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GEOCRONOLOGIA DAS GLACIAÇÕES
CRIOGENIANAS DO BRASIL CENTRAL

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Tese apresentada para obtenção de Título de Livre-Docente junto ao Departamento de Mineralogia e Geotectônica do Instituto de Geociências da Universidade de São Paulo, na área de conhecimento **Geologia Isotópica**.

2011

RESUMO

Mudanças climáticas drásticas ocorreram durante o Neoproterozoico e seu registro pode ser encontrado em vários continentes. De acordo com a literatura, durante o Criogeniano pelo menos dois eventos glaciais podem ter ocorrido, um no Criogeniano Médio (ca. 720 Ma) e outro no Criogeniano Superior (ca. 636 Ma). Neste trabalho serão apresentados dados geocronológicos obtidos em sucessões sedimentares que foram depositadas durante estes dois eventos.

Foram estudados diamictitos da Formação Bebedouro e carbonatos da Formação Sete Lagoas, no cráton do São Francisco, e rochas do Grupo Macaúbas, na Faixa Araçuaí. As idades U-Pb determinadas em cristais de zircão detríticos recuperados de diamictitos que afloram tanto na região cratônica quanto na Faixa Araçuaí, mostram idades máximas de deposição de 874 ± 9 Ma e 900 ± 21 Ma, respectivamente, as quais são similares dentro do erro analítico. Estes dados, associados à idade isocrônica Pb-Pb de 740 ± 22 Ma obtida nos carbonatos de capa da base da Formação Sete Lagoas, sugerem que os depósitos glaciais das formações Bebedouro, Jequitaí, Carrancas, Serra do Catuni e Chapada Acauã, foram depositados durante o evento glacial do Criogeniano Médio, também conhecido como Sturtiano. Além disso, a similaridade das idades máximas de deposição obtidas neste trabalho com aquelas reportadas na literatura para o Grupo Vazante (900-930 Ma) reforçam a possibilidade que os depósitos glaciais que ocorrem na Formação Santo Antônio do Bonito também sejam Sturtianos.

No Cráton Amazônico foram amostrados carbonatos de capa da Formação Mirassol d'Oeste e na Faixa Paraguai Sul foram coletados diamictitos da Formação Puga. Estes diamictitos possuem zircões com idades U-Pb entre 706 Ma e 2100 Ma (apenas 2 cristais arqueanos foram encontrados). As idades variam em função do local amostrado. Diamictitos provenientes do Morro do Puga, área tipo da formação, forneceram zircões com idades entre 970 e 2100 Ma, enquanto que os diamictitos coletados na região da Serra da Bodoquena possuem zircões com idades entre 706 e 1990 Ma, sendo que a principal fonte (mais de 70% dos grãos) tem idade de 1760 Ma. Esses dados permitem sugerir que

as rochas coletadas no Morro do Puga tem sua proveniência ligada às rochas do Cráton Amazônico, enquanto que as da Serra da Bodoquena tiveram suas áreas-fontes no Bloco Rio Apa. A origem da fonte mais jovem, de 700 Ma, ainda é desconhecida. Adicionalmente, os grãos de zircão mais jovens limitam a idade de deposição dos diamictitos da Formação Puga em 706 ± 9 Ma. Carbonatos de capa depositados diretamente sobre os diamictitos da Formação Puga afloram na borda sudoeste do Cráton Amazônico. Estes carbonatos pertencem à Formação Mirassol d'Oeste e forneceram uma idade isocrônica Pb-Pb de 633 ± 25 Ma, a qual é interpretada como a época de deposição. A associação destes dados permite sugerir que os diamictitos da Formação Puga e os carbonatos de capa associados foram depositados durante a Glaciação Marinoana (Criogeniano Superior).

Considerando a tríade das glaciações neoproterozoicas, é importante ressaltar que os principais intervalos de tempo de cada evento glacial são precariamente definidos, principalmente pela carência de rochas adequadas à datação. Se forem considerados os dados mais precisos e com melhor controle estratigráfico disponíveis na literatura, é possível observar que a Glaciação Criogeniana Média (Sturtiana) ocorreu há ca. 716 Ma e que a Glaciação Criogeniana Superior (Marinoana) ocorreu há ca. 636 Ma. Embora os resultados obtidos neste estudo sejam concordantes com estes intervalos, é difícil, com o número restrito de dados disponíveis, sugerir que os eventos glaciais tenham sido de fato globais. Em suma, é óbvia a necessidade de um maior número de dados radiométricos em sucessões glaciogênicas, além de relações estratigráficas claras e melhor definidas, para promover um melhor entendimento sobre o número, duração, sincronia (ou não), extensão, causas e consequências das glaciações neoproterozoicas, principalmente as criogenianas.

ABSTRACT

The Neoproterozoic period witnessed some of the most extreme climatic and biological changes in Earth history. According to the literature, during the Cryogenian at least two glacial events may have occurred, at the Early Cryogenian (ca. 720 Ma) and at the Late Cryogenian (ca. 636 Ma). In this work we will present geochronological data from sedimentary successions which were deposited during these two events.

Diamictites from the Bebedouro Formation and cap carbonates from the Sete Lagoas Formation, São Francisco Craton, as well as rocks from the Macaúbas Group, Araçuaí Fold Belt, were studied. U-Pb SHRIMP ages determined on detrital zircon grains from the diamictites that crop out both on the cratonic area and on the Araçuaí Belt yielded maximum depositional ages of 874 ± 9 Ma and 900 ± 21 Ma, respectively, and they are similar considering the analytical error. These data, associated to the Pb-Pb isochron age of 740 ± 22 Ma determined on cap carbonates from the base of the Sete Lagoas Formation, suggest that the glacial deposits from the Bebedouro, Jequitaí, Carrancas, Serra do Catuni and Chapada Acauã formations were deposited during the Early Cryogenian (Sturtian) Glaciation. Furthermore, the similar maximum depositional ages determined here and those available in the literature for the Vazante Group (900-930 Ma) reinforce the possibility that the glacial deposits from the Santo Antônio do Bonito Formation are also Sturtian.

Cap carbonates from the Mirassol d'Oeste Formation, Amazonian Craton, and diamictites from the Puga Formation, Southern Paraguay Belt, were sampled. Detrital zircons from the Puga Formation yielded U-Pb SHRIMP ages ranging from 706 Ma to 2100 Ma (only 2 Archean crystals were found). The ages vary according to the sample locations. Diamictites from the Puga Hill, the type-section of the Puga Formation, have zircons with ages from 970 e 2100 Ma. Rocks with these ages can be found in the Amazonian Craton, suggesting that this is the most likely source of the sediments. Detrital zircons from the Bodoquena area (about 200 km south of Puga Hill) range from 706 to 1990 Ma. The 1760 Ma source is significantly more important in these samples,

comprising more than 70% of analysed grains, and it could be represented by the adjacent Rio Apa Block. The source of the younger zircons (ca. 700 Ma) is unknown. The youngest zircon was dated at 706 ± 9 Ma, thus constraining the maximum depositional age for the Puga Formation. Cap carbonates deposited directly on top of the diamictites from the Puga Formation crop out at the southwestern margin of the Amazonian Craton. These carbonates belong to the Mirassol d'Oeste Formation and yielded a Pb-Pb isochron age of 633 ± 25 Ma, which is interpreted as its depositional age. The data obtained here allow us to suggest that the diamictites from the Puga Formation and the associated cap carbonates were deposited during the Late Cryogenian (Marinoan) Glaciation.

Considering the triad of the Neoproterozoic glaciations, it is important to mention that the main time interval of each glacial event is poorly constrained, mainly because of the lack of suitable material to date in key sections. If only the most precise data and with good stratigraphic control are considered, it is possible to infer that the Early Cryogenian (Sturtian) Glaciation could have occurred at ca. 716 Ma and the Late Cryogenian (Marinoan) Glaciation at 636 Ma. Although the data obtained in this study are in agreement with these time intervals, due to the restrict geochronological data set it is difficult to suggest that the Cryogenian glacial events were global. In summary, in order to better understand the number, duration, synchronicity, extent, causes and consequences of the Neoproterozoic glaciations, particularly the Cryogenian ones, a larger number of reliable radiometric ages in glacial deposits with good stratigraphic constrain is needed.

AGRADECIMENTOS

Difícil chegar ao fim deste trabalho sem ter que agradecer a muitas pessoas e sem esquecer de ninguém!

Inicialmente eu quero agradecer aos meus colegas e amigos docentes, alunos, técnicos de laboratório e funcionários do Instituto de Geociências da USP, em especial àqueles do Centro de Pesquisas Geocronológicas. Todos, sem exceção, contribuíram para a conclusão deste trabalho e para meu crescimento pessoal e profissional.

Preciso também agradecer aos meus amigos, colegas e alunos da USP, UFMG, UFMT, UFOP, UFPE e UnB, que me acompanharam na busca dos registros das glaciações neoproterozoicas, que me ajudaram a achá-los e a descobrir seus segredos e encantos. Esta procura foi a melhor parte do trabalho!

Os resultados analíticos apresentados nesta tese foram obtidos com o auxílio de várias pessoas e instituições, incluindo a FAPESP, CNPq, CAPES, e colegas dos laboratórios da USP, da China e da Austrália.

Por fim, quero agradecer minha família e meus amigos que nunca deixaram de me dar a mão, o braço, o ombro... e que sempre tiveram uma palavra para me motivar e continuar na luta!

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CAPÍTULO 1

INTRODUÇÃO

O Neoproterozoico é caracterizado por importantes variações climáticas consideradas como cruciais para preparar a biosfera para as “explosões evolutivas” de vida metazoária no Ediacarano e início do Cambriano. Os mais impressionantes registros destes eventos compreendem sucessões glaciais encontradas em praticamente todos os escudos precambrianos. Pelo menos três eventos glaciais têm sido reconhecidos na literatura, sendo que dois destes foram seguidos, quase que imediatamente, pela deposição de capas carbonáticas (*cap carbonates*), interpretadas por vários autores como globalmente contemporâneas, apesar dos escassos dados geocronológicos disponíveis (Hoffman et al., 1998, Kennedy et al., 1998). Os três eventos globais de glaciação no Neoproterozoico (Bowring et al., 2003; Halverson, 2006; Hoffman & Li, 2009; Kendall et al., 2004, 2009) compreendem: o Sturtiano (ca. 720 Ma) e o Marinoano (ca. 636 Ma), de abrangência global, e o Gaskierano (582 Ma), aparentemente mais restrito. Recentemente, Hoffman (2009) propôs que glaciações mais antigas que 726 Ma podem representar uma glaciação distinta, pré-Sturtiana, também conhecida como Glaciação Kaigas (Frimmel et al., 1996). No entanto, as causas, consequências, extensão, duração e a sincronia das glaciações, ainda são controversas, principalmente pela carência de dados geocronológicos em muitas sucessões ao longo do mundo (Kaufman et al., 1997; Kennedy et al., 1998; Hoffman & Schrag, 2002; Jiang et al., 2003; Hoffman, 2009; Xu et al. 2009).

As mudanças climáticas extremas (Hoffman et al., 1998; Hoffman & Schrag, 2002) que ocorreram durante o Criogeniano (850 – 635 Ma) resultaram em fortes variações na

dinâmica e na química dos oceanos, causando anomalias isotópicas que foram registradas nos *cap carbonates* que recobrem os depósitos glaciais em diversos continentes. Estas observações têm permitido o uso de isótopos de C, O e Sr para identificar carbonatos de capa que se formaram imediatamente após os períodos glaciais (Hoffman et al., 1998; Kaufman et al., 1997; Jacobsen & Kaufman, 1999; Halverson et al., 2007, Melezhik et al., 2001; 2009). Recentemente, isótopos de boro, cálcio (Kasemann et al., 2005, 2010; Silva-Tamayo et al., 2010a, b), enxofre (Canfield & Teske, 1996; Fike et al., 2006), ferro (Canfield et al., 2007, 2008; Shen et al., 2008; Halverson et al., 2011) e molibdênio (Moura et al., 2010; Wen et al., 2011) também têm sido empregados visando a identificação destas variações. A utilização destas ferramentas tornou-se indispensável para correlacionar sequências sedimentares em nível global. Contudo, apesar de um crescente número de dados isotópicos obtidos nestes carbonatos, a carência de dados geocronológicos tem dificultado sua correlação global (Kendall et al., 2009), já que carbonatos de idades distintas podem apresentar padrões isotópicos similares (Condon et al., 2005; Kaufman, 2005) resultando em correlações errôneas. Assim, a obtenção de idades absolutas de deposição, tanto nos diamictitos glaciais como nos carbonatos de capa que ocorrem ao redor do globo, é necessária para constatar se estes eventos glaciais foram sincrônicos ou não.

Em função disto, a geocronologia precisa dos depósitos neoproterozoicos representa uma prioridade para que correlações regionais e globais possam ser feitas com maior certeza. Além disto, a geocronologia é uma ferramenta-chave para distinguir entre as diferentes hipóteses propostas para explicar os depósitos glaciais de baixa latitude, entre elas a *Snowball Earth* (Kirschvink, 1992; Hoffman et al., 1998; Hoffman & Schrag, 2002), a *Slushball Earth* (Hyde et al, 2000; Schrag & Hoffman, 2001; Peltier et al., 2007) e a *Pan-glacial* (Hoffman, 2009) que defendem glaciações globais sincrônicas, e a *Zipper-rift*, que sugere que as glaciações foram diacrônicas em função do soerguimento das margens dos rifts gerados pela quebra do supercontinente Rodínia (Eyles & Januszcak, 2004).

1.1. OBJETIVOS

O principal objetivo desta tese é apresentar dados geocronológicos obtidos em diamictitos e carbonatos de capa depositados durante os eventos glaciais criogenianos (850 – 635 Ma) registrados nos crátons do São Francisco e Amazônico e nas faixas Araçuaí e Paraguai, que margeiam os respectivos crátons. As idades foram obtidas (i) pelo método U-Pb SHRIMP em zircões detríticos recuperados dos diamictitos, as quais podem estabelecer a idade máxima de sedimentação e (ii) por isócronas Pb-Pb em carbonatos de capa, que podem fornecer a idade de deposição destas rochas.

As idades obtidas podem auxiliar no estabelecimento dos intervalos de idade nos quais as glaciações ocorreram e contribuir para a discussão sobre a sincronia ou não dos grandes eventos climáticos criogenianos.

1.2. ÁREAS DE ESTUDO

Para este trabalho foram selecionadas quatro áreas de estudo (Figs. 1.1 e 1.2), a saber:

1. Cráton do São Francisco, onde foram amostrados diamictitos da Formação Bebedouro e carbonatos de capa da Formação Sete Lagoas, base do Grupo Bambuí;
2. Faixa Araçuaí, marginal ao Cráton do São Francisco, onde ocorrem os diamictitos das formações Serra do Catuni e Chapada Acauã Inferior, Grupo Macaúbas, os quais têm sido correlacionados com os diamictitos das formações Jequitaí, Bebedouro e Carrancas, depositados sobre o Cráton do São Francisco;
3. Faixa Paraguai Sul, onde afloram os diamictitos da Formação Puga;
4. Craton Amazônico, onde foram amostrados os carbonatos de capa da Formação Mirassol d'Oeste, base do Grupo Araras, e que capeiam os diamictitos da Formação Puga.

1.3. ESTRUTURAÇÃO DA TESE

A tese é constituída por 6 capítulos. O Capítulo 1 contém a introdução da tese e no Capítulo 2 são apresentadas as hipóteses sobre as glaciações neoproterozoicas e os dados geocronológicos disponíveis da literatura sobre as glaciações criogenianas. O Capítulo 3 é constituído por dois artigos (capítulos de livro) que permitem inserir as áreas de estudo dentro de um contexto geológico mais amplo, ou seja, no Paleocontinente São Francisco (Sial et al., 2009) e Paleocontinente Amazônico (Alvarenga et al., 2009). Três artigos compõem o Capítulo 4, onde são apresentados (i) os dados geocronológicos U-Pb obtidos em zircões detríticos de diamictitos da Formação Bebedouro, Cráton do São Francisco (Figueiredo et al., 2009); (ii) idades U-Pb de zircões detríticos do Grupo Macaúbas que ocorrem na Faixa Araçuaí, marginal ao Cráton do São Francisco (Babinski et al., no prelo); e (iii) datações Pb-Pb de carbonatos de capa da Formação Sete Lagoas, Cráton do São Francisco (Babinski et al., 2007). O Capítulo 5 apresenta os dados geocronológicos U-Pb obtidos em zircões detríticos dos diamictitos da Formação Puga, na Faixa Paraguai Sul (Babinski et al., submetido) e a datação do carbonato de capa da Formação Mirassol d'Oeste, no Cráton Amazônico. As conclusões deste estudo são sumarizadas no Capítulo 6.

Figura 1.1. Mapa geológico do Cráton do São Francisco e das faixas Araçuaí, a leste, e Brasília, a oeste, com a localização (estrelas) das áreas de estudo (extraído de Santos et al., 2004).

Figure 1.2. Mapa geológico do sudoeste do Cráton Amazônico e da Faixa Paraguai Norte e Sul, com a localização (estrelas) das áreas de estudo (extraído de Alvarenga et al., 2009).

CAPÍTULO 2

AS GLACIAÇÕES CRIOGENIANAS

Embora não seja o objetivo desta tese discutir as causas das glaciações criogenianas (850 - 635 Ma), serão apresentadas a seguir as principais hipóteses que tentam explicar as drásticas mudanças climáticas ocorridas neste período. Adicionalmente, serão apresentados os dados geocronológicos disponíveis na literatura para os principais registros geológicos das glaciações ao redor do globo e os principais intervalos de idade que restringem as épocas glaciais criogenianas.

2.1. AS HIPÓTESES

Registros geológicos mostrando que a Terra passou por mudanças climáticas importantes durante sua história são antigos e conhecidos (Harland, 1964). Contudo, até o final da década de 1990, poucos cientistas tinham se empenhado em combinar evidências geológicas, sedimentológicas, geoquímicas, paleontológicas e paleomagnéticas para propor hipóteses que explicassem as profundas e expressivas mudanças climáticas que ocorreram no Neoproterozoico.

Algumas hipóteses que surgiram na década de 1970 foram revitalizadas a partir de 1998 quando um grupo coordenado por Paul Hoffman propôs a hipótese da “Snowball Earth” (Hoffman et al., 1998). As hipóteses que tentam explicar as causas das variações climáticas neoproterozoicas são: *Snowball Earth* (Hoffman et al., 1998), *High Obliquity*

(Williams, 1975, 1998, 2000), *Slushball Earth* (Hyde et al., 2000), *Zipper Rift* (Eyles & Januszczak, 2004) e *Pan-Glacial* (Hoffman, 2009).

A hipótese ***Snowball Earth*** (Hoffman et al., 1998) fundamenta-se nas evidências paleomagnéticas observadas por Kirschvink (1992) que mostravam que os diamictitos glaciais da Formação Elatina (Austrália) foram depositados próximos ao paleoequador (15° de latitude). Esta descoberta fez com que Kirschvink (1992) sugerisse uma concentração de massas continentais em baixas e médias latitudes, o que resultaria num albedo bastante alto nos trópicos, diminuindo a temperatura média global. Aliado a isto, o aumento da taxa de intemperismo dos minerais silicáticos resultaria numa diminuição da pressão de CO₂ atmosférico, induzindo uma queda da temperatura. Estes fatores teriam sido responsáveis pelo início das glaciações neoproterozoicas, onde o mar estaria coberto por uma camada de gelo; esta hipótese explicaria os depósitos glaciais de baixa latitude e a ocorrência dos depósitos de ferro (*Banded Iron Formations*), já que os oceanos isolados pela camada de gelo se tornariam anóxicos e ricos em ferro dissolvido (Kirschvink, 1992). Estudos mais sistemáticos realizados pelo grupo de Paul Hoffman nas rochas da Namíbia, incluindo dados sedimentológicos, geoquímicos, isotópicos e paleomagnéticos, resultaram na reedição da hipótese *Snowball Earth* (Hoffman et al., 1998), que defende uma glaciação global, onde os oceanos estariam isolados da atmosfera por extensas capas de neve com centenas de metros de espessura. De acordo com Hoffman et al. (1998), durante a glaciação os vulcões continuariam emitindo CO₂, que não seria consumido já que o intemperismo e a fotossíntese não estariam ocorrendo. Este acúmulo de CO₂ na atmosfera, após alguns milhões de anos, resultaria em uma concentração muito elevada que criaria um efeito *greenhouse*, promovendo um aumento significativo da temperatura, iniciando o degelo das capas de neve e, conseqüentemente, a diminuição do albedo e o recuo das calotas de gelo. Dados de reversões magnéticas nos depósitos glaciais (Solh et al., 1999), modelos de subsidência termal (Hoffman et al., 1998) e anomalias de irídio nos carbonatos de capa (Bodiselitsch et al., 2005) indicam que a duração do evento glacial pode ser de 3 a 12 milhões de anos. O tempo envolvido na inversão da condição *icehouse-greenhouse* ainda é discutível, variando de algumas dezenas de anos (Hoffman et al., 1998) a centenas de milhares de anos (Trindade et al., 2003, Font et al., 2010). O resultado desta inversão seria a ocorrência associada de diamictitos glaciais e carbonatos de capa, interpretados

como resultado do aumento de alcalinidade dos oceanos, provocada pelo intenso intemperismo pós-glacial (Hoffman et al., 1998). Adicionalmente, estes carbonatos de capa apresentam valores de $\delta^{13}\text{C}$ negativos, explicados pela inibição dos organismos fotossintetizantes, principais responsáveis pelo consumo de ^{12}C e fracionamento dos isótopos de C (Kaufman & Knoll, 1995). Um ponto crítico desta teoria é que ela defende que a vida praticamente teria cessado durante os eventos glaciais, o que tem sido questionado por vários pesquisadores (e.g., Corsetti et al., 2006; Corsetti & Lorentz, 2006).

