.1990	0.4840	0.0103	2.0661	0.0439	0.1612	0.0013	0.8270	0.6643	2.545	0.045	2.491	0.017	2.465	0.014	102	103
DKE-30	5.1	C	PMoz	2.1247	0.0408	0.1906	0.0041	5.2477	0.1133	0.0791	0.0008	0.4828	0.3683	1.124	0.022	1.157	0.013	1.177	0.021	97	96
DKE-30	6.1	C	PMoz	6.9900	0.1304	0.3879	0.0082	2.5781	0.0548	0.1317	0.0011	0.4850	0.3522	2.113	0.038	2.110	0.016	2.118	0.014	100	100
DKE-30	7.1	C	PMhhl	1.3806	0.0341	0.1468	0.0039	6.8139	0.1822	0.0662	0.0012	0.9776	0.6847	0.883	0.022	0.881	0.014	0.809	0.039	100	109
DKE-30	8.1	C	PMoz	1.1550	0.0218	0.1291	0.0027	7.7476	0.1647	0.0658	0.0005	0.6942	0.4600	0.783	0.016	0.780	0.010	0.795	0.017	100	98
DKE-30	9.1	C	PMoz	1.0476	0.0202	0.1214	0.0026	8.2394	0.1758	0.0633	0.0006	0.3881	0.2465	0.738	0.015	0.728	0.010	0.715	0.020	101	103
DKE-30	10.1	C	PMoz	1.0464	0.0196	0.1198	0.0025	8.3503	0.1756	0.0641	0.0005	0.1734	0.1056	0.729	0.014	0.727	0.010	0.741	0.017	100	98
DKE-30	11.1	C	PMoz	1.3015	0.0243	0.1423	0.0030	7.0270	0.1483	0.0673	0.0006	0.5242	0.3071	0.858	0.017	0.846	0.011	0.843	0.018	101	102
DKE-30	13.1	C	PMoz	1.1870	0.0220	0.1323	0.0008	7.5568	0.0470	0.0658	0.0009	0.7795	0.4534	0.801	0.005	0.795	0.010	0.795	0.030	101	101
DKE-30	14.1	C	PMhml	5.3101	0.0959	0.3368	0.0014	2.9694	0.0126	0.1157	0.0016	1.3487	0.7787	1.871	0.007	1.870	0.015	1.892	0.024	100	99
DKE-30	16.1	C	PMhhl	1.2417	0.0260	0.1352	0.0017	7.3986	0.0954	0.0659	0.0011	0.5725	0.3271	0.817	0.010	0.820	0.012	0.801	0.034	100	102
DKE-30	17.1	C	PMoz	2.0642	0.0373	0.1933	0.0008	5.1731	0.0225	0.0778	0.0011	0.4649	0.2630	1.139	0.005	1.137	0.012	1.144	0.028	100	100
DKE-30	19.1	C	PMoz	4.9371	0.0934	0.3249	0.0024	3.0781	0.0223	0.1109	0.0017	1.2513	0.6983	1.813	0.011	1.809	0.016	1.817	0.027	100	100
DKE-30	21.1	C	PMoz	1.5473	0.0287	0.1552	0.0006	6.4435	0.0267	0.0728	0.0010	0.1994	0.0930	0.930	0.004	0.949	0.011	1.009	0.029	98	92
DKE-30	22.1	C	PMhll	1.7597	0.0319	0.1716	0.0008	5.8272	0.0270	0.0736	0.0010	0.5459	0.2961	1.021	0.004	1.031	0.012	1.030	0.029	99	99
DKE-30	23.1	C	PMcz	1.2176	0.0225	0.1325	0.0007	7.5464	0.0378	0.0664	0.0010	0.6959	0.3751	0.802	0.004	0.809	0.010	0.814	0.032	99	99
DKE-30	25.1	C	PMoz	12.6921	0.1467	0.4866	0.0051	2.0551	0.0214	0.1913	0.0015	-0.2616	2.6875	2.556	0.022	2.657	0.011	2.757	0.013	96	93
DKE-30	27.1	C	PMcz	1.1876	0.0213	0.1323	0.0015	7.5610	0.0849	0.0663	0.0010	-1.0112	4.3776	0.801	0.008	0.795	0.010	0.812	0.030	101	99
DKE-30	28.1	C	PMoz	1.1712	0.0152	0.1326	0.0014	7.5437	0.0825	0.0647	0.0006	-1.1951	4.0127	0.802	0.008	0.787	0.007	0.760	0.020	102	106
DKE-30	30.1	C	PMoz	1.2929	0.0155	0.1396	0.0014	7.1650	0.0742	0.0686	0.0006	-1.4000	3.2446	0.842	0.008	0.843	0.007	0.884	0.018	100	95
DKE-30	31.1	C	PMoz	1.2522	0.0146	0.1323	0.0014	7.5592	0.0779	0.0685	0.0006	-2.5903	5.2963	0.801	0.008	0.824	0.007	0.881	0.018	97	91
DKE-30	32.1	C	PMoz	1.3237	0.0223	0.1382	0.0015	7.2335	0.0789	0.0698	0.0009	-1.2831	2.2707	0.835	0.009	0.856	0.010	0.921	0.026	98	91
DKE-30	33.1	C	PMoz	1.3371	0.0169	0.1443	0.0015	6.9308	0.0741	0.0667	0.0006	-2.9932	4.7364	0.869	0.009	0.862	0.007	0.824	0.020	101	106
DKE-30	34.1	C	PMoz	1.5959	0.0198	0.1655	0.0018	6.0426	0.0644	0.0702	0.0007	-3.7261	5.3324	0.987	0.010	0.969	0.008	0.932	0.021	102	106
DKE-30	35.1	C	PMoz	1.3066	0.0184	0.1428	0.0016	7.0016	0.0762	0.0668	0.0007	-4.1944	5.4786	0.861	0.009	0.849	0.008	0.827	0.021	101	104
DKE-30	36.1	C	PMoz	13.5998	0.1523	0.5148	0.0052	1.9424	0.0198	0.1923	0.0015	-3.0206	3.6287	2.677	0.022	2.722	0.011	2.766	0.013	98	97
DKE-30	37.1	C	PMoz	11.7090	0.1440	0.4971	0.0056	2.0117	0.0225	0.1717	0.0014	-1.0554	1.1737	2.601	0.024	2.581	0.011	2.574	0.013	101	101
DKE-30	38.1	C	PMoz	10.7364	0.1628	0.4625	0.0048	2.1622	0.0225	0.1716	0.0015	0.4631	0.2856	2.451	0.021	2.501	0.014	2.573	0.015	98	95
DKE-30	39.1	C	PMoz	3.7264	0.0557	0.2754	0.0028	3.6312	0.0372	0.0982	0.0009	0.9052	0.5744	1.568	0.014	1.577	0.012	1.596	0.017	99	98
DKE-30	40.1	C	PMoz	1.0972	0.0168	0.1240	0.0013	8.0632	0.0817	0.0645	0.0006	0.4442	0.2903	0.754	0.007	0.752	0.008	0.754	0.021	100	100
DKE-30	41.1	C	Pmoz/Mez	1.3380	0.0209	0.1440	0.0015	6.9438	0.0716	0.0663	0.0006	0.7423	0.4999	0.867	0.008	0.862	0.009	0.811	0.020	101	107
DKE-30	43.1	C	PMoz	1.1705	0.0196	0.1342	0.0015	7.4530	0.0843	0.0650	0.0006	0.1919	0.1377	0.812	0.009	0.787	0.009	0.771	0.020	103	105
DKE-30	45.1	C	PMoz	1.2438	0.0189	0.1308	0.0014	7.6458	0.0836	0.0679	0.0006	0.1936	0.1486	0.792	0.008	0.821	0.009	0.862	0.019	97	92
DKE-30	46.1	C	PMoz	1.1327	0.0178	0.1265	0.0013	7.9059	0.0822	0.0651	0.0007	0.3247	0.2583	0.768	0.008	0.769	0.008	0.774	0.021	100	99
DKE-30	47.1	C	PMoz	1.0701	0.0162	0.1178	0.0012	8.4855	0.0875	0.0656	0.0006	0.2633	0.2173	0.718	0.007	0.739	0.008	0.789	0.020	97	91
DKE-30	48.1	C	PMoz	1.2112	0.0200	0.1345	0.0015	7.4364	0.0849	0.0648	0.0008	0.1961	0.1682	0.813	0.009	0.806	0.009	0.762	0.026	101	107
DKE-30	49.1	C	PMoz	1.0241	0.0150	0.1155	0.0012	8.6558	0.0880	0.0635	0.0006	0.1577	0.1407	0.705	0.007	0.716	0.007	0.721	0.019	98	98
DKE-30	50.1	C	PMoz	1.4865	0.0370	0.1582	0.0018	6.3222	0.0733	0.0689	0.0013	0.2835	0.2639	0.947	0.010	0.925	0.015	0.894	0.038	102	106
Zircons not included on related diagrams																					
DKE-30	12.1	C	PMoz	1.0611	0.0219	0.1156	0.0025	8.6536	0.1842	0.0669	0.0007	1.5288	0.8306	0.705	0.014	0.734	0.011	0.831	0.022	96	85
DKE-30	15.1	C	PMoz	1.7766	0.0328	0.1648	0.0010	6.0678	0.0373	0.0791	0.0012	0.7986	0.4580	0.983	0.006	1.037	0.012	1.177	0.030	95	84
DKE-30	18.1	R	Mog	1.3415	0.0330	0.1134	0.0013	8.8165	0.1019	0.0855	0.0016	0.5873	0.3299	0.693	0.008	0.864	0.014	1.333	0.037	80	52
DKE-30	20.1B	R	PMoz	1.4931	0.0349	0.1481	0.0014	6.7505	0.0654	0.0735	0.0014	0.5807	0.3221	0.891	0.008	0.928	0.014	1.027	0.038	96	87
DKE-30	20.2N	C	PMzc	1.0129	0.0239	0.1042	0.0008	9.5999	0.0750	0.0690	0.0012	0.6372	0.3504	0.639	0.005	0.710	0.012	0.896	0.037	90	71
DKE-30	24.1	C	PMoz	1.2043	0.0223	0.1371	0.0011	7.2961	0.0571	0.0632	0.0010	0.5093	0.2733	0.828	0.006	0.803	0.010	0.711	0.035	103	117
DKE-30	26.1	R	PMoz	2.6124	0.0337	0.1731	0.0020	5.7779	0.0659	0.1118	0.0012	-0.8581	5.2268	1.029	0.011	1.304	0.009	1.832	0.018	79	56
DKE-30	29.1	C	PMhml	2.7431	0.0387	0.2221	0.0024	4.5029	0.0493	0.0953	0.0008	-2.4944	6.8404	1.293	0.013	1.340	0.010	1.540	0.016	96	84
DKE-30	42.1	C	PMoz	1.1050	0.0710	0.1126	0.0023	8.8791	0.1801	0.0702	0.0035	0.6093	0.4254	0.688	0.013	0.756	0.034	0.932	0.100	91	74
DKE-30	44.1	C	PMoz	3.4404	0.0609	0.2121	0.0023	4.7145	0.0508	0.1171	0.0012	0.7992	0.5899	1.240	0.012	1.514	0.014	1.913	0.018	82	65

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-25	2.1	R	PMoz	0.7318	0.0089	0.0895	0.0012	11.1742	0.1499	0.0603	0.0004	0.6043	0.0797	0.553	0.007	0.558	0.005	0.609	0.014	99	91
DKE-25	3.1	C	PMhml	0.7987	0.0097	0.0972	0.0013	10.2842	0.1386	0.0598	0.0004	0.3713	0.0497	0.598	0.008	0.596	0.005	0.591	0.015	100	101
DKE-25	4.1	C	PMcz	6.8196	0.0791	0.3598	0.0046	2.7796	0.0358	0.1385	0.0007	0.2366	0.0316	1.981	0.022	2.088	0.010	2.204	0.009	95	90
DKE-25	7.1	R	PMoz	0.7954	0.0098	0.0966	0.0013	10.3550	0.1371	0.0604	0.0004	0.7668	0.1047	0.594	0.008	0.594	0.006	0.613	0.013	100	97
DKE-25	8.1	C	PMcz	0.7751	0.0092	0.0932	0.0012	10.7261	0.1391	0.0606	0.0003	0.0344	0.0049	0.575	0.007	0.583	0.005	0.618	0.012	99	93
DKE-25	9.1	C	PMhml	0.7647	0.0093	0.0939	0.0013	10.6493	0.1495	0.0609	0.0005	0.4595	0.0642	0.579	0.008	0.577	0.005	0.629	0.017	100	92
DKE-25	11.1	C	PMoz	6.8697	0.0815	0.3644	0.0048	2.7442	0.0360	0.1382	0.0007	0.2175	0.0306	2.003	0.023	2.095	0.010	2.201	0.009	96	91
DKE-25	13.1	C	PMoz	0.7430	0.0095	0.0911	0.0012	10.9744	0.1467	0.0602	0.0004	0.6770	0.0967	0.562	0.007	0.564	0.006	0.603	0.015	100	93
DKE-25	14.1	C	PMoz	20.6682	0.2518	0.5775	0.0065	1.7315	0.0195	0.2615	0.0019	0.2160	0.0486	2.939	0.027	3.123	0.012	3.253	0.011	94	90
DKE-25	15.1	C	PMoz	6.7844	0.0822	0.3679	0.0041	2.7183	0.0304	0.1347	0.0010	0.3191	0.0735	2.019	0.019	2.084	0.011	2.157	0.012	97	94
DKE-25	17.1	C	PMoz	4.1248	0.0521	0.2833	0.0034	3.5294	0.0417	0.1064	0.0008	0.4141	0.1000	1.608	0.017	1.659	0.010	1.742	0.014	97	92
DKE-25	18.1	C	PMoz	0.7876	0.0131	0.0937	0.0014	10.6676	0.1625	0.0606	0.0008	0.7109	0.1766	0.578	0.008	0.590	0.007	0.621	0.029	98	93
DKE-25	19.1	C	PMoz	6.7037	0.0816	0.3602	0.0041	2.7763	0.0313	0.1358	0.0010	0.1357	0.0344	1.983	0.019	2.073	0.011	2.170	0.013	96	91
DKE-25	21.1	C	PMoz	6.7657	0.0843	0.3658	0.0042	2.7338	0.0317	0.1357	0.0010	0.2184	0.0581	2.010	0.020	2.081	0.011	2.169	0.012	97	93
DKE-25	24.1	C	PMhml	7.5759	0.0907	0.3909	0.0043	2.5582	0.0283	0.1421	0.0010	0.1448	0.0418	2.127	0.020	2.182	0.011	2.249	0.012	97	95
DKE-25	28.1	C	PMhhl	6.4139	0.0835	0.3572	0.0047	2.7997	0.0370	0.1313	0.0014	0.3311	2.0055	1.969	0.022	2.034	0.011	2.113	0.018	97	93
DKE-25	29.1	C	PMoz	8.0073	0.0123	0.0963	0.0014	10.3840	0.1552	0.0613	0.0007	0.2628	1.4678	0.593	0.008	0.601	0.007	0.643	0.024	99	92
DKE-25	30.1	C	PMhhl	0.6947	0.0151	0.0857	0.0014	11.6654	0.1884	0.0585	0.0010	0.3309	1.7149	0.530	0.008	0.536	0.009	0.543	0.038	99	98
DKE-25	31.1	C	PMhhl	6.9949	0.0910	0.3724	0.0049	2.6852	0.0352	0.1376	0.0014	0.2025	0.9788	2.041	0.023	2.111	0.011	2.193	0.018	97	93
DKE-25	32.1	C	PMoz	6.6572	0.0859	0.3552	0.0046	2.8152	0.0365	0.1370	0.0014	0.3115	1.4104	1.959	0.022	2.067	0.011	2.186	0.017	95	90
DKE-25	34.1	C	PMoz	0.8046	0.0112	0.0953	0.0013	10.4946	0.1407	0.0608	0.0007	0.3851	1.5482	0.587	0.008	0.599	0.006	0.626	0.025	98	94
DKE-25	38.1	R	PMoz	0.7700	0.0114	0.0944	0.0013	10.5931	0.1426	0.0589	0.0007	1.0409	3.4185	0.582	0.007	0.580	0.006	0.557	0.027	100	104
DKE-25	39.1	C	PMoz	0.8414	0.0123	0.1005	0.0014	9.9474	0.1337	0.0601	0.0007	0.4161	1.3069	0.618	0.008	0.620	0.007	0.602	0.026	100	103
DKE-25	40.1	C	PMoz	0.7978	0.0113	0.0964	0.0009	10.3762	0.1020	0.0610	0.0005	0.2988	0.0398	0.593	0.006	0.596	0.006	0.633	0.019	100	94
DKE-25	41.1	C	PMoz	6.5370	0.0908	0.3660	0.0036	2.7321	0.0266	0.1311	0.0011	0.2099	0.0281	2.011	0.017	2.051	0.012	2.110	0.014	98	95
DKE-25	42.1	R	PMoz	0.7952	0.0115	0.0954	0.0009	10.4817	0.1026	0.0610	0.0005	0.0885	0.0129	0.587	0.005	0.594	0.006	0.634	0.019	99	93
DKE-25	43.1	C	PMoz	0.7255	0.0108	0.0893	0.0009	11.1974	0.1155	0.0593	0.0005	0.9326	0.1226	0.551	0.005	0.554	0.006	0.571	0.020	100	97
DKE-25	44.1	C	PMoz	0.8157	0.0115	0.0976	0.0010	10.2415	0.1013	0.0607	0.0005	0.4208	0.0554	0.601	0.006	0.606	0.006	0.622	0.019	99	92
DKE-25	45.1	C	PMcz	0.8365	0.0120	0.1003	0.0010	9.9707	0.1003	0.0614	0.0005	0.1185	0.0159	0.616	0.006	0.617	0.007	0.646	0.018	100	95
DKE-25	46.1	C	PMoz	0.7882	0.0115	0.0964	0.0010	10.3690	0.1082	0.0600	0.0005	0.3796	0.0551	0.594	0.006	0.590	0.007	0.597	0.018	101	99
DKE-25	48.1	C	PMoz	6.8118	0.0924	0.3661	0.0035	2.7316	0.0264	0.1366	0.0011	0.2942	0.0382	2.011	0.017	2.087	0.012	2.181	0.013	96	92
DKE-25	50.1	C	PMoz	4.3386	0.0603	0.2950	0.0029	3.3898	0.0338	0.1081	0.0009	1.0720	0.1388	1.666	0.015	1.701	0.011	1.771	0.014	98	94
DKE-25	51.1	C	PMcz	7.8652	0.1068	0.3884	0.0037	2.5747	0.0246	0.1485	0.0012	0.3500	0.0452	2.115	0.017	2.216	0.012	2.325	0.013	95	91
DKE-25	52.1	C	PMhhl	0.8485	0.0181	0.1027	0.0016	9.7348	0.1511	0.0613	0.0011	1.1978	0.1552	0.630	0.009	0.624	0.010	0.646	0.039	101	98
DKE-25	53.1	C	PMhml	0.7560	0.0140	0.0941	0.0016	10.6219	0.1845	0.0588	0.0008	0.6990	0.1948	0.580	0.010	0.572	0.008	0.552	0.028	101	105
DKE-25	55.1	C	PMoz	13.0451	0.1940	0.4902	0.0061	2.0399	0.0252	0.1947	0.0020	0.2495	0.0737	2.572	0.026	2.683	0.014	2.787	0.017	96	92
DKE-25	56.1	C	PMhml	0.8110	0.0128	0.0998	0.0013	10.0237	0.1280	0.0594	0.0006	0.5267	0.1606	0.613	0.007	0.603	0.007	0.575	0.024	102	107
DKE-25	57.1	C	PMoz	0.7823	0.0125	0.0963	0.0013	10.3821	0.1401	0.0598	0.0007	1.0232	0.3228	0.593	0.008	0.587	0.007	0.589	0.024	101	101
DKE-25	58.1	R	PMoz	0.8166	0.0132	0.1017	0.0014	9.8325	0.1326	0.0591	0.0007	0.3517	0.1148	0.624	0.008	0.606	0.007	0.563	0.026	103	111
DKE-25	59.1	R	PMoz	0.8114	0.0130	0.0971	0.0012	10.2982	0.1314	0.0613	0.0007	0.3712	0.1248	0.597	0.007	0.603	0.007	0.643	0.024	99	93
DKE-25	61.1	C	PMoz	0.8245	0.0127	0.0981	0.0012	10.1951	0.1284	0.0613	0.0007	0.3660	0.1333	0.603	0.007	0.611	0.007	0.642	0.023	99	94
DKE-25	62.1	R	PMoz	0.8613	0.0140	0.1020	0.0014	9.8023	0.1307	0.0618	0.0007	0.2755	0.1044	0.626	0.008	0.631	0.008	0.663	0.025	99	94
DKE-25	63.1	C	PMoz	0.8256	0.0129	0.0975	0.0012	10.2550	0.1284	0.0619	0.0007	0.2828	0.1116	0.600	0.007	0.611	0.007	0.664	0.023	98	90
DKE-25	64.1	R	PMhhl	0.7463	0.0133	0.0894	0.0014	11.1844	0.1745	0.0605	0.0007	0.9888	0.4075	0.552	0.008	0.566	0.008	0.616	0.026	98	90
DKE-25	65.1	C	PMoz	0.7726	0.0119	0.0946	0.0012	10.5763	0.1336	0.0599	0.0006	0.2968	0.1279	0.582	0.007	0.581	0.007	0.594	0.023	100	98
Zirconos not included on related diagrams																					
DKE-25	1.1	R	PMoz	0.8023	0.0109	0.0946	0.0013	10.5684	0.1446	0.0619	0.0005	0.2376	0.0318	0.583	0.008	0.598	0.006	0.665	0.017	97	88
DKE-25	5.1	R	PMoz	0.9080	0.0111	0.0992	0.0013	10.0821	0.1325	0.0665	0.0004	0.3533	0.0477	0.610	0.008	0.656	0.006	0.817	0.014	93	75
DKE-25	6.1	C	PMoz	0.8912	0.0106	0.1020	0.0013	9.8059	0.1289	0.0638	0.0004	0.6453	0.0876	0.626	0.008	0.647	0.006	0.730	0.013	97	86
DKE-25	10.1	C	PMcz	6.6855	0.0842	0.3562	0.0048	2.8071	0.0382	0.1380	0.0007	0.3083	0.0431	1.964	0.023	2.071	0.011	2.198	0.009	95	89
DKE-25	12.1	C	PMoz	1.0227	0.0202	0.1062	0.0020	9.4184	0.1733	0.0691	0.0006	0.3805	0.0545	0.650	0.011	0.715	0.010	0.899	0.017	91	72
DKE-25	16.1	R	PMhml	0.8334	0.0107	0.0959	0.0011	10.4228	0.1194	0.0632	0.0005	0.4305	0.1014	0.591	0.006	0.616	0.006	0.710	0.016	96	83
DKE-25	20.1	R	PMoz	0.9542	0.0154	0.0937	0.0011	10.6669	0.1266	0.0761	0.0009	0.5049	0.1306	0.578	0.007	0.680	0.008	1.099	0.024	85	53
DKE-25	22.1	C	PMoz	1.0162	0.0144	0.1123	0.0014	8.9014	0.1117	0.0659	0.0007	0.4659	0.1276	0.686	0.008	0.712	0.007	0.798	0.023	96	86
DKE-25	23.1	R	PMoz	0.7950	0.0105	0.0941	0.0011	10.6235	0.1249	0.0615	0.0005	0.5798	0.1629	0.580	0.007	0.594	0.006	0.6			

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-45	1.1	R	MR/Mog	0.9853	0.0082	0.1152	0.0009	8.6772	0.0644	0.0627	0.0003	0.2160	0.0690	0.703	0.005	0.696	0.004	0.692	0.009	101	102
DKE-45	3.1	R	PMoz	1.0137	0.0087	0.1159	0.0010	8.6318	0.0757	0.0639	0.0004	0.2772	0.0911	0.707	0.006	0.711	0.004	0.734	0.012	99	96
DKE-45	4.1	C	PMhml	1.2316	0.0103	0.1369	0.0010	7.3040	0.0536	0.0660	0.0004	0.3667	0.1217	0.827	0.006	0.815	0.005	0.804	0.014	101	103
DKE-45	6.1	C	PMoz	1.3282	0.0221	0.1453	0.0024	6.8813	0.1154	0.0671	0.0003	0.2848	0.0903	0.875	0.014	0.858	0.010	0.838	0.011	102	104
DKE-45	7.1	C	PMoz	1.2619	0.0214	0.1395	0.0024	7.1701	0.1233	0.0659	0.0004	0.5119	0.1596	0.842	0.014	0.829	0.010	0.798	0.013	102	105
DKE-45	9.1	C	PMzc	5.5552	0.0911	0.3461	0.0058	2.8894	0.0482	0.1180	0.0005	0.0832	0.0251	1.916	0.028	1.909	0.014	1.926	0.008	100	99
DKE-45	10.1	C	PMoz	1.4140	0.0234	0.1483	0.0025	6.7414	0.1130	0.0696	0.0004	0.6568	0.1954	0.892	0.014	0.895	0.010	0.915	0.011	100	97
DKE-45	11.1	R	PMhml	1.1447	0.0192	0.1222	0.0021	8.1859	0.1377	0.0683	0.0005	0.0688	0.0215	0.743	0.012	0.775	0.009	0.876	0.014	96	85
DKE-45	12.1	R	PMoz	1.2646	0.0208	0.1404	0.0023	7.1206	0.1182	0.0662	0.0003	0.3169	0.0914	0.847	0.013	0.830	0.009	0.809	0.011	102	105
DKE-45	13.1	C	PMoz	1.2302	0.0201	0.1360	0.0023	7.3553	0.1221	0.0665	0.0003	0.5301	0.1503	0.822	0.013	0.814	0.009	0.818	0.010	101	100
DKE-45	14.1	R	PMoz	1.3644	0.0229	0.1515	0.0025	6.6022	0.1097	0.0660	0.0004	0.2159	0.0603	0.909	0.014	0.874	0.010	0.803	0.013	104	113
DKE-45	15.1	C	PMoz	1.2237	0.0211	0.1367	0.0023	7.3137	0.1229	0.0659	0.0004	0.8721	0.2401	0.826	0.013	0.811	0.010	0.798	0.012	102	104
DKE-45	16.1	C	PMoz	1.3520	0.0223	0.1494	0.0025	6.6916	0.1115	0.0669	0.0004	0.3469	0.0941	0.898	0.014	0.868	0.010	0.831	0.011	103	108
DKE-45	17.1	C	PMoz	1.1431	0.0207	0.1278	0.0022	7.8219	0.1327	0.0653	0.0006	0.6963	0.1863	0.776	0.012	0.774	0.010	0.781	0.018	100	99
DKE-45	18.1	C	PMoz	1.5874	0.0278	0.1603	0.0026	6.2380	0.1028	0.0697	0.0005	0.1528	0.0403	0.958	0.015	0.965	0.011	0.917	0.014	99	105
DKE-45	18.2	R	PMoz	1.3551	0.0111	0.1465	0.0005	6.8243	0.0212	0.0683	0.0005	0.0144	0.0059	0.882	0.003	0.870	0.005	0.876	0.016	101	101
DKE-45	19.1	R	PMoz	1.2314	0.0065	0.1351	0.0003	7.4024	0.0166	0.0669	0.0004	0.5194	0.1614	0.817	0.002	0.815	0.003	0.830	0.013	100	98
DKE-45	20.1	C	PMoz	1.2257	0.0060	0.1375	0.0002	7.2709	0.0125	0.0657	0.0003	0.6201	0.1915	0.831	0.001	0.812	0.003	0.792	0.011	102	105
DKE-45	21.1	C	PMoz	1.4166	0.0074	0.1539	0.0004	6.4958	0.0149	0.0675	0.0004	0.3330	0.1023	0.923	0.002	0.896	0.003	0.850	0.012	103	109
DKE-45	24.1	C	PMoz	1.2063	0.0081	0.1357	0.0003	7.3666	0.0173	0.0650	0.0004	0.2350	0.0710	0.821	0.002	0.803	0.004	0.769	0.014	102	107
DKE-45	26.1	R	PMoz	1.0177	0.0056	0.1172	0.0003	8.5332	0.0201	0.0639	0.0004	0.2339	0.0702	0.714	0.002	0.713	0.003	0.732	0.014	100	98
DKE-45	27.1	R	PMoz	1.0081	0.0069	0.1154	0.0004	8.6682	0.0318	0.0643	0.0004	0.4231	0.1259	0.704	0.002	0.708	0.004	0.748	0.013	99	94
DKE-45	29.1	C	PMhml	1.5030	0.0144	0.1585	0.0007	6.3080	0.0275	0.0699	0.0007	0.4769	0.1407	0.949	0.004	0.932	0.006	0.925	0.020	102	103
DKE-45	30.1	C	PMoz	0.9759	0.0261	0.1133	0.0019	8.8224	0.1507	0.0604	0.0008	0.2306	0.3230	0.692	0.011	0.692	0.013	0.613	0.027	100	113
DKE-45	31.1	C	PMhml	1.2264	0.0285	0.1366	0.0022	7.3187	0.1201	0.0652	0.0005	0.8182	1.0458	0.826	0.013	0.813	0.013	0.777	0.017	102	106
DKE-45	33.1	C	PMoz	1.3699	0.0309	0.1495	0.0024	6.6888	0.1089	0.0673	0.0005	1.0279	1.1187	0.898	0.014	0.876	0.013	0.842	0.015	103	107
DKE-45	34.1	C	PMcz	1.2867	0.0289	0.1412	0.0023	7.0799	0.1154	0.0669	0.0005	0.6438	0.6523	0.852	0.013	0.840	0.013	0.832	0.016	101	102
DKE-45	35.1	C	PMoz	1.2683	0.0287	0.1425	0.0023	7.0170	0.1152	0.0655	0.0005	1.8059	1.7113	0.859	0.013	0.832	0.013	0.786	0.015	103	109
DKE-45	36.1	C	PMcz	1.2612	0.0281	0.1399	0.0023	7.1469	0.1155	0.0662	0.0005	2.1515	1.9313	0.844	0.013	0.828	0.013	0.808	0.015	102	104
DKE-45	37.1	C	PMzc	1.3532	0.0312	0.1483	0.0024	6.7448	0.1100	0.0669	0.0005	0.9296	0.7801	0.891	0.014	0.869	0.013	0.832	0.016	103	107
DKE-45	38.1	R	PMoz	1.0297	0.0232	0.1168	0.0019	8.5603	0.1391	0.0645	0.0005	1.4828	1.1770	0.712	0.011	0.719	0.012	0.754	0.016	99	94
DKE-45	39.1	C	PMhhl	1.3355	0.0313	0.1458	0.0024	6.8591	0.1118	0.0665	0.0007	0.9903	0.7458	0.877	0.013	0.861	0.014	0.819	0.021	102	107
DKE-45	40.1	R	PMoz	1.1458	0.0265	0.1233	0.0020	8.1088	0.1324	0.0677	0.0006	1.1256	0.8063	0.750	0.012	0.775	0.012	0.857	0.019	97	87
DKE-45	42.1	C	PMoz	1.3032	0.0254	0.1459	0.0024	6.8552	0.1139	0.0668	0.0007	1.3753	0.9984	0.878	0.014	0.847	0.011	0.829	0.023	104	106
DKE-45	43.1	C	PMoz	1.3572	0.0219	0.1514	0.0023	6.6029	0.1004	0.0657	0.0004	1.7468	1.2401	0.909	0.013	0.871	0.009	0.792	0.011	104	115
DKE-45	44.1	C	PMcz	1.2881	0.0211	0.1438	0.0022	6.9545	0.1064	0.0654	0.0004	0.9985	0.7040	0.866	0.012	0.840	0.009	0.782	0.014	103	111
DKE-45	45.1	C	PMoz	1.2811	0.0207	0.1431	0.0022	6.9876	0.1077	0.0658	0.0003	2.7116	1.8984	0.862	0.012	0.837	0.009	0.796	0.009	103	108
DKE-45	46.1	C	PMoz	1.2734	0.0209	0.1418	0.0022	7.0540	0.1075	0.0657	0.0004	3.1432	2.1876	0.855	0.012	0.834	0.009	0.793	0.014	102	108
DKE-45	47.1	C	PMoz	1.3441	0.0233	0.1498	0.0023	6.6771	0.1023	0.0655	0.0005	1.2614	0.8711	0.900	0.013	0.865	0.010	0.788	0.016	104	114
DKE-45	48.1	C	PMoz	1.0308	0.0167	0.1176	0.0018	8.5061	0.1302	0.0643	0.0003	1.8974	1.3014	0.717	0.010	0.719	0.008	0.746	0.011	100	96
DKE-45	49.1	C	PMoz	1.3162	0.0241	0.1474	0.0023	6.7856	0.1045	0.0657	0.0006	1.2147	0.8276	0.886	0.013	0.853	0.011	0.794	0.020	104	112
DKE-45	50.1	C	PMoz	1.1315	0.0196	0.1217	0.0019	8.2173	0.1270	0.0676	0.0005	1.1587	0.7844	0.740	0.011	0.768	0.009	0.853	0.017	96	87
Metamorphic domains (overgrowths)																					
DKE-45	5.1	R	Mog	0.8752	0.0066	0.1040	0.0007	9.6196	0.0661	0.0614	0.0002	0.0792	0.0271	0.638	0.004	0.638	0.004	0.648	0.007	100	98
DKE-45	8.1	R	Mog	0.8272	0.0139	0.0993	0.0017	10.0733	0.1703	0.0608	0.0003	0.0073	0.0035	0.610	0.010	0.612	0.008	0.626	0.011	100	97
DKE-45	22.1B	R	Mog	0.8821	0.0076	0.1056	0.0005	9.4676	0.0405	0.0603	0.0005	0.0287	0.0106	0.647	0.003	0.642	0.004	0.610	0.016	101	106
DKE-45	23.1B	R	Mog	0.8021	0.0055	0.0973	0.0004	10.2809	0.0440	0.0613	0.0004	0.0376	0.0122	0.598	0.002	0.598	0.003	0.642	0.015	100	93
DKE-45	42.2	R	Mog	0.7919	0.0137	0.0947	0.0015	10.5612	0.1675	0.0608	0.0004	0.1393	0.1017	0.583	0.009	0.592	0.008	0.626	0.015	98	93
DKE-45	28.1	R	Mog	0.8974	0.0104	0.1060	0.0005	9.4350	0.0455	0.0620	0.0007	0.1422	0.0426	0.649	0.003	0.650	0.006	0.667	0.023	100	97
Zircons not included on related diagrams																					
DKE-45	32.1	C	Mz/Mez	0.9018	0.0202	0.1092	0.0018	9.1593	0.1484	0.0606	0.0004	0.0332	0.0391	0.668	0.010	0.653	0.011	0.618	0.015	106	109

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (‰)
DKE 43	1.1	C	P Mhml	1.3165	0.0293	0.1438	0.0039	6.9529	0.1887	0.0690	0.0008	0.4714	0.2072	0.866	0.022	0.853	0.013	0.896	0.023	102	97
DKE 43	2.1	C	P Mhml	1.1329	0.0251	0.1267	0.0032	7.8927	0.1993	0.0655	0.0007	0.3320	0.1484	0.769	0.018	0.769	0.012	0.785	0.023	100	98
DKE 43	3.1	C	P Mhml	4.0272	0.0869	0.2841	0.0071	3.5202	0.0879	0.1054	0.0008	0.3360	0.1531	1.612	0.036	1.640	0.017	1.725	0.014	98	93
DKE 43	4.1	C	P Moz	11.3385	0.2285	0.4724	0.0117	2.1166	0.0522	0.1763	0.0006	0.4393	0.2085	2.494	0.051	2.551	0.019	2.619	0.006	98	95
DKE 43	5.1	C	P Moz	1.1367	0.0231	0.1251	0.0031	7.9943	0.1976	0.0668	0.0003	0.1344	0.0637	0.760	0.018	0.771	0.011	0.829	0.008	99	92
DKE 43	6.1	C	P Moz	0.8896	0.0182	0.1058	0.0026	9.4531	0.2356	0.0621	0.0002	0.1750	0.0848	0.648	0.015	0.646	0.010	0.672	0.009	100	97
DKE 43	9.1	C	P Mhml	5.9773	0.1217	0.3575	0.0089	2.7971	0.0699	0.1234	0.0004	0.6563	0.3383	1.970	0.042	1.973	0.018	2.005	0.006	100	98
DKE 43	10.1	C	P Moz	10.0554	0.2046	0.4522	0.0113	2.2112	0.0553	0.1633	0.0005	0.2645	0.1394	2.405	0.050	2.440	0.019	2.488	0.005	99	97
DKE 43	11.1	C	P Moz	1.0881	0.0223	0.1201	0.0030	8.3247	0.2096	0.0662	0.0002	0.1140	0.0615	0.731	0.017	0.748	0.011	0.810	0.008	98	90
DKE 43	12.1	C	P Moz	1.0263	0.0213	0.1198	0.0030	8.3452	0.2101	0.0634	0.0003	0.2314	0.1278	0.730	0.017	0.717	0.011	0.715	0.010	102	102
DKE 43	14.2N	C	P Moz	1.2265	0.0163	0.1383	0.0013	7.2295	0.0661	0.0657	0.0003	0.4257	0.0837	0.835	0.007	0.813	0.007	0.792	0.011	103	105
DKE 43	15.1	C	P Moz	5.0199	0.0611	0.3258	0.0023	3.0695	0.0215	0.1130	0.0005	2.1140	0.4135	1.818	0.011	1.823	0.010	1.850	0.008	100	98
DKE 43	16.1	C	P Moz	1.1525	0.0135	0.1286	0.0008	7.7790	0.0483	0.0660	0.0003	0.4367	0.0850	0.780	0.005	0.778	0.006	0.801	0.008	100	97
DKE 43	17.1	C	P Moz	1.1487	0.0136	0.1288	0.0009	7.7659	0.0517	0.0656	0.0002	0.2931	0.0568	0.781	0.005	0.777	0.006	0.790	0.008	101	99
DKE 43	18.1	C	P Moz	0.9584	0.0121	0.1082	0.0008	9.2387	0.0681	0.0654	0.0004	0.3687	0.0713	0.663	0.005	0.682	0.006	0.784	0.015	97	85
DKE 43	19.1	C	P Mcz	1.0087	0.0117	0.1181	0.0008	8.4689	0.0559	0.0629	0.0003	0.3050	0.0587	0.720	0.004	0.708	0.006	0.698	0.009	102	103
DKE 43	20.1	C	P Mhll	1.7465	0.0197	0.1723	0.0101	5.8044	0.0349	0.0742	0.0003	0.3500	0.0669	1.025	0.006	1.026	0.007	1.049	0.007	100	98
DKE 43	22.1	R	P Moz	1.0639	0.0124	0.1212	0.0008	8.2486	0.0530	0.0645	0.0003	0.1888	0.0358	0.738	0.004	0.736	0.006	0.755	0.009	100	98
DKE 43	23.2N	C	P Moz	0.9214	0.0109	0.1093	0.0007	9.1466	0.0565	0.0618	0.0003	0.3266	0.0616	0.669	0.004	0.663	0.006	0.663	0.011	101	101
DKE 43	24.1	C	P Mhml	6.5430	0.0738	0.3787	0.0023	2.6403	0.0157	0.1269	0.0005	0.5230	0.0979	2.070	0.011	2.052	0.010	2.054	0.006	101	101
DKE 43	24.2B	R	P Mhhl	5.4399	0.0562	0.3389	0.0039	2.9506	0.0339	0.1174	0.0005	0.6414	0.0864	1.881	0.019	1.891	0.009	1.818	0.008	99	98
DKE 43	25.1	C	P Moz	1.2760	0.0161	0.1426	0.0018	7.0121	0.0868	0.0656	0.0004	0.3001	0.0406	0.859	0.010	0.835	0.007	0.791	0.013	103	109
DKE 43	26.1	C	P Mhhl	1.2507	0.0134	0.1390	0.0018	7.1950	0.0930	0.0646	0.0006	0.3036	0.0417	0.839	0.010	0.824	0.006	0.758	0.020	102	111
DKE 43	28.2N	C	P Mcz	1.0949	0.0132	0.1231	0.0015	8.1234	0.0962	0.0650	0.0004	0.3556	0.0481	0.748	0.008	0.751	0.006	0.770	0.012	100	97
DKE 43	28.1B	R	P Moz	1.0847	0.0111	0.1214	0.0014	8.2354	0.0944	0.0651	0.0003	0.2751	0.0372	0.739	0.008	0.746	0.005	0.774	0.010	99	95
DKE 43	29.1	R	P Moz	1.2375	0.0121	0.1372	0.0016	7.2865	0.0830	0.0661	0.0003	0.0707	0.0097	0.829	0.009	0.818	0.005	0.804	0.008	101	103
DKE 43	30.1	C	P Moz	0.9659	0.0116	0.1105	0.0015	9.0526	0.1192	0.0639	0.0003	0.1999	0.0271	0.675	0.008	0.686	0.006	0.733	0.009	98	92
DKE 43	31.1	C	P Mhml	5.7240	0.0581	0.3449	0.0040	2.8997	0.0335	0.1213	0.0005	0.5070	0.0689	1.910	0.019	1.935	0.009	1.974	0.007	99	97
DKE 43	32.1	C	P Mhml	6.2108	0.0757	0.3605	0.0045	2.7742	0.0346	0.1262	0.0006	0.3365	0.0460	1.984	0.021	2.006	0.011	2.044	0.009	99	97
DKE 43	33.1	R	P Moz	1.5268	0.0154	0.1568	0.0018	6.3785	0.0746	0.1713	0.0003	0.3117	0.0424	0.939	0.010	0.941	0.006	0.964	0.009	100	97
DKE 43	34.1	C	P Moz	1.2533	0.0129	0.1361	0.0016	7.3477	0.0855	0.0666	0.0003	0.7382	0.1009	0.823	0.009	0.825	0.006	0.822	0.009	100	100
DKE 43	35.1	C	P Mhml	1.0129	0.0105	0.1166	0.0014	8.5783	0.1009	0.0636	0.0003	0.0673	0.0094	0.711	0.008	0.710	0.005	0.723	0.009	100	98
DKE 43	36.1	C	P Mcz	3.9326	0.2183	0.4264	0.0104	2.3452	0.0572	0.1611	0.0002	0.4574	0.0737	2.290	0.047	2.377	0.021	2.464	0.002	96	93
DKE 43	37.1	C	P Moz	1.0286	0.0245	0.1183	0.0029	8.4534	0.2081	0.0640	0.0003	0.1596	0.0258	0.721	0.017	0.718	0.012	0.736	0.009	100	98
DKE 43	38.1	C	P Moz	0.9372	0.0225	0.1103	0.0028	9.0621	0.2262	0.0622	0.0001	0.5933	0.0961	0.675	0.016	0.671	0.012	0.676	0.004	100	100
DKE 43	39.1	C	P Moz	0.9552	0.0222	0.1108	0.0027	9.0277	0.2196	0.0629	0.0002	1.2896	0.2085	0.677	0.016	0.681	0.011	0.700	0.006	99	97
DKE 43	40.1	C	P Mhml	1.1375	0.0267	0.1280	0.0032	7.8112	0.1924	0.0657	0.0004	0.5686	0.0922	0.777	0.018	0.771	0.013	0.792	0.012	101	98
DKE 43	41.1	C	P Moz	1.0485	0.0243	0.1184	0.0029	8.4434	0.2040	0.0649	0.0002	0.1854	0.0301	0.722	0.016	0.728	0.012	0.767	0.006	99	94
DKE 43	42.2N	C	P Moz	0.9237	0.0223	0.1089	0.0026	9.1810	0.2230	0.0632	0.0005	0.2060	0.0342	0.666	0.015	0.664	0.012	0.710	0.016	100	94
DKE 43	43.1	C	P Moz	1.1647	0.0284	0.1304	0.0033	7.6658	0.1919	0.0650	0.0003	0.2916	0.0477	0.790	0.019	0.784	0.013	0.768	0.009	101	103
DKE 43	59.1	C	P Moz	3.8548	0.0837	0.2851	0.0050	3.5079	0.0613	0.1036	0.0006	0.1885	0.0898	1.617	0.025	1.604	0.017	1.693	0.010	101	95
DKE 43	45.1	C	P Mhll	9.8090	0.2241	0.4496	0.0107	2.2242	0.0531	0.1607	0.0002	0.1898	0.0310	2.393	0.048	2.417	0.021	2.460	0.003	99	97
DKE 43	46.1	R	P Moz	1.0558	0.0241	0.1203	0.0029	8.3149	0.1984	0.0643	0.0002	0.1951	0.0319	0.732	0.016	0.732	0.012	0.748	0.006	100	98
DKE 43	47.1	C	P Moz	1.0873	0.0250	0.1230	0.0030	8.1323	0.1955	0.0643	0.0001	0.1051	0.0172	0.748	0.017	0.747	0.012	0.745	0.004	100	100
DKE 43	48.1	C	P Moz	1.0941	0.0243	0.1253	0.0023	7.9840	0.1438	0.0640	0.0003	0.4468	0.1707	0.761	0.013	0.751	0.012	0.737	0.010	101	103
DKE 43	49.1	R	P Moz	1.1160	0.0245	0.1261	0.0022	7.9302	0.1410	0.0647	0.0003	0.3164	0.1229	0.766	0.013	0.761	0.012	0.762	0.011	101	101
DKE 43	51.1	C	P Moz	0.9192	0.0202	0.1082	0.0019	9.2416	0.1657	0.0624	0.0003	0.5329	0.2179	0.662	0.011	0.662	0.011	0.681	0.011	100	97
DKE 43	52.1	C	P Moz	1.2999	0.0283	0.1424	0.0025	7.0237	0.1237	0.0677	0.0003	0.7105	0.2958	0.858	0.014	0.846	0.012	0.856	0.010	101	100
DKE 43	53.1	C	P Moz	1.2793	0.0291	0.1406	0.0026	7.1118	0.1320	0.0659	0.0004	0.3056	0.1293	0.848	0.015	0.837	0.013	0.798	0.014	101	106
DKE 43	55.1	C	P Moz	1.2735	0.0276	0.1383	0.0024	7.2294	0.1261	0.0671	0.0004	0.3059	0.1346	0.835	0.014	0.834	0.012	0.838	0.011	100	100
DKE 43	56.1	C	P Mcz	1.2418	0.0271	0.1337	0.0024	7.4815	0.1321	0.0681	0.0004	0.5239	0.2349	0.809	0.013	0.820	0.012	0.869	0.011	99	93
DKE 43	57.1	C	P Moz	1.2964	0.0282	0.1395	0.0024	7.1695	0.1248	0.0675	0.0004	0.4113	0.1880	0.842	0.014	0.844	0.012	0.849	0.011	100	99
DKE 43	58.1	C	P Moz	9.7118	0.2318	0.4468	0.0089	2.2380	0.0446	0.1595	0.0008	0.5326	0.2484	2.381	0.040	2.408	0.022	2.447	0.008	99	97
DKE 43	44.1	C	Mz	0.8376	0.0285	0.1030	0.0034	9.7062	0.3233	0.0627	0.0008	0.0177	0.0113	0.632	0.020	0.618	0.016	0.691	0.027	102	91
Zircons not included on related diagrams																					
DKE 43	7.1	C	P Moz	0.9318	0.0191	0.1076	0.0027	9.2970	0.2309	0.0643	0.0003	0.3319	0.1634	0.659	0.016	0.669	0.010	0.746	0.008	99	88
DKE 43	8.1	C	P Mcz	8																	

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)*	Conc. (%)**
DKE-19	1.1	C	PMoz	6.1047	0.1000	0.3574	0.0044	2.7979	0.0344	0.1247	0.0009	0.5132	0.3111	1.970	0.021	1.991	0.014	2.023	0.012	99	97
DKE-19	2.1	C	PMhll	5.3999	0.0885	0.3301	0.0041	3.0295	0.0372	0.1199	0.0008	0.4622	0.2690	1.839	0.020	1.885	0.014	1.955	0.012	98	94
DKE-19	3.1	C	PMcz	3.3006	0.0544	0.2510	0.0031	3.9843	0.0492	0.0963	0.0007	0.2012	0.1126	1.444	0.016	1.481	0.013	1.560	0.013	97	93
DKE-19	4.1	C	PMoz	2.2962	0.0380	0.2037	0.0025	4.9083	0.0607	0.0826	0.0006	0.7501	0.4043	1.195	0.013	1.211	0.012	1.266	0.014	99	94
DKE-19	5.1	C	PMoz	6.7065	0.1101	0.3662	0.0045	2.7308	0.0336	0.1341	0.0010	0.1346	0.0700	2.011	0.021	2.073	0.014	2.149	0.012	97	94
DKE-19	7.1	C	PMoz	5.0059	0.0826	0.3150	0.0039	3.1745	0.0392	0.1162	0.0008	0.7460	0.3635	1.765	0.019	1.820	0.014	1.899	0.013	97	93
DKE-19	8.1	C	PMoz	5.4497	0.0904	0.3321	0.0041	3.0114	0.0376	0.1207	0.0009	2.3005	1.0800	1.848	0.020	1.893	0.014	1.967	0.013	98	94
DKE-19	9.1	C	PMhml	5.1493	0.0846	0.3321	0.0041	3.0109	0.0372	0.1144	0.0008	1.5002	0.6822	1.849	0.020	1.844	0.014	1.873	0.012	100	99
DKE-19	10.1	C	PMhll	2.6891	0.0441	0.2256	0.0028	4.4335	0.0545	0.0873	0.0006	0.2350	0.1036	1.311	0.015	1.325	0.012	1.373	0.014	99	95
DKE-19	12.1	C	PMoz	2.1773	0.0361	0.1993	0.0025	5.0177	0.0625	0.0804	0.0006	0.2578	0.1072	1.172	0.013	1.174	0.011	1.210	0.015	100	97
DKE-19	13.1	C	PMoz	2.2617	0.0372	0.1986	0.0024	5.0362	0.0621	0.0839	0.0006	0.1836	0.0742	1.168	0.013	1.200	0.012	1.295	0.014	97	90
DKE-19	14.1	C	PMoz	6.5989	0.1063	0.3757	0.0047	2.6616	0.0331	0.1290	0.0008	0.5491	0.2405	2.056	0.022	2.059	0.014	2.082	0.011	100	99
DKE-19	16.1	C	PMoz	2.0560	0.0331	0.1873	0.0023	5.3404	0.0663	0.0806	0.0005	0.2687	0.1222	1.106	0.013	1.134	0.011	1.216	0.013	98	91
DKE-19	17.1	C	PMoz	3.2661	0.0524	0.2524	0.0031	3.9613	0.0490	0.0948	0.0006	0.1839	0.0853	1.451	0.016	1.473	0.012	1.531	0.012	99	95
DKE-19	18.1	C	PMoz	3.2491	0.0521	0.2483	0.0031	4.0270	0.0498	0.0958	0.0006	0.2964	0.1401	1.430	0.016	1.469	0.012	1.549	0.012	97	92
DKE-19	19.1	C	PMoz	1.5327	0.0264	0.1594	0.0021	6.2718	0.0833	0.0712	0.0005	1.1278	0.5440	0.954	0.012	0.944	0.011	0.963	0.016	101	99
DKE-19	20.1	C	PMoz	2.3451	0.0379	0.2039	0.0026	4.9039	0.0615	0.0847	0.0005	0.3161	0.1551	1.196	0.014	1.226	0.011	1.314	0.012	98	91
DKE-19	21.1	C	PMoz	4.7827	0.0771	0.3101	0.0039	3.2247	0.0402	0.1134	0.0007	0.3911	0.1966	1.741	0.019	1.782	0.013	1.856	0.011	98	94
DKE-19	22.1	C	PMoz	2.2151	0.0359	0.2004	0.0025	4.9912	0.0620	0.0813	0.0005	0.2286	0.1174	1.177	0.013	1.186	0.011	1.232	0.013	99	96
DKE-19	23.1	C	PMhll	4.3562	0.0723	0.3000	0.0040	3.3339	0.0444	0.1070	0.0009	0.2863	0.1505	1.691	0.020	1.704	0.014	1.752	0.014	99	97
DKE-19	24.1	C	PMoz	2.1722	0.0350	0.1989	0.0025	5.0284	0.0624	0.0805	0.0005	0.1870	0.1004	1.169	0.013	1.172	0.011	1.213	0.013	100	96
DKE-19	25.1	C	PMoz	3.3103	0.0531	0.2566	0.0032	3.8972	0.0482	0.0950	0.0006	0.0784	0.0431	1.472	0.016	1.483	0.012	1.534	0.012	99	96
DKE-19	27.1	C	PMoz	1.6104	0.0132	0.1608	0.0011	6.2172	0.0440	0.0735	0.0003	0.2980	0.1729	0.961	0.006	0.974	0.005	1.029	0.009	99	93
DKE-19	28.1	C	PMoz	2.1823	0.0175	0.1977	0.0014	5.0579	0.0355	0.0810	0.0003	0.3183	0.1750	1.163	0.007	1.175	0.006	1.227	0.008	99	95
DKE-19	29.1	C	PMoz	1.8163	0.0167	0.1726	0.0014	5.7929	0.0474	0.0769	0.0004	0.2245	0.1174	1.027	0.008	1.051	0.006	1.122	0.010	98	92
DKE-19	30.1	C	PMoz	5.5520	0.0453	0.3333	0.0025	3.0007	0.0227	0.1220	0.0006	0.4102	0.2041	1.854	0.012	1.909	0.007	1.985	0.008	97	93
DKE-19	31.1	R	PMoz	4.5651	0.0365	0.3056	0.0022	3.2726	0.0235	0.1099	0.0005	0.9102	0.4321	1.719	0.011	1.743	0.007	1.801	0.007	99	95
DKE-19	33.1	C	PMoz	4.7841	0.0397	0.3081	0.0023	3.2453	0.0239	0.1152	0.0005	1.4057	0.6154	1.732	0.011	1.782	0.007	1.884	0.008	97	92
DKE-19	34.1	C	PMoz	4.3473	0.0335	0.2931	0.0020	3.4122	0.0236	0.1083	0.0004	0.6942	0.2901	1.657	0.010	1.702	0.006	1.775	0.007	97	93
DKE-19	37.1	C	PMoz	6.9645	0.0551	0.3735	0.0027	2.6777	0.0191	0.1367	0.0005	0.3467	0.1294	2.046	0.013	2.107	0.007	2.182	0.006	97	94
DKE-19	38.1	C	PMoz	3.0539	0.0242	0.2359	0.0017	4.2382	0.0305	0.0946	0.0004	0.2683	0.0967	1.366	0.009	1.421	0.006	1.526	0.008	96	90
DKE-19	39.1	C	PMoz	1.6226	0.0136	0.1630	0.0012	6.1350	0.0460	0.0724	0.0003	0.5609	0.1954	0.973	0.007	0.979	0.005	0.996	0.009	99	98
DKE-19	40.1	C	PMoz	14.4862	0.1392	0.5066	0.0070	1.9738	0.0274	0.2109	0.0015	0.4917	0.3253	2.642	0.030	2.782	0.009	2.919	0.012	95	91
DKE-19	41.1	C	PMoz	1.4078	0.0174	0.1503	0.0024	6.6523	0.1058	0.0705	0.0010	1.1808	0.7628	0.903	0.013	0.892	0.007	0.942	0.028	101	96
DKE-19	43.1	C	PMoz	5.3967	0.0508	0.3253	0.0045	3.0744	0.0427	0.1217	0.0009	0.5853	0.3583	1.815	0.022	1.884	0.008	1.981	0.013	96	92
DKE-19	46.1	C	PMcz	14.4760	0.1394	0.5162	0.0073	1.9372	0.0274	0.2064	0.0015	0.8566	0.4894	2.683	0.031	2.781	0.009	2.883	0.012	96	93
DKE-19	47.8	C	PMhml	9.9408	0.0939	0.4411	0.0061	2.2668	0.0316	0.1653	0.0012	0.3733	0.2078	2.356	0.027	2.429	0.009	2.508	0.012	97	94
DKE-19	48.8	C	PMoz	1.5708	0.0161	0.1567	0.0022	6.3799	0.0907	0.0735	0.0005	0.2653	0.1444	0.939	0.012	0.959	0.006	1.029	0.015	98	91
DKE-19	49.1	C	PMoz	5.8437	0.0552	0.3346	0.0047	2.9888	0.0419	0.1280	0.0009	0.8536	0.4545	1.861	0.023	1.953	0.008	2.068	0.013	95	90
DKE-19	50.1	C	PMoz	2.4120	0.0235	0.2084	0.0029	4.7984	0.0675	0.0849	0.0006	0.7025	0.3661	1.220	0.016	1.246	0.007	1.318	0.014	98	93
DKE-19	52.1	C	PMoz	2.0991	0.0198	0.1895	0.0027	5.2770	0.0739	0.0813	0.0006	0.3542	0.1771	1.119	0.014	1.149	0.006	1.232	0.014	97	91
Metamorphic domains (overgrowths)																					
DKE-19	32.1B	R	Mog	0.8116	0.0080	0.0982	0.0009	10.1813	0.0914	0.0605	0.0004	0.0306	0.0173	0.604	0.005	0.603	0.004	0.617	0.014	100	98
Zircons not included on related diagrams																					
DKE-19	6.1	C	PMoz	4.6723	0.0770	0.2748	0.0034	3.6387	0.0452	0.1242	0.0009	0.2096	0.1052	1.565	0.017	1.762	0.014	2.016	0.012	89	78
DKE-19	11.1	C	PMoz	5.5435	0.0907	0.3228	0.0040	3.0976	0.0380	0.1264	0.0009	0.3095	0.1325	1.804	0.019	1.907	0.014	2.046	0.012	95	88
DKE-19	15.1	C	PMcz	2.1040	0.0363	0.1880	0.0024	5.3203	0.0692	0.0820	0.0008	0.4078	0.1821	1.110	0.013	1.150	0.012	1.249	0.019	97	89
DKE-19	26.1	C	PMcz	1.4847	0.0248	0.1486	0.0019	6.6789	0.0846	0.0732	0.0005	0.3742	0.2103	0.893	0.010	0.924	0.010	1.021	0.015	97	88
DKE-19	35.1	C	PMoz	1.4806	0.0191	0.1498	0.0012	6.6738	0.0551	0.0733	0.0007	0.5617	0.2259	0.900	0.007	0.922	0.008	1.022	0.020	98	88
DKE-19	36.1B	R	Mog	0.7792	0.0066	0.0921	0.0007	10.8580	0.0791	0.0620	0.0003	0.0286	0.0112	0.568	0.004	0.585	0.004	0.668	0.009	97	85
DKE-19	42.1B	R	Mog	0.7240	0.0072	0.0871	0.0012	11.4745	0.1631	0.0608	0.0005	0.0182	0.0117	0.539	0.007	0.553	0.004	0.626	0.016	97	86
DKE-19	44.1B	R	Mog	0.9804	0.0102	0.0981	0.0014	10.1947	0.1461	0.0736	0.0006	0.0922	0.0553	0.603	0.008	0.694	0.005	1.032	0.016	87	58
DKE-19	45.1B	R	PMoz	1.3539	0.0127	0.1409	0.0020	7.0983	0.0983	0.0708	0.0005	0.0396	0.0231	0.850	0.011	0.869	0.005	0.950	0.015	98	89
DKE-19	51.1	R	PMoz	11.6195	0.1099	0.4285	0.0060	2.3335	0.0328	0.1986	0.0014	0.5548	0.2831	2.299	0.027	2.574	0.009	2.820	0.012	89	82