Recentemente, Hoffman (2009) descreve que modelos de energia radiativa revelaram que a Terra poderia existir em três estados climáticos discretos: “não-glacial” (continentes não seriam cobertos por gelo), “glacial – interglacial” (lençóis de gelo só ocorreriam em altas latitudes) ou “pan-glacial” (lençóis de gelo ocorreriam em todas as latitudes). Levando isto em consideração, ele propôs para o Neoproterozoico a hipótese do ***Pan-glacial***, onde lençóis de gelo poderiam ocorrer na terra, em todas as latitudes. Na realidade, esta hipótese é muito similar à *Snowball Earth*, mas nesta não há um questionamento profundo sobre os mecanismos de entrada e saída das glaciações.

Uma hipótese alternativa à *Snowball Earth* (Hoffman et al., 1998) é a ***Slushball Earth*** (Hyde et al., 2000), inicialmente formulada a partir de resultados de simulações climáticas computacionais acoplando modelos de circulação global, clima, capas de gelo e paleogeografia. O modelo sugere que poderia haver uma glaciação de baixa latitude, mas que os oceanos ao redor do equador não seriam inteiramente cobertos por capas de gelo. A conexão entre oceano e atmosfera nessas áreas descobertas permitiria a vida de metazoários e organismos fotossintetizantes (Hyde et al., 2000). A favor desta hipótese estão as evidências geológicas mostrando que o ciclo hidrológico não foi totalmente bloqueado durante as glaciações (Arnaud & Eyles, 2002; Condon et al., 2002, Rieu et al., 2007, entre outros), além da alternância de camadas de diamictitos ricas em fragmentos angulosos com camadas destituídas de fragmentos, que sugerem uma oscilação da camada de gelo em um oceano parcialmente aberto (Condon et al., 2002). Por outro lado, dois argumentos questionam a hipótese *Slushball Earth*: (i) a duração dos eventos glaciais, já que somente uma terra completamente coberta de gelo poderia gerar glaciações de longa

duração; e (ii) as anomalias negativas de $\delta^{13}\text{C}$ observadas nos carbonatos de capa, não podem ser explicadas pela teoria *Slushball Earth*.

A hipótese de ***High Obliquity***, ou alta obliquidade, foi criada por George Williams (1975, 1998, 2000), e sugere que as glaciações de baixa latitude são causadas quando a obliquidade orbital (ângulo entre o eixo de rotação e o eixo da eclíptica) for maior que 54° , resultando em temperaturas médias anuais mais baixas em torno do equador do que nos polos. Estudos comparativos de estromatólitos heliotrópicos atuais e antigos indicam que a obliquidade da Terra há 850 Ma era diferente da atual (Vanyo & Awramik, 1982). Isto geraria forte sazonalidade e intensos ventos superficiais que, segundo Williams, explicariam as estruturas de polígonos em cunhas de areia periglaciais encontradas na Formação Elatina, sul da Austrália. O modelo sugere uma grande oscilação climática próximo ao nível do mar (temperatura do ar média variando de -35°C no inverno a menos de 4°C no verão; Williams & Schmidt, 2004). Simulações feitas com modelos atmosféricos de circulação global e reconstruções paleogeográficas para o Neoproterozoico (Donnadieu et al., 2002) mostraram que uma Terra com alta obliquidade não seria capaz de atuar como gatilho para uma glaciação. Adicionalmente, Evans (2006) com base em dados paleomagnéticos mostra que os registros de deposição carbonática e evaporítica estão sempre situados em zonas subtropicais e equatoriais úmidas, o que se contrapõe à hipótese da alta obliquidade.

Um modelo alternativo para as glaciações neoproterozoicas foi proposto por Eyles & Januszczyk (2004) e denominado ***Zipper-Rift***, em alusão aos rifteamentos diacrônicos resultantes da quebra do supercontinente Rodínia, que ocorreu há 750 - 610 Ma atrás. As estruturas geradas teriam um forte controle sobre a ocorrência de geleiras nas ombreiras soerguidas dos rifts, além de preservar os depósitos glaciais. De acordo com os autores, a associação de tilitos e carbonatos seria gerada durante as três fases que constituem a evolução de um rift: fase sin-rift, fase de subsidência térmica e fase pós-rift. Além disso, os autores sugeriram que os carbonatos de capa não fazem parte dos ciclos glaciais, sob o argumento de que estes depósitos estão separados dos depósitos glaciais por sedimentos carbonáticos retrabalhados ou por hiatos temporais significativos. Neste modelo não é discutida a influência das glaciações na explosão de vida cambriana.

2.2. AS GLACIAÇÕES NEOPROTEROZOICAS E AS CORRELAÇÕES GLOBAIS: O QUE AINDA ESTÁ POR FAZER?

Correlações regionais e globais das sucessões neoproterozoicas têm sido dificultadas pela escassez de dados geocronológicos disponíveis e pela quase ausência de rochas adequadas para fornecer idades radiométricas precisas (idades U-Pb em zircão de rochas ígneas intercaladas nos sedimentos). Adicionalmente, fósseis são raros nas rochas desta idade. Consequentemente, os esforços tem sido direcionados para o estabelecimento de uma cronoestratigrafia baseada em quimioestratigrafia isotópica ($\delta^{13}\text{C}$ e $^{87}\text{Sr}/^{86}\text{Sr}$) de carbonatos e na identificação de características litológicas e geoquímicas que permitam distinguir os carbonatos que capeiam os diamictitos glaciogênicos nos diferentes eventos glaciais (Kaufman & Knoll, 1995; Kaufman et al., 1997; Kennedy et al., 1998; Jacobson & Kaufman, 1999; Walter et al., 2000; Halverson et al., 2005, 2007; Halverson, 2006, Hoffman et al., 2007).

As correlações baseadas em quimioestratigrafia e litoestratigrafia comumente assumem glaciações globais (*Snowball Earth*) ou quase-globais (*Slushball Earth*) há 720 Ma (Glaciação Sturtiana) e 636 Ma (Glaciação Marinoana), e outra de extensão ainda incerta há 580 Ma (Glaciação Gaskierana). Contudo, o número, idade, extensão e a duração das Glaciações Neoproterozoicas ainda constituem um assunto de grande debate (Fairchild & Kennedy, 2007, Hoffman, 2009). Assim, a hipótese de três eventos glaciais pode estar incorreta (Kendall et al., 2009). Recentemente, Hoffman (2009) sugeriu que uma glaciação pré-Sturtiana pode ter existido e nela inclui aqueles depósitos glaciais com idades mais antigas que 726 Ma, anteriormente denominada Glaciação Kaigas (Frimmel et al., 1996).

O problema das correlações globais dos depósitos neoproterozoicos feitas com base em quimioestratigrafia ou por comparação entre os registros, sem o suporte de datações radiométricas, foi apontado por Kendall et al. (2006) que demonstraram, através de idades Re-Os em folhelhos negros, que rochas da área-tipo da Glaciação Sturtiana no sul da Austrália são significativamente mais jovens que os diamictitos glaciogênicos de ca. 700 Ma em outros continentes e que foram previamente sugeridos como correlativos.

Químioestratigrafia isotópica de carbono, enxofre e estrôncio é uma ferramenta de grande importância para as correlações, mas é preciso lembrar que ela fornece uma “datação relativa”, que não pode ser interpretada como uma informação cronológica absoluta. Assim, estes dados requerem uma calibração numérica através de dados independentes, sejam eles radiométricos ou bioestratigráficos (Meert, 2007).

A resolução do registro dos isótopos de enxofre para o Neoproterozoico é ainda muito baixa e também exige maior detalhamento, embora tenha mostrado bons resultados em algumas bacias (Fike et al., 2006, Canfield et al., 2007). De qualquer forma, estes isótopos funcionam melhor como marcadores da oxidação dos oceanos e podem refletir variações locais nas composições isotópicas dependendo da posição da seção estudada dentro na bacia.

Os isótopos de carbono (C) foram os pioneiros em estudos químioestratigráficos para correlações de sucessões neoproterozoicas, tanto em escala bacinal quanto global. Eles foram empregados para observar o comportamento do ciclo exógeno do C e a sua relação com variações climáticas extremas e mudanças na biosfera (Derry et al, 1992, Kaufman et al., 1997). A base da químioestratigrafia isotópica do C baseia-se na premissa que a composição isotópica de C (representada por $\delta^{13}\text{C}$) de um carbonato marinho precipitado em equilíbrio com a água do mar se aproxima bastante da composição isotópica do C inorgânico total dissolvido no oceano (Hayes et al., 1999). Além disso, é assumido que os carbonatos retêm sua assinatura primária após o soterramento. Se estas premissas forem respeitadas, os valores de $\delta^{13}\text{C}$ podem ser usados para gerar curvas de variação de C na água do mar. Assim, muitas seções litoestratigráficas foram estudadas onde foram observadas incursões negativas de $\delta^{13}\text{C}$ em carbonatos depositados imediatamente após os períodos glaciais, possivelmente resultantes de uma diminuição da atividade orgânica, e que os valores de $\delta^{13}\text{C}$ se tornavam positivos à medida que a temperatura aumentava. A geração de grande número de dados isotópicos de C entre 1990 e 2000 em seções que não dispunham de idades absolutas resultou em várias curvas de evolução de C criando problemas para as correlações globais, principalmente porque os perfis de C estavam associados a diferentes glaciações (3 Sturtianas e 2 Marinoanas), cujas idades

Figura 2.1. Curvas de referência de $\delta^{13}\text{C}$ obtidas em várias sucessões litoestratigráficas neoproterozoicas e sumarizadas por Melezhik et al. (2001). As curvas de referência representam o intervalo de tempo (a) 850 - 500 Ma, (b) 650 - 500 Ma (todos os dados disponíveis), e (c) 650 - 500 Ma, apenas com os dados obtidos após 1996. Setas em azul indicam a posição das idades das três possíveis glaciações Sturtianas (S1, S2 e S3) e das duas glaciações Varangerianas (atualmente conhecidas como Marinoanas) (V1 e V2) mencionadas por Jacobsen & Kaufman (1999). Setas rosas indicam a posição das idades da glaciações Varangerianas (V1 e V2) usadas por Pelechaty (1998). Setas vermelhas indicam a posição das idades de duas glaciações Sturtianas (S1 e S2) e das duas glaciações Varangerianas (V1 e V2) citadas por Brasier et al. (2000). Setas verdes indicam a posição das idades das glaciações Sturtiana - Rapitan (S) e Marinoana - Varangeriana (V) mencionadas por Walter et al. (2000).

Figura 2.2. Variações isotópicas de C ($\delta^{13}\text{C}$) para carbonatos do Neoproterozoico propostas por Halverson et al. (2007).

eram discutíveis. Melezhik et al. (2001) apresentaram o problema e demonstraram que o uso destas curvas de C (Fig. 2.1), de forma isolada, não poderia ser aceito como parâmetro confiável para as correlações globais, sugerindo que os isótopos de Sr poderiam ser uma ferramenta mais robusta, principalmente se associada aos dados de C. A mesma crítica foi recentemente reiterada por Frimmel (2010). Uma curva de evolução isotópica de C elaborada com dados obtidos em seções estratigráficas de Svalbard, Canadá, Omã e Namíbia (Fig. 2.2), estudadas em detalhe, foi elaborada por Halverson et al. (2007), onde é possível observar uma diminuição substancial das inconsistências levantadas por Melezhik et al. (2001).

O estabelecimento das curvas de evolução isotópica de Sr na água do mar ao longo do tempo baseia-se no princípio que o Sr tem um tempo de residência suficientemente grande na água do mar (ca. 4 Ma) em relação ao tempo de homogeneização inter-oceânica (1500 anos) (Broecker & Peng, 1982). Isto permite que os carbonatos precipitados em

ambiente marinho assumam a composição isotópica de Sr da água do mar da época de sua deposição, permitindo que suas razões $^{87}\text{Sr}/^{86}\text{Sr}$ representem um excelente registro dos eventos ocorridos em oceanos conectados, como as glaciações criogenianas, por exemplo. Contudo, para que este registro seja confiável, as amostras precisam ser selecionadas e somente aquelas que não sofreram processos pós-deposicionais podem ser utilizadas em estudos quimioestratigráficos. Alguns critérios de seleção que podem indicar composições isotópicas de Sr primárias incluem altos teores de Sr (> 1000 ppm), razões $\text{Rb}/\text{Sr} < 0,001$, $\text{Mn}/\text{Sr} < 0,5$, $\text{Fe}/\text{Sr} < 3$ e $\text{Sr}/\text{Ca} < 0,001$, além de valores de $\delta^{18}\text{O} > -10$ ‰ (Jacobsen & Kaufman, 1999; Folling & Frimmel, 2002).

Entre 1990 e 2000, um grande número de dados isotópicos de Sr foram disponibilizados na literatura. Contudo, a falta de um controle geoquímico e geocronológico adequado resultou em inúmeras curvas de evolução que, da mesma forma que ocorreu com os dados isotópicos de C, dificultaram o seu uso em correlações globais. Estes dados foram sumarizados e discutidos por Melezhik et al. (2001) e apresentados na Figura 2.3.

Figura 2.3. Curvas de variação isotópica de Sr em carbonatos neoproterozoicos sumarizadas por Melezhik et al. (2001).

Figura 2.4. Registro da variação isotópica de Sr da água do mar ao longo do Neoproterozoico (símbolos coloridos), sobreposta aos dados isotópicos de C (círculos de cor cinza). Figura extraída de Halverson et al. (2007).

Uma nova curva de evolução isotópica de Sr para a água do mar neoproterozoico, suportada por dados isotópicos de C, foi proposta por Halverson et al. (2007). A curva foi elaborada com base em dados obtidos em sucessões sedimentares ricas em rochas carbonáticas do noroeste do Canadá, Svalbard e norte da Namíbia e complementada com dados de rochas ediacaranas (635-542 Ma) disponíveis da literatura (Fig. 2.4). De acordo com os autores, é possível observar que as razões isotópicas de tornam mais radiogênicas ao longo do Neoproterozoico, partindo de razões $^{87}\text{Sr}/^{86}\text{Sr}$ de 0,7055 no início da era e chegando a razões $^{87}\text{Sr}/^{86}\text{Sr}$ mais elevadas que 0,7085 no limite com o Cambriano, como resposta à quebra do continente Rodínia.

Embora houve um significativo avanço nos últimos anos permitindo a definição de grandes variações isotópicas na água do mar ao longo do Neoproterozoico, as quais podem ser vinculadas aos principais eventos biogeoquímicos e climáticos desta era, a falta de um

controle geocronológico ainda permite que várias críticas sejam feitas quando à possibilidade destas curvas serem usadas, sem discriminação, para correlações globais. Uma crítica à curva proposta por Halverson et al. (2007) é a pequena quantidade de dados isotópicos de Sr utilizados na sua elaboração, o que compromete sua representatividade. Adicionalmente, dados disponíveis na literatura mostram que algumas seções estratigráficas de mesma idade possuem dados isotópicos de carbono distintos e/ou variáveis em função de alteração diagenética, a existência de seções estratigráficas incompletas e variações isotópicas em função das fácies sedimentares estudadas (Melezhik et al., 2001; Fairchild & Kennedy, 2007; Jiang et al., 2007; Giddings & Wallace, 2009). A quimioestratigrafia isotópica de Sr pode ser mais promissora para correlação dos depósitos criogenianos pelo fato das razões isotópicas $^{87}\text{Sr}/^{86}\text{Sr}$ tenderem a ser mais radiogênicas ao longo deste intervalo de tempo, mas essa ferramenta ainda necessita de um maior número de dados para seu aprimoramento (Halverson et al., 2007; Melezhik et al., 2009). Nota-se que composições isotópicas de $^{87}\text{Sr}/^{86}\text{Sr}$ muito mais radiogênicas do que o esperado têm sido encontradas com frequência para os respectivos intervalos de idade (Alvarenga et al., 2004, 2008; Babinski et al., 2007; Figueiredo, 2006, 2010; Kuchenbecker, 2011; Nogueira et al., 2007; Misi et al., 2007; Paula-Santos, 2010; Sawaki et al., 2010, entre outros), e a possível causa estaria vinculada à entrada de água continental durante o degelo (Ohno et al., 2008; Zhao et al., 2009) ou às alterações pós-deposicionais (Sawaki et al., 2010). Desta forma, um controle muito rigoroso deve ser aplicado na seleção de amostras para os estudos isotópicos de Sr.

Em suma, a obtenção de idades radiométricas precisas e acuradas é imprescindível para determinar se as glaciações criogenianas são sincrônicas ou não e para permitir que correlações globais de sucessões sedimentares possam ser feitas com maior confiabilidade, através de curvas de variação isotópica de C e Sr estabelecidas com rochas de idades conhecidas.

2.3. GEOCRONOLOGIA DAS GLACIAÇÕES CRIOGENIANAS: IDADES MÁXIMAS E MÍNIMAS

Como mencionado anteriormente, a determinação de idades que reflitam fielmente a época das glaciações é dificultada pela falta de rochas favoráveis, representadas por níveis vulcânicos intercalados aos diamictitos. Estes níveis têm se mostrado muito raros e uma possível alternativa é a determinação de idades de carbonatos de capa depositados imediatamente após os sedimentos glaciais. Pelo fato da mudança do estado *icehouse-greenhouse* ter sido muito rápida, as idades obtidas nos carbonatos de capa representam idades máximas destes eventos glaciais.

A literatura mostra que existem mais de 70 formações geológicas, ao redor do mundo, que guardam registros de eventos glaciais neoproterozoicos (Fig. 2.5). Várias possuem idades radiométricas, embora a precisão e acurácia sejam variáveis em função do método geocronológico utilizado e o tipo de material empregado na datação. Muitas delas não tem informações sobre a provável idade de deposição e são assumidas como de determinada idade apenas por correlação. Na Tabela 2.1 é apresentado um panorama atualizado das idades disponíveis para formações que apresentam algum controle estratigráfico.

A seguir serão discutidos os dados geocronológicos disponíveis para o Criogeniano com base nas referências da Tabela 2.1 (dados compilados por Xu et al., 2009 e complementado com resultados mais recentes). Serão utilizados os termos Sturtiano (ca. 720 Ma) e Marinoano (ca. 636 Ma) para distinguir os dois grandes eventos glaciais do Criogeniano, como sugerido por Halverson (2006) e seguido por Hoffman (2009).

Figura 2.5. Mapas paleogeográficos (Li et al., 2008) para (A) 635 Ma e (B) 715 Ma, mostrando a distribuição das formações glaciogênicas (estrelas) Marinoanas e Sturtianas, respectivamente. Estrelas são coloridas em função da sucessão pré-glacial: azul = carbonato, verde = mistura de carbonato e siliciclásticas, e branco = vulcânicas ou quando existe um hiato abaixo da formação glaciogênica. Estrelas com bordas pretas representam depósitos que contêm cunhas poligonais de areia e as com contorno vermelho contêm depósitos sedimentares de Fe ou Fe-Mn. Figura extraída de Hoffman & Li (2009).

Tabela 2.1. Compilação de idades radiométricas consideradas relevantes para as glaciações criogenianas.

Localização	Idade (Ma)	Erro	Descrição e estratigrafia	Referências
<i>Idade máxima glaciação Kaigas e equivalentes</i>				
África	765	± 5	U-Pb SHRIMP; Idade máxima estrato glacial de Kundelungu	Key et al. (2001)
África	763	± 6	U-Pb SHRIMP; Idade máxima estrato glacial de Kundelungu	Key et al. (2001)
África	~754	N.I.	U-Pb TIMS; Membro Vulcânico Spitskop (Grupo Stinkfontein)	Hoffmann et al. (2006)
<i>Idade mínima glaciação Kaigas e equivalentes</i>				
África	752	± 6	U-Pb SHRIMP em vulcânicas félsicas sobre diamictito do Grupo Port Nolloth, S Namíbia	Borg et al. (2003)
África	741	± 6	Pb-Pb TIMS no Craton Kalahari; sobre diamictito Kaigas	Frimmel et al. (1996)
África	735	± 5	U-Pb SHRIMP Supergrupo Katanga, Zambia; sobre diamictito Kaigas	Key et al. (2001)
África	~754	N.I.	U-Pb TIMS no Rosh Pina, Supergrupo Hilda, sob Grupo Porth Nolloth	Hoffmann et al. (2006)
<i>Idade máxima para a glaciação Sturtiana</i>				
Laurentia	777	+3; -2	U-Pb TIMS no Grupo Little Dal, sob Grupo Rapitan	Jefferson & Parrish (1989)
Austrália	777	± 7	U-Pb SHRIMP nas vulcânicas Boucat, sul da Austrália	Preiss (2000)
Sul da China	761	± 8	U-Pb SHRIMP no Grupo Danzhou, sob diamictito Chang'an	Li et al. (2003)
África	760	± 1	U-Pb TIMS no Grupo Otavi, sob diamictito Chuos	Halverson et al. (2005)
África	758,5	± 3,5	U-Pb TIMS no Subgrupo Ombombo, Grupo Otavi, sob diamictito Chuos	Hoffman et al. (1996) e Hoffman et al. (1998a)
Laurentia	758	± 12	U-Pb TIMS na Formação Mount Rogers, sob diamictito Konnarock	Aleinikoff et al. (1995)
Sul da China	758	± 23	U-Pb SHRIMP na Formação Xieshuihe, sob diamictito Chang'an	Yin et al. (2003)
Laurentia	755	± 18	U-Pb TIMS de clasto granítico na Formação Sayunei do Grupo Rapitan	Ross & Villeneuve (1997)
Sul da China	758	± 12	U-Pb SHRIMP na Formação Liantuo sob diamictito Chang'an	Ma et al. (1984)
África	746	± 2	U-Pb TIMS no Cráton do Congo, sob diamictito Chuos	Hoffman et al. (1996)
Laurentia	742	± 6	U-Pb TIMS na Formação Kuagunt, sob Formação Sixty Miles	Karlstom et al. (2000)
Laurentia	742	± 2	U-Pb TIMS em riolito de fluxo sob diamictito na Formação Grandfather Mountain	Fetter & Goldberg (1995)
Laurentia	736	+ 23; - 17	U-Pb TIMS na unidade ígnea mais jovem no embasamento da Formação Toby	McDonough & Parrish (1991)
Sul da China	736	± 2	U-Pb TIMS no tufo queratófiro, Grupo Banxi, sob diamictito Chang'an	Gan et al. (1993)
Laurentia	717	± 4	U-Pb SHRIMP em clasto no diamictito Scout Mountain	Fanning & Link (2006)

Tabela 2.1. continuação.

Localização	Idade (Ma)	Erro	Descrição e estratigrafia	Referências
<i>Idade mínima para a Glaciação Sturtiana</i>				
Austrália	680	± 23	U-Pb EPMA em monazitas autigênicas da Fm. Enorama, sobre diamictitos sturtianos	Mahan et al. (2010)
<i>Idades dos diamictitos Sturtianos</i>				
NW China	740	± 7	U-Pb SHRIMP no diamictito Bayisi (base); Grupo Quruqtagh, Bloco Tarim	Xu et al. (2009)
NW China	725	± 10	U-Pb SHRIMP no diamictito Bayisi (topo); Grupo Quruqtagh, Bloco Tarim	Xu et al. (2009)
Laurentia	716,5	± 0,24	U-Pb TIMS em tufo da topo do Grupo Mount Harper, NW Canadá; sin-glacial	Macdonald et al. (2010a)
Oman	711,5	+ 0,3	U-Pb TIMS no diamictito Ghubrah; sin-glacial	Bowring et al. (2007)
Laurentia	709	± 4	U-Pb SHRIMP no Membro Vulcânico Bancroft, sob diamictito Scout Mountain	Fanning & Link (2004)
Laurentia	699	± 3	U-Pb SHRIMP em riolito do Batólito Idaho (Big Creek)	Evans et al. (1997)
Laurentia	685	± 7	U-Pb SHRIMP em riolito de fluxo, Formação Edwardsburg	Lund et al. (2004)
Laurentia	684	± 4	U-Pb SHRIMP em riolito de fluxo, Formação Edwardsburg	Lund et al. (2004)
Austrália	~658	N.I.	U-Pb SHRIMP na Formação Merinjina	Fanning & Link (2006)
<i>Idades entre diamictitos Sturtianos e Marinoanos</i>				
Laurentia	688,6	+ 9,5; - 6,2	U-Pb TIMS em vulcanoclástica félsica das Montanhas Gataga, N.B.C, possivelmente correlato ao Grupo Hyland em Yukon	Ferri et al. (1999)
Laurentia	667	± 2	U-Pb SHRIMP no Membro Superior Scout Mountain, sobre diamictito Scout Mountain	Fanning & Link (2004)
Sul da China	663	± 4	U-Pb TIMS na Formação Datangpo, sob o diamictito Nantuo e sobre o diamictito Tiesiao	Zhou et al. (2004)
Urais	660	± 15	U-Pb TIMS nos Urais, sob o diamictito Lanplandian	Semikhatov (1991)
Austrália	657,2	± 5,4	Isócrona Re-Os da Formação Aralka, Austrália central	Kendall et al. (2006)
Austrália	643	± 2,4	Isócrona Re-Os do Membro Tindelpina, sul da Austrália	Kendall et al. (2006)
Austrália	640,7	± 4,7	Isócrona Re-Os do Dolomito Black River, Grupo Togari, NW Tasmânia	Kendall et al. (2009)
Austrália	592	± 14	Isócrona Re-Os da Formação Aralka, Austrália central	Schaefer & Burgess (2003)
<i>Idade dos diamictitos Marinoanos</i>				
África	635,5	± 1,2	U-Pb TIMS no Grupo Swakop, Namíbia central	Hoffmann et al. (2004)
China	636,3	± 4,9	U-Pb SHRIMP em tufo dentro da base dos diamictitos da Formação Nantuo	Zhang et al., 2008

Tabela 2.1. continuação.

Localização	Idade (Ma)	Erro	Descrição e estratigrafia	Referências
<i>Idades mínimas para glaciação Marinoana ou idades entre diamictitos Marinoanos e Gaskieranos</i>				
Sul da China	635,2	± 0,6	U-Pb TIMS no Doushantuo inferior, sobre diamictito da Formação Nantuo	Condon et al. (2005)
Sul da China	632,5	± 0,5	U-Pb TIMS no Doushantuo inferior, sobre diamictito da Formação Nantuo	Condon et al. (2005)
Sul da China	628,3	± 5,8	U-Pb TIMS no Doushantuo inferior, sobre diamictito da Formação Nantuo	Yin et al. (2005)
Sul da China	621	± 7	U-Pb SHRIMP no Doushantuo, sobre diamictito da Formação Nantuo	Zhang et al. (2005)
Quruqtagh	615	± 6	U-Pb SHRIMP no Zhamoketi, sobre diamictito Tereeken, sob diamictito Hankalchough	Xu et al. (2009)
Laurentia	607,8	± 4,7	Isócrona Re-Os da Formação Old Fort Point	Kendall et al. (2004)
Avalonia	606	+ 3,7; - 2,9	U-Pb TIMS no Grupo Harbour Main, sob diamictito Gaskiers	Krogh et al. (1988)
Laurentia	601,4	± 3,7	U-Pb SHRIMP nas vulcânicas Tayvallich	Dempster et al. (2002)
Sul da China	599	± 4	Isócrona Pb-Pb da Formação Doushantuo, sobre diamictito Nantuo	Barfod et al. (2002)
Avalonia	595,5	± 2	U-Pb TIMS em clastos de tufo no diamictito Squantum	Thompson & Bowring (2000)

2.3.1. Glaciação Sturtiana

De acordo com Hoffman & Li (2009), os registros geológicos da Glaciação Sturtiana (ou Criogeniana Médio) são quase tão comuns quanto os da Glaciação Marinoana (ou Criogeniana Superior) e foram reconhecidos em 14 paleocontinentes (Tabela 2.1, Fig. 2.5). Entretanto, as idades radiométricas disponíveis para estes depósitos variam entre 726 e 670 Ma, resultando numa incerteza muito grande quanto ao número e a duração dos eventos glaciais. Idades U-Pb precisas determinadas em depósitos sin-glaciais têm sido reportadas em apenas 2 sucessões estratigráficas:

1. 723 +16/-10 Ma (SHRIMP) e 711,5 ± 0,3 Ma (ID-TIMS): determinadas em zircão de tufo vulcânico da Formação Ghubrah, em Omã (Brasier et al., 2000; Bowring et al., 2007);

2. $716,33 \pm 0,54$ Ma (ID-TIMS): obtida em zircão de tufo intercalado com diamictitos do topo do Grupo Mount Harper (correlativo do Grupo Rapitan) no Canadá (Macdonald et al., 2010a).

Uma terceira ocorrência de rochas vulcânicas intercaladas com diamictitos foi reportada no Bloco Tarim, noroeste da China, por Xu et al. (2009). Existem dois níveis de vulcânicas, um na base dos diamictitos Bayisi, que forneceu idade U-Pb SHRIMP de 740 ± 7 Ma, e outro próximo ao topo do diamictito, cujos zircões têm uma idade U-Pb de 725 ± 10 Ma. Embora as idades obtidas auxiliem na definição da idade das glaciações neoproterozoicas, a falta de evidências da origem glacial para os diamictitos impossibilita avançar nas correlações globais.

Idades mais jovens foram obtidas por Fanning & Link (2004) na Formação Pocatello, no sudeste de Idaho, Estados Unidos, resultando numa ampla discussão sobre o diacronismo das glaciações criogenianas. Segundo estes autores, naquele local, a idade da glaciação estaria delimitada entre 717 Ma e 667 ± 5 Ma, através de idades determinadas em um clasto vulcânico do diamictito e em tufo vulcânico que ocorre acima do carbonato de capa. Novos dados em clastos do diamictito sugerem que a idade máxima da glaciação seja 701 ± 4 Ma, ou mesmo mais jovem, dependendo de “futuros trabalhos de campo e interpretação da estratigrafia da região” (Fanning & Link, 2008). Adicionalmente, em outro local de Idaho (Oxford Mountains) zircões de uma brecha epiclástica (*epiclastic plagioclase-phyric tuff breccia*) que ocorre abaixo do diamictito do Membro Scout Mountain forneceram uma idade U-Pb (SHRIMP) de 709 ± 5 Ma (Fanning & Link, 2004). Estes mesmos dados foram reavaliados por Fanning & Link (2008) resultando na mudança da idade de 709 ± 5 Ma para 686 ± 4 Ma; contudo, segundo aqueles autores, a exata posição estratigráfica da rocha vulcânica em relação ao diamictito ainda não é bem entendida.

Embora os depósitos glaciais da Austrália tenham sido usados para denominar as glaciações neoproterozoicas, idades radiométricas diretas destas unidades não são disponíveis (Preiss, 2000, Fanning & Link, 2008) e dificultam a correlação com outras sucessões ao redor do globo. Recentemente, Fanning & Link (2008) determinaram uma idade de 659 ± 6 Ma em zircões de uma rocha vulcanoclástica da Formação Wilyerpa

(considerada como sturtiana), no sul da Austrália. Este dado fez com que os autores comparassem esta idade com outras similares, obtidas pelo método Re-Os em folhelhos negros da Formação Tapley Hill (estratigraficamente acima dos diamictitos glaciais), e que variam de 643 a 645 Ma (Kendall et al., 2006). Recentemente, Mahan et al. (2010) obtiveram uma idade de 680 ± 23 Ma em monazitas autigênicas da Formação Enorama (estratigraficamente superior à Formação Tapley Hill), e a interpretaram como a idade mínima de deposição dos depósitos glaciais sturtianos. Estes autores refutam a interpretação dada por Kendall et al. (2006) de que as idades de 640 Ma, obtidas pelo método Re-Os, representem a idade de deposição, e sugerem que sejam interpretadas como da época de percolação de fluidos que afetou as rochas do sul da Austrália (Mahan et al., 2010). Se isto for aceito (e comprovado) a idade mínima para a Glaciação Sturtiana na Austrália é de 680 Ma.

Por fim, idades mais antigas que 726 Ma, variando entre 735 e 754 Ma, têm sido reportadas na literatura (Tabela 2.1) para diamictitos que ocorrem na Namíbia (Grupo Port Nolloth) e na Zâmbia. Contudo, a posição estratigráfica, a origem glacial, ou a precisão dos dados obtidos nestes depósitos requerem uma melhor avaliação (Hoffman & Li, 2009; Macdonald et al., 2010b).

Em suma, os dados disponíveis da literatura mostram um grande intervalo de idades para a Glaciação Sturtiana, contudo é importante lembrar que este amplo intervalo (670 a 726 Ma) é determinado por idades máximas e mínimas de sedimentação. Se forem consideradas apenas as idades obtidas em cristais de zircão de tufos intercalados em depósitos glaciais, com bom conhecimento estratigráfico, o intervalo diminui consideravelmente e se restringe ao período de $711,5 \pm 0,3$ Ma e $716,33 \pm 0,54$ Ma (Bowring et al., 2007; Macdonald et al., 2010a). Se for considerada a idade dos diamictitos Bayisi, na China, este intervalo pode chegar a 740 ± 7 Ma. Contudo, sua origem glaciogênica ainda necessita ser confirmada.

2.3.2. *Glaciação Marinoana*

Considerando a tríade das glaciações neoproterozoicas, a Glaciação Marinoana (Criogeniano Superior) parece ser a melhor datada e aquela para a qual existe maior consenso sobre a sincronia do evento glacial ocorrido há cerca de 636 Ma. Seus registros contêm carbonatos de capa com litologias e sinais isotópicos distintos (Kennedy et al., 1998) e têm sido encontrados em 15 paleocontinentes (Tabela 2.1, Fig. 2.5).

As idades máximas para a Glaciação Marinoana foram determinadas na China e Austrália. Na China, o topo da Formação Datangpo, pré-glacial, tem zircões com idade U-Pb SHRIMP de $654,5 \pm 3,8$ Ma (Zhang et al., 2008). Na Austrália, a idade Re-Os de $643 \pm 2,4$ Ma (Kendall et al., 2006) foi obtida em folhelhos negros do Membro Tindelpina na base da Formação Tapley Hill, pré-Marinoana. Dados obtidos em níveis sin-glaciais incluem a idade U-Pb ID-TIMS de $635,5 \pm 1,2$ Ma em zircões recuperados de um tufo vulcânico da Formação Ghaub (Hoffmann et al., 2004), e a idade U-Pb SHRIMP de $636,3 \pm 4,9$ Ma em zircões de tufos da base da Formação Nantuo (Zhang et al., 2008). O fim do período glacial Marinoano é definido por zircões de camadas de cinzas vulcânicas intercaladas com os carbonatos de capa que recobrem o tilito Nantuo, na Plataforma Yangtze, China, os quais forneceram uma idade U-Pb ID-TIMS de $635,2 \pm 0,6$ Ma (Condon et al., 2008). A combinação destas idades indica uma duração de menos de 10 milhões de anos para esta glaciação que terminou globalmente há 635 Ma. Contudo, é importante ressaltar que ainda não há uma idade disponível para a “verdadeira Glaciação Marinoana” na sua seção tipo (Enorama Creek, Flinders Ranges, Sul da Austrália), cujo carbonato de capa (Formação Nuccallena) é interpretado como a base do Período Ediacarano.

CAPÍTULO 3

OS DEPÓSITOS GLACIAIS CRIOGENIANOS DO BRASIL CENTRAL: CONTEXTO REGIONAL

A geologia regional das áreas estudadas nesta pesquisa encontra-se descrita dentro dos respectivos artigos científicos que constituem os próximos capítulos da tese. Contudo, para mostrar as unidades estudadas dentro de um contexto geológico mais amplo, dois capítulos de livro compõem este Capítulo 3, os quais inserem as formações geológicas pesquisadas dentro das unidades geotectônicas maiores: os crátons do São Francisco e Amazônico e as faixas móveis marginais.

3.1. CRÁTON DO SÃO FRANCISCO E FAIXA ARAÇUAÍ

O capítulo de livro “The São Francisco Palaeocontinent” de autoria de **Sial et al. (2009)**, e que constitui o Capítulo 3 do livro “*Neoproterozoic-Cambrian Tectonics, Global Change e Evolution – A focus on Southwestern Gondwana*”, apresenta o contexto regional dos:

- (i) Diamictitos glaciais da Formação Bebedouro, unidade basal do Grupo Una no Estado da Bahia, e equivalente às formações Jequitaí e Carrancas, base do Grupo Bambuí em Minas Gerais;
- (ii) Diamictitos glaciais das formações Serra do Catuni e Chapada Acauã, parte inferior do Grupo Macaúbas, que ocorrem na Faixa Araçuaí, a leste do Cráton do São Francisco;
- (iii) Carbonatos de capa da base da Formação Sete Lagoas, os quais recobrem os diamictitos glaciais do Grupo Bambuí.

THE SÃO FRANCISCO PALAEOCONTINENT[☆]

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3.1. INTRODUCTION

The São Francisco Craton (Almeida, 1977) and its counterpart in Africa, the Congo Craton (Trompette, 1994), represent sectors of a Neoproterozoic palaeocontinent preserved from the Brasiliano–Pan African orogeny as recorded in their marginal belts. The limits between the São Francisco Craton and the surrounding belts (Brasília, Araçuai, Rio Pardo, Sergipano, Riacho do Pontal and Rio Preto belts; Figure 3.1) are marked by intense folding and overthrusting. In the interior of the craton, the Neoproterozoic sedimentary cover is only gently deformed or horizontal and preserved in isolated basins with distinct names: the São Francisco, Irecê and Una-Utinga basins. The São Francisco basin includes basal assemblages of Palaeo- and Mesoproterozoic age (e.g. Martins-Neto and Alkmim, 2001). However, in this chapter, we restrict to the discussion to the Neoproterozoic sedimentary assemblages comprising the São Francisco Supergroup (Alkmim and Martins-Neto, 2001; Martins-Neto and Alkmim, 2001), which is characterised by a glaciogenic unit at the base Jequitáí Formation and an argillaceous–calcareous–arkosic unit (Bambuí Group) at the top.

The Neoproterozoic evolution of the São Francisco Craton, including the Brasiliano marginal belts and the sedimentary cratonic cover, is reviewed in this chapter. The lithostratigraphy, Neoproterozoic-aged mineralisation (lead, zinc, fluorine, barium and phosphates) and available geochronological and stable isotope data are also presented.

[☆] Sial, A.N., Dardenne, M.A., Misi, A., Pedreira, A.J., Gaucher, C., Ferreira, V.P., Silva Filho, M.A., Uhlein, A., Pedrosa-Soares, A.C., Santos, R.V., Egydio-Silva, M., Babinski, M., Alvarenga, C.J.S., Fairchild, T.R., Pimentel, M.M. 2009. The São Francisco Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds): *Neoproterozoic–Cambrian Tectonics, Global Change and Evolution: a focus on southwestern Gondwana*. Developments in Precambrian Geology, 16, Elsevier, pp. 31–69.

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Figure 3.1 Simplified geological map of the São Francisco Craton showing the extent of Neoproterozoic cover (modified from Misi et al., 2005a,b).

It is well known that the late Neoproterozoic was marked by significant changes in Earth's climate. Clues to these environmental perturbations are recorded in glacial deposits, sometimes formed in equatorial latitudes (Sohl et al., 1999) and often found overlain by dolostones or limestones (so-called cap carbonates; e.g. Kennedy et al., 1998). These strong climatic variations seem to have occurred between 750 and 580 Ma, during at least three distinct glacial epochs, commonly termed Sturtian, Marinoan and Gaskiers glacial events (Halverson et al., 2005). In this chapter, the geological record of Neoproterozoic glacial events and corresponding stable isotopic data for related cap carbonates in the São Francisco Craton are also discussed.

3.2. SYNTHESIS OF THE LITHOSTRATIGRAPHIC UNITS

3.2.1. The Neoproterozoic cover of the São Francisco Craton

3.2.1.1. The Bambuí Group

Over the São Francisco Craton, the Bambuí Group (São Francisco basin; Figure 3.1) is the most important Neoproterozoic sedimentary unit, covering large cratonic areas and occupying, in central Brazil, segments of the eastern side of the Brasília belt. It overlies the Paranoá Group from which it is separated by an unconformity filled

Column	Lithology	Formation	Group
	greenish and reddish arkoses	Três Marias Formation	Bambuí Group
	greenish siltstones greenish shales	Serra da Saudade Formation	
	black oolitic limestones and marls	Lagoa do Jacaré Formation	
	dark grey shales and siltstones with lense of sandstones and limestones	Serra da Santa Helena Formation	
	light grey to pink dolomites grey limestones dark grey limestones dark grey marls greenish and reddish marls pink cap dolomites	Sete Lagoas Formation	
	diamictites	Jeiquitaí Formation	Paranoá Group
	sandstones		
	siltstones		

Figure 3.2 Lithostratigraphic column of the Bambuí Group (Dardenne, 2007a).

by glacial diamictites of the Jeiquitaí Formation (Figure 3.2). The Paranoá sedimentary rocks consist mostly of mature siliciclastic sediments such as quartzites, with intercalations of metasiltstones and minor lenses of limestones and dolostones. This group has been divided into nine lithostratigraphic units (Faria, 1985), beginning with a paraconglomerate, followed by transgressive and regressive siliciclastic-dominated cycles, and ending with pelites and dolostones containing *Conophyton metula* Kirichenko stromatolites (Cloud and Dardenne, 1973; Cloud and Moeri, 1973; Dardenne et al., 1976; Dardenne, 1979). Available geochronological and microfossil data for the Paranoá Group point to an age of 1,170–950 Ma and a source region in the Palaeoproterozoic sialic basement of the craton, suggesting sedimentation on a passive margin (Guimarães, 1997; Pimentel et al., 1999).