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-51	1.1	C	PMhml	1.4501	0.0157	0.1520	0.0021	6.5780	0.0913	0.0693	0.0006	0.3215	0.1529	0.912	0.012	0.910	0.006	0.907	0.018	100	101
DKE-51	2.1	C	PMcz	6.5540	0.0658	0.3701	0.0051	2.7016	0.0370	0.1295	0.0010	0.4156	0.1899	2.030	0.024	2.053	0.009	2.089	0.013	99	97
DKE-51	3.1	C	PMoz	3.4790	0.0370	0.2664	0.0037	3.7538	0.0519	0.0949	0.0008	0.2088	0.0919	1.523	0.019	1.522	0.008	1.533	0.015	100	99
DKE-51	4.1	C	PMoz	4.0888	0.0442	0.2918	0.0041	3.4270	0.0478	0.1011	0.0009	0.4415	0.1873	1.651	0.020	1.652	0.009	1.649	0.016	100	100
DKE-51	5.1	C	PMoz	2.2274	0.0236	0.2037	0.0028	4.9101	0.0676	0.0797	0.0006	0.2462	0.1008	1.195	0.015	1.190	0.007	1.193	0.016	100	100
DKE-51	6.1	C	PMoz	1.9682	0.0230	0.1895	0.0027	5.2784	0.0761	0.0767	0.0007	0.3929	0.1587	1.118	0.015	1.105	0.008	1.116	0.018	101	100
DKE-51	7.1	C	PMoz	9.5134	0.1002	0.4352	0.0062	2.2979	0.0328	0.1599	0.0012	0.3187	0.1226	2.329	0.028	2.389	0.010	2.452	0.013	97	95
DKE-51	8.1	C	PMhml	5.8739	0.0597	0.3527	0.0049	2.8352	0.0394	0.1219	0.0010	0.8952	0.3319	1.948	0.023	1.957	0.009	1.983	0.014	99	98
DKE-51	9.1	C	PMoz	4.3659	0.0761	0.3100	0.0049	3.2258	0.0509	0.1073	0.0018	0.4789	0.1726	1.741	0.024	1.706	0.014	1.757	0.029	102	99
DKE-51	10.1	C	PMoz	2.5662	0.0288	0.2201	0.0031	4.5424	0.0629	0.0851	0.0007	0.5065	0.1767	1.283	0.016	1.291	0.008	1.323	0.016	99	97
DKE-51	11.1	C	PMoz	13.2145	0.1390	0.5153	0.0073	1.9407	0.0275	0.1875	0.0014	0.4040	0.1369	2.679	0.031	2.695	0.010	2.724	0.013	99	98
DKE-51	12.1	C	PMoz	9.6995	0.1103	0.4504	0.0065	2.2204	0.0320	0.1590	0.0015	0.3644	0.1202	2.397	0.029	2.407	0.010	2.442	0.016	100	98
DKE-51	13.1	C	PMoz	1.8730	0.0202	0.1826	0.0026	5.4761	0.0767	0.0759	0.0006	0.5172	0.1658	1.081	0.014	1.072	0.007	1.095	0.017	101	99
DKE-51	14.1	C	PMoz	2.3210	0.0507	0.2122	0.0039	4.7122	0.0863	0.0792	0.0005	0.3384	0.1174	1.241	0.021	1.219	0.015	1.182	0.012	102	105
DKE-51	15.1	C	PMhml	2.2554	0.0492	0.2040	0.0037	4.9017	0.0882	0.0812	0.0006	0.3167	0.1097	1.197	0.020	1.198	0.015	1.231	0.014	100	97
DKE-51	16.1	C	PMoz	1.5581	0.0330	0.1574	0.0028	6.3513	0.1148	0.0717	0.0003	0.4960	0.1712	0.943	0.016	0.954	0.013	0.977	0.009	99	96
DKE-51	17.1	R	PMoz	6.6784	0.1398	0.3767	0.0067	2.6548	0.0475	0.1290	0.0005	0.3684	0.1269	2.061	0.031	2.070	0.018	2.082	0.007	100	99
DKE-51	18.1	C	PMhml	6.9820	0.1465	0.3839	0.0069	2.6046	0.0466	0.1329	0.0005	0.2006	0.0690	2.095	0.032	2.109	0.018	2.133	0.007	99	98
DKE-51	19.1	C	PMhml	7.2634	0.1521	0.3942	0.0071	2.5366	0.0455	0.1340	0.0006	0.3563	0.1223	2.142	0.033	2.144	0.019	2.148	0.008	100	100
DKE-51	20.1	C	PMhll	1.9124	0.0400	0.1841	0.0033	5.4328	0.0966	0.0762	0.0004	0.9738	0.3336	1.089	0.018	1.085	0.014	1.102	0.010	100	99
DKE-51	21.1	C	PMoz	4.5527	0.0948	0.3052	0.0054	3.2770	0.0581	0.1084	0.0004	0.6132	0.2096	1.717	0.027	1.741	0.017	1.776	0.007	99	97
DKE-51	22.1	C	PMoz	2.1451	0.0449	0.1975	0.0035	5.0637	0.0905	0.0783	0.0003	1.0008	0.3414	1.162	0.019	1.163	0.014	1.157	0.009	100	100
DKE-51	23.1	C	PMoz	6.4147	0.1335	0.3627	0.0064	2.7571	0.0490	0.1281	0.0005	0.3723	0.1267	1.995	0.030	2.034	0.018	2.069	0.007	98	96
DKE-51	24.1	C	PMoz	5.7565	0.1193	0.3482	0.0062	2.8718	0.0509	0.1207	0.0005	0.5523	0.1876	1.926	0.029	1.940	0.018	1.966	0.007	99	98
DKE-51	25.1	C	PMoz	1.6970	0.0362	0.1704	0.0031	5.8680	0.1065	0.0735	0.0005	1.0736	0.3641	1.014	0.017	1.007	0.014	1.028	0.015	101	99
DKE-51	26.1	C	PMoz	2.3662	0.0495	0.2118	0.0038	4.7221	0.0843	0.0813	0.0004	0.3736	0.1265	1.238	0.020	1.232	0.015	1.234	0.011	100	100
DKE-51	27.1	C	PMoz	1.4266	0.0340	0.1520	0.0041	6.5801	0.1788	0.0701	0.0006	0.3052	0.1539	0.912	0.023	0.900	0.014	0.930	0.019	101	98
DKE-51	28.1	C	PMoz	4.0646	0.0941	0.2857	0.0076	3.4999	0.0937	0.1028	0.0009	0.4139	0.2018	1.620	0.038	1.647	0.019	1.679	0.016	98	96
DKE-51	29.1	C	PMoz	1.6168	0.0371	0.1619	0.0043	6.1761	0.1633	0.0723	0.0004	0.2279	0.1076	0.967	0.024	0.977	0.014	0.995	0.011	99	97
DKE-51	30.1	C	PMoz	5.6127	0.1278	0.3331	0.0088	3.0019	0.0793	0.1228	0.0006	1.0360	0.4744	1.854	0.042	1.918	0.019	1.996	0.008	97	93
DKE-51	31.1	C	PMoz	6.4315	0.1512	0.3720	0.0100	2.6879	0.0725	0.1287	0.0010	0.4034	0.1793	2.039	0.047	2.037	0.020	2.078	0.014	100	98
DKE-51	32.1	C	PMoz	7.0577	0.1622	0.3702	0.0098	2.7012	0.0717	0.1375	0.0009	0.2660	0.1148	2.030	0.046	2.119	0.020	2.192	0.011	96	93
DKE-51	33.1	C	PMoz	1.4329	0.0331	0.1495	0.0040	6.6880	0.1785	0.0698	0.0004	0.3603	0.1515	0.898	0.022	0.903	0.014	0.920	0.012	100	98
DKE-51	34.1	C	PMoz	11.1904	0.2543	0.4730	0.0125	2.1140	0.0558	0.1724	0.0008	0.5582	0.2274	2.497	0.054	2.539	0.021	2.581	0.007	98	97
DKE-51	35.1	C	PMoz	1.4011	0.0322	0.1470	0.0039	6.8046	0.1800	0.0705	0.0004	0.2978	0.1181	0.884	0.022	0.889	0.014	0.941	0.011	99	94
DKE-51	36.1	C	PMoz	5.6288	0.1282	0.3345	0.0088	2.9900	0.0789	0.1216	0.0005	0.2462	0.0951	1.860	0.043	1.921	0.019	1.979	0.008	97	94
DKE-51	37.1	C	PMoz	1.3982	0.0347	0.1444	0.0038	6.9232	0.1844	0.0696	0.0005	0.5517	0.2078	0.870	0.022	0.888	0.015	0.915	0.016	98	95
DKE-51	38.1	C	PMoz	1.3966	0.0331	0.1484	0.0040	6.7401	0.1803	0.0702	0.0006	0.4473	0.1644	0.892	0.022	0.887	0.014	0.932	0.018	100	96
DKE-51	39.1	C	PMoz	6.0497	0.1780	0.3541	0.0108	2.8243	0.0865	0.1301	0.0026	0.2380	0.0896	1.954	0.051	1.983	0.025	2.097	0.034	99	93

Sample	Spot #	Spot pos.	Spot pos.	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
DKE-53	1.1	C	PMhnl	2.1297	0.0658	0.2004	0.0040	4.9900	0.0996	0.0776	0.0015	-0.1514	0.4199	1.177	0.021	1.158	0.021	1.139	0.038	102	103
DKE-53	2.1	C	PMoz	1.1335	0.0346	0.1271	0.0025	7.8688	0.1548	0.0651	0.0012	-0.0639	0.2029	0.771	0.014	0.769	0.016	0.774	0.040	100	100
DKE-53	3.1	R	PMoz	1.1198	0.0341	0.1269	0.0025	7.8833	0.1549	0.0648	0.0012	-0.0749	0.2780	0.770	0.014	0.763	0.016	0.762	0.040	101	101
DKE-53	4.1	R	PMoz	1.1402	0.0347	0.1266	0.0025	7.8981	0.1552	0.0656	0.0012	-0.0650	0.2906	0.769	0.014	0.773	0.016	0.789	0.040	99	97
DKE-53	5.1	R	PMhhl	5.4228	0.1669	0.3409	0.0068	2.9335	0.0588	0.1171	0.0023	-0.4722	2.6513	1.891	0.033	1.888	0.026	1.914	0.033	100	99
DKE-53	6.1	R	PMoz	2.1949	0.0664	0.1961	0.0039	5.0989	0.1003	0.0817	0.0015	-0.0256	0.1935	1.154	0.021	1.179	0.021	1.242	0.037	98	93
DKE-53	8.1	C	PMoz	4.5147	0.1355	0.3059	0.0060	3.2690	0.0637	0.1079	0.0020	-0.0711	1.7259	1.721	0.029	1.734	0.025	1.768	0.033	99	97
DKE-53	9.1	C	PMcz	1.8866	0.0568	0.1723	0.0034	5.8031	0.1134	0.0809	0.0015	0.0026	0.5955	1.025	0.018	1.076	0.020	1.224	0.037	95	84
DKE-53	10.1	C	PMoz	2.1682	0.0648	0.2014	0.0039	4.9641	0.0969	0.0787	0.0015	0.0408	0.8148	1.183	0.021	1.171	0.021	1.169	0.037	101	101
DKE-53	11.1	C	PMoz	1.0758	0.0324	0.1227	0.0024	8.1497	0.1599	0.0644	0.0012	0.0387	0.4041	0.746	0.014	0.742	0.016	0.750	0.040	101	100
DKE-53	12.1	R	PMoz	1.1225	0.0334	0.1256	0.0024	7.9594	0.1537	0.0658	0.0012	0.0467	0.3303	0.763	0.014	0.764	0.016	0.795	0.039	100	96
DKE-53	13.1	C	PMoz	1.1202	0.0332	0.1246	0.0024	8.0273	0.1547	0.0662	0.0012	0.0503	0.2690	0.757	0.014	0.763	0.016	0.810	0.039	99	93
DKE-53	14.1	C	PMhnl	1.5071	0.0123	0.1549	0.0016	6.4577	0.0648	0.0717	0.0005	0.0234	0.0058	0.928	0.009	0.933	0.005	0.978	0.014	99	95
DKE-53	15.1	C	PMoz	3.3168	0.0239	0.2556	0.0025	3.9119	0.0382	0.0954	0.0006	0.1531	0.0316	1.467	0.013	1.485	0.006	1.541	0.012	99	95
DKE-53	16.1	C	PMhnl	1.3111	0.0088	0.1247	0.0012	8.0195	0.0796	0.0667	0.0005	0.1077	0.0220	0.758	0.007	0.768	0.004	0.825	0.015	99	92
DKE-53	17.1	C	PMhhl	4.1376	0.0401	0.2908	0.0030	3.4383	0.0354	0.1048	0.0010	0.6211	0.1250	1.646	0.015	1.662	0.008	1.714	0.017	99	96
DKE-53	18.1	C	PMhhl	4.3375	0.0752	0.2986	0.0057	3.3490	0.0636	0.1078	0.0019	0.7375	0.1481	1.684	0.028	1.701	0.014	1.766	0.032	99	95
DKE-53	19.1	C	PMoz	1.1345	0.0088	0.1285	0.0013	7.7827	0.0780	0.0651	0.0004	0.0833	0.0165	0.779	0.007	0.770	0.004	0.772	0.014	101	101
DKE-53	20.1	C	PMoz	4.4502	0.0335	0.2989	0.0030	3.3455	0.0332	0.1094	0.0007	0.4468	0.0867	1.686	0.015	1.722	0.006	1.792	0.012	98	94
DKE-53	21.1	C	PMoz	4.1729	0.0326	0.2801	0.0028	3.5699	0.0358	0.1089	0.0007	0.4333	0.0829	1.592	0.014	1.669	0.006	1.784	0.011	95	89
DKE-53	22.1	C	PMhhl	2.6032	0.0333	0.2271	0.0027	4.4036	0.0522	0.0846	0.0011	0.2614	0.0510	1.319	0.014	1.302	0.009	1.312	0.026	101	101
DKE-53	23.1	R	PMoz	1.9781	0.0146	0.1825	0.0018	5.4800	0.0538	0.0793	0.0005	0.2663	0.0497	1.081	0.010	1.108	0.005	1.183	0.013	98	91
DKE-53	24.1	C	PMhnl	7.3549	0.0557	0.3888	0.0039	2.5721	0.0256	0.1386	0.0009	0.2911	0.0537	2.117	0.018	2.156	0.007	2.205	0.011	98	96
DKE-53	25.1	C	PMoz	1.5348	0.0115	0.1590	0.0016	6.2912	0.0627	0.0711	0.0005	0.3199	0.0584	0.951	0.009	0.944	0.005	0.958	0.014	101	99
DKE-53	26.1	C	PMoz	1.1538	0.0092	0.1241	0.0012	8.0556	0.0796	0.0674	0.0005	0.1280	0.0231	0.754	0.007	0.779	0.004	0.846	0.015	97	89
DKE-53	27.1	C	PMhnl	1.0693	0.0250	0.1196	0.0019	8.3603	0.1358	0.0657	0.0006	0.1213	0.0467	0.728	0.011	0.738	0.012	0.794	0.018	99	92
DKE-53	28.1	C	PMoz	2.0244	0.0460	0.1878	0.0030	5.3237	0.0858	0.0778	0.0006	0.5126	0.1979	1.110	0.016	1.124	0.015	1.145	0.015	99	97
DKE-53	29.1	C	PMhhl	13.5378	0.3070	0.5069	0.0083	1.9728	0.0322	0.1958	0.0015	0.5565	0.2162	2.643	0.035	2.718	0.021	2.797	0.012	97	95
DKE-53	30.1	C	PMoz	2.0537	0.0461	0.1872	0.0030	5.3412	0.0857	0.0801	0.0006	0.1539	0.0601	1.106	0.016	1.134	0.015	1.204	0.014	98	92
DKE-53	31.1	C	PMoz	5.0986	0.1144	0.3209	0.0051	3.1160	0.0500	0.1157	0.0008	0.6529	0.2569	1.794	0.025	1.836	0.019	1.892	0.013	98	95
DKE-53	32.1	C	PMoz	1.1599	0.0262	0.1269	0.0020	7.8785	0.1271	0.0667	0.0005	0.1452	0.0575	0.770	0.012	0.782	0.012	0.825	0.016	99	93
DKE-53	33.1	C	PMoz	5.2069	0.1163	0.3185	0.0051	3.1392	0.0504	0.1200	0.0009	1.1612	0.4624	1.783	0.025	1.854	0.019	1.956	0.013	96	91
DKE-53	34.1	C	PMhnl	1.1840	0.0267	0.1304	0.0021	7.6712	0.1231	0.0663	0.0005	0.1146	0.0463	0.790	0.012	0.793	0.012	0.811	0.016	100	97
DKE-53	35.1	C	PMoz	1.9895	0.0457	0.1781	0.0029	5.6157	0.0925	0.0822	0.0006	0.1350	0.0546	1.056	0.016	1.112	0.015	1.255	0.015	95	84
DKE-53	36.1	C	PMoz	4.4880	0.1019	0.3004	0.0049	3.3289	0.0545	0.1086	0.0008	0.7338	0.2984	1.693	0.024	1.729	0.019	1.778	0.014	98	95
DKE-53	37.1	C	PMhhl	1.7498	0.0440	0.1710	0.0029	5.8482	0.0993	0.0786	0.0007	0.3031	0.1245	1.018	0.016	1.027	0.016	1.164	0.018	99	87
DKE-53	38.1	C	PMhhl	2.2636	0.0516	0.2033	0.0033	4.9179	0.0789	0.0802	0.0007	0.5199	0.2143	1.193	0.017	1.201	0.016	1.207	0.017	99	99
DKE-53	39.1	R	PMoz	1.6153	0.0359	0.1626	0.0026	6.1499	0.0984	0.0721	0.0005	0.4563	0.1893	0.971	0.014	0.976	0.014	0.988	0.015	99	98
DKE-53	40.1	C	PMhhl	18.3685	0.3756	0.5848	0.0065	1.7099	0.0191	0.2275	0.0031	-0.9813	6.9650	2.968	0.027	3.009	0.020	3.039	0.021	99	98
DKE-53	42.1	C	PMoz	12.2847	0.2497	0.4827	0.0053	2.0717	0.0228	0.1853	0.0025	-3.4912	13.2764	2.539	0.023	2.626	0.019	2.704	0.022	97	94
DKE-53	43.1	C	PMoz	5.4768	0.1111	0.3492	0.0038	2.8636	0.0314	0.1144	0.0015	-2.8656	8.8444	1.931	0.018	1.897	0.017	1.873	0.023	102	103
DKE-53	44.1	C	PMoz	3.9909	0.0808	0.2928	0.0032	3.4159	0.0378	0.0992	0.0013	-6.0982	15.8379	1.655	0.016	1.632	0.016	1.615	0.025	101	102
DKE-53	45.1	C	PMoz	1.4512	0.0303	0.1472	0.0016	6.7939	0.0757	0.0721	0.0010	-4.7922	10.7429	0.885	0.009	0.910	0.012	0.988	0.029	97	90
DKE-53	46.1	C	PMoz	2.4193	0.0491	0.2149	0.0024	4.6541	0.0519	0.0820	0.0011	-2.3963	4.8056	1.255	0.013	1.248	0.014	1.249	0.026	101	100
DKE-53	47.1	C	PMoz	3.4897	0.0706	0.2615	0.0029	3.8248	0.0425	0.0973	0.0013	-2.8301	4.9813	1.497	0.015	1.525	0.016	1.578	0.024	98	95
DKE-53	48.1	C	PMoz	5.5854	0.1121	0.3444	0.0038	2.9040	0.0317	0.1187	0.0016	-6.0150	9.5599	1.908	0.018	1.914	0.017	1.938	0.023	100	98
DKE-53	49.1	C	PMhnl	2.0281	0.0409	0.1859	0.0020	5.3799	0.0590	0.0793	0.0011	-3.3002	4.7814	1.099	0.011	1.125	0.014	1.183	0.027	98	93
DKE-53	50.1	C	PMhnl	2.5808	0.0517	0.2175	0.0024	4.5984	0.0503	0.0864	0.0011	-4.4705	5.9506	1.268	0.013	1.295	0.015	1.353	0.026	98	95
DKE-53	51.1	C	PMhnl	6.5427	0.1308	0.3655	0.0040	2.7356	0.0300	0.1306	0.0017	-5.7890	7.1267	2.008	0.019	2.052	0.017	2.103	0.023	98	94
DKE-53	52.1	C	PMoz	1.4632	0.0320	0.1427	0.0016	7.0081	0.0807	0.0786	0.0011	-2.0083	2.2998	0.860	0.009	0.915	0.013	1.165	0.027	94	74
DKE-53	53.1	C	PMhnl	3.3320	0.0962	0.2649	0.0084	3.7757	0.1193	0.0923	0.0010	0.2686	0.3522	1.515	0.043	1.489	0.022	1.479	0.020	102	102
DKE-53	54.1	R	PMoz	3.3527	0.0962	0.2420	0.0076	4.1323	0.1305	0.0988	0.0010	0.4283	0.5056	1.397	0.040	1.493	0.022	1.607	0.019	94	87
DKE-53	55.1	C	PMoz	1.9970	0.0591	0.1901	0.0060	5.2610	0.1663	0.0778	0.0010	0.5841	0.6269	1.122	0.032	1.114	0.020	1.145	0.025	101	98
DKE-53	58.1	C	PMoz	1.6835	0.0478	0.1618	0.0051	6.1823	0.1936	0.0762	0.0008	0.2018	0.1703	0.967	0.028	1.002	0.018	1.104	0.020	96	88
DKE-53	59.1	C	PMoz	3.8808	0.1102	0.2669	0.0084	3.7468	0.1173	0.1078	0.0011	0.6609	0.5254	1.525	0.042	1.610	0.023				

DKE-06	14.1	C	PMhml	6.4379	0.0483	0.3565	0.0029	2.8052	0.0230	0.1308	0.0015	0.5898	0.2382	1.965	0.014	2.037	0.007	2.106	0.019	96	93
DKE-06	16.1	C	PMhml	6.8832	0.0477	0.3730	0.0032	2.6810	0.0229	0.1360	0.0015	0.3828	0.1642	2.043	0.015	2.096	0.006	2.173	0.019	97	94
DKE-06	19.1	C	PMoz	6.0673	0.0791	0.3428	0.0047	2.9174	0.0397	0.1299	0.0014	0.3823	0.1807	1.900	0.022	1.986	0.011	2.094	0.019	96	91
DKE-06	23.1	C	PMoz	5.7762	0.0400	0.3312	0.0027	3.0195	0.0245	0.1271	0.0014	0.2715	0.1547	1.844	0.013	1.943	0.006	2.057	0.019	95	90
DKE-06	24.1	C	PMhml	6.7770	0.1150	0.3835	0.0064	2.6076	0.0438	0.1328	0.0016	0.2083	0.1239	2.093	0.030	2.083	0.015	2.132	0.020	100	98
DKE-06	25.1	C	PMoz	6.8990	0.0545	0.3675	0.0033	2.7208	0.0243	0.1350	0.0015	0.2222	0.1379	2.018	0.015	2.099	0.007	2.161	0.019	96	93
DKE-06	26.1	C	PMoz	7.0756	0.1647	0.4008	0.0076	2.4951	0.0473	0.1327	0.0019	0.0917	0.0881	2.173	0.035	2.121	0.021	2.131	0.024	102	102
DKE-06	27.1	C	PMoz	6.2043	0.0946	0.3465	0.0032	2.8862	0.0263	0.1304	0.0017	0.4806	0.4345	1.918	0.015	2.005	0.013	2.101	0.022	96	91
DKE-06	28.2N	C	PMoz	9.2539	0.1428	0.4245	0.0041	2.3558	0.0229	0.1587	0.0020	0.4292	0.3496	2.281	0.019	2.363	0.014	2.439	0.022	97	94
DKE-06	29.1	C	PMoz	6.5338	0.1041	0.3612	0.0038	2.7688	0.0291	0.1338	0.0017	0.2747	0.2132	1.988	0.018	2.050	0.014	2.145	0.022	97	93
DKE-06	32.1	C	PMoz	7.1723	0.2413	0.3875	0.0118	2.5807	0.0784	0.1415	0.0020	0.4934	0.3357	2.111	0.054	2.133	0.030	2.241	0.024	99	94
DKE-06	33.1	C	PMoz	7.4754	0.1108	0.3986	0.0036	2.5086	0.0227	0.1368	0.0017	0.2988	0.1951	2.163	0.017	2.170	0.013	2.183	0.022	100	99
DKE-06	34.1	C	PMoz	6.6818	0.1051	0.3735	0.0039	2.6773	0.0277	0.1317	0.0017	0.9333	0.5860	2.046	0.018	2.070	0.014	2.117	0.022	99	97
DKE-06	35.1	C	PMoz	5.9921	0.0901	0.3486	0.0032	2.8684	0.0263	0.1262	0.0016	0.0742	0.0449	1.928	0.015	1.975	0.013	2.044	0.022	98	94
DKE-06	40.1	C	PMoz	7.0304	0.1025	0.3839	0.0044	2.6048	0.0297	0.1348	0.0009	0.4515	0.2993	2.095	0.020	2.115	0.013	2.158	0.011	99	97
DKE-06	42.1	C	PMoz	6.3441	0.0866	0.3554	0.0035	2.8138	0.0273	0.1299	0.0009	0.5730	0.4115	1.960	0.016	2.025	0.012	2.094	0.012	97	94
DKE-06	44.1	C	PMoz	7.0995	0.0962	0.3790	0.0037	2.6383	0.0261	0.1373	0.0009	0.2778	0.2162	2.072	0.018	2.124	0.012	2.190	0.011	98	95
DKE-06	45.1	C	PMhml	6.8089	0.0939	0.3684	0.0036	2.7142	0.0262	0.1347	0.0010	0.5157	0.4232	2.022	0.017	2.087	0.012	2.156	0.013	97	94
DKE-06	46.1	C	PMhml	6.4206	0.0850	0.3587	0.0034	2.7875	0.0261	0.1304	0.0009	1.1054	0.9523	1.976	0.016	2.035	0.012	2.101	0.011	97	94
DKE-06	47.1	C	PMoz	6.0778	0.1560	0.3440	0.0088	2.9073	0.0744	0.1286	0.0011	0.2025	0.1842	1.906	0.042	1.987	0.022	2.077	0.015	96	92
DKE-06	48.1	C	PMoz	6.2306	0.0849	0.3549	0.0035	2.8176	0.0276	0.1290	0.0010	0.3407	0.3261	1.958	0.017	2.009	0.012	2.082	0.013	97	94
DKE-06	49.1	C	PMoz	6.6097	0.0875	0.3587	0.0034	2.7875	0.0267	0.1351	0.0009	0.1338	0.1356	1.976	0.016	2.061	0.012	2.161	0.012	96	91
DKE-06	50.1	C	PMoz	6.6165	0.0875	0.3589	0.0034	2.7866	0.0266	0.1351	0.0009	0.1260	0.1357	1.977	0.016	2.062	0.012	2.162	0.012	96	91
DKE-06	51.1	C	PMoz	6.6088	0.1482	0.3697	0.0076	2.7050	0.0555	0.1318	0.0011	0.3615	0.0853	2.028	0.036	2.061	0.020	2.119	0.015	98	96
DKE-06	52.1	C	PMoz	7.3742	0.1657	0.3880	0.0080	2.5771	0.0529	0.1387	0.0011	0.1821	0.0431	2.114	0.037	2.158	0.020	2.207	0.014	98	96
DKE-06	53.1	C	PMhml	6.8720	0.1528	0.3701	0.0075	2.7023	0.0550	0.1362	0.0011	0.2575	0.0613	2.030	0.035	2.095	0.020	2.175	0.014	97	93
DKE-06	55.1	C	PMhml	7.1835	0.1602	0.3785	0.0077	2.6420	0.0537	0.1383	0.0011	0.0643	0.0155	2.069	0.036	2.134	0.020	2.202	0.014	97	94
DKE-06	56.1	C	PMoz	6.7787	0.1517	0.3770	0.0077	2.6523	0.0542	0.1309	0.0011	0.5009	0.1206	2.062	0.036	2.083	0.020	2.107	0.014	99	98
DKE-06	57.1	C	PMoz	6.2954	0.1389	0.3648	0.0074	2.7414	0.0554	0.1271	0.0010	0.3419	0.0825	2.005	0.035	2.018	0.019	2.056	0.014	99	98
DKE-06	58.1	C	PMoz	8.9769	0.1977	0.4316	0.0087	2.3172	0.0468	0.1529	0.0013	0.2214	0.0537	2.313	0.039	2.336	0.020	2.375	0.014	99	97
DKE-06	59.1	C	PMoz	7.2614	0.1598	0.3830	0.0077	2.6108	0.0526	0.1385	0.0011	0.1888	0.0459	2.090	0.036	2.144	0.019	2.205	0.014	97	95
DKE-06	60.1	C	PMoz	6.4293	0.1421	0.3522	0.0071	2.8397	0.0572	0.1333	0.0011	0.2645	0.0646	1.945	0.034	2.036	0.019	2.139	0.014	96	91
DKE-06	61.1	C	PMoz	6.2870	0.1376	0.3586	0.0072	2.7889	0.0560	0.1295	0.0011	0.2074	0.0509	1.975	0.034	2.017	0.019	2.089	0.014	98	95
DKE-06	62.1	C	PMoz	6.5704	0.1437	0.3703	0.0074	2.7008	0.0542	0.1307	0.0011	0.4443	0.1095	2.031	0.035	2.055	0.019	2.104	0.014	99	96
Metamorphic domains (overgrowths)																					
DKE-06	15.1B	R	Mog	0.7701	0.0073	0.0939	0.0008	10.6529	0.0921	0.0601	0.0008	0.0180	0.0076	0.578	0.005	0.580	0.004	0.601	0.027	100	96
DKE-06	21.1 B1	R	Mog	0.7615	0.0074	0.0920	0.0008	10.8654	0.0904	0.0608	0.0008	0.0729	0.0371	0.568	0.005	0.575	0.004	0.627	0.027	99	91
DKE-06	21.2 B2	R	Mog	0.7024	0.0077	0.0878	0.0008	11.3897	0.1066	0.0587	0.0008	0.0114	0.0076	0.543	0.005	0.540	0.005	0.549	0.031	100	99
Zircons not included on related diagrams																					
DKE-06	1.1	C	PMoz	5.0196	0.0690	0.3005	0.0013	3.3276	0.0149	0.1236	0.0009	0.4468	0.0839	1.694	0.007	1.823	0.012	2.007	0.013	93	84
DKE-06	13.1	C	PMoz	4.7904	0.0683	0.2857	0.0016	3.5006	0.0191	0.1239	0.0010	0.3189	0.0539	1.620	0.008	1.783	0.012	2.012	0.014	91	81
DKE-06	17.1	C	PMoz	5.2650	0.0443	0.3027	0.0026	3.3031	0.0288	0.1286	0.0015	0.1036	0.0459	1.705	0.013	1.863	0.007	2.077	0.020	92	82
DKE-06	18.1	C	PMoz	4.9585	0.0330	0.2850	0.0023	3.5089	0.0287	0.1280	0.0014	0.1994	0.0911	1.616	0.012	1.812	0.006	2.068	0.019	89	78
DKE-06	20.1	C	PMoz	4.7369	0.0366	0.2750	0.0024	3.6366	0.0311	0.1258	0.0014	0.3081	0.1501	1.566	0.012	1.774	0.006	2.038	0.019	88	77
DKE-06	22.1	C	PMoz	4.2217	0.0337	0.2610	0.0024	3.8309	0.0350	0.1238	0.0014	0.1758	0.0962	1.495	0.012	1.678	0.007	2.010	0.020	89	74
DKE-06	28.1B	R	Mog	0.6457	0.0139	0.0787	0.0009	12.7017	0.1430	0.0613	0.0012	0.0859	0.0750	0.489	0.005	0.506	0.009	0.644	0.043	97	76
DKE-06	30.1	C	PMoz	5.2915	0.0818	0.3054	0.0030	3.2743	0.0318	0.1269	0.0016	0.6252	0.4635	1.718	0.015	1.867	0.013	2.053	0.022	92	84
DKE-06	31.1	C	PMoz	6.3863	0.1718	0.3320	0.0060	3.0120	0.0548	0.1354	0.0019	0.4188	0.2988	1.848	0.029	2.030	0.023	2.165	0.024	91	85
DKE-06	36.1	C	PMoz	2.1921	0.0390	0.1546	0.0019	6.4681	0.0804	0.1020	0.0013	0.7185	0.4192	0.927	0.011	1.179	0.012	1.666	0.024	79	56
DKE-06	37.1	C	PMoz	6.0924	0.1039	0.3423	0.0039	2.9212	0.0331	0.1325	0.0017	0.2316	0.1305	1.898	0.019	1.989	0.015	2.128	0.022	95	89
DKE-06	38.1	C	PMoz	6.2518	0.0960	0.3404	0.0038	2.9378	0.0324	0.1362	0.0014	0.9243	0.5693	1.889	0.018	2.012	0.013	2.176	0.017	94	87
DKE-06	39.1	C	PMoz	3.9238	0.0536	0.2696	0.0026	3.7095	0.0359	0.1058	0.0008	0.8682	0.5542	1.539	0.013	1.619	0.011	1.733	0.013	95	89
DKE-06	41.1	C	PMoz	4.3305	0.0621	0.2659	0.0027	3.7615	0.0387	0.1188	0.0008	0.6252	0.4310	1.520	0.014	1.699	0.012	1.939	0.012	89	78
DKE-06	43.1	C	PMoz	6.5065	0.0886	0.3502	0.0035	2.8554	0.0281	0.1356	0.0009	0.5419	0.4060	1.936	0.016	2.047	0.012	2.168	0.012	95	89
DKE-06	54.1	C	PMoz	5.2497	0.1175	0.3118	0.0063	3.2074	0.0653	0.1240	0.0011	0.1208	0.0289	1.749	0.031	1.861	0.019	2.013	0.015	94	87
DKE-06	63.1	C	PMoz	5.0830	0.1123	0.3098	0.0063	3.2282	0.0652	0.1203	0.0010	0.1333	0.0330	1.740	0.031	1.833	0.019	1.960	0.014	95	89