The Jeiquitaí Formation (Figure 3.3a and b), at the base of the Bambuí Group, consists of clast-supported, greenish or grey diamictites (quartzite, granite, gneiss, limestone, dolostone and siltstone) with clay-rich matrix, sometimes marly, and minor siltstone and sandstone lenses. This unit crops out discontinuously in central Brazil and overlies both the Paranoá rocks and the basement (Dardenne, 1978a, 1979, 2000; Karfunkel and Hoppe, 1988; Uhlein, 1991, 2004; Uhlein et al., 1999) and represents syn-glacial deposition over a wide area of the São Francisco Craton. Locally, a thin horizon of a light-grey to pink cap dolostone overlies the glaciogenic diamictites.

Deposits of the Jeiquitaí Formation, probably Sturtian in age (~750 Ma), are well exposed at the margins of Serra do Cabral and Agua Fria areas in the southeastern portion of the São Francisco Craton. This formation, up to 150 m thick, overlies quartzites of the Palaeo-Mesoproterozoic Espinhaço Supergroup with a slight unconformity in the Serra do Cabral area. The tops of these quartzites show subglacial erosion structures such as grooved and striated pavements with striae-oriented ENE-WSW. The Jeiquitaí Formation consists of massive and stratified diamictites with granules, pebbles and boulders of gneisses, granites, quartzites and carbonates (Uhlain et al., 2004). The diamictites are massive in their lower part and in their upper part exhibit alternating clast-rich and

Figure 3.3 (a) Lithofacies variations of the Jequitai Formation near Serra do Cabral, Minas Gerais, and Cristalina, Goiás (Uhlein et al., 1999); (b) diamictite of the Jequitai Formation at Serra da Água Fria, Minas Gerais.

clast-poor beds (Figure 3.3). This formation also contains lenticular sandstones, siltstones and a few laminated siltstone–mudstone intercalations. Clast-poor diamictites overlying striated pavement suggest two phases of the glacial event. The first phase, probably on the continent, produced ice erosion (striated pavement) and the second one a clast-poor diamictite deposited by gravity flows during ice retreat and sea-level rise.

The Jequitai Formation, also named the Carrancas Formation, is correlative to the Bebedouro Formation that underlies carbonates of the Salitre Formation (Una Group) in the northeastern part of the São Francisco cratonic area. In the Onça do Pitangui map sheet, in the state of Minas Gerais, layers of unmetamorphosed varvites (Figure 3.4) of the Carrancas Formation form a 30 m-thick deposit with dropstones (informally named the Moema sequence by Rocha-Campos et al., 2007) towards the top of the Jequitai Formation.

Figure 3.4 (a and b) Unmetamorphosed rhythmite deposit of the Carrancas Formation. Locally, dropstones are observed in this deposit. Onça do Pitangui map sheet, Minas Gerais, eastern Brazil.

The post-glacial transgression flooded the craton, during which time pelitic and carbonate sediments of the lower Bambuí Group were deposited (Figure 3.2) over at least 300,000 km² (Dardenne, 1978a,b, 1979, 2000; Misi and Kyle, 1994). This sedimentary association, which follows the Jequitai glaciation, is repeated in three regressive megacycles. Each of these megacycles begins with a regional marine transgression associated with basin subsidence, the evidence for which is seen in deep pelitic marine facies, passing upwards into shallow platform, subtidal and supratidal facies.

From base to top, these megacycles are arranged as follows. *Megacycle I* is pelitic–calcareous, corresponding to the Sete Lagoas Formation, forming a coarsening-upward sequence with calcilutites in the basal portion and passing into limestones and dolostones at the top. At the base, characteristic pink dolomitic calcilutites with green argillaceous films are generally observed. They pass progressively to dark-grey and black-laminated calcilutites. The limestones, generally dark-grey to black and microcrystalline, are well stratified and homogeneous, but lenticular in large scale. To the top, carbonate facies are predominantly dolomitic, distinguished by intraclasts and oolites. The first cycle ends with an extensive subaerial exposure marked by tepee structures, moulds of sulphate nodules, dissolution breccias, and laminated and columnar stromatolites (Dardenne, 1978a,b, 1979; Dardenne and Freitas-Silva, 1999).

Megacycle II begins with the Serra de Santa Helena Formation, which is dominantly pelitic at the base and signals sudden but broad subsidence of the basin. The pelites are followed by dark-grey platform limestones of the Lagoa do Jacaré Formation, which were deposited in an environment dominated by storms and tidal currents.

Megacycle III is pelitic to sandy and represented by the pelitic Serra da Saudade Formation at the base, deposited in deep-platform environment with episodic influence of storms, and by the Três Marias Formation at the top, dominated by arkose and deposited in a shallow-platform environment influenced by storm currents with episodic tidal to supratidal facies. The pelites are distinctly greenish in colour, containing abundant detrital mica grains in bedding planes, while the greenish or reddish arkoses are fine-grained and rich in plagioclase and mica. Some contain volcanic lithic fragments derived from the erosion of a Brasiliano magmatic arc to the east and are regarded as a foreland basin molasse.

Along the São Francisco River, between the cities of Januária, Itacarambi, Montalvânia and Manga, well-developed and continuous carbonate horizons (Figure 3.5) have been named the Januária and Nhandutiba formations (Dardenne, 1978a,b). The former represents a typical coarsening-upward regressive megacycle, consisting, from bottom to top, of dark-grey calcilutites, dark-grey to grey limestones with lamellar breccias or oolites, pink saccharoidal dolarenites and beige lithographic dolostones. Evidence of emergence is sometimes observed between saccharoidal dolarenites and lithographic dolostones, indicating discontinuity or unconformity in the upper portion of the Januária Formation. This unconformity underlines the impermeable level of lithographic dolostone that has controlled the percolation of basinal fluids responsible for development of secondary dolomitisation, formation of saccharoidal dolarenite and development of Mississippi valley-type Pb–Zn–F mineralisation.

The carbonate and pelite facies are considerably thicker in the region between the Januária–Itacarambi cratonic area and Vazante (Figure 3.6), reflecting regional subsidence induced by differential movement along N–S faults during deposition, generating significant westward deepening of the basin (Alvarenga and Dardenne, 1978; Dardenne, 1978a,b, 1979, 2000).

The cap dolostone at the base of the Sete Lagoas Formation overlying Carrancas diamictites has been dated at ca. 740 ± 22 Ma (whole-rock Pb–Pb isochron on carbonates; Babinski et al., 2007) and this age, along with C and Sr isotope data for carbonates of the Bambuí Group (Santos et al., 2000, 2004; Vieira et al., 2007), supports a Sturtian age for the Carrancas glaciation and a Cryogenian age for at least the base of the Bambuí Group.

Figure 3.5 Lithostratigraphy of the Bambuí Group along the São Francisco valley (Dardenne, 1978b, 1979).

3.2.1.2. The Una Group

The carbonate and siliciclastic units of the Una Group are present in four now disconnected “sub-basins” in the northeastern sector of the São Francisco Craton. The Irecê, Campinas, Una-Utinga and Ituaçu “sub-basins” were probably connected before the end of the Pan African/Brasiliano orogeny, forming a large sedimentary basin

Figure 3.6 Thickness variations of the Bambuí Group in the São Francisco Craton (after Dardenne, 1978b, 1979).

covering much of the craton. Possible correlative successions (although with separate nomenclature) are present in the passive margin basins of the Sergipano (Vasa Barris Group) and Rio Pardo belts, bordering the cratonic area (Figure 3.7). The Una Group is composed of two megasequences (Figure 3.8):

- Glaciogenic megasequence*: This interval is represented by the Bebedouro Formation that unconformably and variably overlies Palaeoproterozoic gneisses and migmatites and Mesoproterozoic metasedimentary rocks of the Chapada Diamantina Group. The Bebedouro Formation consists of diamictites, pelites and sandstones with a variety of lithofacies that are grouped into four associations (Guimarães, 1996): (i) aeolian (extraglacial), (ii) ice-contact, (iii) pro-glacial and (iv) iceberg melting.
- Carbonate megasequence*: This package is dominantly carbonate, with subordinate siliciclastic units, and unconformably overlies the Bebedouro Formation. Referred as the Salitre Formation, the megasequence is subdivided into five informal units (Misi and Souto, 1974; Misi, 1979), each of which can be correlated with the Bambuí Group. From top to bottom:

Unit A1 is a black organic-rich, cross-stratified and rippled oolitic and pisolitic limestone that correlates with the Lagoa do Jacaré Formation.

Unit A comprises interbedded marls, shales and siltstones with local lime grainstone beds. It correlates with the Serra de Santa Helena Formation.

Unit B1 consists mainly of grey to reddish dolostone with tepee structures, replaced, nodular evaporates and intraformational breccia. It correlates with the dolomitic facies at the top of the Sete Lagoas Formation.

Unit B is a grey-laminated limestone and dolomitic limestone that grades upward into the more dominantly dolomitic Unit B1. It correlates with the Sete Lagoas Formation beneath the upper dolomitic facies.

Unit C is a pink argillaceous dolostone that unconformably overlies the Bebedouro Formation and correlates with the base of the Sete Lagoas Formation.

A composite stratigraphic section of the Una Group is represented in Figure 3.8, while Table 3.1 shows the possible correlations between the Una (Vieira et al., 2007) and Bambuí groups.

Figure 3.7 Simplified geological map showing the Neoproterozoic “sub-basins” of the Una Group in the central area of Bahia state, Brazil (modified from Guimarães, 1996).

3.2.2. The surrounding belts

3.2.2.1. The Brasília belt

In the western margin of the São Francisco Craton, a thick Meso–Neoproterozoic sedimentary/metasedimentary pile forms the Brasília belt (Araí, Paranoá, Serra da Mesa, Araxá, Ibiá, Vazante, Canastra and Bambuí) that extends for more than 1,000 km (Figure 3.9) and is part of a passive margin association. These rocks are mostly undeformed and unmetamorphosed over the craton and increasingly deformed and metamorphosed westward, reaching amphibolite and granulite facies conditions in the central part of the belt (Dardenne, 1978a,b, 2000; Fuck et al., 1994; Pimentel et al., 2000; Piuzana et al., 2003). The evolution of the deformation and metamorphism reflects the vergence of this belt towards the São Francisco Craton. The tectonic zonation proposed by Costa and Angeiras (1971) was reformulated by Fuck et al. (1994) who identified an internal zone in the west, a cratonic zone in the east and an external zone in between.

A WNW–ESE lineament at about the latitude of Brasília allows a subdivision of the Brasília belt into two segments with distinct geotectonic histories. In the northern segment, sedimentary units were metamorphosed to greenschist facies or are unmetamorphosed and the well-preserved stratigraphy allows the reconstitution of the palaeogeography and depositional systems. In the area to the north of Brasília, where sequences of the Paranoá and Araí groups occur, the compressional tectonics is expressed in right lateral transcurrent faults and thrusts that locally have affected the sedimentary cover.

The southern segment exhibits distinct tectonic features when compared to the northern one. Deformation and metamorphism are very intense and have obliterated stratigraphic relationships between the various units. The Araxá, Canastra, Ibiá and Vazante groups are involved in a complex, imbricated system of nappes and thrusts

Figure 3.8 Composite stratigraphic cartoon showing the two megasequences of the Una Group (modified from Misi et al., 2007).

Table 3.1 Correlations between the Una and Bambuí groups at the formational level.

Group	Bambuí (Formations)	Una (Formations)
Lithotypes		
Arkose, siltstone	Três Marias	—
Siltstone, pelites	Serra da Saudade	—
Black oolitic limestone	Lagoa do Jacaré	Unit A1
Marl, shales	Serra Santa Helena	Unit A
Dolostone with tepee	Sete Lagoas	Unit B1
Laminated limestone		Unit B
Pink dolostone		Unit C
Diamictite, arkose, pelitic rocks	Jeiquitaí, Carrancas	Bebedouro

indicating a large amount of tectonic transport, on the order of tens to hundreds of kilometres. The contacts between the various involved assemblages correspond to low-angle shear zones, with sheath folds and lateral ramps. The external zone is characterised by a well-developed thin-skinned system of folding and overthrusting from west to east, with increasing deformation and metamorphism towards the internal zone to the west. From east to west, it is possible to recognise successive rock assemblages belonging to the Bambuí, Vazante and Ibiá-Araxá groups.

Lithostratigraphic units of the Bambuí Group are easily identified on the flanks of anticlines and synclines, frequently faulted, with N-S to N20E trending axes, as in the Formosa and Bezerra areas to the east of Brasília. In the core of anticlines, one observes sandstones, arkoses, rhythmites and stromatolitic dolomites of the Paranoá Group, overlain in erosional unconformity by Bambuí sedimentary rocks of the Jeiquitaí, Sete Lagoas, Serra de Santa Helena, Lagoa do Jacaré, Serra da Saudade and Três Marias formations (Figure 3.2).

In the Cristalina area (Figure 3.1), diamictites of the Jeiquitaí Formation occur in the flanks of a large brachianticline, disconformably overlying quartzites of the Paranoá Group (Faria, 1985; Cukrov et al., 2004; Uhlein et al., 2004). In this area, phyllites of the Canastra Group overthrust Jeiquitaí diamictites.

Figure 3.9 Simplified geological map of the Brasília belt showing the Palaeo-Meso-Neoproterozoic units, according to Dardenne (2000), Pimentel et al. (2001) and Valeriano et al. (2004). Pb-Zn deposits: (1) Vazante; (2) Morro Agudo; (3) Fagundes; (4) Ambrosia (Cunha et al., 2007).

Figure 3.10 Schematic geological column of the Lagoa Formation (Seer et al., 1989; Baptista, 2004).

In the southwestern portion of the São Francisco Craton, unusual conglomeratic facies occur in the Lagoa Formosa, Samburá and Carmo do Rio Claro formations of the upper Bambuí Group. The Lagoa Formosa Formation (Seer, 2001) crops out for over 300 km between the towns of Tiros and Presidente Olegário (Serra do Abaeté Mountains), where it shows little deformation with only a weak schistosity indicating northeastern vergence. This assemblage (Figure 3.10) probably correlates with the Serra da Saudade Formation and comprises two interbedded sequences: (i) stratified and massive diamictites (500–1,000 m thick) and (ii) a well-stratified 1,000–2,000 m-thick sequence of fine-grained diamictites, siltstones, sandstones, greywackes, jaspillites and limestones (Uhlein et al., 2004). Pebbles and boulders are predominantly of intrabasinal arkoses and siltstones floating in a greenish grey pelitic to sandy-pelitic matrix. Some clasts are composed of limestone, chert, quartzite and granite-gneiss. These paraconglomerates are interpreted as gravitational debris flows originated by regional uplift of the internal zone of the Brasília belt. This interpretation is supported by a northeasterly decrease in the size of clasts in the diamictites from boulders to granules. At the base of the Lagoa Formosa Formation, thin horizons of phosphorite are intercalated with green pelites, with abundant potassium-rich illite, and known as the Verdete facies. Carbonate beds (calcirudites, calcarenites and columnar stromatolites up to 10 m thick) are interbedded with siltstones in the lower portion of the unit and may correlate with either the top of the Lagoa do Jacaré Formation (Dardenne, 1979) or the base of the Lagoa Formosa Formation (Lima and Uhlein, 2005).

The Samburá Formation (Branco, 1957) is the name given to various minor occurrences of conglomerates, generally isolated within pelites of the upper portion of the Bambuí Group, showing similar characteristics. Most of these conglomerates are clast-supported and composed mainly of granitic, rhyolitic, arkosic and quartzitic pebbles. They are interpreted to have been deposited in eastward-tapering alluvial fans (Figure 3.11) (Castro and Dardenne, 1996).

The Carmo do Rio Claro Formation near the Furnas dam, described by Heilbron et al. (1987), is a combination of conglomerates, diamictites, arkoses and siltstones that correlate with the Samburá Formation, as well as slates, phyllites and marbles. The diamictites, which contain rounded gneiss, metapelite and metasandstone

Figure 3.11 Schematic geological column of the Samburá Formation near Samburá hydroelectric dam (after Castro and Dardenne, 1996).

pebbles, are considered to be derived from the west. The sequence is overthrust by rocks of Canastra and Araxá groups, which form the Passos Nappe. The Carmo do Rio Claro Formation is stratigraphically and genetically equivalent to the Lagoa Formosa Formation.

The Vazante Group metasediments are a thick sequence of marine pelites and dolostones that crop out for over 300 km along a north–south trend in northwestern Minas Gerais state, spanning the cities of Coromandel, Lagamar, Vazante, Paracatu and Unaí. Misi et al. (2007) have argued that the Vazante and Bambui groups are correlative, based on equivalent Sr isotope signatures in both successions (~ 0.7074) and the interpretation of seismic profiles by Romeiro-Silva and Zalan (2005). This controversial argument conflicts with the contrasting C isotope records between the two units and new Re-Os ages (unpublished) of organic shales of the Serra do Garrote, Serra do Poço Verde and Serra da Lapa formations that imply a late Mesoproterozoic age (A.J. Kaufman and K. Azmy, oral communications).

The Vazante Group is subdivided into seven formations (Figure 3.12), from base to top: Santo Antônio do Bonito, Rocinha, Lagamar, Serra do Garrote, Serra do Poço Verde, Morro do Calcario and Lapa formations (Dardenne, 2001).

The Santo Antônio do Bonito Formation (Souza, 1997), considered to be the basal formation, consists of beds of white quartzites, sometimes conglomeratic, intercalated with slates. In the Santo Antônio do Bonito and Santo Inacio rivers, this formation is characterised by the presence of diamictite horizons within the quartzites that contain limestone, dolostone, metasiltstone and granitic pebbles floating in a pelitic matrix, which is locally phosphatic. Larger concentrations of phosphate are found in the slaty facies and in phospharenite layers, rich in

Figure 3.12 Lithostratigraphic column of the Vazante Group (Dardenne, 2007b).

intraclasts and pellets (Phosphorite 1). The diamictites represent debris flows deposited in relatively deep waters in a glacial environment.

The Rocinha Formation (Dardenne et al., 1997) is composed of a basal rhythmic sandy and pelitic sequence that is gradational with the underlying Santo Antônio do Bonito Formation. The upper Rocinha Formation is a thick package of regularly intercalated slates and metasilstones. It passes upward into dark-grey marly and pyritic slates, with thin phosphatic laminations that transition into the intraclast- and pellet-rich phospharenites (Phosphorite 2), forming the Rocinha phosphate deposit. In the upper part of the formation, rhythmic sediments (quartzites and siltstones) contain the Lagamar phosphate deposit composed almost entirely of phospharenites (Phosphorite 3).

The Lagamar Formation is a psammite–pelite–carbonate unit represented in its basal portion by alternations of conglomerate, quartzite, metasiltstone and slate. The conglomerates, ascribed to the Arrendido Member, consist of a framework of quartzite, metasiltstone and dark-grey limestones clasts. The psammites are overlain by dolomitic intraformational breccias passing upward into dark-grey, well-stratified limestones with intercalations of lamellar breccias followed by stromatolitic dolostones, which comprise beige to pale pink bioherms composed of microbialaminated dolostones, oncolitic dolarenites and dolorudites, and columnar stromatolites with convex and conical laminations classified as *C. metula* Kirichenko and *Jacutophyton* type (Dardenne et al., 1976). Laterally and vertically, these bioherms interfinger with marly metasiltstones and slates.

The Serra do Garrote Formation (Madalosso and Valle, 1978; Madalosso, 1980), which is about 400 m thick, comprises thick, dark-grey to greenish grey slates, sometimes rhythmic, carbonaceous and pyritic, with fine-grained quartzite intercalations. The Serra do Poço Verde Formation (Dardenne, 1979) is dominantly dolomitic and is divided into four members (Rigobello et al., 1988), described successively from the base to the top. The 500 m-thick Lower Morro do Pinheiro Member is composed of light-grey- and/or pink-laminated dolostones with cyanobacteria mats, intercalated with oncolitic dolarenites and intraformational breccias and lenses of dolostones with columnar stromatolites. The Upper Morro do Pinheiro Member is composed of medium- to dark-grey microbialaminated dolostones with common birdseye fenestrae, intercalated with layers of dolarenites, lamellar breccias and carbonaceous shales (thickness from 300 to 500 m). The occurrence of *Conophyton cylindricus* Maslov or *C. metula* Kirichenko (Moeri, 1972; Cloud and Dardenne, 1973) near the Cabeludo village is probably related to this unit. The Lower Pamplona Member is 100–200 m thick and composed of grey, green and purple siltstones intercalated with micritic and microbialaminated dolostones with small lenses of fine-grained to conglomeratic sandstones. The ~400 m-thick Middle Pamplona Member is composed of light-grey to pink microbialaminated dolostones with locally developed mud cracks and nodular barite, intercalated with beds of dolarenite, lamellar breccia and columnar stromatolites, and with lenses of shales. The 200–300 m-thick Upper Pamplona Member is characterised by the presence of light-grey to pink stromatolitic dolostones constituting biostromes and bioherms with convex-laminated columns and associated with oolitic and oncolitic dolarenites and dolorudites. In the region of Morro Agudo, Paracatu and Unaí, this unit corresponds to the Morro do Calcario Formation, which is here >900 m thick and composed mainly of dolorudites derived from reworked but still partially preserved stromatolitic bioherms and associated oolitic and oncolitic dolarenites. In this region, the Morro do Calcario and Serra do Poço Verde formations cannot be easily subdivided as they are in the Vazante region.