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1σ	²⁰⁶ Pb/ ²³⁸ U	±1σ	²³⁸ Pb/ ²⁰⁶ U	±1σ	²⁰⁷ Pb/ ²⁰⁶ U	±1σ	²⁰⁸ Pb/ ²⁰⁶ U	±1σ	²⁰⁶ Pb/ ²³⁸ U	±1σ	²⁰⁷ Pb/ ²³⁵ U	±1σ	²⁰⁷ Pb/ ²⁰⁶ U	±1σ	Conc. (%)	Conc. (%)
DKE-56	1.1	C	PMoz	9.8996	0.0720	0.4482	0.0036	2.2310	0.0178	0.1620	0.0008	0.8088	0.7135	2.387	0.016	2.425	0.007	2.474	0.008	98	96
DKE-56	2.1	C	PMoz	6.6111	0.0500	0.3565	0.0029	2.8052	0.0230	0.1361	0.0007	1.3817	1.2833	1.965	0.014	2.061	0.007	2.174	0.008	95	90
DKE-56	3.2N	C	PMoz	7.4496	0.0542	0.3964	0.0032	2.5227	0.0202	0.1376	0.0006	0.6740	0.6999	2.152	0.015	2.167	0.006	2.194	0.008	99	98
DKE-56	5.1	C	PMoz	7.3581	0.0544	0.3916	0.0032	2.5533	0.0207	0.1376	0.0007	0.2929	0.3660	2.130	0.015	2.156	0.007	2.194	0.009	99	97
DKE-56	6.1	C	PMoz	6.2579	0.0458	0.3549	0.0028	2.8175	0.0225	0.1294	0.0006	0.6600	0.8972	1.958	0.013	2.013	0.006	2.087	0.008	97	94
DKE-56	7.1	C	PMoz	10.6287	0.0781	0.4470	0.0036	2.2372	0.0179	0.1734	0.0008	0.9947	1.4656	2.382	0.016	2.491	0.007	2.591	0.008	96	92
DKE-56	8.1	C	PMoz	6.4421	0.0467	0.3639	0.0029	2.7482	0.0218	0.1298	0.0006	0.4314	0.6938	2.001	0.014	2.038	0.006	2.092	0.008	98	96
DKE-56	9.2	C	PMoz	6.3347	0.0472	0.3604	0.0029	2.7747	0.0222	0.1295	0.0006	0.4244	0.8353	1.984	0.014	2.023	0.007	2.088	0.009	98	95
DKE-56	10.1	C	PMhml	6.8521	0.0499	0.3823	0.0030	2.6159	0.0208	0.1313	0.0006	0.9195	2.0376	2.087	0.014	2.092	0.006	2.113	0.008	100	99
DKE-56	11.1	C	PMoz	6.3203	0.0305	0.3634	0.0008	2.7515	0.0060	0.1282	0.0006	0.7837	0.2944	1.998	0.004	2.021	0.004	2.070	0.008	99	97
DKE-56	12.1	C	PMoz	7.1753	0.0348	0.3856	0.0008	2.5933	0.0056	0.1362	0.0006	0.2749	0.1029	2.102	0.004	2.133	0.004	2.176	0.008	99	97
DKE-56	13.1	R	PMoz	6.7884	0.0356	0.3815	0.0009	2.6215	0.0064	0.1304	0.0006	0.8025	0.2995	2.083	0.004	2.084	0.005	2.100	0.008	100	99
DKE-56	14.1	C	PMoz	7.1075	0.0345	0.3841	0.0008	2.6033	0.0057	0.1358	0.0006	0.5337	0.1985	2.096	0.004	2.125	0.004	2.171	0.008	99	97
DKE-56	15.1	C	PMhml	6.4615	0.0344	0.3625	0.0012	2.7586	0.0089	0.1310	0.0006	1.3750	0.5098	1.994	0.006	2.041	0.005	2.108	0.008	98	95
DKE-56	16.1	C	PMoz	10.1971	0.0489	0.4614	0.0011	2.1672	0.0051	0.1628	0.0007	0.5296	0.1957	2.446	0.005	2.453	0.004	2.482	0.007	100	99
DKE-56	17.1	C	PMoz	8.2718	0.0395	0.4017	0.0010	2.4896	0.0060	0.1507	0.0007	0.5161	0.1902	2.177	0.004	2.261	0.004	2.350	0.008	96	93
DKE-56	18.1	C	PMoz	5.9621	0.0282	0.3412	0.0007	2.9307	0.0061	0.1284	0.0006	0.1773	0.0653	1.893	0.003	1.970	0.004	2.073	0.008	96	91
DKE-56	19.1	C	PMoz	6.4233	0.0303	0.3615	0.0008	2.7662	0.0059	0.1305	0.0006	0.1396	0.0511	1.989	0.004	2.035	0.004	2.102	0.008	98	95
DKE-56	20.1	C	PMoz	6.3154	0.0306	0.3591	0.0008	2.7850	0.0062	0.1289	0.0006	0.4029	0.1470	1.978	0.004	2.021	0.004	2.081	0.008	98	95
DKE-56	21.1	C	PMoz	6.3240	0.0384	0.3567	0.0014	2.8031	0.0109	0.1303	0.0006	0.6544	0.2380	1.967	0.007	2.022	0.005	2.100	0.008	97	94
DKE-56	22.1	C	PMoz	6.9011	0.0412	0.3905	0.0016	2.5605	0.0102	0.1297	0.0006	0.7106	0.2576	2.125	0.007	2.099	0.005	2.091	0.008	101	102
DKE-56	23.1	C	PMoz	6.3565	0.0298	0.3497	0.0007	2.8592	0.0060	0.1335	0.0006	0.4488	0.1621	1.933	0.004	2.026	0.004	2.141	0.007	95	90
DKE-56	24.1	C	PMhll	6.1074	0.0511	0.3443	0.0017	2.9044	0.0146	0.1302	0.0006	0.3519	0.1450	1.907	0.008	1.991	0.007	2.098	0.007	96	91
DKE-56	25.1	C	PMoz	6.3277	0.0533	0.3590	0.0018	2.7854	0.0143	0.1294	0.0006	0.8220	0.3494	1.978	0.009	2.022	0.007	2.087	0.008	98	95
DKE-56	27.1	C	PMoz	7.2498	0.0621	0.3993	0.0021	2.5041	0.0129	0.1332	0.0006	0.3337	0.1513	2.166	0.009	2.143	0.008	2.137	0.008	101	101
DKE-56	28.2	C	PMoz	6.5074	0.0547	0.3650	0.0019	2.7401	0.0140	0.1305	0.0006	0.5376	0.2612	2.006	0.009	2.047	0.007	2.102	0.008	98	95
DKE-56	29.1B	R	PMoz	5.9668	0.0508	0.3409	0.0018	2.9337	0.0155	0.1284	0.0006	0.3770	0.1890	1.891	0.009	1.971	0.007	2.074	0.008	96	91
DKE-56	29.2	C	PMoz	6.5821	0.0561	0.3718	0.0019	2.6895	0.0138	0.1292	0.0006	0.5925	0.3101	2.038	0.009	2.057	0.007	2.085	0.008	99	91
DKE-56	30.1	C	PMhml	7.3386	0.0616	0.3893	0.0020	2.5684	0.0131	0.1379	0.0006	0.5123	0.2789	2.120	0.009	2.154	0.007	2.197	0.007	98	96
DKE-56	31.1	C	PMoz	6.3173	0.0531	0.3555	0.0018	2.8127	0.0144	0.1300	0.0006	0.3216	0.1824	1.961	0.009	2.021	0.007	2.095	0.007	97	94
DKE-56	32.1	C	PMhml	7.3248	0.0768	0.3946	0.0033	2.5340	0.0211	0.1354	0.0006	0.6709	0.3972	2.144	0.015	2.152	0.009	2.166	0.008	100	99
DKE-56	33.1	C	PMoz	6.7033	0.0587	0.3781	0.0021	2.6449	0.0145	0.1294	0.0006	0.7848	0.4857	2.067	0.010	2.073	0.008	2.088	0.008	100	99
DKE-56	34.1	C	PMhhl	6.4343	0.0611	0.3676	0.0023	2.7201	0.0168	0.1284	0.0007	1.1463	0.7435	2.018	0.011	2.037	0.008	2.074	0.009	99	97
DKE-56	46.1	C	PMhml	6.7659	0.0967	0.3808	0.0036	2.6262	0.0252	0.1308	0.0008	0.4125	0.1000	2.080	0.017	2.081	0.013	2.105	0.011	100	99
DKE-56	47.1	C	PMoz	6.5353	0.0926	0.3672	0.0035	2.7230	0.0259	0.1304	0.0008	0.3067	0.0725	2.016	0.016	2.051	0.012	2.100	0.010	98	96
DKE-56	48.1	C	PMhml	7.4121	0.1044	0.3939	0.0037	2.5385	0.0240	0.1377	0.0008	0.2256	0.0521	2.141	0.017	2.162	0.013	2.195	0.010	99	98
DKE-56	49.1	C	PMoz	6.0590	0.0890	0.3479	0.0035	2.8744	0.0290	0.1281	0.0008	0.3957	0.0895	1.925	0.017	1.984	0.013	2.070	0.011	97	93
DKE-56	50.1N	C	PMoz	6.4503	0.0925	0.3638	0.0035	2.7485	0.0267	0.1300	0.0008	0.6681	0.1478	2.000	0.017	2.039	0.013	2.096	0.010	98	95
DKE-56	50.2B	R	PMoz	6.3486	0.0900	0.3543	0.0035	2.8222	0.0277	0.1309	0.0008	0.1064	0.0234	1.955	0.017	2.025	0.012	2.107	0.010	97	93
DKE-56	51.1	C	PMoz	7.1630	0.1010	0.3847	0.0037	2.5994	0.0248	0.1364	0.0008	0.2494	0.0531	2.098	0.017	2.132	0.012	2.178	0.010	98	96
DKE-56	52.1	C	PMoz	6.3943	0.0900	0.3615	0.0034	2.7666	0.0262	0.1296	0.0007	0.3735	0.0777	1.989	0.016	2.031	0.012	2.090	0.010	98	95
DKE-56	53.1	C	PMoz	9.8156	0.1425	0.4643	0.0046	2.1538	0.0213	0.1567	0.0010	1.0664	0.2176	2.459	0.020	2.418	0.013	2.416	0.011	102	102
DKE-56	54.1	C	PMoz	6.4889	0.0926	0.3655	0.0035	2.7357	0.0262	0.1300	0.0008	0.2656	0.0532	2.008	0.016	2.044	0.012	2.095	0.010	98	96
DKE-56	55.1	C	PMoz	6.5676	0.0925	0.3700	0.0035	2.7031	0.0258	0.1299	0.0007	0.5374	0.1054	2.029	0.017	2.055	0.012	2.094	0.010	99	97
DKE-56	56.1	C	PMoz	7.4255	0.1048	0.3942	0.0038	2.5369	0.0243	0.1378	0.0008	0.2356	0.0454	2.142	0.017	2.164	0.013	2.195	0.010	99	98
DKE-56	57.1	C	PMoz	7.0363	0.0991	0.3767	0.0036	2.6544	0.0253	0.1362	0.0008	0.2713	0.0513	2.061	0.017	2.116	0.012	2.176	0.010	97	95
DKE-56	60.1	C	PMoz	6.4930	0.0308	0.3637	0.0017	2.7495	0.0127	0.1303	0.0008	0.6128	0.2384	2.000	0.008	2.045	0.004	2.099	0.011	98	95
DKE-56	62.1	C	PMoz	7.1484	0.0327	0.3852	0.0016	2.5964	0.0106	0.1357	0.0008	0.0987	0.0373	2.100	0.007	2.130	0.004	2.170	0.010	99	97
DKE-56	63.1	C	PMoz	6.4754	0.0310	0.3675	0.0015	2.7210	0.0112	0.1291	0.0008	0.2101	0.0784	2.018	0.007	2.043	0.004	2.083	0.010	99	97
DKE-56	64.1	C	PMoz	6.7421	0.0316	0.3677	0.0015	2.7195	0.0111	0.1343	0.0008	0.2189	0.0806	2.019	0.007	2.078	0.004	2.151	0.010	97	94
DKE-56	65.1	C	PMoz	6.3704	0.0297	0.3647	0.0015	2.7423	0.0110	0.1280	0.0007	0.2210	0.0801	2.004	0.007	2.028	0.004	2.069	0.010	99	97
DKE-56	66.1	C	PMoz	7.3545	0.0337	0.3911	0.0016	2.5572	0.0104	0.1377	0.0008	0.2213	0.0792	2.128	0.007	2.155	0.004	2.195	0.010	99	97
DKE-56	68.1	C	PMoz	6.3254	0.0305	0.3618	0.0016	2.7640	0.0119	0.1282	0.0007	0.1260	0.0440	1.991	0.007	2.022	0.004	2.071	0.010	98	96
DKE-56	69.1	C	PMoz	6.4506	0.0316	0.3651	0.0016	2.7392	0.0117	0.1296	0.0008	0.2608	0.0897	2.006	0.007	2.039	0.004	2.090	0.010	98	96
DKE-56	70.2N	C	PMhml	6.0798	0.0542	0.3516	0.0017	2.8444	0.0140	0.1253	0.0007	0.4012	0.1496	1.942	0.008	1.987	0.008	2.031	0.010		

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
NCEB-351	1.1	C	PMoz	7.1395	0.0772	0.3919	0.0027	2.5514	0.0174	0.1343	0.0013	0.2107	0.0305	2.132	0.012	2.129	0.010	2.152	0.016	101	99
NCEB-351	2.1	C	PMoz	7.1007	0.0868	0.3841	0.0032	2.6033	0.0219	0.1358	0.0013	0.2435	0.0352	2.096	0.015	2.124	0.011	2.171	0.017	104	97
NCEB-351	3.1	C	PMoz	6.3539	0.0707	0.3643	0.0026	2.7450	0.0192	0.1283	0.0012	0.2117	0.0309	2.002	0.012	2.026	0.010	2.072	0.016	103	97
NCEB-351	4.1	C	PMoz	6.8435	0.0732	0.3961	0.0026	2.5248	0.0166	0.1260	0.0012	0.5965	0.0885	2.151	0.012	2.091	0.009	2.040	0.017	95	105
NCEB-351	5.1	C	PMoz	14.5873	0.1563	0.5286	0.0035	1.8917	0.0127	0.2025	0.0019	0.4016	0.0601	2.736	0.015	2.789	0.010	2.852	0.015	104	96
NCEB-351	6.1	C	PMoz	6.7503	0.0723	0.3810	0.0026	2.6247	0.0177	0.1302	0.0012	0.3432	0.0521	2.081	0.012	2.079	0.009	2.098	0.016	101	99
NCEB-351	7.1	C	PMhml	6.2465	0.0704	0.3604	0.0026	2.7745	0.0197	0.1283	0.0012	0.5203	0.0800	1.984	0.012	2.011	0.010	2.073	0.016	104	96
NCEB-351	9.1	C	PMhml	11.7793	0.1258	0.4832	0.0032	2.0696	0.0137	0.1803	0.0017	0.8096	0.1278	2.541	0.014	2.587	0.010	2.658	0.016	105	96
NCEB-351	12.1	R	PMoz	6.8251	0.0736	0.3759	0.0025	2.6601	0.0176	0.1325	0.0012	0.3211	0.0529	2.057	0.012	2.089	0.010	2.128	0.016	103	97
NCEB-351	17.1	C	PMoz	6.6866	0.1378	0.3638	0.0095	2.7484	0.0714	0.1353	0.0008	0.2945	0.1512	2.000	0.045	2.071	0.018	2.164	0.010	108	92
NCEB-351	20.1	C	PMhml	6.4281	0.1348	0.3572	0.0094	2.7992	0.0738	0.1289	0.0008	0.0687	0.0307	1.969	0.045	2.036	0.018	2.081	0.011	106	95
NCEB-351	21.1	C	PMhml	6.5267	0.1347	0.3677	0.0096	2.7195	0.0713	0.1300	0.0007	0.6466	0.2724	2.019	0.045	2.050	0.018	2.095	0.010	104	96
NCEB-351	23.1	C	PMoz	6.9835	0.1485	0.3764	0.0100	2.6569	0.0708	0.1346	0.0008	0.3556	0.1376	2.059	0.047	2.109	0.019	2.156	0.011	105	96
NCEB-351	24.1	C	PMoz	4.7001	0.0984	0.3149	0.0083	3.1758	0.0841	0.1094	0.0007	0.5066	0.1883	1.765	0.041	1.767	0.017	1.792	0.012	102	98
NCEB-351	27.1	C	PMoz	6.3908	0.1079	0.3629	0.0049	2.7559	0.0371	0.1272	0.0008	0.1689	0.0448	1.996	0.023	2.031	0.015	2.057	0.010	103	97
NCEB-351	28.1	C	PMoz	19.7208	0.3351	0.5952	0.0082	1.6800	0.0231	0.2418	0.0014	0.3051	0.0803	3.011	0.033	3.078	0.016	3.134	0.009	104	96
NCEB-351	30.1	C	PMoz	6.3193	0.1042	0.3587	0.0048	2.7876	0.0374	0.1272	0.0007	0.1602	0.0415	1.976	0.023	2.021	0.014	2.057	0.010	104	96
NCEB-351	33.1	R	PMoz	6.1044	0.0997	0.3503	0.0047	2.8550	0.0379	0.1270	0.0007	0.1616	0.0410	1.936	0.022	1.991	0.014	2.054	0.010	106	94
NCEB-351	37.1	C	PMoz	3.6932	0.0618	0.2719	0.0036	3.6773	0.0493	0.0973	0.0006	0.6608	0.1622	1.551	0.018	1.570	0.013	1.579	0.012	102	98
NCEB-351	38.1	C	PMoz	6.5024	0.1075	0.3569	0.0048	2.8015	0.0380	0.1311	0.0008	0.2539	0.0619	1.968	0.023	2.046	0.014	2.110	0.010	107	93
NCEB-351	41.1	C	PMoz	6.8454	0.1139	0.3758	0.0039	2.6607	0.0275	0.1331	0.0015	0.2906	0.0530	2.057	0.018	2.092	0.015	2.136	0.019	104	96
NCEB-351	50.1	C	PMzc	6.3193	0.0791	0.3535	0.0031	2.8291	0.0246	0.1310	0.0009	0.2249	0.0269	1.951	0.015	2.021	0.011	2.108	0.011	108	93
NCEB-351	59.1	C	PMoz	12.2677	0.1414	0.4876	0.0038	2.0507	0.0160	0.1870	0.0012	0.2884	0.0272	2.560	0.016	2.625	0.011	2.719	0.011	106	94
NCEB-351	60.1	C	PMoz	6.9354	0.0843	0.3738	0.0032	2.6754	0.0232	0.1345	0.0008	0.2665	0.0317	2.047	0.015	2.103	0.011	2.155	0.011	105	95
NCEB-351	61.1	C	PMoz	7.1994	0.0835	0.3788	0.0030	2.6399	0.0207	0.1378	0.0009	0.3614	0.0431	2.071	0.014	2.136	0.010	2.196	0.011	106	94
NCEB-351	43.2N	C	PMoz	6.7874	0.1094	0.3619	0.0036	2.7632	0.0273	0.1354	0.0015	0.2878	0.0527	1.991	0.017	2.084	0.014	2.165	0.019	109	92
NCEB-351	45.2N	C	PMoz	13.4201	0.2114	0.4899	0.0047	2.0412	0.0197	0.2002	0.0022	0.6457	0.1190	2.570	0.020	2.710	0.015	2.833	0.018	110	91
NCEB-351	47.1N	C	PMoz	11.6773	0.1938	0.4663	0.0053	2.1443	0.0244	0.1824	0.0020	0.1480	0.0275	2.468	0.023	2.579	0.015	2.677	0.019	108	92
Metamorphic domains (overgrowths)																					
NCEB-351	8.1	R	Mog	0.7652	0.0110	0.0957	0.0008	10.4546	0.0834	0.0604	0.0006	0.0135	0.0025	0.589	0.004	0.577	0.006	0.612	0.022	104	96
NCEB-351	13.1	R	Mog	0.8210	0.0087	0.0999	0.0007	10.0144	0.0668	0.0603	0.0006	0.0082	0.0016	0.614	0.004	0.609	0.005	0.609	0.020	99	101
NCEB-351	14.1	R	Mog	0.8308	0.0170	0.1004	0.0026	9.9579	0.2566	0.0606	0.0004	0.0166	0.0103	0.617	0.015	0.614	0.009	0.619	0.013	100	100
NCEB-351	16.1	R	Mog	0.7729	0.0165	0.0936	0.0024	10.6782	0.2769	0.0598	0.0005	0.0188	0.0060	0.577	0.014	0.581	0.009	0.589	0.018	102	98
NCEB-351	35.1	R	Mog	0.7390	0.0134	0.0887	0.0012	11.2735	0.1562	0.0596	0.0004	0.0137	0.0038	0.548	0.007	0.562	0.008	0.581	0.014	106	94
NCEB-351	36.1	R	Mog	0.8542	0.0143	0.1028	0.0014	9.7250	0.1310	0.0596	0.0004	0.0004	0.0026	0.631	0.008	0.627	0.008	0.582	0.015	92	108
NCEB-351	39.1	R	Mog	0.8868	0.0151	0.1052	0.0015	9.5603	0.1331	0.0613	0.0005	0.0085	0.0030	0.645	0.009	0.645	0.008	0.642	0.017	100	100
NCEB-351	40.1	R	Mog	0.8189	0.0138	0.1002	0.0011	9.9818	0.1109	0.0600	0.0007	0.0096	0.0018	0.615	0.007	0.607	0.008	0.596	0.024	97	103
NCEB-351	42.1	R	Mog	0.7649	0.0132	0.0934	0.0011	10.7081	0.1236	0.0595	0.0007	0.0091	0.0017	0.576	0.006	0.577	0.008	0.581	0.024	101	99
NCEB-351	44.1	R	Mog	0.7643	0.0133	0.0943	0.0009	10.6073	0.1061	0.0596	0.0008	0.0023	0.0028	0.581	0.006	0.576	0.008	0.583	0.028	100	100
NCEB-351	47.2	R	Mog	0.7492	0.0131	0.0928	0.0011	10.7713	0.1226	0.0595	0.0007	0.0093	0.0020	0.572	0.006	0.568	0.008	0.579	0.025	101	99
NCEB-351	48.1	R	Mog	0.8226	0.0133	0.0976	0.0010	10.2470	0.1012	0.0616	0.0007	0.0069	0.0015	0.600	0.006	0.609	0.007	0.655	0.025	109	92
NCEB-351	49.1	R	Mog	0.7662	0.0107	0.0912	0.0007	10.9657	0.0847	0.0604	0.0005	0.0020	0.0019	0.563	0.004	0.578	0.006	0.613	0.019	109	92
NCEB-351	51.1	R	Mog	0.7015	0.0088	0.0877	0.0007	11.4075	0.0976	0.0587	0.0004	0.0044	0.0007	0.542	0.004	0.540	0.005	0.548	0.015	101	99
NCEB-351	52.1	R	Mog	0.8417	0.0104	0.0999	0.0007	10.0136	0.0741	0.0605	0.0005	0.0006	0.0015	0.614	0.004	0.620	0.006	0.616	0.017	100	100
NCEB-351	56.1	R	Mog	0.8064	0.0096	0.0965	0.0008	10.3663	0.0826	0.0608	0.0004	0.0069	0.0013	0.594	0.005	0.600	0.005	0.628	0.014	106	95
NCEB-351	45.1B	R	Mog	0.7948	0.0126	0.0966	0.0009	10.3538	0.0987	0.0605	0.0007	0.0112	0.0024	0.594	0.005	0.594	0.007	0.615	0.025	104	97
NCEB-351	43.1B	R	Mog	0.8557	0.0138	0.1028	0.0010	9.7314	0.0945	0.0608	0.0007	0.0037	0.0014	0.631	0.006	0.628	0.008	0.627	0.024	99	101
NCEB-351	45.3B	R	Mog	0.7002	0.0114	0.0874	0.0009	11.4441	0.1174	0.0592	0.0007	0.0007	0.0012	0.540	0.005	0.539	0.007	0.569	0.025	105	95
Zircons not included on related diagrams																					
NCEB-351	18.1	C	PMoz	5.6601	0.1165	0.3296	0.0086	3.0338	0.0790	0.1257	0.0007	0.7843	0.3815	1.837	0.042	1.925	0.018	2.037	0.010	111	90
NCEB-351	10.1	C	PMoz	4.4358	0.0540	0.2639	0.0021	3.7894	0.0298	0.1236	0.0										