The Lapa Formation constitutes the upper portion of the Vazante Group described in the region of Paracatu as Serra do Velosinho and Serra da Lapa members. These are represented by carbonaceous phyllites, marly metasiltstones, dolomitic lenses and quartzite layers. The dolomitic lenses show various facies, including microbialaminated dolostones, columnar stromatolites and intraformational breccias, interfingering with the pelitic sequence that regionally overlies the dominantly dolomitic formations of Morro do Calcario and Serra do Poço Verde. In the Unaí region, the Lapa Formation comprises a rhythmic turbidite sequence of siltstones and pelite, locally with lithic sandstones and conglomerates intercalated with dark-grey slates. The clasts are fragments of phyllites, fine-grained quartzites and cherts.

The presence of *C. cylindricus* Maslov and *C. metula* Kirichenko in the dolomitic bioherms near Cabeludo and Lagamar implies a correlation with the Paranoá Group (Dardenne, 2000). Carbon and Sr isotope signatures for dolomitic rocks of these groups are fairly distinct from signatures for the Bambuí Group carbonates and are more consistent with a Mesoproterozoic age (Santos et al., 2000, 2004). Consequently, the glaciogenic diamictite of the Santo Antônio do Bonito and the overlying sedimentary sequence of Vazante Group have an unknown age, but are inferred to be Kaigas-equivalent (c. 750 Ma) or older.

Recently, various workers (Olcott et al., 2005; Azmy et al., 2005) have proposed the occurrence of a regional glacial event spanning most of the Upper Pamplona Member and Lapa Formation in the Vazante area, and Morro do Calcario, Mocambo and Lapa Formations near The Morro Agudo mine. This contention is based on the presence of a layer of brecciated rocks with dropstones, interpreted as possible glacial diamictites with a negative $\delta^{13}\text{C}$ excursion in overlying dolostones. These breccias may be alternatively interpreted as intraformational dolomitic breccias, possibly of debris flow origin (Dardenne, 2007b). If confirmed, the glacial episode recorded in the Lapa Formation would be related to the Sturtian glaciation, represented regionally by the Jequitai Formation. This suggests that the Vazante Group constitutes a stratigraphic equivalent of the upper portion of the Paranoá Group, deposited in a passive margin setting, as previously proposed by Campos Neto (1984) and Dardenne (2007a,b).

3.2.2.2. The Araçuaí belt

The name Araçuaí was formerly applied only to the arcuate marginal belt located along the east and southeast edges of the São Francisco Craton (Almeida, 1977; Figure 3.1). The Araçuaí orogen was defined after the identification of several geotectonic components (e.g. ophiolites, pre-collisional magmatic arc and syn-collisional

Figure 3.13 Geological map of the Araçuaí orogen (modified from Pedrosa-Soares et al., 2008) and its location in relation to the São Francisco–Congo Craton (see inset, modified from Alkmim et al., 2006b: WCB, West-Congo belt; Neoproterozoic aulacogens: PA, Paramirim; PI, Pirapora; SA, Sangha).

granites) related to that marginal belt (Pedrosa-Soares et al., 2001, 2008). This orogen is 700 km long and up to 500 km wide and encompasses the entire region between the São Francisco Craton and the Atlantic continental margin, between 15°S and 21°S in eastern Brazil (Figure 3.13), including the southeastern limit of the São Francisco Craton. The Araçuaí orogen and its counterpart located in southwestern Africa (the West-Congo belt; Trompette, 1994; Tack et al., 2001) form a confined orogen evolved into an embayment carved into the São Francisco–Congo palaeocontinent (Pedrosa-Soares et al., 2001, 2008; Alkmim et al., 2006b).

The basement of the Araçuaí orogen, not described here, consists of Archaean to Mesoproterozoic units, including the rift-sag Espinhaço Supergroup and the related anorogenic Borrachudos suite (Uhlein et al., 1998b; Martins-Neto, 2000; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Noce et al., 2007).

The Araçuaí orogen comprises several Neoproterozoic stratigraphic units, such as the Macaúbas Group, which includes glacially influenced deposits and deep marine sedimentation associated with meta-mafic–ultramafic ophiolite slivers, tonalitic batholiths making up a pre-collisional magmatic arc (G1 suite), orogenic metasedimentary successions (e.g. Salinas Formation and Nova Venécia Complex) and a large sector dominated by syn-collisional granites (G2 suite). The belt is intruded by Cambrian-age post-collisional plutons (Figure 3.13).

Two magmatic episodes yield age constraints for the rift stage of the precursor basin of the Araçuaí orogen. The older age from a meta-dolerite dyke of Pedro Lessa, located ca. 50 km to the southeast of Diamantina, was dated at 906 ± 2 Ma (U-Pb, zircon and baddeleyite; Machado et al., 1989). The A-type granites of Salto da Divisa yielded a zircon U-Pb age of 875 ± 9 Ma (Silva et al., 2007).

The precursor rift basin was filled by units of the proximal Macaúbas Group (Figure 3.13). The oldest rift deposits consist of metamorphosed sandstones, arkoses, conglomerates and pelites that lack evidence of glaciation and represent fluvial sedimentation (Karfunkel and Hoppe, 1988; Noce et al., 1997; Martins, 2006). U-Pb ages of detrital zircon grains constrain the maximum sedimentation age for this early rift stage at around 900 Ma (Pedrosa-Soares et al., 2008). These pre-glacial deposits of the Macaúbas basin are covered by metamorphosed, generally massive diamictites with minor sandstones, pelites and rhythmities, representing glacio-terrestrial to gravitational glacio-marine sedimentation (Pedrosa-Soares et al., 1992; Uhlein et al., 1998b, 1999; Martins-Neto et al., 2001; Martins, 2006). This lower diamictite-rich unit was thrust over the Neoproterozoic carbonate-pelite cover of the São Francisco Craton and can be correlated with the glaciogenic Jequitai Formation of the cratonic region (Karfunkel and Hoppe, 1988; Uhlein et al., 1999; Martins-Neto and Hercos, 2002). The eastern part of the proximal Macaúbas Group includes a thick pile of sediments with layered diamictites, pebble-bearing iron formations, graded sandstones and pelites with dropstones, representing glaciogenic debris flows and turbidites deposited in submarine fan environment (Pedrosa-Soares et al., 1992; Uhlein et al., 1999; Martins-Neto et al., 2001). This succession includes mafic volcanic rocks, metamorphosed to greenschist facies, with pillow structures and other features of subaqueous flows and is derived from tholeiitic basalt protoliths with a dominant within-plate signature (Gradim et al., 2005). Inherited zircon grains from older felsic rocks yielded U-Pb ages from Archaean to Late Mesoproterozoic (the youngest at $1,156 \pm 21$ Ma), with a whole-rock Sm-Nd T_{DM} model age of ca. 1.5 Ga (Babinski et al., 2005). However, some samples with oceanic signatures and slightly positive $\epsilon Nd_{(900\text{ Ma})}$ of +0.23, together with the inherited zircon grains, suggest a transitional mafic magma that migrated through a thinned continental crust. Accordingly, these greenschists provide strong evidence of volcanism in an extensional marine basin floored by thin continental crust, during the very late rift stage of the Macaúbas basin (Gradim et al., 2005; Pedrosa-Soares et al., 2008).

The upper succession of the proximal Macaúbas Group consists of interlayered sandstones and pelites, metamorphosed from the greenschist to lower amphibolite facies, representing the proximal passive margin sedimentation of the Macaúbas basin. The youngest detrital zircon grain from a sandstone layer constrains the maximum sedimentation age of this shelf succession to around 864 Ma (Pedrosa-Soares et al., 2000a, 2008).

The distal part of the Macaúbas Group is a passive margin sequence free of diamictite, known as the Ribeirão da Folha Formation (Figure 3.13). This unit was formerly considered to be part of the Salinas Formation (Pedrosa-Soares et al., 1992, 1998, 2001), but since the redefinition presented by Lima et al. (2002), it has been separated (Pedrosa-Soares et al., 2008). The Ribeirão da Folha Formation was metamorphosed progressively from west to east to the garnet, staurolite, kyanite and sillimanite zones (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Queiroga et al., 2006). The western part of the Ribeirão da Folha Formation comprises banded quartz-mica schists (fine-grained turbidites) with thin calc-silicate (marl) and marble lenses. The eastern Ribeirão da Folha Formation is associated with thrust slices of oceanic meta-mafic and meta-ultramafic rocks, constituting a tectonically dismembered ophiolite complex. This eastern succession is rich in peraluminous mica schists (pelagic pelites) with intercalations of sulphide-bearing cherts and diopsidites associated with massive sulphide bodies, oxide, silicate and sulphide-type banded iron formations, and graphite schists (Pedrosa-Soares et al., 1992, 1998, 2001; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Suita et al., 2004; Queiroga et al., 2006). The meta-mafic rocks associated with the Ribeirão da Folha Formation are coarse-grained massive ortho-amphibolites, representing gabbroic protoliths, and medium- to fine-grained banded ortho-amphibolites that seem to be derived from dolerites and basalts (Pedrosa-Soares et al., 1998, 2008; Suita et al., 2004). These ortho-amphibolites have a geochemical signature akin to modern ocean-floor mafic rocks and are comparable to similar rocks of other Neoproterozoic ophiolites (Pedrosa-Soares et al., 1998, 2001; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Suita et al., 2004). The U-Pb (LA-ICPMS) age of 660 ± 29 Ma obtained from zircon crystals of a plagiogranite vein suggests the epoch of magmatic crystallisation of at least part of the oceanic magmatic rocks (Queiroga et al., 2007).

The Jequitinhonha Complex comprises a thick paragneiss (kinzigitic) association, consisting of migmatized biotite gneisses, with variable contents of garnet, cordierite, sillimanite and graphite, thick intercalations of graphite-rich gneiss, and minor quartzite and calc-silicate rocks. The paragneiss protoliths comprise marine arkosic greywackes to peraluminous pelites, deposited under oxidising conditions, with horizons of carbonaceous black mud that were deposited in a restricted marine environment. The sedimentary protoliths are interpreted to have been deposited during the passive margin stage of the Macaúbas basin (Pedrosa-Soares and Wiedemann-Leonardos, 2000b).

The western external domain of the Araçuaí orogen, adjacent to the southeastern edge of the São Francisco Craton, is a west-verging fold-thrust belt, with increasing intensity of metamorphism from west to east.

The northern segment of the external tectonic domain bends to the east, outlining the northern curvature of the orogen, where it shows north-verging structures and metamorphic grade increasing from north to south (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Pedrosa-Soares et al., 2001, 2008; Uhlein, 2004; Alkmim et al., 2006b).

Along the western external domain, the major normal faults of the precursor rift basin, active during the Tonian, were reactivated as thrust faults during the contractional orogenic event (Uhlein et al., 1999; Martins-Neto and Hercos, 2002; Uhlein, 2004). The main deformational event, related to the syn-collisional stage, is characterised by zones with asymmetrical folds showing westward vergence separated by mylonitic belts of brittle–ductile shear zones. An east–southeast dipping S1 cleavage or mylonitic schistosity with down-dip stretching lineation and kinematic indicators such as S–C surfaces, patterns of deformed and boudinaged veins, asymmetric deformation around porphyroclasts and clasts, and sheath folds indicate a westward tectonic transport towards the São Francisco Craton (Uhlein, 1991; Pedrosa-Soares et al., 1992; Uhlein et al., 1995, 1998b; Cunningham et al., 1998; Alkmim et al., 2006b). Thrusts and reverse faults are defined by mylonites in which S–C fabrics or porphyroclast tails indicate top-up-to-the-west movement. Pebbles within Macaúbas diamictite layers were stretched with their long axis parallel to the E–W mineral lineation (Uhlein, 1991; Pedrosa-Soares et al., 1992). Rocks of the eastern proximal Macaúbas Group show asymmetric extensional crenulation cleavage (S_2) as the axial surface of F_2 folds. These structures show top-down-to-the-east tectonic transport and indicate a regional-scale normal-sense shear zone probably related to the extensional (gravitational) collapse of the Araçuaí orogen (Marshak et al., 2006).

The internal tectonic domain is the high-grade metamorphic core of the orogen, which includes the pre-collisional magmatic arc, paragneiss complexes and a huge anatectic zone dominated by syn-collisional S-type granites and post-collisional intrusions (Figure 3.13). The internal tectonic domain is generally west-verging, but the back-arc zone between 17°30'S and 20°S shows syn-collisional tectonic transport to the east (Pedrosa-Soares et al., 2001, 2006, 2008; Alkmim et al., 2006b; Vauchez et al., 2007). The southern sector of the Araçuaí orogen and its connection with the Ribeira orogen are marked by syn-collisional west-vergent thrusts cut by late, dextral, strike–slip shear zones (Cunningham et al., 1998; Heilbron et al., 2004; Peres et al., 2004; Alkmim et al., 2006b). High-grade metamorphism and partial melting are syn-kinematic with the regional foliation and took place at ca. 585–560 Ma. The internal domain shows widespread granitoids of the G1 and G2 suites (Figure 3.13). In general, G1 and G2 batholiths display high-temperature (600–700°C) solid-state foliation parallel to the magmatic orientation, both concordant with the regional foliation of the country rocks, suggesting synchronous deformation of the magmatic rocks and their host rocks (Uhlein et al., 1998a; Pedrosa-Soares et al., 2001, 2006; Vauchez et al., 2007).

The Araçuaí orogen records four evolutionary stages whose timing is constrained by U–Pb ages of the pre-collisional (ca. 630–585 Ma), syn-collisional (ca. 585–560 Ma), late collisional (ca. 560–530 Ma) and post-collisional (ca. 530–490 Ma) magmatic episodes (Pedrosa-Soares et al., 2001, 2008; Silva et al., 2005).

The pre-collisional stage (ca. 630–585 Ma) is mainly represented by the I-type G1 plutonic suite. This suite represents the roots of a magmatic arc formed in a continental active margin setting (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Nalini-Junior et al., 2000b; Pedrosa-Soares et al., 2001, 2008). It consists of deformed batholiths and stocks composed of tonalite, granodiorite and minor diorite, with dioritic to mafic enclaves. Chemical compositions from many samples of several G1 plutons outline a consistent metaluminous, calc-alkaline, volcanic arc signature (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Nalini-Junior et al., 2000b, 2005; Pedrosa-Soares et al., 2001; Campos et al., 2004; Martins et al., 2004b). U–Pb zircon and monazite ages constrain the evolution of the G1 suite to between ca. 630 and ca. 585 Ma (Noce et al., 2000; Nalini-Junior et al., 2000b; Whittington et al., 2001; Silva et al., 2005). The location of this magmatic arc relative to the zone with oceanic slivers suggests subduction to the east and a suture zone roughly located along the 42°W meridian, south of 17°S (Pedrosa-Soares et al., 1998, 2001, 2008).

The forearc-and-arc rock assemblage includes orogenic units that represent the volcano-sedimentary pile of the magmatic arc and/or contain sediments supplied by it (e.g. Rio Doce Group; Vieira, 2007), together with strongly deformed tonalitic bodies, thrust slices of basement rocks and probable passive margin deposits (Pedrosa-Soares et al., 2008).

The Nova Venécia Complex comprises pelite-rich sediments deposited in the back-arc basin of the Araçuaí orogen (Noce et al., 2004). This complex contains migmatized, high-grade paragneisses rich in biotite, garnet, cordierite and/or sillimanite, with lenses of calc-silicate rocks (Pedrosa-Soares et al., 2006, and references therein). U–Pb ages for detrital zircons constrain the maximum age of sedimentation to around 608 ± 18 Ma (revised age by Pedrosa-Soares et al., 2008, after data from Noce et al., 2004). High-grade metamorphism and partial melting are syn-kinematic with the regional foliation and took place at around 580–570 Ma, but the complex was also involved in late orogenic episodes of granite genesis (Noce et al., 2004; Munhá et al., 2005; Pedrosa-Soares et al., 2006).

To the northwest of the magmatic arc is found the Salinas Formation (formerly included in the Macaúbas Group; e.g. Pedrosa-Soares et al., 2001; Lima et al., 2002), a metamorphosed greywacke–pelite succession with intercalations of clast-supported conglomerates locally rich in clasts of volcanic rocks, and calc-silicate (meta-marl) lenses, named. U–Pb data for detrital zircons from greywacke samples suggest a maximum sedimentation age of 588 ± 24 Ma for the Salinas Formation (revised age by Pedrosa-Soares et al., 2008, after data from Lima et al., 2002). Furthermore, zircons from cobbles of felsic volcanic rocks of the Salinas conglomerates yielded U–Pb magmatic ages around 630–600 Ma (Pedrosa-Soares et al., unpublished data). As the magmatic arc of the Araçuaí orogen is the most likely source for these volcanic clasts, as well as for the ca. 588 Ma detrital zircons, the Salinas Formation is considered to be an orogenic sedimentary pile (Lima et al., 2002; Santos, 2007; Pedrosa-Soares et al., 2008). The Salinas Formation records the regional deformation of the Araçuaí orogen, but in low-strain zones, well-preserved sedimentary features, such as convoluted bedding and intraformational breccias suggesting sedimentation in a tectonically active marine environment (Lima et al., 2002; Santos, 2007).

The Capelinha Formation is a metamorphosed sandstone succession, with intercalations of pelite and detrital iron formation and is also interpreted to be sourced from the adjacent orogen (Queiroga et al., 2006; Pedrosa-Soares et al., 2008).

The G2 suite represents the syn-collisional stage (ca. 585–560 Ma). It mainly consists of foliated, S-type, biotite $\pm$ garnet $\pm$ cordierite $\pm$ sillimanite granites and two-mica granites, with common mylonitic features. Xenoliths and roof pendants of metasedimentary rocks in several stages of assimilation and of variable size are abundant (Celino et al., 2000; Nalini-Junior et al., 2000a,b; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Pedrosa-Soares et al., 2001, 2006; Campos et al., 2004). U–Pb ages suggest that the main episode of G2 granite generation took place at around 575–560 Ma (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Campos et al., 2004; Silva et al., 2005; Pedrosa-Soares et al., 2006), although some G2 granites crystallised at around 582 Ma (Nalini-Junior et al., 2000a). Large batholiths dominated by mylonitic garnet–biotite granites reveal well-preserved magmatic features with similar magmatic crystallisation ages (Pedrosa-Soares et al., 2006; Vaucher et al., 2007).

The late collisional stage apparently lasted from ca. 560 to 530 Ma, but these age limits are poorly constrained geochronologically. The S-type G3 suite is late to post-collisional. It consists of garnet- and/or cordierite-rich leucogranite, which generally occur as veins and small intrusions cutting G2 granites. Zircon and monazite U–Pb data from G3 leucogranite samples have yielded ages from ca. 550 to 520 Ma (Whittington et al., 2001; Campos et al., 2004; Silva et al., 2005; Pedrosa-Soares et al., 2006).

The post-collisional stage (530–490 Ma) is related to the gravitational collapse of the Araçuaí orogen (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Pedrosa-Soares et al., 2001). G4 and G5 are plutonic suites associated with this stage and both lack the regional foliation (Figure 3.13). The G4 suite includes relatively shallow (5–15 km), S-type granitic intrusions with variable contents of biotite, muscovite and/or garnet. They have circular surface outcrops and sometimes form clusters of amalgamated plutons with locally preserved pegmatoidal roofs (Pedrosa-Soares and Wiedemann-Leonardos, 2000b, and references therein). The G5 suite predominantly consists of I-type granitic intrusions, which may include charnockitic, enderbitic and/or mafic portions, and minor mafic bodies with subordinate granitic and/or charnockitic facies. Magma mingling and mixing features are very common. The granitic rocks are generally porphyritic to subporphyritic and have a high-K calc-alkaline signature (Wiedemann et al., 2002; Campos et al., 2004; Martins et al., 2004b).

The evolution of the precursor basin and orogenic stages of the Araçuaí orogen are presented in Table 3.2. This orogenic belt defines the southeastern limit for the São Francisco Craton, but it does not represent a common plate margin orogen. The Araçuaí orogen (together with its counterpart located in Africa, the West-Congo belt) is better characterised as a confined orogen because it developed inside an embayment (like a large gulf or a Red Sea-type basin) carved into the São Francisco–Congo palaeocontinent. Its precursor basin was not completely ensialic but was filled by continental rift (including glaciogenic) and passive margin sediments. The subsequent orogeny emplaced slivers of ophiolite and a pre-collisional magmatic arc, evidence that rifting proceeded to ocean-floor production (Pedrosa-Soares et al., 2001, 2008).

Regional structural studies provide explanations for the tectonic mechanisms underlying the different stages of the Araçuaí orogen (Uhlein, 2004; Alkmim et al., 2006b). The so-called “nutcracker tectonics” model explains this orogen as a series of kinematically distinct deformation stages: (1) early north-verging deformation; (2) development of the west-verging fold-thrust belt against the São Francisco Craton and an east-vergent zone in the back-arc region; (3) lateral escape to the south along dextral strike–slip faults; (4) late-stage gravitational collapse with development of extensional sites and intrusion of post-collisional plutons (Alkmim et al., 2006b).

3.2.2.3. The Rio Pardo Basin

The Rio Pardo Group is a Neoproterozoic–Early Palaeozoic sedimentary fill of the Rio Pardo basin in the southeastern portion of the São Francisco Craton (Figure 3.14). The first reference to rocks of this group was

Table 3.2 Evolution of the precursor basin and orogenic stages of the Araçuaí orogen, southeastern Brazil (age references in text).

Age	Stage; environment and/or process	Stratigraphic unit and/or structural feature
< 900 Ma	Early continental rift; pre-glacial, fluvial	Diamictite-free, lowermost proximal Macaúbas Group
ca. 875 Ma	Continental rift; pre-glacial, fluvial to marine	Salto da Divisa anorogenic suite; Diamictite-free proximal Macaúbas Group
< 875 Ma	Late continental rift; mainly gravitational glacio-marine sedimentation and transitional mafic volcanism	Diamictite-rich and diamictite-turbidite sequences with transitional basalts of the proximal Macaúbas Group
< 864 Ma	Proximal passive margin; continental shelf, marine transgression	Diamictite-free, sand-pelite unit of the uppermost proximal Macaúbas Group
ca. 816 Ma to ?	Distal passive margin and ocean spreading; continental slope to ocean floor	Ribeirão da Folha Formation, Dom Silvério Group and mafic-ultramafic oceanic rocks (ophiolite slivers)
> 630 Ma	Start of oceanic lithosphere subduction from west to east	Nucleation of north-verging structures (cf. nutcracker tectonics model; see text)
630–585 Ma	Pre-collisional; continental magmatic arc and related basins	G1 suite, Rio Doce Group, Nova Venécia Complex and Salinas Formation
585–560 Ma	Syn-collisional; fold-and-thrust tectonics, syn-kinematic metamorphism and partial melting	G2 suite and migmatization of other units; west- and east-verging thrust-and-folding with low-angle dip shear zones
560–500 Ma	Late- to post-collisional; adiabatic post-kinematic S-type anatexis, partially coeval to lateral escape to the south	G3 and G4 suites; NNE to NE-trending, high-angle dip, dextral, strike-slip shear zones controlled the lateral escape
520–490 Ma	Post-collisional; gravitational collapse, rising of mantelic magma and partial melting of deep crust	G5 suite; extensional sites along NE and NW trends of lineaments

made in 1870 (Hartt, 1941) when conglomeratic schist and fine-grained sandstones were described and the impression of a plant resembling *Asterophyllites scutigera* Dawson (previously known from the Upper Devonian of St. John, New Brunswick, Canada) was found. The Baixo Rio Pardo rocks were thus assigned a Lower Palaeozoic age. The interval for deposition of rocks of this group is still poorly constrained; Karmann et al. (1989) suggested a depositional range from 1,100 to 600 Ma.

Until the middle 1960s, studies of this basin were at the reconnaissance level, and the stratigraphy was subdivided into the Rio Pardo and Salobro formations (Oliveira and Leonardos, 1940; Almeida, 1968). Later, the low-grade metasedimentary rocks of the basin were divided into six formations, formalised by Pedreira et al. (1969): the Panelinha, Camacan, Salobro, Água Preta, Serra do Paraíso and Santa Maria formations, with the Rio Pardo Formation, previously used by Almeida (1968), elevated to group status. The “layer cake” stratigraphy was somewhat modified in subsequent studies (Andrade and Nunes, 1974; Siqueira et al., 1978) that placed the Salobro Formation at the top of the group and introduced other minor changes. Karmann et al. (1989) once again revised the stratigraphy and the interpretation of the basin’s tectonic evolution, from an aborted-rift, giving way to a thermally subsiding basin, followed by a foreland basin.

A stratigraphic chart for the Rio Pardo Basin and its basement is shown in Figure 3.15. The first column shows the ages (palaeontological or geochronological) of the distinct units; the second, third and fourth columns show the relationships among the lithostratigraphic units (group, subgroup and formations); in the central panel the same relationships are in a two-dimensional display, in a north–south section and, in the last two columns, the possible tectonic evolution and depositional environments for the units.

Rocks of the Rio Pardo Basin were deposited on Archaean basement (Moraes Filho and Lima, 2007) that consists of charnockites, enderbites and granulitic trondhjemitic, as well as enderbitic-trondhjemitic orthogneisses and meta-gabbro-norite (Ibicaraí Complex). These rocks were intruded by rocks of the Pau Brasil suite, which include charnockites, monzo-diorites, granites, quartz-monzonites and tonalities, the Anuri Syenite, and diabase dykes (Souto et al., 1972). The Itapetinga Complex is composed of syeno-granitic to tonalitic orthogneisses, intruded by anorogenic granitoids that crop out in the southern region of the basin.

The basal unit of the Rio Pardo Group, the Panelinha Formation, is observed as discontinuous outcrops in the northern, northwestern and western borders of the basin and its lower contact is non-conformable on the

Figure 3.14 Simplified geological map of the Rio Pardo basin, eastern São Francisco Craton (modified from Pedreira, 1999).

Figure 3.15 Stratigraphic chart for the Rio Pardo Group (modified from Pedreira, 1999).

crystalline basement. Its upper contact is with sedimentary rocks of the Camacan, Água Preta, Serra do Paraíso and Salobro formations (Figure 3.14). The Panelinha Formation is about 200 m thick (Karmann et al., 1989) and its lower portion is composed of immature breccias with coarse-grained matrix, superposed by polymictic metaconglomerates with pebbles of basement rocks, including syenite gneisses (Pedreira et al., 1969). These coarse clastics grade upwards into massive and laminated metagreywackes with conglomerate intercalations and disseminated pyrite, followed by metarkoses with conglomeratic intercalations; the metarkoses show parallel lamination, cross-bedding and graded bedding.

The Itaimbé Subgroup (Figure 3.14) was proposed by Karmann et al. (1989) to encompass the Camacan, Água Preta, Santa Maria Eterna and Serra do Paraíso Formations, whose contacts are lateral and separated by possible unconformity from the Panelinha Formation and unconformably overlain by the Salobro Formation.

The Camacan Formation (Pedreira et al., 1969) crops out in the northern border of the basin, along a belt between the towns of Camacan and Santa Luzia, and has a thickness of 200 m. The northern limit of the belt is the São Pedro River and the southern one is found in the Chororão, Rochedo and Lapão ranges, in the domain of the Salobro Formation (Figure 3.14). The lower and upper contacts of the Camacan Formation are unconformable, respectively, with the Panelinha and Salobro formations. The Camacan Formation is composed of metasilstones and slates, argillaceous metalimestones and dark-grey dolomites, pyrite-bearing metagreywackes and fine-grained sandstones. The base of the formation consists of limestones and dolostones with intraclasts, followed by a thick interval where fine-grained sandstones alternate with argillite, shale and slate. Approximately in the middle part of this formation, a bed of greywacke separates this lower succession from an upper one of similar composition. Costa Pinto (1977) suggested a transition from continental environment, as evidenced by evaporation of the base of this formation, to open and then restricted marine, during which time carbonates were deposited.

The Água Preta Formation (Pedreira et al., 1969) crops out in the central part of the basin, in an area limited by the town of Pau Brasil and the villages of Gurupá-Mirim, Itaimbé and Teixeira do Progresso, as well as in erosional windows, east and northeast of Teixeira do Progresso village. The northeastern contact of the Água Preta Formation with the Salobro Formation is represented by a contractional fault (Rio Pardo–Água Preta Fault; Karmann et al., 1989). The contact with the carbonate facies of the Serra do Paraíso Formation is interfingered. The minimum thickness of the Água Preta Formation is of about 2,300 m.

In the Água Preta Formation, laminated phyllites and metasilstones predominate, followed by micaceous and calcareous sandstones; subordinate argillaceous and dolomitic metalimestones and metadolostones also occur. This formation contains a few lenses of carbonate rocks southeast of the town of Pau Brasil. Costa Pinto (1977) analysed C and O isotopes in one of these lenses and found isotopic signatures similar to those in the Serra do Paraíso Formation.

The Santa Maria Eterna Formation (Lima et al., 1981) consists of massive dolomitic marbles with intercalations of grey metadolostone and micaceous quartzite. It crops out in the southeastern portion of the basin and is mostly overlain by the Neogene Barreiras Formation. It is in fault contact with the Água Preta Formation and interfingers with the Serra do Paraíso Formation (Karmann et al., 1989). Owing to the lack of continuous outcrops, its thickness has never been estimated.

The Serra do Paraíso Formation (Pedreira et al., 1969) makes up the southern and western borders of the basin. It is bound below by the basement and the Água Preta Formation. The former contact is generally in the form of reverse faults while the latter is gradational and interfingered. In the type-section, south of Itaimbé village, the formation is 1,575 m thick. North of the town of Pau Brasil, this formation consists of metadolomites and calcareous metadolomites. Tepee structures occur close to its base, indicating subaerial exposure. On the road between Pau Brasil and the Pardo River, it consists of micaceous metalimestones, metasilstones and phyllites. In the region of the Serra do Paraíso–Gurupá Mirim, it is predominantly sand-carbonate metarhytmities, metalimestones and metarenites, while along the BR-101 road, it begins with calc-schists at the base, followed by alternating metalimestones, metadolomites and quartzites.

The Salobro Formation crops out in the northeastern sector of the Rio Pardo Basin, in an area of high relief formed by ranges such as the Chororão, Rochedo, Pacuípe, and Lapão. Its contact to the southwest with Água Preta Formation is a reverse fault, with tectonic transport to northeast (Rio Pardo–Água Preta Fault). The northern contact is a significant unconformity with the Camacan Formation. To northeast, the Salobro Formation is covered by the Barreiras Formation. Karmann et al. (1989) determined a thickness of about 5,000 m.

The Salobro Formation consists of immature metarenites, fine- and coarse-grained greywackes, metarkoses, metasilstones, slates and metaconglomerates (both clast- and matrix-supported) rhythmically interstratified. Conglomerate lenses occur throughout this formation, the thickest being the Lapão conglomerate, which crops out in the northeastern sector of the formation, east of the town of Santa Luzia, and was formerly washed for diamonds. It is poorly sorted, with clasts whose diameter ranges from 1 to 50 cm; the clasts are from the basement (granulites, syenite gneisses), chert, metadolomite and metalimestone. The other conglomerate lenses have similar

composition, with clasts of phyllites and fine-grained quartzites. The Serra do Lapão conglomerate, in the lower Salobro Formation, occurs south of the town of Santa Luzia and comprises metadolostone pebbles with a calcarenite matrix.

The source for the Salobro Formation sediments was represented by two areas of high topographic relief (granulites and syenites from the north and sedimentary rocks from the south). The burial must have been fast, as indicated by the immaturity of the sediments. The interfingering facies of these conglomerates occurred on tectonic ramps, as demonstrated elsewhere by Steidtmann and Schmitt (1988).

The sedimentation in the Rio Pardo Basin started with the deposition of alluvial fans in graben-type basins, limited by extensional faults. The alluvial fans were formed by the conglomerates and breccias of the Panelinha Formation (Figure 3.14). Distally, these fans grade to fluvial systems (Figure 3.15) with abandoned channels and swampy areas evidenced by horizons of conglomerate intercalated with pyritic greywackes in the upper part of the formation.

Rifting ceased before development of oceanic crust, and the Camacan, Água Preta, Serra do Paraíso and Santa Maria Eterna formations (Itaimbé Subgroup) were deposited on the Panelinha Formation under marine conditions during the subsequent thermal subsidence phase. In the deeper parts of the basin, turbidites of the Água Preta Formation were deposited. Carbonate lenses (Mutuns creek, Nancy, Gurupá Mirim and others) in gradational contact with these turbidites were probably deposited in shallower environments.

The Água Preta Formation grades into Camacan, Serra do Paraíso and Santa Maria Eterna formations towards the shallower parts of the basin. Evidence for shallow deposition is found in all of these units: ripple marks and features that indicate subaerial exposure, such as intraclasts and mud cracks, occur in the Camacan Formation; intraclasts and tepees are found in the Santa Maria Eterna Formation; and stromatolites at the base of the Serra do Paraíso Formation indicate deposition under inter- to subtidal conditions. These formations can be interpreted, respectively, to have been deposited in muddy tidal flat, littoral and carbonate platform environments (Figure 3.15). The balance between the siliciclastics influx and the shape of the coastline can account for the distinct facies within these formations. Facies variability is most pronounced in the Serra do Paraíso Formation, whose depositional environment grades from outer platform to supratidal. Delgado et al. (2003) interpreted these formations as deposited on a passive margin during a retrograding depositional cycle.

After lithification of the limestones, sands and clays, tectonic reactivation uplifted both the basement north of the basin and the sedimentary rocks south of the Rio Pardo–Água Preta Fault. Both are source areas for the sediments of the Salobro Formation, as indicated by the pebble composition of conglomerates in the unit.

This brief description of the Rio Pardo Group demonstrates that the sedimentary environment and tectonic setting during deposition of the Rio Pardo Group sediments evolved from an intracontinental rift through a phase of thermal subsidence and into a foreland basin.

3.2.2.4. The Sergipano belt

The Sergipano is a Brasiliano/Pan-African fold-and-thrust belt (Figures 3.11–3.13) located on the northeastern edge of the São Francisco Craton. The structure and lithology of this belt were compared to the Ndjolé Series of northern Gabon by Allard and Hurst (1969). Later, Cordani (1973) proposed a comparison with the sequences of Mbalmayo-Bengbis, Dja and Sembe Ouesso of southern Cameroon and the extreme north of the Congo. According to Trompette (1994), the Sergipano belt represents the Brazilian continuation of the Oubanguides and, together, they form a roughly E-W elongated mega-orogen more than 5,000 km long. More recently, the Sergipano belt has been correlated with the Yaoundé belt (Cameroon, Africa) by Oliveira et al. (2006) based on palaeocontinent reconstructions, lateral variation of rock units and tectonic histories.

This belt was formed by continental collision between the Congo–São Francisco Craton and the Pernambuco–Alagoas massif during the Brasiliano/Pan-African orogeny (Brito Neves et al., 1977). Initially, it was interpreted as a geosyncline (Humphrey and Allard, 1968; Silva Filho and Brito Neves, 1979) and later as a collage of lithostratigraphic domains (Davison and Santos, 1989; Silva Filho, 1998) or as a Neoproterozoic fold-and-thrust belt produced by inversion of a passive margin basin located at the northeastern edge of the ancient São Francisco plate (D'el Rey Silva, 1999).

The São Francisco Craton, to the south of the Sergipano belt, is a granite-greenstone belt sequence with high-grade terranes with ages ranging from 3,400 to 2,080 Ma (Oliveira et al., 2006). The Pernambuco–Alagoas massif, to the north of Sergipano belt, comprises Palaeoproterozoic to Mesoproterozoic high-grade gneisses and migmatites intruded by Neoproterozoic granitoids (Brito Neves et al., 1982; Silva Filho et al., 2002). The Sergipano belt consists, from north to south, of six lithostratigraphic domains (Figures 3.16–3.18) separated by major shear zones: Canindé, Poço Redondo, Marancó, Macururé, Vaza Barris and Estância (Santos and Souza, 1988; Davison and Santos, 1989; Silva Filho, 1998). The Macururé, Vaza Barris and Estância are composed mostly of metasedimentary rocks with metamorphic grade, varying from unmetamorphosed in the Estância domain, through greenschist grade in the Vaza Barris, to amphibolite facies in the Macururé domain.

Figure 3.16 Domains of the Sergipano belt (modified from D'el Rey Silva, 1995). MSZ, BMJSZ, SMASZ and ISZ represent, respectively, the Macururé, Belo Monte-Jeremoabo, São Miguel do Aleixo and Itaporanga shear zones.

Figure 3.17 Simplified geological map of the eastern portion of the Sergipano belt (modified from D'el Rey Silva, 1999).

Figure 3.18 Simplified geological map of the western portion of the Sergipano belt (modified from Jordan, 1973).

Silva Filho and Torres (2002) and Silva Filho et al. (2003) suggested three additional domains: Rio Coruripe, Viçosa and Pernambuco-Alagoas. It is possible that the Rio Coruripe domain, metamorphosed to granulite facies and later retrograded to amphibolite and greenschist grade, is a high-grade counterpart of the Macururé domain, judging from the lateral continuity of similar rock types (Oliveira et al., 2006). All these domains, except for the Estância and Vaza Barris, were intruded by Brasiliano/Pan-African granitoids. Additional descriptions of all of the domains in the Sergipano can be found in Davison and Santos (1989) and Santos et al. (1998), and a summary on the geochronological data, including recent geochronology, is found in Oliveira et al. (2006).

The Canindé domain (Figure 3.16) comprises (a) an elongated pink granite sheet (Garrote unit); (b) a metavolcano-sedimentary sequence (Novo Gosto unit) represented by fine-grained amphibolite, marble, graphite schist, micaschist and metagreywacke; (c) subvolcanic microgabbro-quartz diorite complex (Gentileza unit); (d) the Canindé gabbro-leucogabbro complex (gabbro, gabbro-norite, peridotite and pegmatitic gabbro) and (e) granitic plutons (tonalites, granodiorites and quartz-syenites). The ages of the rocks in this domain range from about 715 to 609 Ma (Silva Filho et al., 1997; Nascimento et al., 2005; Oliveira et al., 2006).

The Poço Redondo domain (Figure 3.16) is composed of biotite gneisses, migmatites and granitic intrusions, such as the Serra Negra augen-gneisses, the Sítios Novos granodiorite and leucogranite sheets. Grey gneisses from the palaeosome of migmatites yielded ages slightly older (960–980 Ma U-Pb SHRIMP ages; Carvalho et al., 2005) than the Serra Negra augen-gneisses (952 Ma U-Pb SHRIMP age; Carvalho et al., 2005). The Sítios Novos granodiorite has been dated at 651 ± 11 Ma (U-Pb TIMS; Carvalho et al., 2005). The Maracó domain is mainly formed of a metavolcano-sedimentary sequence (quartzite, conglomerate, micaschist, phyllites and lenses of andesite, dacite and quartz porphyry) with peridotite and amphibolite lenses (Figure 3.16). The presence of andesites and dacites, dated at about 603 Ma (Carvalho et al., 2005) and conformably interleaved with phyllites, and of granodiorite stocks dated at 595 ± 11 Ma (Silva Filho et al., 1997), points to an Ediacaran

Figure 3.19 Diamictite of the Juetê Formation, overlain by carbonates of the Acauã Formation, at Serra da Borracha, about 20 km west from Patamutê village, state of Bahia.

depositional age. The Macururé domain consists mainly of garnet micaschists and phyllites (Figure 3.16) with some quartzite and marble. These metasedimentary rocks were intruded by granitic plutons dated at about 625–630 Ma (Bueno et al., 2005; Long et al., 2005).

The Vaza Barris domain (Figure 3.16) shows two cycles of sedimentation, both with a continental to shallow-marine, basal siliciclastic megasequence, overlain by a carbonate sequence (D’el Rey Silva, 1995, 1999). The lower siliciclastic megasequence is represented by the Juetê (sandstone, diamictite; Figure 3.19), Ribeirópolis (silty phyllites, metagreywackes, pebbly phyllites, diamictites; Figure 3.20) and Itabaiana (conglomerate, quartzite, metasilite) formations. The lower carbonate megasequence is known as the Jacoca Formation, which is stratigraphically equivalent to the Acauã Formation. These two megasequences form the Estância-Miaba Group. The upper siliciclastic megasequence (Simão Dias Group) is represented by the Lagarto-Palmares, the Jacaré and the Frei Paulo formations and is overlain by the Vaza Barris Group that comprises diamictites of the Palestina Formation and the upper carbonate megasequence, the Olhos D’Água Formation. All of these rocks spread continuously across the craton margin into the Sergipano belt where they occur around the Itabaiana and Simão Dias basement domes and are overlain by a metadiamictite and metacarbonate belonging to the upper megasequence. According to D’el Rey Silva (1999), both the basement and cover experience the same Neoproterozoic compressive deformation under subgreenschist facies.