Sample	Spot #	Spot pos.	Interp. CL	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²³⁰ Pb/ ²³⁰ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	²⁰⁸ Pb/ ²⁰⁶ U	±1s	²⁰⁶ Pb/ ²³⁸ U	±1s	²⁰⁷ Pb/ ²³⁵ U	±1s	²⁰⁷ Pb/ ²⁰⁶ U	±1s	Conc. (%)	Conc. (%)
NCEB-427	2.1	R	Mog	1.9366	0.0284	0.1803	0.0015	5.5465	0.0456	0.0776	0.0004	0.0474	0.0168	1.069	0.008	1.094	0.010	1.139	0.011	98	107
NCEB-427	3.1	C	PMoz	7.6458	0.1073	0.3901	0.0026	2.5635	0.0171	0.1412	0.0008	0.3808	0.1328	2.123	0.012	2.190	0.013	2.238	0.010	97	105
NCEB-427	4.1	C	PMoz	6.9282	0.0962	0.3687	0.0024	2.7125	0.0178	0.1363	0.0008	0.2593	0.0894	2.023	0.011	2.102	0.012	2.177	0.010	96	108
NCEB-427	6.1	C	PMoz	12.2458	0.1730	0.4686	0.0032	2.1339	0.0146	0.1881	0.0011	0.5891	0.1983	2.477	0.014	2.623	0.013	2.729	0.010	94	110
NCEB-427	8.1	C	PMoz	2.1134	0.0367	0.1895	0.0015	5.2779	0.0412	0.0814	0.0009	1.7581	0.5719	1.119	0.008	1.153	0.012	1.236	0.022	97	110
NCEB-427	9.1	C	PMoz	4.3024	0.0647	0.2940	0.0022	3.4016	0.0250	0.1083	0.0009	0.5332	0.1721	1.661	0.011	1.694	0.012	1.774	0.015	98	107
NCEB-427	10.1	C	PMoz	7.3895	0.1045	0.3865	0.0027	2.5871	0.0179	0.1407	0.0008	0.4000	0.1273	2.107	0.012	2.160	0.013	2.232	0.010	98	106
NCEB-427	11.1	C	PMoz	7.2381	0.1155	0.3911	0.0036	2.5568	0.0239	0.1352	0.0013	0.3485	0.1101	2.128	0.017	2.141	0.014	2.163	0.016	99	102
NCEB-427	12.1	C	PMoz	2.5889	0.0366	0.2213	0.0015	4.5181	0.0302	0.0863	0.0005	0.2509	0.0782	1.289	0.008	1.297	0.010	1.351	0.012	99	105
NCEB-427	13.1	C	PMoz	6.6553	0.0730	0.3658	0.0025	2.7336	0.0188	0.1336	0.0006	0.2017	0.0601	2.010	0.012	2.067	0.010	2.143	0.007	97	107
NCEB-427	15.1	R	PMoz	6.6997	0.0650	0.3743	0.0019	2.6714	0.0135	0.1320	0.0005	0.3308	0.1002	2.050	0.009	2.073	0.009	2.121	0.007	99	103
NCEB-427	16.1	C	PMhhl	5.5123	0.0560	0.3282	0.0019	3.0474	0.0173	0.1242	0.0007	0.3939	0.1205	1.829	0.009	1.903	0.009	2.016	0.009	96	110
NCEB-427	17.1	C	PMoz	7.1572	0.0723	0.3866	0.0021	2.5866	0.0139	0.1353	0.0007	0.4423	0.1364	2.107	0.010	2.131	0.009	2.164	0.008	99	103
NCEB-427	18.1	C	PMoz	5.2824	0.0533	0.3303	0.0017	3.2072	0.0160	0.1176	0.0006	0.3568	0.1110	1.840	0.008	1.866	0.009	1.921	0.008	99	104
NCEB-427	19.1	R	PMoz	7.5735	0.0740	0.3919	0.0020	2.5519	0.0132	0.1424	0.0006	0.3285	0.1029	2.131	0.009	2.182	0.009	2.252	0.007	98	106
NCEB-427	20.1	C	PMoz	8.2213	0.0836	0.4105	0.0023	2.4360	0.0138	0.1457	0.0006	0.3901	0.1234	2.217	0.011	2.256	0.009	2.292	0.008	98	103
NCEB-427	22.1	C	PMoz	2.6812	0.0263	0.2282	0.0012	4.3827	0.0232	0.0857	0.0003	0.1471	0.0474	1.325	0.006	1.323	0.007	1.336	0.008	100	101
NCEB-427	23.1	R	PMoz	5.8449	0.0575	0.3378	0.0017	2.9604	0.0153	0.1264	0.0005	0.3903	0.1268	1.876	0.008	1.953	0.008	2.047	0.007	96	109
NCEB-427	24.1	R	PMhhl	5.7130	0.0569	0.3418	0.0020	2.9254	0.0167	0.1223	0.0006	0.3513	0.7715	1.895	0.009	1.933	0.009	1.990	0.008	98	105
NCEB-427	26.1	R	PMoz	7.9167	0.1427	0.4007	0.0042	2.4954	0.0265	0.1429	0.0014	0.1477	0.0336	2.172	0.020	2.222	0.016	2.258	0.017	98	104
NCEB-427	27.1	C	PMoz	4.2301	0.0775	0.2936	0.0033	3.4064	0.0386	0.1052	0.0011	0.5408	0.1208	1.659	0.017	1.680	0.015	1.721	0.019	99	104
NCEB-427	28.1	R	PMoz	4.5742	0.0825	0.3109	0.0033	3.2163	0.0343	0.1071	0.0011	0.4279	0.0942	1.745	0.016	1.745	0.015	1.754	0.018	100	101
NCEB-427	29.1	R	PMoz	4.6577	0.0844	0.3165	0.0034	3.1598	0.0341	0.1057	0.0011	0.4184	0.0901	1.773	0.017	1.760	0.015	1.731	0.019	101	98
NCEB-427	30.1	C	PMoz	6.5957	0.1161	0.3709	0.0038	2.6958	0.0277	0.1293	0.0012	0.3721	0.0787	2.034	0.018	2.059	0.015	2.086	0.016	99	103
NCEB-427	31.1	R	PMoz	4.4300	0.0794	0.3004	0.0032	3.3294	0.0356	0.1064	0.0011	0.4596	0.0963	1.693	0.016	1.718	0.015	1.743	0.018	99	103
NCEB-427	33.1	C	PMhhl	6.6263	0.1175	0.3684	0.0038	2.7148	0.0283	0.1311	0.0012	0.2486	0.0507	2.022	0.018	2.063	0.016	2.109	0.016	98	104
NCEB-427	34.1	C	PMoz	1.8620	0.0335	0.1797	0.0020	5.5649	0.0628	0.0754	0.0008	0.6539	0.1317	1.065	0.011	1.068	0.012	1.081	0.022	100	101
NCEB-427	35.1	C	PMoz	3.4570	0.0624	0.2644	0.0028	3.7817	0.0400	0.0953	0.0009	0.3815	0.0760	1.512	0.014	1.517	0.014	1.540	0.018	100	102
NCEB-427	36.1	C	PMoz	8.2675	0.1447	0.4005	0.0041	2.4966	0.0257	0.1501	0.0014	0.2980	0.0586	2.172	0.019	2.261	0.016	2.343	0.016	96	108
NCEB-427	37.1	C	PMoz	7.8321	0.1405	0.4013	0.0044	2.4921	0.0273	0.1419	0.0013	0.2880	0.0560	2.175	0.020	2.212	0.016	2.247	0.016	98	103
NCEB-427	38.1	C	PMoz	8.0319	0.2116	0.4028	0.0051	2.4829	0.0315	0.1452	0.0024	0.1783	0.0385	2.182	0.023	2.235	0.024	2.286	0.028	98	105
NCEB-427	39.1	C	PMoz	14.0576	0.3716	0.5255	0.0069	1.9029	0.0251	0.1963	0.0031	0.2250	0.0493	2.722	0.029	2.754	0.025	2.801	0.026	99	103
NCEB-427	44.1	C	PMoz	14.0840	0.3638	0.5128	0.0065	1.9502	0.0247	0.2003	0.0031	0.0951	0.0231	2.668	0.028	2.755	0.024	2.834	0.026	97	106
NCEB-427	45.1	C	PMoz	7.8761	0.2028	0.4005	0.0050	2.4968	0.0314	0.1426	0.0023	0.1465	0.0367	2.171	0.023	2.217	0.023	2.255	0.027	98	104
NCEB-427	46.1	C	PMhhl	7.1455	0.1854	0.3883	0.0049	2.5751	0.0326	0.1355	0.0022	0.1121	0.0288	2.115	0.023	2.130	0.023	2.167	0.028	99	102
NCEB-427	48.1	R	PMoz	4.3578	0.1118	0.2986	0.0038	3.3490	0.0425	0.1067	0.0017	0.3192	0.0855	1.684	0.019	1.704	0.021	1.747	0.028	99	104
NCEB-427	50.1	C	PMoz	7.4471	0.0519	0.3980	0.0039	2.5123	0.0247	0.1374	0.0012	0.2893	0.0841	2.160	0.018	2.167	0.006	2.191	0.014	100	101
NCEB-427	51.1	R	PMoz	4.8426	0.0315	0.3169	0.0031	3.1554	0.0304	0.1120	0.0009	0.3229	0.0927	1.775	0.015	1.792	0.005	1.834	0.014	99	103
NCEB-427	53.1	C	PMoz	2.0827	0.0153	0.1948	0.0019	5.1334	0.0505	0.0787	0.0007	0.3518	0.0975	1.147	0.010	1.143	0.005	1.167	0.017	100	102
NCEB-427	54.1	C	PMoz	2.7707	0.0194	0.2298	0.0022	4.3509	0.0423	0.0885	0.0007	0.3418	0.0936	1.334	0.012	1.348	0.005	1.400	0.016	99	105
NCEB-427	55.1	C	PMoz	1.5936	0.0110	0.1611	0.0016	6.2078	0.0604	0.0719	0.0006	0.1770	0.0480	0.963	0.009	0.968	0.004	0.984	0.017	99	102
NCEB-427	56.1	C	PMoz	2.2805	0.0163	0.2046	0.0020	4.8872	0.0474	0.0814	0.0007	0.4595	0.1230	1.200	0.011	1.206	0.005	1.236	0.017	99	103
NCEB-427	57.2	C	PMoz	12.6929	0.0839	0.4946	0.0048	2.0217	0.0196	0.1869	0.0015	0.2371	0.0621	2.591	0.021	2.657	0.006	2.718	0.014	97	105
NCEB-427	58.1	C	PMoz	1.6484	0.0112	0.1626	0.0016	6.1495	0.0595	0.0733	0.0006	0.3640	0.0943	0.971	0.009	0.989	0.004	1.023	0.017	98	105
NCEB-427	49.2N	C	PMhml	2.1384	0.0548	0.1924	0.0025	5.1970	0.0666	0.0808	0.0013	0.2744	0.0780	1.134	0.013	1.161	0.018	1.220	0.031	98	108
NCEB-427	52.2N	C	PMoz	6.7916	0.0470	0.3733	0.0037	2.6792	0.0265	0.1335	0.0011	0.1128	0.0319	2.045	0.017	2.085	0.006	2.141	0.014	98	105
NCEB-427	60.1N	C	PMoz	2.5531	0.0166	0.2147	0.0021	4.6575	0.0450	0.0867	0.0007	0.1717	0.0436	1.254	0.011	1.287	0.005	1.360	0.016	97	108
NCEB-427	7.1N	C	PMoz	4.2117	0.0613	0.2857	0.0019	3.5005	0.0237	0.1069	0.0007	0.7461	0.2486	1.620	0.010	1.676	0.012	1.752	0.012	97	108
Metamorphic domains (overgrowths)																					
NCEB-427	1.1	R	Mog	0.6478	0.0129	0.0830	0.0006	12.0449	0.0941	0.0580	0.0005	0.0011									

ANEXO II

(análises SHRIMP U-Th-Pb em zircões ígneos)

sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
273A-1.1	4.33	214	74	0.36	28.07	0.0695	17.8	1.3945	18.2	0.1456	2.7	876.0	21.6
273A-2.1	8.20	106	69	0.67	15.27	0.0714	31.1	1.5000	31.8	0.1525	3.3	914.8	27.6
273A-3.1	1.80	969	63	0.07	87.07	0.0601	6.8	0.8497	7.3	0.1026	2.3	629.8	13.8
273A-4.1	3.07	776	66	0.09	69.66	0.0647	10.1	0.9013	10.5	0.1011	2.3	620.6	13.7
273A-5.1	6.05	155	54	0.36	20.88	0.0709	21.3	1.4318	21.8	0.1465	2.8	881.6	22.8
273A-5.2	1.17	1044	226	0.22	94.57	0.0634	5.2	0.9109	5.8	0.1042	2.3	638.9	14.0
273A-6.1	5.23	136	81	0.62	18.31	0.0653	19.5	1.3341	20.1	0.1482	2.9	891.0	23.8
273A-7.1	8.80	86	25	0.30	12.31	0.0693	27.9	1.4370	28.8	0.1505	3.3	904.0	27.6
273A-7.2	1.25	1184	265	0.23	104.53	0.0591	4.7	0.8266	5.3	0.1014	2.3	622.9	13.5
273A-8.1	8.05	77	26	0.35	10.99	0.0974	24.3	2.0416	24.9	0.1518	3.4	911.2	28.6
273A-8.2	3.79	328	195	0.61	41.28	0.0636	15.3	1.2290	15.7	0.1403	2.5	846.2	20.0
273A-9.1	1.81	927	75	0.08	85.25	0.0607	6.7	0.8781	7.2	0.1050	2.4	643.4	15.0
273A-10.1	4.99	154	95	0.63	20.32	0.0704	17.5	1.4088	18.0	0.1451	2.7	873.3	22.3
273A-11.1	5.40	144	55	0.39	19.60	0.0651	23.8	1.3356	24.3	0.1489	3.0	894.8	24.8
273A-11.2	2.91	688	149	0.22	57.51	0.0604	10.1	0.7852	10.5	0.0943	2.4	581.1	13.2
273A-12.1	5.24	198	81	0.43	25.27	0.0758	15.5	1.4657	16.0	0.1403	2.7	846.4	21.1
273A-12.2	2.55	655	155	0.24	57.57	0.0582	13.0	0.7979	13.3	0.0995	2.4	611.6	14.0
273A-12.3	9.63	122	59	0.50	11.88	0.0636	42.5	0.8953	43.2	0.1021	3.7	626.7	21.7
273A-13.1	7.39	76	31	0.43	10.68	0.0908	24.2	1.8921	24.8	0.1509	3.4	906.1	27.9
273A-14.1	4.33	191	77	0.42	25.57	0.0697	19.1	1.4237	19.5	0.1483	2.8	891.3	22.6
273A-15.1	7.00	119	44	0.38	16.60	0.0721	24.0	1.4863	24.7	0.1495	3.1	898.2	25.8
273A-16.1	8.35	96	30	0.32	14.56	0.0735	36.9	1.6274	37.7	0.1606	3.6	959.8	31.0
273A-17.1	9.54	94	26	0.28	12.70	0.0692	37.5	1.3402	38.4	0.1405	3.5	847.4	27.3
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
125A-1.1	4.15	376	193	0.53	35.09	0.0629	16.6	0.9008	17.0	0.1039	2.5	637.2	15.4
125A-2.1	2.21	664	485	0.76	62.35	0.0601	13.6	0.8853	13.9	0.1068	2.5	654.3	15.2
125A-3.1	1.57	742	477	0.66	67.65	0.0634	6.9	0.9126	7.4	0.1044	2.3	640.0	14.1
125A-4.1	3.58	354	189	0.55	32.92	0.0671	14.4	0.9632	14.8	0.1041	2.4	638.6	14.9
125A-5.1	2.30	549	348	0.66	50.80	0.0629	9.9	0.9119	10.3	0.1051	2.6	644.3	15.9
125A-6.1	3.16	390	226	0.60	37.12	0.0640	14.7	0.9458	15.0	0.1071	2.4	656.1	15.2
125A-7.1	3.93	364	189	0.54	33.91	0.0625	18.3	0.8950	18.6	0.1038	2.6	636.9	15.4
125A-8.1	5.54	224	117	0.54	21.85	0.0674	26.9	0.9945	27.3	0.1070	3.0	655.5	18.4
125A-9.1	2.83	490	378	0.80	45.83	0.0620	12.0	0.9020	12.3	0.1055	2.4	646.7	14.6
125A-10.1	3.21	439	307	0.72	41.65	0.0649	14.8	0.9538	15.2	0.1067	3.1	653.4	19.5
125A-11.1	2.00	610	349	0.59	56.26	0.0628	9.5	0.9106	9.9	0.1051	2.4	644.2	14.4
125A-12.1	2.26	604	443	0.76	57.01	0.0618	10.4	0.9128	10.7	0.1072	2.4	656.2	14.8
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
273B-1.2	3.96	514	88	0.18	47.66	0.0678	18.5	0.9665	18.8	0.1034	2.9	634.1	17.5
273B-2.1	8.96	116	21	0.19	14.36	0.0684	33.0	1.2305	33.8	0.1305	3.3	790.6	24.4
273B-2.2	2.06	872	219	0.26	74.17	0.0598	6.9	0.7983	7.3	0.0969	2.3	596.2	13.1
273B-3.1	6.66	258	78	0.31	23.56	0.0622	22.0	0.8477	22.5	0.0988	2.7	607.4	15.7
273B-4.1	4.75	111	26	0.24	12.44	0.0973	18.4	1.6641	18.7	0.1239	3.1	752.8	21.5
273B-5.1	10.28	88	32	0.37	12.15	0.0807	32.1	1.5995	33.0	0.1436	3.8	865.0	29.7
273B-6.1	9.59	175	42	0.25	18.44	0.0791	35.4	1.2034	36.1	0.1102	4.1	674.1	25.5
273B-7.1	2.26	632	118	0.19	56.22	0.0583	8.5	0.8133	9.0	0.1011	2.6	621.0	15.5
273B-8.1	2.04	681	142	0.22	60.49	0.0591	7.1	0.8247	7.6	0.1012	2.3	621.3	13.7
273B-9.1	2.70	630	125	0.20	55.37	0.0592	9.2	0.8109	9.7	0.0994	2.3	610.6	13.6
273B-10.1	2.59	554	153	0.29	49.03	0.0575	10.9	0.7951	11.3	0.1003	2.4	616.0	13.9
273B-11.1	7.45	152	35	0.24	16.94	0.0674	36.3	1.1117	36.9	0.1197	4.0	728.7	26.9
273B-11.2	2.77	799	102	0.13	66.05	0.0623	8.1	0.8020	8.6	0.0934	2.3	575.5	12.9
273B-12.1	8.15	151	39	0.26	17.42	0.0662	29.9	1.1165	30.6	0.1225	3.3	744.7	22.6
273B-13.1	1.66	933	136	0.15	83.31	0.0607	6.9	0.8547	7.4	0.1021	2.3	626.8	13.7
273B-14.1	1.99	714	122	0.18	65.18	0.0601	7.3	0.8610	7.8	0.1040	2.3	637.6	14.1
273B-15.1	0.68	1765	591	0.35	152.30	0.0596	2.9	0.8188	3.7	0.0997	2.2	612.8	13.1
273B-16.1	3.01	535	85	0.17	49.46	0.0629	14.2	0.9039	14.5	0.1042	2.5	639.0	15.0
273B-17.1	3.01	381	242	0.66	49.49	0.0728	13.2	1.4682	13.6	0.1462	2.5	879.6	20.4
273B-18.1	3.79	500	60	0.12	44.62	0.0632	17.9	0.8688	18.3	0.0998	2.6	613.0	14.9
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
211-1.1	0.69	552	377	0.71	47.80	0.0594	4.3	0.8202	5.1	0.1003	2.8	614.9	16.6
211-2.1	1.66	308	255	0.85	28.64	0.0620	7.3	0.9075	7.8	0.1065	2.4	650.7	15.2
211-3.1	4.44	336	228	0.70	31.81	0.0624	19.1	0.9014	19.5	0.1058	2.6	642.9	17.3
211-4.1	5.55	283	181	0.66	26.59	0.0612	28.1	0.8692	28.5	0.1036	2.4	632.5	18.0
211-5.1	6.12	218	96	0.45	21.30	0.0660	24.6	0.9684	25.1	0.1062	2.5	652.2	16.8
211-6.1	4.48	314	322	1.06	30.76	0.0604	21.5	0.9048	21.9	0.1101	2.4	664.7	16.9
211-8.1	2.87	649	421	0.67	57.56	0.0597	19.2	0.8237	19.5	0.1010	2.4	615.4	15.7
211-9.1	3.53	408	228	0.58	36.98	0.0599	14.1	0.8387	14.5	0.1029	2.3	623.6	14.7
211-10.1	4.25	327	182	0.58	30.51	0.0592	20.5	0.8471	20.8	0.1040	2.4	636.9	15.5
211-11.1	3.03	421	357	0.88	39.52	0.0608	13.4	0.8864	13.7	0.1065	2.3	648.0	14.9
211-12.1	3.62	362	295	0.84	33.69	0.0591	12.2	0.8501	12.7	0.1057	2.4	639.4	15.0
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
277-1.1	3.05	481	451	0.97	44.66	0.0662	11.1	0.9601	11.3	0.1051	2.3	641.7	14.6
277-2.1	2.20	550	443	0.83	50.87	0.0666	4.2	0.9723	4.9	0.1059	2.3	645.0	14.8
277-3.1	1.27	707	588	0.86	65.51	0.0654	4.0	0.9602	4.6	0.1064	2.3	651.4	14.3
277-4.1	1.66	895	557	0.64	82.70	0.0668	2.5	0.9827	3.4	0.1067	2.3	647.8	14.2
277-5.1	1.79	502	274	0.56	47.09	0.0610	6.2	0.8970	6.6	0.1066	2.3	655.4	14.7
277-6.1	2.28	619	515	0.86	56.76	0.0648	6.6	0.9322	7.0	0.1043	2.3	638.3	14.3
277-7.1	1.17	1519	1169	0.79	140.04	0.0673	1.8	0.9934	2.9	0.1070	2.2	649.2	13.9
277-8.1	2.10	620	376	0.63	58.06	0.0654	5.4	0.9662	5.9	0.1071	2.3	653.3	14.5
277-9.1	2.02	684	559	0.84	62.49	0.0691	3.5	1.0020	4.2	0.1051	2.3	638.6	14.1
277-10.1	1.36	930	776	0.86	85.57	0.0660	3.8	0.9670	4.4	0.1063	2.3	646.8	14.1
277-11.1	3.34	464	273	0.61	43.77	0.0678	9.8	0.9999	10.1	0.1070	2.3	649.0	14.8
277-12.1	2.62	564	325	0.59	54.13	0.0630	7.0	0.9466	7.4	0.1090	2.3	664.3	14.9

sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
269-1.1	0.36	176	99	0.59	20.63	0.0705	2.5	1.2856	4.1	0.1363	3.0	823.5	23.2
269-1.2	1.13	176	47	0.28	15.81	0.0636	3.0	0.8180	7.9	0.1036	2.8	635.5	17.2
269-3.1	2.34	393	68	0.18	43.38	0.0623	11.0	1.0775	15.8	0.1254	2.8	761.5	19.9
269-3.2	0.59	243	41	0.18	22.08	0.0584	3.0	0.8393	4.8	0.1049	2.7	643.0	16.6
269-4.1	0.49	297	185	0.64	26.00	0.0656	2.3	0.8399	5.0	0.1014	2.7	622.4	15.9
269-5.1	0.94	147	66	0.47	13.73	0.0653	3.4	0.8862	7.2	0.1076	3.0	658.5	18.5
269-6.1	0.67	180	156	0.90	15.97	0.0646	3.0	0.7854	8.5	0.1025	2.8	628.9	16.9
269-7.1	1.84	38	56	1.52	3.43	0.0675	8.0	0.9569	13.4	0.1023	4.7	627.8	27.9
269-8.1	1.03	618	184	0.31	70.61	0.0670	1.7	1.1506	5.0	0.1315	2.6	796.3	19.5
269-9.1	0.97	175	41	0.24	15.66	0.0608	3.4	0.8107	7.4	0.1028	2.8	630.8	16.9
269-10.1	0.60	177	166	0.97	15.79	0.0667	3.3	0.8762	5.3	0.1029	2.8	631.5	16.7
269-11.1	0.11	278	195	0.72	24.37	0.0621	2.5	0.8841	3.8	0.1018	2.7	625.0	16.0
269-12.1	0.35	291	66	0.23	25.44	0.0622	2.5	0.8502	3.9	0.1015	2.7	623.1	16.2
269-13.1	2.17	146	25	0.18	12.24	0.0635	5.8	0.7023	14.0	0.0956	3.0	588.4	17.0
269-14.1	0.87	124	77	0.64	11.05	0.0607	4.3	0.8626	7.7	0.1030	2.9	632.1	17.7
269-15.1	0.69	222	184	0.86	18.68	0.0579	3.7	0.8087	5.8	0.0974	2.8	599.1	15.8
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
200A-1.1	0.62	380	314	0.85	34.39	0.0687	2.0	0.9064	4.0	0.1047	2.7	642.0	16.2
200A-2.1	0.13	969	651	0.69	88.55	0.0645	1.4	0.8785	3.0	0.1062	2.6	650.8	15.9
200A-3.1	0.05	648	460	0.73	59.91	0.0645	1.7	0.9083	3.1	0.1076	2.6	658.6	16.3
200A-4.1	0.29	380	254	0.69	37.07	0.0608	2.4	0.9163	4.0	0.1131	2.7	690.8	17.5
200A-5.1	0.63	364	276	0.78	35.19	0.0641	2.8	0.9163	4.7	0.1118	2.7	683.0	17.7
200A-6.1	0.65	285	212	0.77	26.77	0.0657	2.5	0.8938	4.3	0.1086	2.7	664.4	17.1
200A-7.1	0.22	604	419	0.72	56.44	0.0655	1.6	0.9215	3.4	0.1085	2.6	664.0	16.4
200A-7.2	1.67	596	314	0.55	37.63	0.0442	7.9	0.5464	7.7	0.0723	2.7	449.8	11.6
200A-8.1	0.78	337	210	0.64	30.76	0.0643	2.3	0.8714	5.3	0.1054	2.7	646.2	16.5
200A-9.1	0.27	457	259	0.59	41.90	0.0638	2.2	0.9157	3.8	0.1065	2.6	652.3	16.3
200A-10.1	0.34	663	501	0.78	60.61	0.0662	1.7	0.8720	3.6	0.1060	2.6	649.3	16.0
200A-11.1	0.19	694	479	0.71	63.84	0.0632	1.6	0.8720	3.4	0.1069	2.7	654.7	16.6
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
221-1.1	0.98	86	61	0.74	10.21	0.0800	3.5	1.2591	7.3	0.1366	4.1	825.6	31.7
221-2.1	1.27	94	70	0.78	11.22	0.0659	4.0	1.1800	9.5	0.1375	3.0	830.6	23.2
221-3.1	0.57	100	48	0.50	11.78	0.0642	3.7	1.2020	6.2	0.1361	2.9	822.4	22.5
221-4.1	0.98	80	41	0.52	9.82	0.0718	3.6	1.2499	6.5	0.1410	3.0	850.3	23.7
221-5.1	0.74	92	50	0.56	11.08	0.0692	5.9	1.3918	6.6	0.1398	3.1	843.5	24.2
221-6.1	1.07	164	108	0.68	19.58	0.0720	2.7	1.2323	7.0	0.1373	3.0	829.6	23.3
221-7.1	0.71	210	140	0.69	25.33	0.0740	3.2	1.2083	5.8	0.1391	2.7	839.8	21.5
221-8.1	1.60	87	52	0.62	10.01	0.0751	3.9	1.1706	11.5	0.1314	3.1	795.6	23.0
221-9.1	0.67	136	88	0.67	16.60	0.0706	2.9	1.3122	5.0	0.1409	2.9	850.0	22.8
221-10.1	1.03	72	45	0.65	8.79	0.0709	3.9	1.2381	10.9	0.1410	3.1	850.2	24.7
221-11.1	0.86	94	51	0.56	11.90	0.0724	3.4	1.3634	6.4	0.1463	2.9	880.1	24.1
221-12.1	1.15	75	42	0.57	8.97	0.0665	4.6	1.2347	8.0	0.1366	3.1	825.2	23.6
221-13.1	0.71	256	149	0.60	29.79	0.0722	3.4	1.2252	5.6	0.1344	2.7	812.7	20.9
221-14.1	2.23	108	55	0.53	13.03	0.0689	6.8	1.0809	15.3	0.1377	3.1	831.6	24.1
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
231-1.1	0.37	172	76	0.45	14.40	0.0618	3.6	0.8719	5.7	0.0972	2.8	597.7	16.0
231-2.1	3.96	164	96	0.60	14.84	0.0584	11.8	0.9190	15.7	0.1009	3.3	619.9	19.4
231-3.1	1.40	117	52	0.46	10.08	0.0669	4.1	0.8511	12.8	0.0984	3.0	605.2	17.5
231-4.1	1.60	95	42	0.46	8.53	0.0746	3.8	0.8636	8.3	0.1022	3.0	627.6	18.0
231-5.1	0.50	147	39	0.27	14.41	0.0606	3.6	0.9510	5.6	0.1132	2.8	691.1	18.6
231-6.1	0.23	527	5	0.01	46.70	0.0599	1.8	0.8463	3.4	0.1028	2.8	631.0	16.7
231-6.2	1.16	154	57	0.39	14.33	0.0648	3.5	0.8877	8.3	0.1070	2.9	655.1	18.2
231-7.1	0.26	199	98	0.51	17.76	0.0643	2.7	0.8783	5.0	0.1038	2.7	636.9	16.7
231-8.1	0.60	290	104	0.37	25.74	0.0613	2.5	0.8289	5.6	0.1026	2.7	629.6	16.2
231-9.1	1.22	108	41	0.40	10.10	0.0681	4.0	0.9237	8.7	0.1073	3.1	657.3	19.1
231-10.1	0.37	259	145	0.58	22.54	0.0620	3.9	0.8197	5.5	0.1009	2.7	620.0	16.1
231-11.1	0.83	175	66	0.39	15.65	0.0614	3.3	0.8276	6.2	0.1030	2.8	631.9	16.9
231-12.1	0.86	92	36	0.41	8.20	0.0583	5.6	0.8923	7.1	0.1028	3.0	630.8	18.1
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
125E-1.1	0.68	186	151	0.84	16.30	0.0669	2.8	0.9453	3.9	0.1025	2.8	623.6	16.5
125E-2.1	0.70	937	288	0.32	81.18	0.0625	1.5	0.8655	3.0	0.1004	2.6	614.9	15.2
125E-3.1	0.17	1073	206	0.20	95.07	0.0606	1.2	0.8615	2.8	0.1031	2.6	631.8	15.4
125E-4.1	2.32	905	201	0.23	81.79	0.0604	4.7	0.8533	5.3	0.1025	2.6	629.9	15.6
125E-5.1	2.94	817	116	0.15	69.81	0.0612	5.0	0.8173	5.6	0.0969	2.6	593.2	14.9
125E-6.1	0.39	808	174	0.22	71.26	0.0620	1.5	0.8766	3.0	0.1025	2.6	627.4	15.5
125E-7.1	2.35	274	64	0.24	24.84	0.0620	5.0	0.8822	5.7	0.1032	2.7	632.1	16.6
125E-8.1	0.26	654	81	0.13	58.63	0.0616	2.1	0.8853	3.4	0.1042	2.6	638.6	15.9
125E-9.1	1.22	628	93	0.15	55.61	0.0640	2.5	0.8998	3.6	0.1020	2.6	624.9	15.6
125E-10.1	0.45	821	184	0.23	72.10	0.0637	1.5	0.8981	3.0	0.1023	2.6	624.4	15.4
125E-11.1	6.19	781	177	0.23	72.86	0.0612	13.3	0.8545	13.6	0.1013	2.6	623.0	15.9
125E-12.1	2.11	1088	212	0.20	98.58	0.0614	3.8	0.8747	4.6	0.1033	2.6	632.5	15.6
sample/zircon	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	± (%)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s
170A-1.1	0.90	109	52	0.49	9.73	0.0537	6.6	0.8971	6.1	0.1019	3.0	633.1	17.9
170A-2.1	0.43	222	93	0.43	21.59	0.0608	2.9	0.9263	4.2	0.1127	2.7	687.7	17.8
170A-3.1	1.66	427	36	0.09	37.56	0.0645	3.8	0.9127	6.8	0.1005	2.7	618.2	16.0
170A-4.1	0.59	285	41	0.15	26.06	0.0612	2.8	0.8780	4.2	0.1061	3.0	649.1	18.4
170A-5.1	0.50	266	131	0.51	25.23	0.0611	2.9	0.9160	4.5	0.1099	2.7	671.6	17.2
170A-6.1	0.97	73	42	0.60	6.48	0.0694	5.7	1.0345	7.7	0.1023	3.2	630.9	19.2
170A-7.1	2.19	88	85	1.00	7.83	0.0754	4.6	0.8888	12.6	0.1030	3.0	622.6	18.7
170A-8.1	0.19	486	201	0.43	46.48	0.0647	2.3	0.9699	4.2	0.1112	3.3	678.8	21.3
170A-9.1	1.24	146	15	0.10	13.47	0.0591	4.1	0.8209	9.4	0.1064	2.8	649.6	17.9
170A-10.1	1.49	151	90	0.61	14.09	0.0661	3.6	0.8621	9.9	0.1082	2.8	656.2	18.1
170A-11.1	0.67	180	123	0.70	16.68	0.0631	3.4	0.9060	5.9	0.1073	2.8	655.6	17.6
170A-12.1	0.72	418	66	0.16	40.02	0.0651	2.6	0.9220	5.1	0.1113	2.6	676.2	17.0