The thick Olhos D’Água Formation is composed of a sequence of marbles and interbedded green, calcareous chlorite schists and silty phyllites; it overlies diamictites and pebbly metagreywackes of the Palestina Formation

Figure 3.20 Diamictite of the Ribeirópolis Formation in sharp contact with dolostones of the Jacoca Formation, Capitão Farm, state of Sergipe. Dolostone shows wavy contact and some bedding is observed in the diamictite. Some clasts seem to be concentrated and salient at the top of diamictite, which suggests lithification and erosion prior to deposition of the dolostone.

Figure 3.21 Pebbly metagreywacke of the Palestina Formation, sometimes with dropstones (about 12 km from Rosario village, state of Bahia).

(Figure 3.21). Marble beds are interbedded with blue to black, fine-grained metalimestone and grey metadolostone. The thick carbonates around the Simão Dias dome (Figure 3.17) pass upward into supertidal–intertidal facies with oolites, and wave-reworked structures indicating a near-shore environment (D’el Rey Silva, 1995). The Jacoca Formation (older) and Olhos D’Água Formation (younger) carbonates have been interpreted as cap carbonates by Sial et al. (2000a, 2006).

An acidic tuff within the Ribeirópolis diamictites in an outcrop on the Ribeirópolis-Gloria road, 7 km from Ribeirópolis town, was dated at around 730 Ma (zircon, U–Pb; B.B. Brito Neves, written communication), suggesting that the deposition of the overlying Jacoca carbonates happened right after the Sturtian glaciation event. The youngest U–Pb age of zircon grains (615 Ma) from the Frei Paulo metagreywackes above the Olhos D’Água carbonates constrains the deposition of Jacoca and Olhos D’Água to the interval 730–615 Ma.

The Estância domain (Figure 3.16) comprises, from base to top, sandstones and argillites of the Juetê Formation, dolostones and limestones of the Acauã Formation, sandstones and conglomerate lenses of the Largato Formation, and sandstone and minor conglomerate lenses of the Palmares Formation (Silva Filho and Brito Neves,

1979). Although D'el Rey-Silva (1999) assumed that deposition in the Estância and Vaza Barris domains occurred during two cycles of sedimentation on a passive continental margin of the ancient São Francisco plate (detritus sources from the São Francisco Craton), Oliveira et al. (2006) proposed instead deposition in foreland basins, based on detrital zircon populations with ages between 570 and 657 Ma and an inferred provenance from the Sergipano belt and Borborema province to the north.

3.2.2.5. The Riacho do Pontal and Rio Preto belts

The Riacho do Pontal and Rio Preto belts border the northern portion of the São Francisco Craton (Figure 3.1) and are part of the southern portion of the Borborema Province (Brito Neves, 1983; Brito Neves et al., 2000). The Riacho do Pontal belt consists of the southern frontal zone of the Borborema Province that overthrusts directly onto the basement of the São Francisco Craton and represents a subautochthonous metasedimentary sequence (Jardim de Sá et al., 1992; Oliveira, 1998a). The Rio Preto belt is located at the northwestern border of the São Francisco Craton and is interpreted as an intracontinental belt (Egydio-Silva, 1987; Egydio-Silva et al., 1990; Andrade Filho et al., 1994; Trompette, 1994).

The Riacho do Pontal belt stretches from the southern region of the states of Piauí and Pernambuco to the northern region of the state of Bahia. This belt can be divided into four zones or structural domains, from south to north: (a) cratonic domain, (b) southern zone, (c) central zone and (d) internal or northern zone (Figure 3.22). The cratonic domain consists of biotite-hornblende gneisses, migmatites, gneissic batholiths of granodioritic-tonalitic composition and of patches of metasedimentary rocks. The rocks in this domain are Archaean to Palaeoproterozoic in age, but affected by the Pan-African/Brasiliano cycle. The boundary between the Riacho do Pontal belt and the São Francisco Craton is somewhat unclear in this domain. A southward thrust overriding the craton was part of a Pan-African/Brasiliano tectonic event that also deformed carbonates of the Una Group (Alkmim et al., 1993; Brito Neves, 1983).

The southern zone is represented by a thrust-fold belt with southern vergence onto the São Francisco Craton (Angelim, 1988; Gomes, 1990; Jardim de Sá et al., 1992; Oliveira, 1998a,b). The rocks that form the Riacho do Pontal belt are variably referred as the Salgueiro or Casa Nova groups, depending on the location. These rocks are terrigenous-shelf metasedimentary rocks represented by garnet micaschists, quartzites, marbles and gneisses. Deep-sea, flysch-like metagreywackes with carbone and/or feldspathic clasts are regarded as turbidites. Syn-tectonic granitoids are interwoven into the metasedimentary rocks, as are foliated sheets of two-mica monzogranitic tonalites, known as Rajada orthogneisses. The syn-tectonic Serra da Boa Esperança granite and quartz-syenite yielded an age of 555 ± 10 Ma (Rb-Sr, Pb-Pb; Jardim de Sá et al., 1996). The WSW-ESE foliation shows NNW-SSE stretching lineation and sheath folds with axes subparallel to that lineation and resulted from the same event that has generated the nappes. The thrust deformation induced a decollement of the metasedimentary sequence over the cratonic massif (Jardim de Sá et al., 1992).

The central zone consists of metabasalts, amphibolites, chlorite-, talc-, tremolite- and actinolite-bearing rocks regarded as ultramafic intercalations in schists, and metachert. This zone contains the Monte Orebe and Brejo Seco mafic-ultramafic massifs with metasedimentary intercalations (Marimon, 1990; Moraes, 1992). The mafic rocks are Neoproterozoic tholeiitic ocean-floor basalts (Moraes, 1992), believed to be an ophiolitic complex.

Figure 3.22 Cross-section for the Riacho do Pontal belt that can be divided into four zones or structural domains, from south to north: (a) cratonic domain, (b) southern zone, (c) central zone and (d) internal or northern zone (Uhlein et al., 2008).

Figure 3.23 Cross-section for the Rio Preto belt (Egydio-Silva, 1987).

The internal zone is bounded by ductile right-lateral shear zone and corresponds to a tectonically complex region where subvertical mylonitic foliation shows recumbent S structures with right-lateral movements. This zone consists of remnants of Brasiliano metasedimentary sequences, intrusive granitoids and slices from the basement that form stretched amygdaloidal bodies. The mylonitic zone is up to 1 km thick with proto-mylonites through ultramylonites (Vauchez and Egydio-Silva, 1992; Vauchez et al., 1995).

The Rio Preto belt is considered a marginal fold belt forming a westward extension of the Riacho do Pontal and Sergipano belts. This belt is of great structural and lithostratigraphic interest because it displays a transition between cratonic facies, in the south, to metasedimentary rocks at the central part, and gneisses of the basement towards the northern border. The metasedimentary sequences are equivalent to the São Francisco Supergroup in the state of Bahia.

From south to north, the Rio Preto belt can be subdivided into three structural units, each with distinct lithologic, structural and metamorphic characteristics, separated by disrupted structures (Figure 3.23). In the southern unit, in the portion of the São Francisco Craton and adjacent to this belt (the São Desidério region), the sequence begins with microcrystalline limestone with argillaceous intercalations, characterised by 50 m-thick horizontal layers, and shows no signs of metamorphism or deformation (cratonic domain). This unit was named the São Desidério Formation (Egydio-Silva, 1987; Egydio-Silva et al., 1989) and probably correlates with the Sete Lagoas Formation of the Bambuí Group in Minas Gerais (Figure 3.24).

To the north, cratonward from the fold belt, these limestones are overlain by clastic rocks with intercalations of marls and limestones. This sequence (Serra da Mamona Formation) has been correlated with the Santa Helena Formation (Egydio-Silva, 1987; Egydio-Silva et al., 1989). North of Barreiras, they are overlain by meta-arkoses, fine- to medium-grained quartzites, locally feldspathic, metagreywackes and rare carbonate intercalations. This sequence (Riachão das Neves Formation) is considered partially equivalent to the Três Marias Formation of the Bambuí Group (Egydio-Silva, 1987; Trompette, 1994).

In the central portion of the belt, the lower part of the succession is more complete with the occurrence of boulder diamictites, with clasts up to 50 cm in diameter of gneiss, quartzite, metasiltstone, marble and schist. Referred here as the Canabrinha Formation (Figure 3.25), this unit has been correlated to the Bebedouro Formation, although a glaciogenic origin remains unsubstantiated. Further north, in the region of Formosa do Rio Preto (Figure 3.23), the Rio Preto Group contains predominantly micaceous quartzites and schists with a few intercalations of amphibolites and itabirites. It has been correlated with the Chapada Diamantina Group (Inda and Barbosa, 1978).

The general structure of the Rio Preto belt is of an asymmetric fan, probably associated with late ENE-WSW shearing that also occurred in the Borborema Province. Structural analysis of this belt defined three structural units (Egydio-Silva, 1987; Egydio-Silva et al., 1990; Andrade Filho et al., 1994). The central and southern ones, well developed between the craton and the Cariparé thrust, show a single deformation phase with southern vergence

Figure 3.24 Limestones of the São Desidério Formation near São Desidério, Rio Preto belt, probably correlated with the Sete Lagoas Formation of the Bambuí Group.

Figure 3.25 Diamictites of the Canabrinha Formation, Rio Preto belt.

towards the São Francisco Craton. The northern unit is metamorphosed to the greenschist facies, is polyphase and is thrust northwards over the Piauí basement. The absence of granitic intrusions and rare volcanism favour an ensialic origin for this belt (Egydio-Silva, 1987; Trompette, 1994).

The Riacho do Pontal belt appears to have participated in the collision between the São Francisco Craton and the Pernambuco–Alagoas massif. The basin can be schematically reconstructed with terrigenous-shelf sediments and deep-sea sediments (turbidites) in a passive margin context. Mafic–ultramafic massifs (Monte Orebe and Brejo Seco) may be remnants of a narrow oceanic crust-floored basin (Oliveira, 1998a,b). The Brasiliano collisional deformation is complex, with thrusts towards the south over the São Francisco Craton associated with greenschist to amphibolite metamorphism (Jardim de Sá et al., 1992). In the internal zone, a continental-scale subvertical dextral ductile shear zone is the main structural feature.

The Rio Preto belt is a good example of an intracontinental fold belt whose central structure seems to have been controlled by the geometry of a Mesoproterozoic trough and by the intervention of late shears. The sedimentary basin, as reconstructed, is composite, with a northern trough of Mesoproterozoic age filled by the Rio Preto Group and juxtaposed against a younger subsiding southern basin. In this younger basin, a mostly clastic and thickened equivalent of the late Proterozoic Una Supergroup was deposited. The migration of the depocenter of the basins towards the south implies the intervention, before deposition of the Bebedouro Group, of

epirogeny uplifting northern highlands against which the Neoproterozoic basin was juxtaposed (Egydio-Silva, 1987; Trompette, 1994; Andrade Filho et al., 1994).

3.3. ISOTOPE CHEMOSTRATIGRAPHY

C and O isotope studies were performed across different sections of some sequences in the southwestern portion of the São Francisco Craton (Santos et al., 2000, 2004; Alvarenga et al., 2007). The most complete profile is at the Serra de São Domingos region, where over 700 m of Bambuí Group sediments was deposited atop siliciclastic rocks of the Paranoá Group. These two groups are separated by a karstic limestone breccia infilled with well-sorted sand grains that may be the lateral equivalent of Jequitai Formation diamictites. In the base of this section, dolostones of the upper Paranoá Group have $\delta^{13}\text{C}$ values close to 0‰ (VPDB), similar to values observed in other areas where this unit is exposed. The Paranoá–Bambuí transition is characterised by a spike of low $\delta^{13}\text{C}$ values ($\sim -5\%$), followed by limestones with slightly positive values. The uppermost dolostones of the Conselheiro Mata Group (Espinhaço Supergroup) and dolostones of the Rio Pardo Grande Formation (lowermost portion of the Macaúbas Group) display $\delta^{13}\text{C}$ values around 0‰ (Figure 3.26; Santos et al., 2000), typical of Mesoproterozoic to early Neoproterozoic carbonates. The most remarkable aspect of these profiles is a positive $\delta^{13}\text{C}$ excursion in the upper Sete Lagoas Formation that continues through the Serra de Santa Helena and Lagoa do Jacaré Formation limestones (Figure 3.26). This excursion, identified across different profiles of the northern and southern reaches of the Bambuí basin and used to reconstruct the palaeogeography of the Bambuí basin (Santos et al., 2004), is a robust stratigraphic marker related to a Neoproterozoic positive carbon isotope excursion recognised worldwide.

A sedimentological and C–O isotope study carried out by Vieira et al. (2007) in nine sections of the Sete Lagoas cap carbonate sequence at its classical outcropping area, in the southern tip of the São Francisco Craton, helped refine its stratigraphy. According to these authors, the Sete Lagoas Formation comprises two shallowing-upward megacycles, each cycle bound by a flooding surface and amalgamated with a third-order sequence boundary. The first megacycle is represented by deep-platform deposits with abundant crystal fans (aragonite pseudomorphs) that show negative $\delta^{13}\text{C}$ values (-4.5%) that grade upward to storm wave and tide-influenced layers with values around 0‰. A second megacycle is characterised by a thick, mixed substorm wave-base succession deposited during another transgression that drowned the platform, with lime mudstone–pelite rhythmite that grades to crystalline limestone rich in organic matter with $\delta^{13}\text{C}$ values up to $+14\%$.

Vieira et al. (2007) proposed that the very high $\delta^{13}\text{C}$ values observed in the second megacycle, coupled with geochronological data, support correlation with post-Sturtian sequences. Some differences in the depositional record for the Sete Lagoas and other post-Sturtian units described in North America, Australia and Namibia are perhaps due to the deposition of the Sete Lagoas carbonates in shallower settings, preserving a thick record of storm- and wave-influenced sedimentation, not found elsewhere. Alternatively, these differences may be ascribed to diachronous deposition of the post-Sturtian cap carbonate sequences.

Chemostratigraphic studies of the Una Group have been carried out by Torquato and Misi (1977), Misi and Kyle (1994) and Misi and Veizer (1998). Sulphur, carbon and oxygen isotope analysis, as well as Sr isotope determinations, confirmed the evolution of the sedimentary sequences as proposed by Misi and Souto (1974) and by Misi (1979). In the study by Misi and Veizer (1998), samples for isotope analysis were collected along the eastern border of the Irecê “basin” covering the stratigraphic section of carbonate lithofacies from the laminated limestones of Unit B to the black oolitic limestones of Unit A1. Based on petrographic studies and trace element determinations, the best preserved samples (micrites, with $\text{Mn}/\text{Sr} < 0.09$ and $\text{Sr} > 300$ ppm) were selected for isotope analysis. For the best preserved samples, there is a progressive increase in $\delta^{13}\text{C}$ ($+0.3\%$ to $+8.7\%$), decline in $^{87}\text{Sr}/^{86}\text{Sr}$ (0.70789 – 0.70745) and a relatively uniform $\delta^{18}\text{O}$ signal (-5% to -7.5% VPDB) upsection (Figure 3.27).

Previous C and O isotope studies by Torquato and Misi (1977) of 63 whole-rock carbonate samples from the same stratigraphic section demonstrated the same increase in $\delta^{13}\text{C}$ values in the upper section (Unit A1, ca. $+8\%$) and showed significant negative shifts in the pinkish dolostones of Unit C, in the lower section (around -6% ; $n = 12$). Considering only the best preserved samples and considering the data of Torquato and Misi (1977) from the same section, it is possible to show the variation of the isotopic composition through the Una Group (Figure 3.27).

$^{87}\text{Sr}/^{86}\text{Sr}$ variation, when compared with the seawater evolution curves through the Neoproterozoic (Jacobsen and Kaufman, 1999), suggests an age of sedimentation between 750 and 600 Ma. $\delta^{13}\text{C}$ variation, with a negative shift at the base and a high positive excursion in the upper section, is compatible with that observed in the Bambuí Group (Misi et al., 2007).

A preliminary C and O isotope study on carbonates of the Rio Pardo Group was carried out by Costa Pinto (1977), but as the 120 samples involved lack of stratigraphic control, they are not used in our compilations. Costa

Figure 3.26 C and O isotope profiles for carbonates of the Bambuí Group at Serra de São Domingos, southeastern portion of the São Francisco Craton and for carbonates of the Domingas Formation (Macaúbas Group) and Rio Pardo Grande Formation of the Espinhaço Supergroup (modified from Santos et al., 2004).

Figure 3.27 C and Sr isotope variation of carbonates from the Una Group (modified from Misi and Veizer, 1998).

Pinto (1977) concluded that this sequence was deposited in a continental to restricted marine to open-marine sedimentary environment. Camacan Formation carbonates show $\delta^{13}\text{C}$ values from -5.7‰ to $+4.2\text{‰}$ and $\delta^{18}\text{O}$ values from -10‰ to -5‰ , while carbonates from the Salobro Formation have $\delta^{13}\text{C}$ from -2‰ to $+2\text{‰}$ and $\delta^{18}\text{O}$ from -19.7‰ to -8‰ . Água Preta Formation carbonates display $\delta^{13}\text{C}$ values from -1.3‰ to $+6.2\text{‰}$ and $\delta^{18}\text{O}$ from -11.7‰ to -5.8‰ , while carbonates of the Serra do Paraíso Formation yielded $\delta^{13}\text{C}$ values from -1.7‰ to $+8.9\text{‰}$ and $\delta^{18}\text{O}$ from -8.4‰ to -1.8‰ . Although a true chemostratigraphic profile is not available for these carbonates, the data imply large $\delta^{13}\text{C}$ fluctuations for the Camacan (-5.7‰ to $+4\text{‰}$) and the Serra do Paraíso (-1.7‰ to $+8.9\text{‰}$) formations, of a similar magnitude to those observed in cap carbonates. The Serra do Paraíso organic-rich limestones exhibit $\delta^{13}\text{C}$ variation somehow similar to that of Olhos D'Água Formation carbonates in the Sergipano belt. C and O isotope values suggest a transition from continental to marine environment for the Camacan Formation (Costa Pinto, 1977). These marine environments continued southwards as expressed in the Água Preta and Serra do Paraíso formations, suggesting deposition of the Rio Pardo Group on a passive margin.

Thirty samples from the Serra do Paraíso Formation at Toca da Onça locality (Mg/Ca $\sim 0.5\text{--}0.6$) show $\delta^{13}\text{C}$ values from 0.1‰ to $+2.2\text{‰}$ and $\delta^{18}\text{O}$ from -4.7‰ to -4.2‰ (A.N. Sial, unpublished data) within the range reported by Costa Pinto (1977). Four dark, organic matter-rich, Mg-carbonates of the Água Preta Formation (Nancy locality) display $\delta^{13}\text{C}$ values of $+4.1\text{‰}$ to $+8.2\text{‰}$ (A.N. Sial, unpublished data).

For carbonate formations of the Sergipano belt, Sial et al. (2000a, 2006) reported $\delta^{13}\text{C}$ values for the Jacoca Formation, from a section along the Vaza Barris River, of around -4‰ and $\delta^{18}\text{O}$ values around -8‰ (Figure 3.28). Strong oscillations are also observed in Mg/Ca ratios and Si, Fe, Mn and Rb concentrations in samples from the lowest portion of this profile. To the north of the town of Simão Dias, the Olhos D'Água Formation is represented by marbles, interbedded calcareous chlorite-schists and silty phyllites and overlies diamictites, pebbly metagreywackes of the Palestina Formation. Sial et al. (2006) produced a chemostratigraphic profile ($n = 120$)

Figure 3.28 C and O isotope and chemical stratigraphy of carbonates of the Jacoca Formation at the Capitão farm, along the Vaza Barris River, state of Sergipe.

Figure 3.29 C, O and Sr isotope and chemical stratigraphy of carbonates of the Olhos D'Água Formation at Serra do Capitão, state of Sergipe.

starting from the contact between the Olhos D'Água Formation and metagreywackes of the Palestina Formation. Basal marly and dolomitic carbonates display $\delta^{13}\text{C}$ values of -4‰ to -5‰ , increasing upsection (0 – 1‰). Towards the top, $\delta^{13}\text{C}$ changes dramatically to values of $\sim +8\text{‰}$ to $+10\text{‰}$ and the same behaviour is observed in a section at Serra do Capitão (Figure 3.29) and in another section near Rosario village in Bahia. This vigorous isotope fluctuation suggests contrasting, base to top, environmental conditions during deposition of this sequence, allowing for an enormous C-isotope oscillation, typical of Sturtian cap carbonates (Kennedy, 1996; Hoffman and Schrag, 2002). $\delta^{18}\text{O}$ values vary from -7‰ to -10‰ , a typical for Cryogenian carbonates.

Figure 3.30 C and O isotope and chemical stratigraphy of carbonates of the Acauã Formation at Serra da Borracha, ca. 20 km to the NW of Patamutê village, state of Bahia.