ANEXO III

(análises SHRIMP U-Th-Pb em zircões metamórficos)

sample/spot	zircon domain	²⁰⁶ Pb _c (%)	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁷ Pb*/ ²³⁵ U	± (%)	²⁰⁶ Pb*/ ²³⁸ U	± (%)	corr.	²⁰⁶ Pb*/ ²³⁸ U (Ma)	1s (Ma)
S506-10_AUS	rim	0.33	174	3.7	0.02	14.68	0.78644	2.25185	0.09774	0.86	0.38	601.2	5.0
S506-7_AUS	rim	0.02	691	31.5	0.05	58.54	0.83602	1.11939	0.09858	0.93	0.84	606.1	5.4
S506-1_AUS	rim	0.18	208	6.1	0.03	17.70	0.79856	2.48951	0.09871	0.83	0.33	606.8	4.8
S506-6_AUS	rim	0.05	330	12.2	0.04	28.03	0.82045	1.43160	0.09886	1.00	0.70	607.7	5.8
S506-5_AUS	rim	0.11	586	22.0	0.04	49.92	0.81867	1.06296	0.09906	0.74	0.69	608.9	4.3
S506-11.1_AUS	rim	0.27	274	7.1	0.03	23.46	0.79781	1.74308	0.09944	0.79	0.45	611.1	4.6
S506-9_AUS	rim	0.04	310	8.5	0.03	26.55	0.84341	1.24602	0.09960	0.77	0.62	612.1	4.5
S506-8_AUS	rim	0.33	208	4.7	0.02	17.92	0.79326	2.66708	0.09997	0.83	0.31	614.2	4.9
S506-3_AUS	rim	0.22	249	6.0	0.02	21.42	0.81390	1.66560	0.10001	0.79	0.47	614.5	4.6
S506-11_AUS	rim	0.04	384	16.4	0.04	33.01	0.84750	1.18551	0.10006	0.76	0.64	614.7	4.4
S506-12_AUS	rim	0.06	734	34.6	0.05	63.39	0.84123	1.01577	0.10050	0.72	0.71	617.4	4.2
S506-2.1_BR	rim	0.58	190	13.9	0.08	16.47	0.82802	4.33362	0.10021	2.86	0.66	615.7	16.8
S506-4.1_BR	rim	0.68	180	44.5	0.25	15.44	0.81487	4.24201	0.09895	2.77	0.65	608.3	16.1
S506-6.1_BR	rim	0.66	216	19.9	0.09	18.79	0.84177	5.26507	0.10046	2.80	0.53	617.1	16.5
S506-8.1_BR	rim	0.52	200	4.9	0.03	16.82	0.80244	3.86679	0.09743	2.82	0.73	599.4	16.1
S506-13_AUS	core	0.01	256	108.1	0.44	28.98	1.33189	1.13500	0.13199	0.78	0.69	799.2	5.8
S506-14_AUS	core	0.01	206	82.5	0.41	31.76	2.63303	1.42166	0.17956	1.03	0.72	1064.6	10.1
S506-4_AUS	core	0.04	423	144.6	0.35	71.74	2.74438	1.27923	0.19729	1.14	0.89	1160.7	12.1
S506-2_AUS	core	0.02	579	70.1	0.13	130.44	3.87541	0.79478	0.26219	0.72	0.90	1501.0	9.6
DKE-350-7_AUS	rim	4.24	14	1.1	0.08	1.19	0.77615	6.23487	0.09843	2.55	0.41	586.1	18.0
DKE-350-8_AUS	rim	1.56	25	3.0	0.12	2.08	0.84060	3.84511	0.09750	1.77	0.46	590.9	10.9
DKE-350-4_AUS	rim	0.68	25	0.7	0.03	2.05	0.78988	4.16956	0.09616	1.82	0.44	592.1	10.9
DKE-350-3_AUS	rim	1.36	42	0.4	0.01	3.51	0.78093	3.36610	0.09744	1.69	0.50	592.6	10.0
DKE-350-15_AUS	rim	0.80	10	0.5	0.06	0.82	0.73654	7.04301	0.09855	2.41	0.34	604.7	14.9
DKE-350-14_AUS	rim	1.27	34	1.0	0.03	2.89	0.82712	3.51817	0.09954	1.59	0.45	605.3	9.4
DKE-350-5_AUS	rim	1.50	61	2.1	0.04	5.24	0.86952	2.86351	0.09987	1.44	0.50	605.4	8.6
DKE-350-13_AUS	rim	0.29	85	1.5	0.02	7.19	0.80987	2.34287	0.09881	1.32	0.56	606.4	7.7
DKE-350-9_AUS	rim	0.29	58	1.0	0.02	5.00	0.82547	2.94509	0.10001	1.71	0.58	616.7	10.1
DKE-350-11_AUS	rim	0.23	14	0.3	0.03	1.19	0.83386	5.84969	0.10107	2.19	0.38	623.8	13.1
DKE-350-10_AUS	rim	0.31	38	0.9	0.02	3.37	0.89163	3.98347	0.10338	1.98	0.50	637.0	12.1
DKE-350-17_AUS	rim	0.21	76	0.8	0.01	6.46	0.80105	2.67534	0.09952	1.48	0.55	613.2	8.7
DKE-350-20_AUS	core	0.23	152	57.8	0.39	14.82	0.97249	2.06768	0.11325	1.27	0.61	691.3	8.3
DKE-350-18_AUS	core	0.34	136	68.8	0.52	13.33	1.00969	2.40740	0.11384	1.94	0.80	693.4	12.8
DKE-350_16_AUS	core	0.17	94	50.2	0.55	9.40	1.00109	2.26052	0.11624	1.49	0.66	708.5	10.1
DKE-350-22_AUS	core	0.01	210	89.4	0.44	21.01	0.99318	1.74634	0.11613	1.31	0.75	708.6	8.8
DKE-350-21_AUS	core	0.35	92	45.3	0.51	9.26	1.00403	2.20669	0.11727	1.32	0.60	713.6	8.9
DKE-107-4_AUS	type-2	0.12	145	3.1	0.02	12.32	0.85379	1.90898	0.09914	0.87	0.45	609.4	5.0
DKE-107-1.2_AUS	type-2	0.33	179	8.1	0.05	15.28	0.80576	2.17034	0.09925	0.85	0.39	610.0	4.9
DKE-107-6_AUS	type-2	0.11	141	4.4	0.03	12.17	0.80946	2.40666	0.10004	0.87	0.36	614.6	5.1
DKE-107-8.2_AUS	type-2	0.57	78	1.3	0.02	6.78	0.78169	2.71473	0.10015	0.97	0.36	615.3	5.7
DKE-107-3.2_AUS	type-2	0.59	88	1.6	0.02	7.67	0.78767	4.19445	0.10114	0.99	0.24	621.1	5.9
DKE-107-7.1_AUS	type-2	0.35	138	2.3	0.02	12.04	0.81262	2.46452	0.10125	0.87	0.35	621.7	5.1
DKE-107-9.1_AUS	type-2	0.51	21	2.3	0.11	1.80	0.96347	5.54829	0.10200	1.65	0.30	626.1	9.8
DKE-107-5.1_BR	type-2	3.13	78	1.3	0.02	6.97	0.85206	18.64411	0.10112	2.11	0.11	621.0	12.3
DKE-107-6.1_BR	type-2	2.47	78	1.1	0.01	6.91	0.83268	17.09344	0.10073	2.08	0.12	618.7	12.2
DKE-107-7.2_BR	type-2	1.58	93	4.6	0.05	7.98	0.80595	9.79254	0.09804	2.77	0.28	602.9	15.9
DKE_107-8.2_BR	type-2	1.64	78	1.1	0.01	7.02	0.88019	8.11347	0.10320	1.65	0.20	633.1	9.9
DKE-107-9.2_AUS	type-1	0.12	164	7.7	0.05	14.43	0.85283	1.82913	0.10221	0.85	0.47	627.3	5.1
DKE-107-1.1_AUS	type-1	0.78	52	5.4	0.11	4.46	0.77761	5.77470	0.09974	1.45	0.25	612.9	8.5
DKE-107-7.2_AUS	type-1	0.95	15	2.3	0.16	1.31	0.84358	9.17651	0.10106	1.93	0.21	620.6	11.4
DKE-107-9.1_BR	type-1	0.90	130	4.5	0.04	11.59	0.86005	6.05695	0.10303	1.49	0.25	632.1	8.9
DKE-107-10.1_BR	type-1	1.51	86	2.6	0.03	7.67	0.81746	10.96214	0.10221	1.79	0.16	627.4	10.6

ANEXO IV

(análises LA-ICP-MS Lu-Hf em zircões ígneos)

sample	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-125E 1.1	0.282622	0.000046	0.000883	0.000013	623	8.1	0.99
DKE-125E 2.1	0.282666	0.000044	0.000921	0.000005	614	9.4	0.90
DKE-125E 3.1	0.282563	0.000034	0.001583	0.000014	631	5.9	1.14
DKE-125E 4.1	0.282583	0.000035	0.001189	0.000012	629	6.7	1.09
DKE-125E 5.1	0.282678	0.000035	0.000901	0.000024	593	9.4	0.89
DKE-125E 6.1	0.282656	0.000023	0.000966	0.000004	627	9.3	0.92
DKE-125E 7.1	0.282588	0.000027	0.001649	0.000025	632	6.7	1.09
DKE-125E 10.1	0.282495	0.000045	0.001162	0.000002	624	3.5	1.29
DKE-125E 11.1	0.282555	0.000033	0.001026	0.000006	623	5.6	1.15
DKE-125E 12.1	0.282623	0.000033	0.001231	0.000014	632	8.1	1.00
DKE-125E 8.1	0.282674	0.000034	0.001065	0.000025	638	10.2	0.87
DKE-125E 9.1	0.282692	0.000030	0.001006	0.000003	624	10.5	0.84
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-125A 4.1	0.282241	0.000028	0.001715	0.000016	638	-5.5	1.87
DKE-125A 5.1	0.282195	0.000038	0.002080	0.000016	644	-7.1	1.98
DKE-125A 6.1	0.282234	0.000037	0.001911	0.000023	656	-5.4	1.88
DKE-125A 7.1	0.282202	0.000045	0.001677	0.000046	636	-6.9	1.95
DKE-125A 8.1	0.282251	0.000048	0.001809	0.000018	655	-4.8	1.84
DKE-125A 11.1	0.282161	0.000039	0.002055	0.000022	644	-8.3	2.05
DKE-125A 12.1	0.282112	0.000043	0.002235	0.000036	656	-9.9	2.16
DKE-125A 1.1	0.282347	0.000033	0.001946	0.000026	637	-1.8	1.64
DKE-125A 10.1	0.282150	0.000039	0.002201	0.000016	653	-8.6	2.07
DKE-125A 2.1	0.282266	0.000031	0.001971	0.000035	654	-4.3	1.81
DKE-125A 3.1	0.282078	0.000037	0.002674	0.000017	640	-11.6	2.25
DKE-125A 9.1	0.282140	0.000039	0.002354	0.000025	646	-9.1	2.10
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-170 1.1	0.282526	0.000050	0.001803	0.000043	633	4.5	1.23
DKE-170 2.1	0.282639	0.000036	0.000627	0.000015	687	10.2	0.91
DKE-170 3.1	0.282495	0.000050	0.001390	0.000054	618	3.3	1.30
DKE-170 4.1	0.282590	0.000033	0.000941	0.000017	649	7.5	1.05
DKE-170 9.1	0.282535	0.000033	0.000676	0.000007	649	5.7	1.17
DKE-170 10.1	0.282741	0.000048	0.000482	0.000001	656	13.2	0.69
DKE-170 11.1	0.282548	0.000057	0.001731	0.000021	655	5.8	1.17
DKE-170 12.1	0.282563	0.000041	0.000695	0.000032	676	7.2	1.09
DKE-170 5.1	0.282471	0.000052	0.001134	0.000008	671	3.7	1.31
DKE-170 6.1	0.282507	0.000031	0.000824	0.000010	630	4.2	1.25
DKE-170 7.1	0.282598	0.000036	0.000806	0.000002	622	7.2	1.05
DKE-170 8.1	0.282504	0.000034	0.000787	0.000030	678	5.1	1.23
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-200A 1.1	0.282291	0.000054	0.000289	0.000007	642	-3.0	1.71
DKE-200A 2.1	0.282389	0.000040	0.000301	0.000006	650	0.7	1.49
DKE-200A 3.1	0.282349	0.000044	0.000218	0.000001	658	-0.5	1.57
DKE-200A 4.1	0.282262	0.000029	0.000246	0.000002	690	-2.9	1.75
DKE-200A 5.1	0.282413	0.000056	0.000188	0.000002	683	2.3	1.41
DKE-200A 6.1	0.282397	0.000034	0.000187	0.000001	664	1.3	1.46
DKE-200A 7.1	0.282226	0.000041	0.000246	0.000001	664	-4.8	1.85
DKE-200A 10.1	0.282251	0.000027	0.000216	0.000000	649	-4.2	1.80
DKE-200A 11.1	0.282428	0.000044	0.000245	0.000001	654	2.2	1.40
DKE-200A 8.1	0.282289	0.000040	0.000227	0.000001	646	-2.9	1.71
DKE-200A 9.1	0.282288	0.000044	0.000195	0.000001	652	-2.8	1.71

sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-211 1.1	0.282103	0.000037	0.000553	0.000001	614	-10.4	2.16
DKE-211 2.1	0.282028	0.000038	0.000491	0.000005	650	-12.2	2.30
DKE-211 3.1	0.281794	0.000040	0.000478	0.000002	642	-20.7	2.82
DKE-211 5.1	0.281080	0.000035	0.001189	0.000012	652	-46.1	4.38
DKE-211 9.1	0.282131	0.000041	0.000542	0.000005	623	-9.2	2.09
DKE-211 10.1	0.282209	0.000040	0.000587	0.000002	636	-6.2	1.91
DKE-211 12.1	0.282214	0.000043	0.000520	0.000001	639	-5.9	1.89
DKE-211 11.1	0.282149	0.000032	0.000623	0.000002	648	-8.0	2.04
DKE-211 4.1	0.282091	0.000039	0.000523	0.000001	632	-10.4	2.17
DKE-211 6.1	0.282044	0.000048	0.001026	0.000051	664	-11.6	2.27
DKE-211 8.1	0.282314	0.000038	0.000549	0.000003	615	-2.9	1.69
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-221 1.1	0.282807	0.000237	0.000454	0.000005	825	19.3	0.43
DKE-221 2.1	0.282617	0.000098	0.001325	0.000014	830	12.1	0.89
DKE-221 3.1	0.281580	0.000171	0.000881	0.000008	822	-24.5	3.20
DKE-221 9.1	0.282455	0.000099	0.001714	0.000022	850	6.6	1.26
DKE-221 10.1	0.282502	0.000099	0.001403	0.000005	850	8.5	1.15
DKE-221 13.1	0.282561	0.000121	0.001986	0.000020	812	9.4	1.05
DKE-221 14.1	0.282614	0.000070	0.001134	0.000035	831	12.2	0.89
DKE-221 11.1	0.282650	0.000078	0.001230	0.000006	880	14.4	0.78
DKE-221 12.1	0.282413	0.000064	0.001445	0.000037	825	4.8	1.36
DKE-221 5.1	0.282280	0.000350	0.001176	0.000014	843	0.5	1.65
DKE-221 7.1	0.282455	0.000078	0.001033	0.000010	839	6.8	1.25
DKE-221 8.1	0.282729	0.000100	0.001052	0.000012	795	15.5	0.65
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-231 1.1	0.282110	0.000045	0.000000	0.000000	597	-10.3	2.14
DKE-231 2.1	0.281979	0.000065	0.000004	0.000001	619	-14.4	2.42
DKE-231 3.1	0.282154	0.000054	0.000009	0.000001	605	-8.5	2.03
DKE-231 4.1	0.282097	0.000083	0.000006	0.000001	627	-10.1	2.15
DKE-231 5.1	0.281515	0.000338	0.000000	0.000000	691	-29.3	3.39
DKE-231 8.1	0.282073	0.000169	0.000002	0.000001	629	-10.9	2.20
DKE-231 11.1	0.281982	0.000082	0.000003	0.000000	631	-14.0	2.40
DKE-231 10.1	0.282192	0.000056	0.000016	0.000002	620	-6.9	1.94
DKE-231 12.1	0.282207	0.000087	0.000005	0.000001	630	-6.1	1.90
DKE-231 6.2	0.282641	0.000053	0.000006	0.000001	655	9.8	0.91
DKE-231 7.1	0.281848	0.000162	0.000004	0.000001	636	-18.7	2.70
DKE-231 9.1	0.282151	0.000041	0.000000	0.000001	657	-7.5	2.01
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-269 1.1	0.282685	0.000064	0.001128	0.000009	823	14.5	0.73
DKE-269 3.1	0.282020	0.000044	0.001300	0.000029	761	-10.5	2.28
DKE-269 3.2	0.282523	0.000030	0.000071	0.000000	643	5.4	1.18
DKE-269 5.1	0.282323	0.000051	0.000043	0.000001	658	-1.4	1.63
DKE-269 6.1	0.282461	0.000047	0.000119	0.000002	628	2.8	1.33
DKE-269 14.1	0.282406	0.000033	0.000068	0.000008	632	1.0	1.46
DKE-269 15.1	0.282385	0.000032	0.000146	0.000013	599	-0.5	1.53
DKE-269 1.2	0.282428	0.000042	0.000161	0.000001	635	1.8	1.41
DKE-269 10.1	0.282375	0.000033	0.000065	0.000001	631	-0.1	1.53
DKE-269 4.1	0.282486	0.000038	0.000143	0.000008	622	3.5	1.28
DKE-269 8.1	0.281805	0.000060	0.002386	0.000075	796	-17.9	2.77
DKE-269 9.1	0.282360	0.000036	0.000093	0.000001	630	-0.7	1.56

sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-273A 6.1	0.282192	0.000061	0.002460	0.000039	891	-2.3	1.86
DKE-273A 8.1	0.282237	0.000043	0.002115	0.000028	911	-0.1	1.74
DKE-273A 10.1	0.282248	0.000048	0.002089	0.000038	873	-0.5	1.73
DKE-273A 13.1	0.282147	0.000045	0.002482	0.000010	906	-3.6	1.96
DKE-273A 14.1	0.282160	0.000046	0.003188	0.000054	891	-3.8	1.96
DKE-273A 15.1	0.282272	0.000049	0.002589	0.000016	898	0.6	1.68
DKE-273A 16.1	0.282384	0.000041	0.001974	0.000046	959	6.2	1.37
DKE-273A 12.1	0.282145	0.000037	0.002442	0.000063	846	-4.9	1.99
DKE-273A 5.1	0.282262	0.000047	0.002218	0.000005	881	0.1	1.70
DKE-273A 5.2	0.282357	0.000025	0.000680	0.000007	638	-0.9	1.58
DKE-273A 7.1	0.282280	0.000045	0.001867	0.000059	904	1.5	1.63
DKE-273A 7.2	0.282393	0.000027	0.000547	0.000010	622	0.1	1.50
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-273B 2.1	0.282333	0.000045	0.002326	0.000005	790	14.5	0.73
DKE-273B 2.2	0.282501	0.000018	0.000612	0.000014	596	-10.5	2.28
DKE-273B 3.1	0.282173	0.000046	0.001811	0.000025	607	5.4	1.18
DKE-273B 8.1	0.282422	0.000030	0.000549	0.000015	621	-1.4	1.63
DKE-273B 9.1	0.282403	0.000027	0.000435	0.000011	610	2.8	1.33
DKE-273B 10.1	0.282590	0.000035	0.000911	0.000025	610	1.0	1.46
DKE-273B 12.1	0.282166	0.000056	0.002696	0.000042	744	-0.5	1.53
DKE-273B 14.1	0.282697	0.000036	0.000894	0.000015	637	1.8	1.41
DKE-273B 15.1	0.282538	0.000028	0.000751	0.000021	612	-0.1	1.53
DKE-273B16.1	0.282595	0.000029	0.000581	0.000013	639	3.5	1.28
DKE-273B 4.1	0.282437	0.000056	0.002244	0.000012	752	-17.9	2.77
DKE-273B 7.1	0.282635	0.000031	0.000542	0.000007	621	-0.7	1.56
sample/zircon	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm$ (2s)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm$ (2s)	Age (Ma)	eHf (t)	T_{DM1} (Ga)
DKE-277 1.1	0.282348	0.000028	0.000775	0.000006	641	-1.2	1.60
DKE-277 2.1	0.282309	0.000034	0.000432	0.000002	645	-2.3	1.67
DKE-277 8.1	0.282322	0.000031	0.000513	0.000003	653	-1.7	1.64
DKE-277 9.1	0.282201	0.000039	0.001151	0.000017	638	-6.6	1.94
DKE-277 10.1	0.282274	0.000027	0.000700	0.000001	646	-3.7	1.76
DKE-277 11.1	0.282275	0.000032	0.000440	0.000002	649	-3.5	1.75
DKE-277 12.1	0.282331	0.000031	0.000473	0.000003	664	-1.2	1.62
DKE-277 3.1	0.282269	0.000040	0.000700	0.000002	651	-3.7	1.77
DKE-277 4.1	0.282328	0.000031	0.000627	0.000004	647	-1.7	1.64
DKE-277 5.1	0.282348	0.000032	0.000528	0.000003	655	-0.8	1.58
DKE-277 6.1	0.282342	0.000034	0.000608	0.000012	638	-1.4	1.61
DKE-277 7.1	0.282345	0.000030	0.000668	0.000006	649	-1.1	1.60

ANEXO V

(análises SHRIMP Oxigênio $^{18}\text{O}/^{16}\text{O}$ em zircões ígneos)

Sample DKE-43	ratio (‰)	error (‰)	Sample DKE-45	ratio (‰)	error (‰)	Sample DKE-221	ratio (‰)	error (‰)
5.1	5.16	0.11	10.1	6.71	0.07	1.1	5.47	0.05
16.1	4.83	0.09	6.1	6.00	0.09	2.1	5.67	0.12
17.1	4.48	0.08	7.1	7.27	0.06	3.1	6.15	0.07
18.1	5.25	0.07	11.1	12.50	0.07	9.1	6.21	0.06
22.1	7.25	0.11	12.1	5.72	0.09	10.1	5.45	0.09
25.1	4.83	0.11	16.1	5.70	0.10	13.1	6.17	0.11
29.1	4.49	0.09	20.1	5.78	0.10	14.1	5.91	0.06
30.1	5.89	0.08	24.1	6.78	0.13	11.1	6.05	0.07
35.1	4.60	0.07	26.1	7.67	0.12	12.1	5.98	0.07
6.1	4.87	0.09	29.1	5.25	0.07	5.1	5.09	0.1
56.1	4.40	0.07	30.1	7.73	0.10	7.1	5.84	0.1
52.2	7.24	0.07	31.1	6.26	0.06	8.1	6.24	0.09
47.1	8.11	0.07	37.1	6.46	0.11	4.1	6.04	0.08
38.1	5.84	0.11	38.1	7.12	0.09	6.1	5.67	0.11
37.1	3.64	0.13	3.1	5.58	0.07			
34.1	5.30	0.06	34.1	6.90	0.14			
26.1	4.22	0.08	35.1	7.49	0.11			
19.1	5.24	0.14	36.1	6.28	0.06			
33.1	5.28	0.09	4.1	5.09	0.08			
41.1	4.68	0.05						
48.1	4.78	0.12						
49.1	6.15	0.08						
51.1	6.82	0.07						

Sample DKE-273A	ratio (‰)	error (‰)	Sample DKE-273B	ratio (‰)	error (‰)	Sample DKE-211	ratio (‰)	error (‰)
6.1	6.07	0.10	2.1	5.45	0.10	1.1	7.92	0.05
8.1	6.44	0.09	2.2	8.05	0.11	2.1	8.23	0.09
10.1	5.25	0.06	3.1	9.10	0.08	3.1	7.78	0.11
13.1	5.57	0.09	8.1	7.65	0.10	5.1	7.75	0.10
14.1	6.15	0.08	9.1	7.42	0.09	9.1	8.00	0.07
15.1	5.80	0.11	10.1	7.87	0.08	10.1	7.40	0.11
16.1	6.00	0.11	12.1	5.01	0.08	12.1	8.07	0.12
12.1	6.31	0.04	14.1	6.72	0.04	11.1	8.58	0.07
5.1	5.59	0.08	15.1	7.06	0.08	4.1	7.07	0.09
5.2	7.59	0.06	16.1	6.93	0.06	6.1	7.73	0.11
7.1	5.85	0.08	4.1	7.27	0.09	8.1	8.03	0.06
7.2	7.69	0.06	7.1	7.15	0.09	7.1	8.18	0.07
1.1	5.60	0.08	1.1	5.06	0.07			
3.1	8.12	0.08	1.2	6.70	0.08	Sample DKE-277		
4.1	7.87	0.08	5.1	4.89	0.06	2.1	5.92	0.09
8.2	5.71	0.09	6.1	6.41	0.15	8.1	6.24	0.07
9.1	8.17	0.05	11.1	5.53	0.10	9.1	5.48	0.08
11.1	5.85	0.08	11.2	6.51	0.03	10.1	5.61	0.14
11.2	7.69	0.06	12.2	5.45	0.10	11.1	6.25	0.10
12.2	7.57	0.08	13.1	7.61	0.08	12.1	5.90	0.09
12.3	7.88	0.07	17.1	4.64	0.07	3.1	5.79	0.11
2.1	5.96	0.12	18.1	5.94	0.08	4.1	5.54	0.16
20.1	5.20	0.04				5.1	6.01	0.09
20.2	5.84	0.09				6.1	5.72	0.09
						7.1	5.79	0.08

Sample DKE-170	ratio (‰)	error (‰)	Sample DKE-200A	ratio (‰)	error (‰)	Sample DKE-269	ratio (‰)	error (‰)
1.1	8.27	0.15	1.1	7.8	0.11	1.1	6.24	0.08
2.1	6.85	0.06	2.1	7.27	0.1	3.1	10.34	0.05
3.1	7.65	0.06	3.1	7.81	0.06	3.2	9.44	0.08
4.1	8.02	0.08	4.1	8.19	0.08	5.1	8.74	0.07
9.1	7.3	0.07	5.1	6.79	0.07	6.1	10.82	0.08
10.1	5.94	0.1	6.1	6.73	0.06	14.1	10.33	0.06
11.1	7.88	0.07	7.1	6.93	0.09	15.1	10.27	0.07
12.1	7.86	0.05	10.1	7.55	0.08	10.1	9.66	0.07
5.1	9.06	0.07	11.1	7.63	0.09	4.1	9.91	0.1
6.1	7.86	0.05	8.1	7.51	0.07	8.1	10	0.11
7.1	7.39	0.06	9.1	7.93	0.09	9.1	10.01	0.1
8.1	7.95	0.05				11.1	9.58	0.1
						12.1	8.81	0.06
						13.1	9.3	0.11
						2.1	8.9	0.1
						7.1	8.69	0.11

ANEXO VI

(análises LA-ICP-MS ETR em zircões)

Sample	Zircon type	Y	Nb	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Ta	Pb	Th	U	Ti
510-1	igneous core	1662	2.95	< 0.01	4.36	0.06	1.41	5.83	0.57	39.9	13.4	171	60.8	271	53.5	467	81.9	1.50	2.93	41.1	116	28.0
510-2	igneous core	1329	2.50	< 0.01	4.01	0.06	1.59	5.31	0.63	33.8	11.0	137	47.9	211	43.0	367	64.3	1.32	2.91	33.5	106	24.4
510-3	igneous core	1300	1.97	< 0.01	3.43	0.03	0.81	3.46	0.50	26.4	9.26	126	46.4	214	44.3	385	68.1	1.32	2.07	34.0	99.4	32.2
510-4	igneous core	1084	1.63	< 0.01	4.26	0.04	1.06	3.04	0.29	22.1	7.60	97	35.8	169	34.1	301	52.4	1.01	4.56	65.1	165	24.1
510-5	igneous core	1520	3.15	< 0.01	4.70	0.08	2.14	6.17	0.72	38.7	12.9	158	55.8	245	49.7	433	72.9	1.34	3.36	40.3	111	26.7
510-6	igneous core	709	2.39	< 0.01	3.04	0.03	0.63	2.01	0.24	14.4	5.02	66	24.3	114	24.8	228	40.8	3.30	2.00	24.5	73.3	16.7
510-24	igneous core	722	1.32	< 0.01	3.54	0.04	0.92	3.15	0.47	18.3	5.81	71	25.3	112	23.0	196	33.5	0.70	1.67	23.2	94.5	20.4
510-25	igneous core	1487	3.91	< 0.01	8.50	0.04	1.25	4.69	0.49	31.6	10.9	140	51.2	234	48.1	418	70.7	2.47	5.85	79.7	152	19.1
510-26	rim	234	0.30	20.3	4.76	8.77	51.7	17.8	6.14	36.1	5.93	42.7	8.73	21.2	2.26	9.67	1.24	0.05	0.08	0.52	40.1	13.0
510-27	rim	216	0.26	0.09	3.46	0.08	0.97	0.68	0.41	2.06	0.42	3.60	0.90	2.80	0.46	3.30	0.44	0.04	0.18	2.96	45.9	12.4
510-28	rim	14.2	0.23	< 0.01	1.12	0.00	0.09	0.44	0.27	1.25	0.21	1.92	0.40	1.47	0.24	2.10	0.26	0.04	< 0.01	0.26	24.9	7.83
510-12	rim	15.9	0.23	0.15	0.75	0.04	0.28	0.44	0.31	1.48	0.25	2.24	0.51	1.66	0.24	1.98	0.30	0.04	0.03	0.41	83.0	7.90
510-13	rim	10.6	0.27	< 0.01	0.63	< 0.01	0.08	0.25	0.14	0.81	0.16	1.41	0.34	1.18	0.15	1.29	0.21	0.04	0.03	0.28	33.2	7.20
510-14	rim	18.5	0.30	< 0.01	0.01	< 0.01	0.00	0.00	0.01	0.12	0.08	1.26	0.54	2.80	0.77	8.55	2.26	0.12	< 0.01	0.04	4.01	5.01
510-21	rim	14.5	0.30	< 0.01	0.18	< 0.01	0.00	0.02	0.09	0.67	0.18	1.78	0.45	1.54	0.32	3.19	0.67	0.06	< 0.01	0.16	5.63	6.48
510-18	rim	16.3	0.23	0.12	1.15	0.04	0.28	0.28	0.19	0.96	0.23	2.06	0.51	1.67	0.29	2.21	0.35	0.04	0.11	0.94	90.5	8.70
510-19	rim	13.5	0.24	0.03	0.53	0.01	0.10	0.26	0.18	1.10	0.23	1.92	0.42	1.32	0.22	1.65	0.25	0.04	0.06	0.57	78.4	10.8
510-8	rim	11.5	0.29	0.10	0.88	0.01	0.12	0.21	0.17	0.88	0.18	1.58	0.34	1.20	0.19	1.46	0.26	0.05	0.04	0.71	101	8.26
510-22	rim	30.3	0.26	< 0.01	1.00	0.02	0.44	0.77	0.45	2.61	0.54	4.17	1.02	3.23	0.51	4.09	0.60	0.05	0.10	1.72	169	17.6
510-23	rim	15.1	0.25	0.02	0.83	0.02	0.23	0.40	0.25	1.20	0.24	1.81	0.50	1.52	0.22	1.84	0.27	0.05	0.09	1.30	147	10.3
510-10	rim	25.9	0.30	0.01	1.10	0.01	0.14	0.53	0.35	2.04	0.40	3.38	0.81	2.56	0.40	3.02	0.46	0.05	0.06	1.03	119	10.7
510-7	rim	15.6	0.28	< 0.01	1.78	0.01	0.22	0.37	0.25	1.22	0.27	2.03	0.50	1.52	0.27	2.06	0.35	0.04	0.16	2.84	33.3	11.2
510-9	rim	18.7	0.30	< 0.01	1.45	< 0.01	0.08	0.23	0.16	0.80	0.18	1.81	0.56	3.36	1.04	20.1	4.50	0.04	0.14	2.17	28.9	9.55
510-16	rim	15.2	0.27	0.04	0.64	0.02	0.07	0.21	0.13	0.92	0.24	1.90	0.50	1.61	0.32	2.59	0.51	0.04	0.05	0.86	18.3	14.7
510-17	rim	29.6	0.25	< 0.01	1.73	0.02	0.54	0.89	0.54	2.44	0.54	4.61	1.03	2.98	0.51	3.49	0.53	0.03	0.36	5.32	40.5	10.6
510-29	rim	25.6	0.23	< 0.01	1.08	< 0.01	0.25	0.77	0.43	2.18	0.43	3.45	0.84	2.61	0.40	3.32	0.47	0.04	0.06	1.26	116	12.1
510-30	rim	23.1	0.25	< 0.01	0.60	< 0.01	0.11	0.32	0.30	1.73	0.36	2.94	0.80	2.38	0.41	3.31	0.45	0.05	0.03	0.31	72.1	9.81
510-11	homg. rounded	13.2	0.28	< 0.01	3.05	0.03	0.71	0.91	0.46	1.57	0.25	1.83	0.40	1.28	0.21	1.47	0.22	0.04	0.14	2.80	29.1	14.9
510-15	homg. rounded	13.6	0.28	< 0.01	4.09	0.03	0.32	0.55	0.29	1.30	0.21	2.01	0.41	1.35	0.21	1.71	0.24	0.04	0.26	4.70	33.8	10.2
510-20	homg. rounded	13.2	0.29	< 0.01	3.94	0.02	0.28	0.37	0.23	1.03	0.22	1.82	0.38	1.39	0.23	1.77	0.27	0.05	0.37	6.94	33.0	12.8
350-14	igneous core	2654	11.9	< 0.01	9.57	0.09	2.11	6.49	1.13	45.2	17.3	243	95.3	452	95.2	861	145	3.99	2.96	43.0	92.6	5.28
350-15	igneous core	117	0.44	< 0.01	1.61	0.01	0.29	2.16	1.73	17.2	3.89	24	3.68	8.7	1.22	9	1.31	0.14	0.03	0.79	73.1	5.15
350-16	igneous core	2099	8.15	< 0.01	6.00	0.07	1.76	5.22	1.03	35.2	13.33	189	74.9	359	76.0	700	124	3.08	2.77	41.3	79.6	4.86
350-17	igneous core	3564	5.63	0.01	5.44	0.21	4.02	9.92	2.00	67.4	24.65	341	129	596	122	1076	184	2.31	2.80	41.7	78.7	4.94
350-18	igneous core	2688	17.8	< 0.01	13.3	0.08	1.87	6.20	0.99	45.2	17.94	252	97.5	463	98.2	884	154	7.05	3.31	49.4	124	4.75
350-19	igneous core	3372	8.00	< 0.01	6.41	0.16	3.68	9.56	1.64	61.2	22.53	317	122	578	119	1075	182	3.22	3.20	48.2	102	6.19
350-20	igneous core	2485	5.65	< 0.01	5.62	0.11	2.83	8.00	1.27	47.3	17.10	236	89.0	428	91.2	832	140	2.57	2.62	39.6	97.3	8.30
350-21	igneous core	1013	3.28	< 0.01	3.06	0.03	0.71	2.51	0.67	18.5	6.71	91	35.1	169	36.4	343	61.2	1.29	0.64	10.8	49.4	19.3

ANEXO VII

(análises LA-ICP-MS em rutilo)

Sample	Al (ppm)	P (ppm)	Sc (ppm)	V (ppm)	Cr (ppm)	Mn (ppm)	Fe (ppm)	Cu (ppm)	Zn (ppm)	Y (ppm)	Zr (ppm)	Nb (ppm)	Mo (ppm)	Hf (ppm)	Ta (ppm)	W (ppm)	Zr-in T (°C)
RT506-01	35.40	18.60	1.58	3153	81.30	0.02	1979	16.09	9.28	0.05	355.2	297.1	10.45	16.92	17.60	5.19	751
RT506-04	43.80	18.30	1.28	3890	56.50	6.40	1710	15.38	9.00	0.05	263.0	249.5	9.48	13.78	13.47	4.99	724
RT506-05	75.80	14.80	2.36	4400	117.90	12.10	2570	15.40	10.21	0.07	297.5	228.2	10.36	14.03	10.26	5.42	735
RT506-06	27.67	20.10	2.24	2308	196.60	0.04	1635	14.68	9.22	0.06	574.7	384.3	11.12	20.86	20.89	20.64	796
RT506-07	18.78	16.30	2.20	1315	548.00	0.02	1717	15.11	8.06	0.06	292.9	591.0	13.31	15.42	37.92	40.96	734
RT506-09	28.26	17.60	2.14	1135	1120.00	0.04	980	14.26	7.89	0.06	347.6	415.8	12.82	15.90	20.26	47.90	749
RT506-10	19.37	23.50	2.00	1499	337.70	1.86	1588	15.21	9.16	0.06	203.4	443.6	11.39	12.17	35.43	32.58	703
RT506-11	22.20	17.10	1008.00	3114	169.10	0.12	1013	14.58	8.37	0.07	224.6	311.3	11.21	13.05	15.55	5.52	711
RT506-12	18.50	15.20	2.39	1412	1375.00	0.00	1680	15.29	9.30	0.06	419.9	1494.0	14.03	18.92	151.60	82.40	766
RT506-13	53.30	15.80	1.88	3150	68.80	0.96	1787	15.94	7.62	0.14	321.2	272.0	9.39	15.84	13.03	5.23	742
RT506-14	14.95	16.30	2.64	1424	735.00	64.90	2679	13.10	17.80	0.07	115.8	426.5	9.58	10.48	26.25	47.30	658
RT506-15	28.08	19.30	1.55	3089	63.60	0.11	1553	13.89	7.62	0.06	415.8	252.2	9.51	18.00	19.90	5.42	765
RT514-01	130.00	19.80	1625.00	1894	101.00	24.27	2384	13.76	9.06	0.06	212.8	397.3	8.74	11.55	32.14	10.68	706
RT514-02	17.02	17.90	1.54	2118	75.50	0.00	1604	14.28	8.07	0.06	274.8	371.8	8.43	13.48	23.23	9.15	728
RT514-03	114.20	19.60	2.06	1840	112.60	0.01	1655	13.55	7.99	0.05	304.0	385.5	8.19	13.40	20.50	8.58	737
RT514-04	95.00	19.00	2.25	2247	138.80	0.00	1858	13.97	7.90	0.06	535.8	344.0	8.53	21.10	25.95	7.29	789
RT514-05	53.50	14.30	2.09	2076	101.70	0.06	1557	13.78	8.34	0.07	298.4	396.8	8.95	15.15	26.64	8.60	735
RT514-06	239.00	15.90	1.79	1855	115.30	0.11	1616	12.55	8.27	0.06	242.9	381.6	8.65	12.37	20.47	8.30	717
RT514-07	88.90	17.90	1.59	1998	105.30	0.02	1912	13.16	7.34	0.06	234.5	459.3	8.54	12.87	26.88	9.73	714
RT514-08	49.00	16.70	1.85	2034	105.20	0.11	1484	14.00	18.20	0.06	248.7	346.7	8.73	13.85	19.25	8.78	719
RT514-09	65.30	18.00	1.73	1886	101.90	0.03	1722	350.00	7.40	0.06	243.6	418.2	8.76	15.37	33.00	8.48	718
RT514-10	290.00	18.30	2111.00	1809	105.70	0.00	1957	13.76	8.02	0.06	250.0	342.3	7.99	12.36	15.75	9.43	720
RT514-11	332.00	19.10	2.63	1667	183.20	0.03	1741	13.96	8.11	0.07	279.9	430.4	8.44	13.68	19.11	10.53	730
RT514-12	19.55	15.50	2.31	1804	97.30	0.00	1589	12.48	7.45	0.07	325.0	370.3	8.29	16.38	17.10	7.87	743
RT514-13	78.60	16.10	1633.00	1729	112.60	116.00	1519	28.40	7.59	0.22	120.6	460.5	8.82	9.85	33.05	10.79	661
RT514-14	26.86	17.30	1.19	1878	107.80	0.21	1688	12.87	8.40	0.05	148.3	377.0	8.93	11.60	19.64	8.37	677
RT514-15	74.20	17.90	2.41	1894	143.90	0.00	1750	12.67	9.24	0.04	248.0	415.8	9.06	13.93	20.12	9.19	719
RT350-1	78.80	13.80	1.95	1957	88.40	-0.05	1769	9.01	9.00	0.06	222.5	405.7	7.99	12.29	27.60	8.59	722
RT350-2	78.30	13.40	2.22	1972	95.80	0.10	1787	9.10	6.29	0.04	202.6	391.3	8.33	11.21	26.87	9.97	715
RT350-3	48.00	17.50	1.78	2146	96.30	0.07	1838	8.92	6.90	0.04	300.8	406.4	8.26	14.50	23.75	9.05	748
RT350-4	135.30	17.50	1.31	2156	56.80	0.13	1520	9.27	8.30	0.05	250.1	352.6	8.07	14.80	23.47	8.79	732
RT350-5	73.00	18.80	2.36	1861	85.50	0.07	1801	9.97	6.01	0.05	198.7	387.4	8.28	11.65	25.24	9.18	713
RT350-6	245.30	15.10	1.68	1953	55.90	0.13	1639	9.04	6.57	0.05	268.5	366.0	8.15	13.57	25.60	9.66	738
RT350-7	77.20	15.10	1.96	1962	65.40	3.18	1997	9.64	9.42	0.05	211.9	324.3	7.93	11.92	15.94	9.36	718
RT350-8	139.20	16.90	2.16	1986	94.90	0.08	2021	9.18	6.86	0.05	174.3	392.3	8.34	11.07	21.25	10.12	703
RT350-9	176.30	17.50	1.98	1738	494.00	-0.01	1937	9.99	9.10	0.17	160.0	1192.0	10.41	6.10	65.61	11.38	696
RT350-10	60.90	17.60	1.34	1853	81.00	0.72	1547	9.74	7.77	0.26	179.6	342.3	8.50	10.98	23.72	9.62	705
RT350-11	23.18	20.50	2.27	2020	76.10	0.03	1494	8.97	5.90	0.05	263.2	352.7	8.82	14.28	21.77	8.59	736
RT350-12	400.00	15.00	1.26	1606	564.20	3.40	2384	9.27	8.13	0.13	53.8	919.0	9.94	2.99	53.09	9.05	618
RT350-13	86.10	14.70	1.75	2120	63.80	0.34	2062	9.19	11.63	0.06	282.9	371.2	7.65	13.95	20.36	7.90	743

ANEXO VIII

(análises de microsonda em onfacita, granada e fengita)

Sample	S-514A	S514A'	S-514B	S-514C	DKE-350AD	DKE-350BD	DKE-350C	DKE-350DD	DKE-350E
Garnet									
SiO ₂	38.81	38.73	37.69	37.58	39.23	38.40	39.83	39.12	38.70
TiO ₂	0.20	0.34	0.20	0.05	0.03	0.21	0.05	0.09	0.14
Al ₂ O ₃	22.11	21.95	21.68	21.88	22.66	21.37	22.73	21.67	21.57
FeO	22.84	22.86	23.25	22.29	24.32	25.31	21.22	25.18	24.85
MnO	0.24	0.24	0.22	0.24	0.46	0.64	0.29	0.92	0.69
MgO	8.83	8.63	9.04	8.67	7.10	5.16	8.40	5.48	6.85
CaO	7.54	8.16	7.57	8.41	7.96	10.11	9.36	9.56	8.21
Cr ₂ O ₃	0.08	0.04	0.00	0.02	0.00	0.03	0.05	0.03	0.07
Total	100.65	100.96	99.64	99.14	101.76	101.24	101.93	102.04	101.09
Si	2.95	2.95	2.91	2.91	2.97	2.97	2.98	2.99	2.97
Al	1.98	1.97	1.98	2.00	2.02	1.95	2.00	1.95	1.95
Ti	0.01	0.02	0.01	0.00	0.00	0.01	0.00	0.01	0.01
Fe ²⁺	1.45	1.45	1.50	1.45	1.54	1.64	1.33	1.61	1.60
Mn	0.02	0.02	0.02	0.02	0.03	0.04	0.02	0.06	0.05
Mg	1.00	0.98	1.04	1.00	0.80	0.59	0.94	0.62	0.78
Ca	0.62	0.67	0.63	0.70	0.65	0.84	0.75	0.78	0.68
Cr	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cations	8.04	8.05	8.09	8.08	8.01	8.04	8.02	8.03	8.04
X _{Alm}	0.47	0.47	0.47	0.46	0.51	0.53	0.44	0.52	0.51
X _{Sps}	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.02	0.01
X _{Prp}	0.33	0.31	0.33	0.32	0.27	0.19	0.31	0.20	0.25
X _{Grs}	0.20	0.21	0.20	0.22	0.21	0.27	0.25	0.25	0.22
Omphacite									
SiO ₂	57.61	55.33	54.32	53.96	56.99	55.97	56.44	56.38	54.70
TiO ₂	0.13	0.13	0.10	0.12	0.05	0.04	0.05	0.05	0.04
Al ₂ O ₃	15.63	14.89	15.10	14.26	9.84	8.00	10.86	8.75	8.23
Cr ₂ O ₃	0.03	0.03	0.03	0.03	0.05	0.25	0.06	0.14	0.38
FeO	2.59	2.67	2.42	2.46	3.55	6.12	4.11	5.54	4.78
MnO	0.00	0.02	0.02	0.02	0.01	0.01	0.01	0.01	0.01
MgO	7.15	6.93	6.74	7.21	10.25	9.81	8.79	9.54	9.91
CaO	10.91	11.13	10.53	11.22	15.41	15.28	14.23	14.78	15.16
Na ₂ O	8.36	8.02	8.12	7.73	5.54	5.50	6.53	5.84	5.38
Total	102.41	99.15	97.39	97.00	101.69	100.99	101.07	101.03	98.61
Si	1.97	1.96	1.95	1.95	2.00	1.99	1.98	2.00	1.99
Al(IV)	0.03	0.04	0.05	0.05	0.00	0.01	0.02	0.00	0.01
Al(VI)	0.60	0.58	0.59	0.56	0.40	0.33	0.43	0.36	0.34
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ³⁺	0.00	0.01	0.02	0.03	0.00	0.05	0.02	0.03	0.04
Fe ²⁺	0.07	0.07	0.06	0.05	0.10	0.13	0.10	0.13	0.10
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.37	0.37	0.36	0.39	0.54	0.52	0.46	0.50	0.54
Ca	0.40	0.42	0.41	0.43	0.58	0.58	0.54	0.56	0.59
Na	0.56	0.55	0.57	0.54	0.38	0.38	0.45	0.40	0.38
Cr	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01
Cations	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00	4.00
Phengite									
SiO ₂	53.03	53.04	50.17	50.18	53.86	53.29	51.01	51.99	51.14
TiO ₂	0.54	0.54	0.77	0.79	0.24	0.28	0.24	0.26	0.23
Al ₂ O ₃	27.15	27.16	26.41	26.64	26.47	26.26	27.68	25.97	25.42
Cr ₂ O ₃	0.02	0.03	0.03	0.05	0.05	0.05	0.06	0.07	0.08
FeO	1.18	1.18	1.08	1.18	1.76	1.44	1.55	1.55	1.72
MnO	0.00	0.00	0.00	0.01	0.00	0.02	0.00	0.00	0.01
MgO	4.66	4.67	4.14	4.19	4.58	4.41	3.64	4.29	4.18
CaO	0.02	0.02	0.01	0.00	0.00	0.04	0.01	0.05	0.11
Na ₂ O	0.36	0.36	0.42	0.40	0.41	0.65	0.72	0.63	0.75
K ₂ O	10.19	10.20	10.49	10.50	10.35	10.71	9.84	10.81	9.88
BaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	97.15	97.20	93.51	93.95	97.72	97.14	94.75	95.63	93.52
Si	6.88	6.87	6.80	6.77	6.96	6.95	6.80	6.91	6.92
Al(IV)	1.13	1.13	1.20	1.23	1.04	1.05	1.20	1.09	1.08
Al(VI)	3.02	3.02	3.01	3.01	2.99	2.98	3.14	2.97	2.98
Ti	0.05	0.05	0.08	0.08	0.02	0.03	0.02	0.03	0.02
Cr	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.01
Fe ²⁺	0.13	0.13	0.12	0.13	0.19	0.16	0.17	0.17	0.20
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.90	0.90	0.84	0.84	0.88	0.86	0.72	0.85	0.84
Ca	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.02
Na	0.09	0.09	0.11	0.11	0.10	0.16	0.19	0.16	0.20
K	1.69	1.69	1.81	1.81	1.71	1.78	1.67	1.83	1.71
Cations	13.89	13.89	13.98	13.98	13.90	13.98	13.93	14.03	13.97
T (°C)	715.17	721.28	631.00	578.69	621.14	667.28	754.19	686.59	622.35
P (Kbar)	32.84	32.84	29.46	28.72	28.35	29.75	29.28	29.46	27.76

ANEXO IX

(análises ICP-MS Sr-Nd em rochas ígneas)

Sample	Rb (ppm)	Sr (ppm)	Sm (ppm)	Nd (ppm)	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	(⁸⁷ Sr/ ⁸⁶ Sr) _i	eSr _(T1)	eNd _(T1)	T ₁ (Ma)
DKE-170	148.0	131.8	6.0	30.8	3.2583	0.7331	0.1169	0.5124	0.7028	-13.19	1.70	663
DKE-200	322.5	397.6	8.9	52.0	2.3530	0.7325	0.1034	0.5120	0.7106	97.19	-5.45	654
DKE-211	134.6	688.1	11.8	65.8	0.5662	0.7108	0.1083	0.5121	0.7056	26.61	-4.25	637
DKE-221	49.2	203.1	2.6	14.8	0.7018	0.7109	0.1084	0.5124	0.7026	-13.63	4.98	831
DKE-269	186.7	1493.8	20.7	155.4	0.3619	0.7141	0.0806	0.5116	0.7109	100.88	-10.75	628
DKE-231	258.9	598.4	6.9	42.1	1.2544	0.7256	0.0995	0.5117	0.7144	150.99	-9.65	627
DKE-273A	73.6	211.7	2.6	12.6	1.0075	0.7148	0.1265	0.5124	0.7020	-20.80	3.84	892
DKE-273B	107.8	369.3	3.8	23.6	0.8458	0.7154	0.0979	0.5121	0.7079	58.83	-3.55	618
DKE-277	65.8	503.5	4.7	24.6	0.3784	0.7091	0.1153	0.5120	0.7056	27.01	-5.87	648

ANEXO X

(análises geoquímicas – granitóides)

Sample	KE-267	KE-273A	KE-233A	KE-263B	KE-270	KE-254	KE-221	KE-170A	KE-171A	KE-200A	KE-259	KE-276	KE-277	KE-278
Unit	LC	LC	LC	LC	LC	LC	LC	LC	LC	LC	BO	BO	BO	BO
SiO ₂ (wt%)	67.68	68.2	65.21	67.61	67.97	66.17	65.38	55.4	55.5	62.21	67.07	68.56	69	69.05
Al ₂ O ₃ (wt%)	13.82	14.33	15.04	14.28	15.33	16.32	15.24	16.01	17.44	17.12	15.78	15.01	15.46	14.99
CaO (wt%)	2.13	1.96	2.81	2.35	1.73	2.92	2.55	5.22	7.67	3.47	2.03	2.18	2.45	1.98
K ₂ O (wt%)	5.33	5.17	4.14	3.99	5.98	3.42	2.94	3.71	1.77	4.86	7.39	2.41	3.7	4.91
Na ₂ O (wt%)	3.13	3.68	3.78	3.93	3.85	4.18	3.84	3.89	3.99	4.22	2.61	4.75	4.03	3.29
FeO (wt%)	2.43	2.91	3	1.89	1.62	2.46	3.48	3.51	3.52	3.5	2.31	3.15	1.99	2.32
Fe ₂ O ₃ (wt%)	3.76	4.64	4.84	3.39	3.73	4.04	5.8	8.52	9.77	6.01	4.05	4.99	3.19	2.99
MgO (wt%)	1.36	0.76	0.95	1.16	0.07	1.12	1.23	3.05	4.41	1.78	0.79	0.93	0.9	1.05
MnO (wt%)	0.09	0.1	0.12	0.06	0.06	0.06	0.12	0.16	0.12	0.12	0.03	0.17	0.08	0.07
TiO ₂ (wt%)	0.66	0.73	0.77	0.48	0.19	0.45	0.75	1.58	1.2	1.1	1.09	0.8	0.37	0.39
P ₂ O ₅ (wt%)	0.21	0.21	0.2	0.15	0.19	0.16	0.24	0.62	0.35	0.37	0.28	0.24	0.18	0.07
LOI	0.31	0.1	0.08	0.89	0.24	0.31	0.22	0.32	0.35	0.37	0.83	0.21	0.3	0.16
Sum	98.51	99.88	97.94	98.28	99.34	99.16	98.26	98.46	102.6	101.63	101.96	100.28	99.67	98.95
Ba (ppm)	836	1384	1782	892	1230	860	1091	1158	862	1865	3141	479	868	741
Be (ppm)	4.5	1.7	2.5	3.4	1.4	3.9	1.5	3.2	0.8	4	1.3	4	3.6	0.8
Cs (ppm)	3.61	1.22	2.62	1.68	1.04	3.82	1.55	3.05	1.49	3.65	0.32	7.72	4.62	2.65
Ga (ppm)	23.5	26.7	26	25.6	23.2	25.2	19.7	17.9	17.9	31.9	20.9	28.7	19	22.1
Hf (ppm)	5.89	10.33	8.97	2.92	14.34	4.9	9.09	6.58	2.86	5.39	19.04	9.63	2.47	3.75
Nb (ppm)	14.25	21.65	20.13	8.26	2.11	3.44	6.48	11.63	6.7	19.39	4.23	29.97	3.67	5.95
Rb (ppm)	184.4	127.7	127.2	115.1	126	144.9	84.4	124.6	56.8	187.4	160.7	133.5	92.4	159.3
Sn (ppm)	1.4	0.5	1.6	1.7	4.6	1.1	3.5	5.7	4	4.6	3.3	2.6	2.7	2.1
Sr (ppm)	399.7	221.2	318.9	587	272.7	404.9	535.9	675.1	536.8	745.3	1472.2	175.8	239.9	152
Ta (ppm)	0.89	2.42	0.98	2.05	0.05	0.05	0.05	1.47	0.05	8.55	0.05	2.27	0.33	0.23
Th (ppm)	12.6	9.3	6.3	10.1	2.7	13.9	13.4	4.2	0.1	5.1	31.8	9.7	3.6	5.7
U (ppm)	2.84	1	0.95	2.22	1.04	1.74	1.11	2.16	0.86	2.73	1.75	3.49	1.31	0.85
W (ppm)	0.1	0.2	0.1	0.5	2	0.1	0.1	0.1	0.1	8	0.1	0.1	0.7	0.1
Y (ppm)	25.8	43.87	37.77	11.34	9.7	7.01	14.76	20.08	18.92	20.18	11.02	48.84	10.95	27.4
Zr (ppm)	216.8	506.1	465.7	116.9	677	191.9	314.5	259	100.3	345.8	884.2	471.8	93.8	148.4
Co (ppm)	10.9	6.3	5.1	8.8	1.9	7.7	9.4	15.6	15.1	14	4.4	5.1	5.2	6.7
Cr (ppm)	43	9	9	18	5	15	26	36	26	24	6	10	4	41
Cs (ppm)	3.55	1.13	2.38	1.59	0.24	3.67	1.33	3.2	1.56	3.73	0.27	7.98	4.27	2.48
Cu (ppm)	22.1	18	9.7	6.7	2.2	17.4	36.7	19.3	8.9	29.4	22.4	7.8	3.1	7
Ga (ppm)	7.8	7.9	7	5.9	3.6	6.8	9.2	8.2	5.9	10.1	7	8.7	4.4	6.3
Hf (ppm)	0.31	0.17	0.19	0.1	0.06	0.14	0.09	0.21	0.18	0.24	0.12	0.09	0.05	0.07
La (ppm)	49.8	66.5	53.5	29.4	6.3	38.6	22.3	45	13.8	52	141.4	40.4	18.4	17.8
Li (ppm)	30	24	25	38	1	38	20	33	41	46	26	89	38	31
Nb (ppm)	1.46	1.38	1.53	0.65	0.25	0.5	0.87	0.75	0.28	1.24	0.97	2.52	0.73	1.22
Ni (ppm)	23.3	8.1	5.4	17	2.7	7.5	13.3	19.4	17.1	15.9	2.3	6.1	2.3	19.2
Pb (ppm)	3.5	2.2	2.5	4.3	2.6	2.9	4.6	2.1	2.3	4.2	5.7	2.8	1.1	1.5
Rb (ppm)	137.8	60.6	71.5	81.3	9.7	101.2	74.9	134	67.6	141.6	64	123.7	61.5	100.4
Sc (ppm)	6	10.7	8.1	3.1	0.6	1.5	4.6	4	5.9	6.7	1.3	11	2.4	5.6
Sn (ppm)	2.2	2.8	3.6	1.2	0.6	0.8	1.7	1.3	1	1.5	9.9	3.4	1.1	3.4
Sr (ppm)	33.8	12.4	15.6	44.7	10.2	13.9	21.6	24.8	28.6	48.2	78	7.1	9.3	7.2
Th (ppm)	22.9	13.9	9.8	16.2	1.5	19.2	13.1	9.3	2.1	9.9	51.6	9	4.9	4.5
U (ppm)	2.03	0.37	0.49	1.79	0.15	0.77	0.31	1.87	0.61	1.47	0.91	1.85	1	0.2
V (ppm)	38	27	23	31	35	36	48	66	88	77	28	17	21	32
Y (ppm)	17.9	29.28	31.17	7.33	7.95	5.6	10.32	14.13	10.78	11.09	6.3	24.37	7.36	6.39
Zn (ppm)	65	66	62	45	20	50	71	63	41	84	70	91	38	32
Zr (ppm)	6.2	3.3	4.1	2.2	3	5	1.1	1.7	2.3	5.6	1	1.8	0.7	1.7
Ce (ppm)	96.2	133.9	96.6	54.4	15.6	93.7	116.5	142.3	54.1	118.9	345.9	124.2	33.9	72.2
Dy (ppm)	5.09	7.87	7.23	2.11	1.66	1.8	3.83	6.43	4.41	3.86	3.8	9.31	1.85	5.62
Er (ppm)	2.33	4.21	3.71	0.95	1.24	0.71	1.98	2.95	2.43	1.64	0.92	4.79	1.19	2.72
Eu (ppm)	1.88	2.17	2.66	0.97	1.82	0.62	1.43	2.74	1.46	2.52	2.11	1.94	0.88	0.77
Gd (ppm)	6.87	10.16	8.67	2.59	2.06	2.72	5.69	8.36	4.67	6.48	8.21	10.51	2.46	5.67
Ho (ppm)	0.88	1.56	1.4	0.34	0.34	0.19	0.66	1.09	0.8	0.63	0.45	1.57	0.33	1
La (ppm)	46.6	66	52	30.7	7.9	51.5	62	71.8	31.4	62.5	206.3	62.7	23.8	32.6
Lu (ppm)	0.15	0.7	0.58	0.19	0.26	0.27	0.43	0.58	0.62	0.2	0.17	0.57	0.09	0.4
Nd (ppm)	47	57.6	46.6	21.6	8.6	34.8	49.9	72.7	27.3	54.1	132.8	55.9	15	33.2
Pr (ppm)	12.1	15.08	11.64	5.46	1.84	10.19	13.5	18.83	6.95	14.66	37.12	15.57	3.8	8.46
Sm (ppm)	9	11.1	8.9	3.5	2.2	4.3	8.9	12.4	5.1	8.6	17.6	11.3	2.6	6.3
Tb (ppm)	0.92	1.48	1.13	0.39	0.26	0.21	0.62	0.94	0.6	0.96	0.62	1.53	0.37	0.96
Tm (ppm)	0.28	0.55	0.47	0.13	0.18	0.05	0.19	0.34	0.3	0.29	0.06	0.64	0.13	0.34
Yb (ppm)	2	3.5	3.9	1	1.1	0.7	1.5	2.4	2.4	1.9	0.5	4.1	1	2.4
Eu/Eu*	0.74	0.63	0.93	0.99	2.63	0.56	0.62	0.83	0.92	1.04	0.72	0.55	1.07	0.4
(La/Sm) _n	2.84	3.26	3.2	4.81	1.97	6.57	3.82	3.18	3.38	3.99	3.94	3.04	5.02	2.84
(La/Yb) _n	14.12	11.43	8.08	18.61	4.35	44.59	25.05	18.13	7.93	19.94	91.52	9.27	14.42	8.23
(Tb/Yb) _n	1.96	1.8	1.23	1.66	1.01	1.28	1.76	1.67	1.06	2.15	5.11	1.59	1.57	1.7
Tot REE	231.3	315.88	245.49	124.33	45.06	201.76	267.13	343.86	142.54	277.24	125.63	304.63	87.4	172.64
ASI	0.93	0.94	0.95	0.95	0.96	1.03	1.08	0.8	0.78	0.92	1.06	1.04	1.02	1.05

Sample Unit	KE-279 BO	KE-176B ST	KE-180 ST	KE-191A ST	KE-191B ST	KE-194A ST	KE-195 ST	KE-205A ST	KE-208 ST	KE-209 ST	KE-211 ST	KE-212 ST	KE-213 ST	KE-214 ST	KE-244A ST
SiO ₂ (wt%)	69.45	58.76	60	61	61.09	61.67	61.96	62.68	62.97	62.98	63.02	63.27	63.33	63.45	65.77
Al ₂ O ₃ (wt%)	15.08	17.47	14.88	13.45	14.85	16.56	14.89	15.65	14.78	16.7	16.53	15.84	15.98	15.97	16.95
CaO (wt%)	1.91	5.29	4.12	3.41	3.84	4.27	3.37	3.25	3.3	2.89	4.25	3.87	3.56	3.2	2.39
K ₂ O (wt%)	3.77	3.06	4.32	6.66	4.76	3.45	5.35	5.56	4.7	3.09	2.44	3.73	5.17	4.57	4.94
Na ₂ O (wt%)	4.09	5.17	4.21	2.99	3.28	3.62	3.23	3.44	3.12	4.66	4.08	3.62	3.75	3.71	4.46
FeO (wt%)	1.51	3.47	3.34	2.84	3.07	3.97	2.68	3.83	3.36	4.08	3.25	2.87	3.06	3.88	2.33
Fe ₂ O ₃ (wt%)	2.1	6.72	7.35	5.48	5.63	5.87	5.48	5.6	5.34	6.64	5.6	6.11	5.96	5.59	3.69
MgO (wt%)	0.63	3.35	2.62	4.21	2.81	1.89	2.26	2.27	2.2	1.38	1.55	1.63	2.62	1.63	0.99
MnO (wt%)	0.03	0.15	0.14	0.08	0.12	0.08	0.09	0.09	0.13	0.2	0.11	0.09	0.1	0.11	0.05
TiO ₂ (wt%)	0.27	1.05	1.51	0.83	1.1	0.83	0.98	1.02	0.96	1.18	0.63	0.93	1.06	1.05	0.5
P ₂ O ₅ (wt%)	0.08	0.35	0.58	0.51	0.46	0.35	0.35	0.34	0.36	0.35	0.16	0.24	0.41	0.36	0.18
LOI	0.28	0.87	0.5	0.3	0.36	0.6	0.33	0.54	0.71	0.23	0.47	0.2	0.36	0.34	0.51
Sum	98.68	102.26	100.24	98.96	98.31	99.19	98.31	100.44	98.58	100.28	98.84	99.54	102.31	99.99	100.42
Ba (ppm)	745	1374	1223	2307	1120	1963	1189	1071	1195	608	809	1630	1028	1584	1735
Be (ppm)	2	3.1	3.4	9.8	4.9	2.7	5.6	5.7	3.9	4.1	2	2.2	5.1	3.8	2.1
Cs (ppm)	1.35	2.16	1.45	16.81	5.44	1.41	4.47	5.88	1.51	5.76	1.44	1.03	3.21	2.36	2.12
Ga (ppm)	19.3	29.6	29.9	24.8	20.3	28.8	19.8	27.8	26.4	31.9	22.2	19.3	27.8	29.3	26.4
Hf (ppm)	3.04	3.65	6.14	13.54	8.9	4.81	10.98	7.88	7.2	10.87	4.33	10.46	8.83	8.06	6.7
Nb (ppm)	6.87	6.17	25.66	17.58	10.93	5.82	12.55	16.74	14.5	25.76	6.33	3.87	10.54	21.64	4.94
Rb (ppm)	106.5	70.6	150.6	318.7	195.7	109.8	271.9	246.7	140.1	160	81	90.3	195.2	140.1	87.1
Sn (ppm)	2	8.6	5.7	6	9.8	1.5	6.5	4.3	2.8	2.2	2.3	2.8	1.6	2.3	1.3
Sr (ppm)	329.6	1118.8	638.4	702.9	501.8	916.7	560.4	505.7	541.6	224.7	436.1	419.3	452.2	666.7	604.4
Ta (ppm)	0.22	0.39	7.68	0.42	1.49	2.82	0.79	1.48	0.73	0.99	0.05	0.05	0.05	2.18	0.05
Th (ppm)	6.5	2.8	17.7	52.4	22.9	5.8	22.9	15.1	13.3	9.1	0.9	0.1	8.1	7.9	3.2
U (ppm)	1.06	0.51	1.58	9.63	5.54	0.81	4.03	3.22	0.95	1.84	0.61	0.89	1.55	2.34	0.76
W (ppm)	0.1	0.1	0.9	0.1	0.1	1.6	0.1	0.1	0.1	0.2	0.1	0.1	0.1	0.1	0.1
Y (ppm)	5.53	15.55	27.32	25.57	22.33	17.18	21.8	17.8	32.89	40.13	18.13	18.37	18.76	35.32	5.99
Zr (ppm)	115	181.4	298	540.4	316.5	257.2	349.5	321.8	277.5	628.1	155.7	472.4	328.2	383.1	273.7
Co (ppm)	2.7	15.7	17.3	14.2	12.3	11.1	10.1	15.6	12.2	8.1	10.9	8.4	15.1	14.1	6.9
Cr (ppm)	8	39	39	139	60	15	50	47	58	14	11	11	48	18	7
Cs (ppm)	1.08	2.17	1.46	19.06	5.63	1.53	3.38	5.44	1.47	5.58	1.63	1.09	3.24	2.53	2.07
Cu (ppm)	2.6	3.8	68.4	76.5	11.2	12.3	9.9	45.1	14.1	19.3	16.2	22.4	19.4	19.8	7.4
Ga (ppm)	4	9.5	9.9	11.5	8.4	11.2	7.2	10.1	9.2	10.4	7.3	5.5	9.8	10.9	7.3
Hf (ppm)	0.12	0.11	0.36	0.23	0.42	0.11	0.3	0.35	0.25	0.19	0.08	0.08	0.18	0.33	0.09
La (ppm)	15.5	7.9	43.4	90.3	46.5	55.6	52	53.1	93.6	49.9	6.2	10.4	53.7	42.2	23.8
Li (ppm)	11	32	24	55	38	24	31	37	21	66	29	23	28	45	24
Nb (ppm)	0.52	0.3	1.96	1.74	1.17	0.34	1.8	1.3	0.8	1.55	0.65	0.63	0.78	0.98	0.39
Ni (ppm)	2.8	36.2	26.5	74.3	30.7	5	25.7	29.9	27.9	9.5	4.4	5.8	32.7	12.8	5.6
Pb (ppm)	3.7	1.8	3	48.9	4	1.8	4.5	5.9	2.9	2.3	3.4	3	4	3	2.2
Rb (ppm)	58.2	79.3	120.4	194.6	168.2	108.5	153.8	203.2	109	156.1	75.3	64.8	143.4	124.4	61.8
Sc (ppm)	1.9	6.5	6.8	2	4.2	6.6	4	6	9.2	11.4	5.6	4.2	7.4	7.4	3.9
Sn (ppm)	1.8	1.6	2.6	2.4	2.7	3.3	2.5	3.6	2.9	2.6	1.4	1.3	2.5	1.9	1.2
Sr (ppm)	11.5	67.6	35.1	54	25.9	49.2	30.5	36.2	47.5	9	19.1	15	19.9	55.4	21.1
Th (ppm)	10	1.2	24.2	94.2	28	9.6	38.6	19.3	16.4	11	1.7	1.1	15.2	7.2	4.3
U (ppm)	0.42	0.11	0.64	9.73	3.55	0.37	3.06	2.14	0.44	1.18	0.19	0.34	0.41	0.95	0.2
V (ppm)	16	77	88	41	59	47	48	69	60	32	54	59	65	75	30
Y (ppm)	3.88	7.66	17.48	14.4	11.88	7.72	11.97	11.76	19.72	36.78	12.47	11.09	9.82	15.91	5.55
Zn (ppm)	38	69	74	35	64	77	50	67	62	101	49	44	71	86	55
Zr (ppm)	2.6	2.7	6.7	3.2	7	1.8	4.8	7.6	5	4.3	1.2	1.3	3.2	8.4	2.7
Ce (ppm)	53.9	52.6	119.3	191.8	153	103.1	146.9	117.4	188.7	120.1	47.1	69.8	151.4	150.4	49.2
Dy (ppm)	1.24	3.21	5.32	6.17	5.39	3.44	4.86	3.91	6.95	7.83	3.78	4.48	5.31	7.46	1.55
Er (ppm)	0.49	1.25	2.72	2.77	2.31	1.55	2.74	1.69	3.25	4.14	2.01	2.18	2.24	3.6	0.46
Eu (ppm)	0.82	1.5	2.28	2.05	1.78	1.72	1.57	1.61	2.21	1.74	0.91	1.11	1.45	2.7	1.61
Gd (ppm)	2.91	4.89	7.82	9.78	7.53	5.74	8.12	5.5	9.87	10.2	3.65	5.28	7.52	10.35	2.62
Ho (ppm)	0.26	0.56	0.91	1.04	0.97	0.48	1	0.57	1.18	1.43	0.79	0.89	0.89	1.3	0.18
La (ppm)	30.2	27.3	57.2	107.5	79.3	56	78.2	62.1	95.8	58	27.7	38.5	85.5	63.6	28.1
Lu (ppm)	0.13	0.32	0.5	0.49	0.4	0.26	0.61	0.31	0.41	0.37	0.35	0.5	0.5	0.29	0.08
Nd (ppm)	24.2	29.8	56.7	90	68.4	44.7	63.8	45.8	77.2	57.3	23.4	35.2	69.8	73.9	20.1
Pr (ppm)	6.79	6.91	14.48	23.25	18.32	11.65	16.85	12.79	21.15	13.79	5.75	8.75	18.07	19.12	5.2
Sm (ppm)	4.2	6	10.3	14.3	11.9	7.1	10.6	7.4	13.2	11.2	4.6	5.6	11.4	13	3.3
Tb (ppm)	0.24	0.62	0.91	1.06	0.9	0.67	0.81	0.66	1.36	1.5	0.41	0.69	0.77	1.34	0.3
Tm (ppm)	0.05	0.18	0.43	0.3	0.31	0.19	0.35	0.25	0.45	0.47	0.2	0.32	0.21	0.49	0.06
Yb (ppm)	0.2	1.3	2.7	2.2	2.2	1.4	2.5	1.4	2.8	3.3	1.8	2.1	1.7	2.9	0.4
Eu/Eu*	0.54	0.85	0.78	0.53	0.58	0.83	0.52	0.78	0.6	0.5	0.68	0.63	0.48	0.72	1.68
(La/Sm) _n	6.43	2.5	3.05	4.12	3.66	4.33	4.05	4.6	3.98	2.84	3.3	3.77	4.11	2.68	4.67
(La/Yb) _n	250.06	12.73	12.84	29.61	21.85	24.24	18.96	26.88	20.74	10.65	9.33	11.11	30.48	13.29	42.58
(Tb/Yb) _n	5.28	2.03	1.43	2.05	1.74	2.04	1.38	2.01	2.07	1.93	0.97	1.4	1.93	1.97	3.19
Tot REE	756.56	136.44	281.57	452.71	352.71	238	338.91	261.39	424.53	291.37	122.45	175.4	356.76	350.45	113.16
ASI	0.99	0.82	0.78	0.73	0.85	0.95	0.86	0.89	0.91	1.03	0.97	0.93	0.88	0.95	1

Sample	KE-244A	KE-246	KE-282	KE-283	KE-292	KE-294	KE-313	KE-315	KE-317	KE-318	KE-319	KE-199	KE-203	KE-205B	KE-215A
Unit	ST	ST	ST	ST	ST	ST	ST	ST	ST	ST	ST	TB	TB	TB	TB
SiO ₂ (wt%)	65.77	66.12	69.65	70.05	70.51	70.53	71.46	71.7	72.46	74.13	75.43	62.04	62.6	62.74	63.53
Al ₂ O ₃ (wt%)	16.95	16.87	14.32	14.08	14.89	15.83	15.54	14.13	14.45	13.36	14.51	16.22	16.33	15.41	14.93
CaO (wt%)	2.39	3.25	3.59	1.83	1.99	1.81	0.76	2.55	1.38	1.2	0.92	3.29	5.58	3.31	2.96
K ₂ O (wt%)	4.94	3.79	1.89	4.45	3.66	5.38	7.42	1.99	6.19	4.09	6.41	4.65	2.09	5.84	4.88
Na ₂ O (wt%)	4.46	4.19	3.73	3.47	3.9	4.16	3.43	4.24	3.16	4.15	2.98	3.79	3.47	3.13	4.15
FeO (wt%)	2.33	3.47	3.44	2.38	2.37	0.98	0.66	2.3	1.77	1.81	0.93	3.16	3.43	3.65	2.27
Fe ₂ O ₃ (wt%)	3.69	4.78	5.74	2.91	3.31	1.37	0.97	3.01	2.62	2.41	1.38	5.17	6.92	5.54	5.16
MgO (wt%)	0.99	1.2	1.67	0.78	1.05	0.18	0.09	0.96	0.35	0.28	0.16	1.58	3.26	2.73	1.5
MnO (wt%)	0.05	0.11	0.13	0.08	0.06	0.02	0.03	0.07	0.04	0.04	0.02	0.08	0.11	0.09	0.09
TiO ₂ (wt%)	0.5	0.61	0.45	0.63	0.48	0.09	0.05	0.38	0.31	0.17	0.08	1.03	0.73	0.93	1.01
P ₂ O ₅ (wt%)	0.18	0.16	0.11	0.19	0.12	0.03	0.03	0.1	0.09	0.04	0.04	0.38	0.13	0.33	0.34
LOI	0.51	0.22	0.24	0.49	0.18	0.21	0.24	0.19	0.14	0.07	0.1	0.49	0.27	0.4	0.27
Sum	100.42	101.31	101.53	99.19	100.16	99.61	100.04	99.34	101.19	99.95	102.03	98.71	101.5	100.45	98.82
Ba (ppm)	1735	1522	478	1099	1128	1437	823	457	860	589	1326	1828	628	1102	1635
Be (ppm)	2.1	2.8	0.2	1.4	1.7	2.5	1.9	0.7	1.5	1.6	0.9	2.9	1.3	3.4	2
Cs (ppm)	2.12	2.38	2.87	2.05	1.73	1.34	4.5	1.26	1.58	1.98	1.5	1.83	2.12	20.41	2.9
Ga (ppm)	26.4	21.4	12	23.6	23.6	18.3	25.4	17.5	19.4	22.6	9.6	20.7	20.3	24.7	29.4
Hf (ppm)	6.7	4.64	2.19	8.32	4.4	2.42	1.29	3.34	6.51	4.91	2.68	8.31	5.63	8.32	6.7
Nb (ppm)	4.94	12.14	1.87	12.04	3.81	0.54	0.78	6.37	4.91	9.97	0.05	7.28	2.93	16.46	22.54
Rb (ppm)	87.1	118.4	60.8	104.8	84.2	138.3	234.7	48.5	221.1	100.9	121.5	152.3	73	319.4	146.2
Sn (ppm)	1.3	0.6	1.8	0.3	0.3	3.9	0.3	1.1	7.9	2.7	9.5	6.4	2.8	4.3	6.7
Sr (ppm)	604.4	189.6	274.4	230.2	523	311.9	172.9	223.1	105.6	253.1	606.3	284	421	654.7	
Ta (ppm)	0.05	0.53	0.05	0.05	0.05	0.05	0.06	2.74	2.14	0.66	0.05	1.21	0.05	1.89	6.28
Th (ppm)	3.2	8	0.1	8.3	7.1	0.1	5.5	3.3	15.3	6.3	0.1	4.1	6.1	14.7	9.2
U (ppm)	0.76	1.43	0.62	1.03	0.79	0.71	2.8	0.87	2.73	1.42	0.62	1.19	1.24	5.71	1.92
W (ppm)	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.7	0.1	0.1	0.1	0.1	0.1	1.1	0.1
Y (ppm)	5.99	13.83	19.15	18.3	8.04	2.86	14.75	12.69	15.11	53.52	2.2	23.08	21.12	20.74	39.49
Zr (ppm)	273.7	197.7	82.6	431.3	178.3	60.3	26.6	153.8	276.7	143.6	81.9	344.9	211.4	358.4	320.2
Co (ppm)	6.9	8.5	8.4	4.5	7.7	1.4	0.6	6.9	2.6	2.3	1.1	10.3	12.3	15.1	10.3
Cr (ppm)	7	5	15	9	24	13	3	17	5	8	7	18	40	78	13
Cs (ppm)	2.07	2.63	2.98	1.99	1.91	0.57	0.65	1.33	1.14	1.7	0.57	1.67	2.25	21.16	2.4
Cu (ppm)	7.4	20.4	2.9	5.2	6.1	4.1	1.4	3.3	3.4	2.4	3.2	16.8	25.2	31.9	17.6
Ga (ppm)	7.3	7.4	5.9	7	6.6	1.9	1.5	5.3	4	4.6	1.7	9.3	5.6	10.3	7.5
Hf (ppm)	0.09	0.2	0.09	0.07	0.11	0.18	0.05	0.05	0.32	0.05	0.08	0.23	0.15	0.39	0.3
La (ppm)	23.8	51	7.4	101.6	48	5	11	19.7	32.5	17.8	3.1	53.9	22.6	51.9	63.6
Li (ppm)	24	27	32	24	28	7	10	47	14	31	2	30	18	47	29
Nb (ppm)	0.39	1.07	0.39	0.83	0.43	1.08	0.52	0.66	1.93	3.98	0.56	1.06	0.63	1.54	2.41
Ni (ppm)	5.6	2.4	4	3.9	17.1	1.4	1.4	10.7	2.4	3.4	1.8	10.2	21.7	37.8	10
Pb (ppm)	2.2	2.1	2	2.3	3	4	5.6	2.3	1.8	1.2	6.1	8.5	2	7.2	2.4
Rb (ppm)	61.8	115.6	75.6	63.4	73.6	16.4	10.8	51.7	52.8	37.7	19.1	119.8	76.5	281.4	77.4
Sc (ppm)	3.9	8.9	7.8	7.7	4.9	0.4	0.6	4	1.9	6.1	0.8	5.8	4.3	5.9	3.9
Sn (ppm)	1.2	1.2	1.5	1	1.2	2.1	1.6	0.9	2	3	1.3	1.9	1.4	5.3	2.2
Sr (ppm)	21.1	9.6	10.4	15.1	19.1	9.4	5.6	7.8	12.3	4.9	7.1	32.8	13.9	38.8	32.2
Th (ppm)	4.3	14.3	2.9	14.3	10.8	2.5	8.5	4.7	21.4	5.1	2.8	11	12.1	24.1	10.3
U (ppm)	0.2	0.66	0.26	0.42	0.33	0.19	1.92	0.63	0.98	0.31	0.33	0.5	0.73	4.13	0.82
V (ppm)	30	44	61	17	31	11	<1	28	12	6	7	62	77	64	52
Y (ppm)	5.55	9.41	8.54	19.47	8.29	2.21	3.71	7.6	8.99	8.04	1.83	11.37	7.65	14.21	22.48
Zn (ppm)	55	55	53	50	54	11	9	33	28	53	14	77	40	56	66
Zr (ppm)	2.7	5.6	0.6	1.9	3.2	2.7	1	0.7	9.2	0.7	1.6	4.6	1.3	5.1	5.9
Ce (ppm)	49.2	78.6	23.2	187.8	85.9	16.9	22.9	33.5	109.7	59.4	11.6	168.1	58.4	103.7	166.3
Dy (ppm)	1.55	2.58	3.57	4.24	1.78	0.55	2.84	2.39	4.18	9.86	0.37	6.58	4.08	4.39	7.6
Er (ppm)	0.46	1.1	2.71	1.96	0.81	0.32	1.14	1.28	1.7	5.46	0.12	2.72	2.51	2.05	3.96
Eu (ppm)	1.61	1.38	0.79	2.29	1.15	0.32	0.97	0.64	0.76	0.71	0.27	2.39	0.86	1.55	2.53
Gd (ppm)	2.62	3.22	3.73	8.02	3.42	0.71	2.76	2.07	4.87	9.3	0.52	9.26	4.59	6.06	10.23
Ho (ppm)	0.18	0.45	0.83	0.68	0.32	0.1	0.48	0.36	0.78	1.89	0.05	1.07	0.94	0.77	1.39
La (ppm)	28.1	43	15.4	98.9	47.2	11.2	11	18.6	58.4	27.6	8	77.6	32.8	49.1	68.3
Lu (ppm)	0.08	0.19	0.58	0.21	0.05	0.22	0.05	0.1	0.4	0.73	0.05	0.36	0.65	0.32	0.5
Nd (ppm)	20.1	30.1	12.4	69.9	31.8	6.1	9.6	12.6	39.8	32.5	4.6	89.5	26.3	47.2	79.2
Pr (ppm)	5.2	8.33	3.74	19.67	8.82	1.85	2.6	3.48	13.28	7.51	0.98	23.96	6.81	12.19	20.56
Sm (ppm)	3.3	4.8	3.4	10.4	4.8	0.9	2.5	2.5	8.3	8.7	0.7	15.1	5.9	8.6	13.6
Tb (ppm)	0.3	0.5	0.5	1.05	0.45	0.05	0.46	0.36	0.76	1.5	0.05	0.99	0.61	0.79	1.37
Tm (ppm)	0.06	0.16	0.33	0.17	0.13	0.05	0.13	0.23	0.23	0.75	0.05	0.36	0.31	0.23	0.5
Yb (ppm)	0.4	0.9	2.8	1.4	0.7	0.1	1.2	1.3	1.4	4.5	0.1	2.6	2.3	1.7	3.1
Eu/Eu*	1.68	1.08	0.68	0.77	0.87	1.23	1.14	0.87	0.37	0.24	1.38	0.62	0.51	0.66	0.66
(La/Sm) _n	4.67	4.91	2.48	5.22	5.39	6.83	2.41	4.08	3.86	1.74	6.27	2.82	3.05	3.13	2.75
(La/Yb) _n	42.58	28.96	3.33	42.81	40.87	67.88	5.56	8.67	25.28	3.72	48.48	18.09	8.64	17.5	13.35
(Tb/Yb) _n	3.19	2.36	0.76	3.19	2.74	2.13	1.63	1.18	2.31	1.42	2.13	1.62	1.13	1.98	1.88
Tot REE	113.16	175.31	73.98	406.69	187.33	39.37	58.63	79.41	244.56	170.41	27.46	400.59	147.06	238.65	379.14
ASI	1	1	0.97	1	1.06	0.99	1.03	1.03	1	0.99	1.07	0.94	0.9	0.88	0.85

Sample	KE-218	KE-222B	KE-230	KE-232	KE-233B	KE-235	KE-236	KE-237A	KE-258A	KE-261	KE-263A	KE-262	KE-272	KE-273B
Unit	TB	TB	TB	TB	TB	TB	TB	TB	TB	TB	TB	TB	TB	TB
SiO ₂ (wt%)	64.1	64.71	64.94	65.13	65.28	65.64	65.64	65.68	66.27	67.39	67.4	67.4	68.11	68.36
Al ₂ O ₃ (wt%)	15.92	15.87	16.73	14.81	16.81	15.39	15.39	15.66	16.73	16.98	15.58	14.86	13.66	16
CaO (wt%)	2.61	3.66	3.23	2.22	3.04	2.69	2.69	2.36	3.08	2.83	2.74	2.64	1.82	2.1
K ₂ O (wt%)	4.65	1.69	5.12	4.78	6.52	5.05	5.05	5.01	3.98	6.37	3.4	2.97	5.34	5.64
Na ₂ O (wt%)	4.06	4.31	4.54	3.54	4.46	3.61	3.61	3.8	4.68	3.89	4.09	3.87	3.09	4.02
FeO (wt%)	2.41	3.78	2.63	2.7	2.69	2.66	2.66	2.79	2.74	2.44	2.22	2.8	2.07	2.46
Fe ₂ O ₃ (wt%)	3.93	5.46	4.76	4.08	4.7	4.27	4.27	4.53	4.22	3.08	3.43	3.84	3.16	3.24
MgO (wt%)	0.97	1.27	1.64	1.32	1.58	1.29	1.29	1.15	1.2	0.68	0.91	1.35	0.93	1.21
MnO (wt%)	0.07	0.1	0.1	0.05	0.08	0.07	0.07	0.07	0.1	0.04	0.1	0.07	0.05	0.07
TiO ₂ (wt%)	0.63	0.49	0.77	0.82	0.83	0.8	0.8	0.85	0.64	0.76	0.41	0.52	0.46	0.56
P ₂ O ₅ (wt%)	0.18	0.14	0.21	0.27	0.39	0.33	0.33	0.28	0.22	0.19	0.18	0.14	0.15	0.19
LOI	0.07	0.21	0.4	0.67	0.4	0.54	0.54	0.21	0.36	0.68	0.4	0.36	0.31	0.45
Sum	97.2	97.91	102.44	97.69	104.09	99.67	99.67	99.6	101.48	102.89	98.63	98.03	97.08	101.85
Ba (ppm)	1968	438	1001	1033	2027	1342	1342	1343	911	1016	702	1025	841	1154
Be (ppm)	3	1.3	2.5	1.3	2.8	3.4	3.4	1.2	5.5	6.5	4.9	1.5	2.8	4.6
Cs (ppm)	1	1.93	1.4	1.83	2.59	1.07	1.07	1.39	5.04	5.5	5.86	2.87	1.57	3.73
Ga (ppm)	20.3	18.6	23.4	20.2	20.7	23.8	23.8	22.8	23.2	30.7	22.1	25	23.4	25.5
Hf (ppm)	7.01	2.5	4.58	9.69	8.35	6.88	6.88	11.06	4.97	5.14	2.68	3.44	4.75	3.75
Nb (ppm)	8.34	6.44	9.05	4.74	8.01	4.34	4.34	9.98	10.24	19.76	4.53	6.9	9.75	15.94
Rb (ppm)	129.8	85.2	95.9	128.7	167.2	151.8	151.8	172.8	137.3	266.2	101.5	103.5	208.1	158.6
Sn (ppm)	3.9	3.7	2.3	3.6	3.8	3.5	3.5	2.8	1.4	2.5	1.8	0.3	1.7	4.8
Sr (ppm)	735.2	231.9	501.5	435.8	579.8	463.4	463.4	489.5	423.2	684.1	245.2	682.8	383.3	505
Ta (ppm)	0.05	0.07	0.5	0.05	0.71	0.05	0.05	1.43	1.5	2.1	1.6	0.11	0.05	2.81
Th (ppm)	7.5	1.6	8.6	5.1	7.2	17	17	10.8	21.8	10.7	0.5	7.3	25.6	10.1
U (ppm)	1.21	0.89	0.61	0.75	3.87	1.3	1.3	1.53	3.13	5.07	0.96	0.79	2.01	1.57
W (ppm)	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.5	0.2	0.1	0.1	0.1	1.5
Y (ppm)	16.61	21.33	15.13	12.47	18.54	19.66	19.66	30.01	15.27	9.18	9.65	5.67	15.59	21.84
Zr (ppm)	246.4	76.9	237.1	365.2	276.7	284.4	284.4	424.9	305.9	281.1	134.9	133.3	163.4	163.4
Co (ppm)	5.6	7.3	8.9	8.9	8.8	8.4	8.4	7.7	6.8	5.5	6.8	12	6.3	9.1
Cr (ppm)	10	5	20	19	16	17	17	14	14	6	5	27	30	39
Cs (ppm)	0.85	1.84	1.42	1.48	2.66	1.09	1.09	1.28	5.66	5.09	6.22	3.18	1.26	3.7
Cu (ppm)	5	3.7	5.6	2.9	4.9	10.3	10.3	15.2	2.3	5.6	4.9	94.6	4.8	6.1
Ga (ppm)	5.3	6.1	6.7	8.7	8.7	8.4	8.4	7.8	6.6	8.1	5.3	8.5	5.5	7.5
Hf (ppm)	0.16	0.12	0.14	0.11	0.3	0.2	0.2	0.23	0.23	0.2	0.06	0.1	0.24	0.12
La (ppm)	67	6.6	43.2	65.3	60	95.3	95.3	58.2	43.1	64.3	6.1	57.2	69.1	48.3
Li (ppm)	23	29	18	20	28	21	21	30	26	35	48	31	17	34
Nb (ppm)	1.21	0.34	0.55	0.59	0.93	0.87	0.87	1.75	1.14	1.83	0.27	0.58	1.13	1.07
Ni (ppm)	5	1.8	12.9	13.1	9.1	8.9	8.9	7.3	10.2	3.7	3.1	22.5	13.1	20.6
Pb (ppm)	3.9	1.8	1.8	3	6.8	4.9	4.9	4.4	2.6	5.3	1.6	1.8	3.3	2.7
Rb (ppm)	56.1	94.5	73.2	70.6	113.2	91.1	91.1	108.8	112.9	151.9	81.4	112.3	98.3	105.7
Sc (ppm)	2	7.3	4.1	5.6	5.9	6.4	6.4	4.6	4.5	1.7	4.6	2.8	3.5	2.6
Sn (ppm)	1.6	1.3	2.2	1.9	2.1	1.7	1.7	2.3	3.6	2.1	3.4	1.2	2.1	1.7
Sr (ppm)	23.2	8.8	22.6	19.8	30.5	29.5	29.5	28.2	18.9	28.3	10	52.1	31.3	24.7
Th (ppm)	17.2	2.5	13.7	13.3	15.6	28.7	28.7	18.3	40.1	23.1	1.6	13.7	44.3	13.9
U (ppm)	0.58	0.6	0.22	0.21	2.75	0.56	0.56	0.57	2.49	2.78	0.73	0.38	1.09	0.9
V (ppm)	35	42	44	41	51	52	52	48	32	7	29	48	27	31
Y (ppm)	9.81	15.13	11.06	11.23	12.34	10.27	10.27	16.39	15.35	4.9	7.98	6.02	10.72	13.66
Zn (ppm)	48	51	47	57	66	63	63	68	41	81	43	53	33	48
Zr (ppm)	4.2	0.8	4.7	2.4	5.8	4.4	4.4	3.7	8.5	5.3	1.2	3.7	6.9	3
Ce (ppm)	138.7	22.7	85.5	169.8	171.5	275.2	275.2	177.1	69.3	99.3	9.9	87.5	102.6	78
Dy (ppm)	4.06	3.96	3.03	3.47	5.99	4.8	4.8	7.16	2.98	2.19	2.11	1.15	2.92	3.77
Er (ppm)	2.15	2.92	1.59	0.99	2.18	1.92	1.92	3.7	1.53	0.85	0.93	0.43	1.33	2.11
Eu (ppm)	1.49	0.95	1.02	1.4	2.09	1.78	1.78	1.87	0.99	1.23	0.72	0.98	1.01	1.22
Gd (ppm)	5.69	3.37	3.89	6.15	6.38	6.72	6.72	8.96	3.38	4.03	1.96	2.21	3.88	4.1
Ho (ppm)	0.57	0.88	0.51	0.53	0.9	0.87	0.87	1.23	0.52	0.28	0.35	0.19	0.41	0.75
La (ppm)	85.5	13	43.9	89.4	88	152.4	152.4	84.3	36.9	56.9	5.5	47.2	62.3	49.5
Lu (ppm)	0.44	0.64	0.31	0.26	0.32	0.22	0.22	0.68	0.31	0.07	0.07	0.05	0.25	0.33
Nd (ppm)	56.7	13.1	31.5	73.2	82.9	94.1	94.1	84.6	25.8	38.2	5.3	28.6	34.1	28
Pr (ppm)	15.9	2.92	8.79	19.51	21.06	30.1	30.1	22.21	6.95	10.55	1.08	8.39	10.36	8.19
Sm (ppm)	8.4	3.6	5.3	10	10.1	12.7	12.7	13.3	4.8	6.1	1.4	3.6	5.6	4.9
Tb (ppm)	0.63	0.53	0.59	0.51	0.84	0.67	0.67	1.19	0.52	0.49	0.28	0.26	0.51	0.65
Tm (ppm)	0.17	0.4	0.17	0.12	0.28	0.22	0.22	0.46	0.19	0.13	0.13	0.06	0.19	0.27
Yb (ppm)	1.6	2.7	1.3	1	2.2	1.6	1.6	3.3	1.5	0.5	0.9	0.4	1.3	1.8
Eu/Eu*	0.66	0.84	0.69	0.55	0.8	0.59	0.59	0.53	0.76	0.76	1.34	1.07	0.67	0.84
(La/Sm) _n	5.58	1.98	4.54	4.9	4.78	6.58	6.58	3.48	4.22	5.12	2.15	7.19	6.1	5.54
(La/Yb) _n	32.39	2.92	20.47	54.18	24.24	57.73	57.73	15.48	14.91	68.97	3.7	71.52	29.04	16.67
(Tb/Yb) _n	1.68	0.84	1.93	2.17	1.62	1.78	1.78	1.53	1.48	4.17	1.32	2.77	1.67	1.54
Tot REE	322	71.67	187.4	376.34	394.74	583.3	583.3	410.06	155.67	220.82	30.63	181.02	226.76	183.59
ASI	0.97	1.02	0.89	0.99	0.84	0.94	0.94	0.98	0.95	0.92	1.01	1.03	0.96	0.97

LC-Lagoa Caiçara; BO-Boi; ST-Santa Quitéria; TB-Tamboril