Sial et al. (2006) reported a C isotope profile for dolostones of the Acauã Formation (stratigraphically equivalent to the Jacoca Formation) overlying diamictites of the Juetê Formation at the São Gonçalo Farm near the town of Euclides da Cunha in the state of Bahia. These dolostones have laminations whose thickness gradually decreases upsection, indicating deposition during a transgression. The diamictite contains boulders of granite, orthogneiss, phyllite, quartz, some of which are dropstones, and an iron-rich claystone layer within a matrix-supported granular wackestone. Values of $\delta^{13}\text{C}$ start around -5‰ and gradually increase to -3‰ at about 15 m in height, while $\delta^{18}\text{O}$ varies from around -4.5‰ to about -6.5‰ upsection. At Serra da Borracha locality, about 20 km west of Patamutê village, a similar C-isotope profile is observed in a carbonate section that overlies 1-m-thick diamictite bed resting on the basement. $\delta^{13}\text{C}$ values are around -5‰ while $\delta^{18}\text{O}$ values are around -10‰ (Figure 3.30).

In summary, in proximal or distal sections of the Jacoca/Acauã carbonates, $\delta^{13}\text{C}$ values are virtually always negative, mostly clustering around -5‰ , whereas three sections of the Olhos D'Água carbonates show distinct C-isotope profiles, with strong fluctuations between negative (-5‰) and highly positive ($+8\text{‰}$ to $+10\text{‰}$) values.

The Jacoca and Olhos D'Água formations display similar $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (on average, 0.707974 and 0.707922, respectively) in the eastern portion of this belt. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for Acauã carbonates at Borracha Hill vary from 0.70717 to 0.70732, at Patamutê they range from 0.70751 to 0.70937, and at São Gonçalo Farm, from 0.70942 to 0.71099. For the Olhos D'Água carbonates from the western portion of the belt, ratios vary from 0.70753 to 0.70828. The average $^{87}\text{Sr}/^{86}\text{Sr}$ for Olhos D'Água Formation carbonates is in good agreement with values reported for seawater at ca. 720 Ma (Jacobsen and Kaufman, 1999). The average $^{87}\text{Sr}/^{86}\text{Sr}$ for the Jacoca Formation, however, is slightly too high for the seawater at ca. 740 Ma. The Sr temporal variation curve compiled by Jacobsen and Kaufman (1999), however, is not well constrained for the interval around 740 Ma.

3.4. GEOCHRONOLOGY AND PROVENANCE: SYNTHESIS OF U-Pb AND Sm-Nd DATA

The Neoproterozoic successions deposited on the São Francisco Craton contain basal glaciogenic units (Jeiquitá Formation and correlative units), which are capped by carbonates and then followed by siliciclastic sediments of the Bambuí Group. The available geochronological constraints were obtained on detrital zircons recovered from the matrix of the glacial diamictites that provide maximum depositional ages, and on the cap carbonates and overlying pelitic rocks of the Bambuí Group.

Nd isotope data indicate a Palaeoproterozoic source region for sediments of the Paranoá, Canastra and Vazante Groups (Pimentel et al., 2001). The Paranoá and Canastra Groups show the oldest Nd model ages that range between 2.3 and 2.1 Ga and, consequently, a provenance of clastic material from Palaeoproterozoic sources in the São Francisco Craton and passive margin tectonic setting. The Vazante Group, in turn, has yielded Nd model ages between 1.9 and 1.7 Ga, which may indicate a transition between the Paranoá and Bambuí Groups. Araxá and Ibiá sedimentary rocks present a bimodal Nd model age pattern (1.8–2.1 and 1.0–1.3 Ga). The Bambuí Group exhibits more uniform Nd model ages (1.4–1.9 Ga). The younger Nd model ages for these last units are interpreted as a juvenile source related to the Goiás magmatic arc, in the west of the Brasília belt.

U-Pb SHRIMP ages obtained on detrital zircon grains recovered from sandstones and conglomerates of the Vazante Group have given ages from 2.3 to 0.95 Ga, while detrital zircon grains from arkoses and conglomerates of the Bambuí Group yielded younger ages of ca. 650 Ma (Dardenne et al., 2003).

3.4.1. Geochronology of the glacial deposits

U-Pb SHRIMP ages were obtained on detrital zircon recovered from a glacial succession from the Bebedouro Formation in the Irecê basin, Bahia (Figure 3.1). This unit starts with shale with dropstones, interbedded with discontinuous layers of sandstone and overlain by a diamictite. The succession is capped by carbonates. Detrital zircon from the laminated facies and the diamictite range in age from 3,050 to 875 Ma ($n = 19$), indicating multiple source terranes (Babinski et al., 2004). The older ages are from the Archaean to Palaeoproterozoic basement; Mesoproterozoic zircon grains may be either from the Espinhaço Supergroup volcano-sedimentary sequence or from the ca. 1,000 Ma dykes that cut both the basement and the Espinhaço rocks. The source of the youngest zircon grains has been recently identified and precisely dated at 875 ± 9 Ma (Silva et al., 2007). It represents the anorogenic Salto da Divisa granites, which record continental rifting in the Araçuaí Orogen (Pedrosa-Soares et al., 2008).

Detrital zircons from diamictites of the Jequitai Formation, sampled at the Água Fria Range (Figure 3.1), were dated by Pb-Pb evaporation and yielded a maximum age of ca. 900 Ma (Buschwaldt et al., 1999). Correlation of glacial deposits from surrounding belts on the craton provides additional, albeit interpretive, age constraints. U-Pb SHRIMP ages were reported for detrital zircon recovered from sandstones from the uppermost part of the Chapada Acauã Formation (Macaúbas Group), a post-glacial shelf succession of interlayered sandstones and pelites (Pedrosa-Soares et al., 1992; Noce et al., 1997). Most detrital zircon grains yielded ages between 1,000 and 950 Ma, but the almost concordant youngest zircon grain constrains the maximum sedimentation age for the upper Chapada Acauã Formation at 864 ± 30 Ma (Pedrosa-Soares et al., 2000a). U-Pb SHRIMP age determinations of granitic pebbles from the Santo Antônio do Bonito (Vazante Group) and Cubatão (Ibiá Group) diamictites have given Palaeoproterozoic ages (2.3 Ga), providing no useful age constraints, but indicating a cratonic source for these formations (Dardenne et al., 2003).

A recent geochronological study on detrital zircon from different units of the Macaúbas Group, including pre-glacial and glacial rocks, allowed identification of source areas with ages ranging from 900 to 2,740 Ma. The maximum depositional age for the glacial deposits was defined by the youngest detrital zircon dated at 900 ± 21 Ma (Babinski et al., 2007; Pedrosa-Soares et al., 2008). A U-Pb zircon age of 730 Ma (B.B. Brito Neves, written communication) on an acidic tuff within the Ribeirópolis diamictites underlying the Jacoca carbonates, coupled with U-Pb age of zircon grains (615 Ma) from Frei Paulo metagreywackes above the Olhos D'Água carbonates, constrains the deposition of Jacoca and Olhos D'Água Formations to the interval 730–615 Ma.

3.4.2. Geochronology of the Bambuí Group

Data for Bambuí Group sedimentary rocks were obtained by the Rb-Sr method on clays and whole-rock samples and by the K-Ar method on clays (Bonhomme, 1976; Thomaz Filho and Bonhomme, 1979; Parenti Couto et al., 1981; Bonhomme et al., 1982; Thomaz Filho et al., 1998). Rb-Sr ages range from 695 ± 12 Ma ($R_0 = 0.7077$) to 560 ± 40 Ma ($R_0 = 0.7110$) and K-Ar ages on fine fraction clays range from 662 ± 18 to ~ 478 Ma. The older ages were considered as minimum depositional ages, and the younger ones as a result of later thermal events related to the Brasiliano orogeny. A Pb-isotopic study on the carbonate rocks from the Sete Lagoas and Lagoa do Jacaré formations was done by Babinski et al. (1999) and a minimum depositional age of 686 ± 69 Ma was suggested for these rocks. It was also suggested by those authors that a large-scale fluid percolation event affected most of the rocks from the southern part of the São Francisco basin at ca. 520 Ma. The same event could have been responsible for the remagnetisation of these carbonates (D'Agrella Filho et al., 2000; Trindade et al., 2004).

Recently, radiometric data were determined for deep-platform rocks organised in centimetre-scale cycles of lime mudstone and calcite crystal fans, interpreted as aragonite pseudomorph, that crop out at the Samba Quarry,

Figure 3.31 Pb-Pb isochron for Sete Lagoas Formation carbonates at the Sambra quarry, Minas Gerais (from Babinski et al., 2007).

near Sete Lagoas (Figure 3.1) at the southern part of the São Francisco Craton. These aragonite-pseudomorph crystals are black to dark-grey with acicular morphology, laterally connected to thin, millimetric cement-crusts. Crystal layers are covered by light-grey or red lime mudstones showing parallel to undulating lamination often truncated by stylolites. These carbonates yielded a Pb-Pb isochron age of 740 ± 22 Ma (Babinski et al., 2007). Because the rocks of this outcrop are extremely well preserved and undeformed, suggesting that the Pb-isotopic system was not disturbed by later events, this value was interpreted as the depositional age for this cap carbonate (Figure 3.31).

Vieira et al. (2007) have identified two depositional sequences in the Sete Lagoas Formation, the lower one essentially consisting of carbonates and the upper one of pelite-calclutite in the base, overlain by black limestone. The age of the upper sequence is younger than 610 Ma as recently constrained by a U-Pb (LA-MC-ICP-MS) age for detrital zircon grains from siltstones (Rodrigues and Pimentel, 2008). Thus, geochronological constraints available for the Neoproterozoic rocks deposited on the craton suggest that while the lower Bambuí Group may record the Sturtian glacial event, the upper Bambuí Group is Ediacaran in age.

3.5. SYNTHESIS OF NEOPROTEROZOIC MINERALISATIONS: LEAD, ZINC, FLUORINE, BARIUM AND PHOSPHATES

Neoproterozoic successions of the São Francisco Craton host a variety of mineral deposits, some of them of great economic importance. Among the known mineral deposits are the following:

- (i) Rapitan-type Fe-Mn formation in the Porteirinha in the Rio Vacaria valley is associated with glacial sedimentation of the Macaúbas Group in the Araçuaí belt, state of Minas Gerais. The source of iron and manganese is probably related to exhalative processes and mafic volcanism, as suggested by Dardenne and Schobbenhaus (2000).
- (ii) Phosphorites deposits (Figures 3.32 and 3.33) are associated with the lower Vazante Group at Coromandel, Rocinha and Lagamar (Dardenne et al., 1986, 1997; Da Rocha Araújo et al., 1992; Misi et al., 2005a,b), to the base of the Sete Lagoas Formation of the Bambuí Group at Campos Belos (Dardenne et al., 1986), to the lower portion of the Serra da Saudade Formation of the Bambuí Group at Cedro do Abaeté (Dardenne et al., 1986) and to stromatolitic beds of the Salitre Formation in the Irecê basin (Misi and Kyle, 1994). These phosphate accumulations are stratigraphically controlled and present in both, cratonic non-deformed strata and marginal fold belts (see Dardenne et al., 1986, 1997; Da Rocha Araújo et al., 1992; Misi and Kyle, 1994; Misi et al., 2005b). According to Misi et al. (2005b), primary phosphate concentrations are related to global episodes of phosphatisation associated with glacial events.
- (iii) Mississippi valley-type CaF_2 -Ba-Pb-Zn deposits are associated with the upper portion of the first carbonate megacycle of the Bambuí Group near Januária–Itacarambi, Montalvânia, Serra do Ramalho and Salitre basin

Figure 3.32 Phosphate deposits of the Neoproterozoic sedimentary basins of the São Francisco Craton. (a) Phosphorite of Rocinha, Vazante Group with lenses of micritic limestone; (b) phosphorite of Lagamar, Vazante Group; (c) stromatolitic and intraclastic phosphorite, Una Group, Irecê “basin”; (d) columnar stromatolitic phosphorite, Una Group, Irecê “basin” forming a continuous bed.

(Dardenne, 1979; Misi et al., 1997, 1999a,b; Dardenne and Freitas-Silva, 1999). All the data indicate migration of mineralising fluids from the Brasília belt towards the palaeogeographic high of the São Francisco Craton and the incorporation of radiogenic lead lixiviated from basement granite-gneisses.

- (iv) Irish-type Pb–Zn deposits occur in the dolomitic facies of the Vazante Group at Morro Agudo mine (Misi et al., 1997, 1999a,b, 2005a,b; Dardenne and Freitas-Silva, 1999; Bettencourt et al., 2001; Monteiro, 2002). This mineralisation (Figure 3.9) was initiated by progressive connate fluid expulsion from the basal sedimentary sequences under compression, related to uplift of Brasília belt, and subsequent channelisation along large normal fault systems.
- (v) Hydrothermal Zn deposits, such as Vazante Mine, are associated with major faults (Figure 3.9). The mineralisation is related to the evolution of connate basinal and metamorphic fluids (Dardenne and Freitas-Silva, 1999; Misi et al., 2000, 2005a,b, 2007; Bettencourt et al., 2001; Monteiro, 2002).

3.6. CONCLUSIONS AND REGIONAL ANALYSIS

A regional picture of the Neoproterozoic evolution of the São Francisco palaeocontinent depicted from a review of pertaining geological literature, coupled with relevant unpublished information and syntheses, leads to the following major conclusions:

- (1) The correlation between the Vazante and Bambuí Groups is still a matter of debate. Sr isotopic data of carbonate fluorapatite and of associated micritic limestones (~ 0.7074 in both successions; see Misi et al., 2007) may support this correlation. Nevertheless, ages of ca. 650 and 950 Ma on detrital zircon grains from Bambuí and Vazante clastic units, respectively (Dardenne et al., 2003, 2007a,b; Rodrigues, 2007), and new unpublished Re–Os ages suggest that the Vazante Group may be older as previously thought.

One glacial event has been recorded at the base of the Vazante Group by Dardenne (2001, 2006), represented by diamictites of the Santo Antônio do Bonito Formation. Azmy et al. (2006) have recognised the

Figure 3.33 Macro and microscopic features in the Zn-Pb mineralisation of the Neoproterozoic basins of the São Francisco Craton. (a) Massive stratiform mineralisation at the N ore body, Morro Agudo deposit, Vazante Group, formed by millimetric layers of chert, dolomite (Dol) and fine-grained sphalerite (Sph), with patches of galena (Gn) and pyrite (Py) and with calcite veins (Cc). (b) Massive coarse-grained sphalerite replacing oolitic dolarenite, JKL orebody, Morro Agudo mine, Vazante Group. (c) Complex nodules within dolostone (Dol) with an inner core of pyrite (Py) surrounded by coarse-crystalline sphalerite (Sph) at Irecê basin, Una Group. Sphalerite also occurs as pore-filling crystals in dolostone. Microcline detrital grains (Mi) in blue colour. Cathodoluminescence photomicrograph by Misi and Kyle (1994). (d) Coarse crystalline sphalerite (Sph) within microquartz (Qz: length-slow quartz) in contact with carbonate fluorapatite (Ap) in dolostone in the Una Group, Irecê basin. Probably a sulphate nodule replaced by microquartz and by sphalerite post-dating the apatite formation. Transmitted light photomicrograph (Misi et al., 2005a).

record of a second glaciation at the base of the Serra da Lapa Formation, at the top of the Vazante Group, based on the presence of diamictites and black shales with dropstones and on the presence of a negative $\delta^{13}\text{C}$ excursion in the overlying carbonates. The veracity of this second glaciation, however, is a matter of debate and requires further investigation.

- (2) There is no unambiguous geochronological control on most carbonate formations. The Pb-Pb isochron age of 740 ± 22 Ma at the base of the Sete Lagoas Formation (Babinski et al., 2007) is the only successful age on a cap carbonate to date. It is, if correlations are correct, an indirect age on the Sturtian glaciation (Jequitaí/Carrancas). However, the age of the carbonates and, consequently, of the Bambuí Group remains debatable considering that detrital zircon grains from the Serra de Santa Helena and Três Marias formations have yielded ages of ca. 650 Ma.
- (3) In the light of similar $\delta^{13}\text{C}$ profiles, it is assumed that the Sete Lagoas Formation, the Salitre Formation (Una Group) and Olhos D'Água Formation (Sergipano belt) are probably correlative cap carbonates, deposited in the aftermath of the Sturtian snowball event. In saying so, a long-standing Sturtian interval is assumed representing either different glaciations or a protracted interval encompassing almost 80 Ma (Babinski et al., 2007).

All these carbonate sequences start with negative $\delta^{13}\text{C}$ (around -5%) that increase gradually to a pronounced positive excursion ($> +8\%$ to 10% or even higher in the case of the Sete Lagoas Formation) that has proven useful in regional palaeogeographic investigations (e.g. identification of palaeohighs in the Serra do Cabral and Jequitaí regions; Santos et al., 2004). Carbonates of the Agua Preta Formation (Rio Pardo belt) show $\delta^{13}\text{C}$ values within the range of the Olhos D'Água Formation, despite limited C isotope data.

The Jacoca Formation carbonates (older than the Olhos D'Água Formation), overlying 730 Ma-old Ribeiropolis diamictites in the Sergipano belt (B.B. Brito Neves, written communication), display distinct $\delta^{13}\text{C}$ signatures, where virtually all values are negative in sections examined to date. The age of 615 Ma for metagreywackes of the Frei Paulo Formation on top of the Olhos D'Água Formation constrains the deposition of Jacoca (older) and Olhos D'Água (younger) Formations to the age range of 730–615 Ma.

- (4) The speculation that the Agua Preta (Rio Pardo belt; $\delta^{13}\text{C}$ values as high as +8‰) and São Desidério Formation carbonates (Rio Preto belt) were deposited concomitantly with the Sete Lagoas awaits confirmation from further geological and isotopic investigations.
- (5) Primary phosphate concentrations (phosphorites) in the Neoproterozoic of the São Francisco Craton are related to global episodes of phosphatisation, probably associated with glaciogenic events. Most of these mineralisations are unequivocally above glacial horizons; some others are not, but occur immediately above negative $\delta^{13}\text{C}$ excursions, suggesting they may nevertheless be related to glacial events.

ACKNOWLEDGEMENTS

W.R. Van Schmus, C. Riccomini, R.A. Fuck and Galen P. Halverson reviewed an early version of this chapter and greatly contributed to improve it.

3.2. CRÁTON AMAZÔNICO E FAIXA PARAGUAI SUL

O capítulo de livro “The Amazonian Palaeocontinent” de autoria de **Alvarenga et al. (2009)**, e que constitui o Capítulo 2 do livro “*Neoproterozoic-Cambrian Tectonics, Global Change e Evolution – A focus on Southwestern Gondwana*”, apresenta o contexto regional dos:

- (i) Diamictitos glaciais da Formação Puga, sotopostos às rochas do Grupo Corumbá, na Faixa Paraguai Sul, no Estado do Mato Grosso do Sul;
- (ii) Carbonatos de capa da Formação Mirassol d’Oeste, os quais recobrem os diamictitos glaciais da Formação Puga no sudoeste do Cráton Amazônico, Estado do Mato Grosso.

THE AMAZONIAN PALAEOCONTINENT[☆]

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2.1. INTRODUCTION

Neoproterozoic rocks were deposited on the southern margin of the Amazonian Craton/Rio Apa Block (Figure 2.1). Rock deformation increases from incipient on the border of the craton to tight folds within the compressional belt (Figure 2.2) (Almeida, 1964; Alvarenga and Trompette, 1993). The belt was deformed by the end of the Brasiliano Orogeny between 550 and 500 Ma (Trompette, 1994; Pimentel et al., 1996; Trindade et al., 2003; Tohver et al., 2006), followed immediately by post-orogenic granite magmatism at 500 Ma (Almeida and Mantovani, 1975). A branch of the southern Paraguay Belt extends from Corumbá into Bolivia as a large syncline known as the Tucavaca Belt (Litherland et al., 1986). These Neoproterozoic rocks have been geographically separated by Neogene sediments of the Pantanal Basin, and they are subdivided into the northern and southern Paraguay belts, which include outcrops in the Corumbá and Serra da Bodoquena regions (Figure 2.1).

Glacially derived deposits have been identified in this belt and were interpreted within a model of the global distribution with carbonate rocks overlying the glaciogenic deposits. The Puga Formation, Jacadigo Group and some lithofacies of the Cuiabá Group have been correlated with the Marinoan–Ghaub glaciations (~635 Ma) based on cap-dolomite lithofacies, palaeomagnetic data, and associated $\delta^{13}\text{C}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ isotope trends (Trompette et al., 1998; Nogueira et al., 2003; Trindade et al., 2003; Alvarenga et al., 2004, 2008; Allen and Hoffman, 2005). The record of a second glaciation in the northern Paraguay Belt was found above the post-Marinoan–Ghaub carbonates and was related to the 580 Ma Gaskiers glaciation (Alvarenga et al., 2007).

[☆] Alvarenga, C.J.S. de, Boggiani, P.C., Babinski, M., Dardenne, M.A., Figueiredo, M.F., Santos, R.V., Dantas, E.L., 2009. The Amazonian Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.): Neoproterozoic–Cambrian Tectonics, Global Change and Evolution: a focus on southwestern Gondwana. *Developments in Precambrian Geology*, 16, Elsevier, pp.15–28.

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Figure 2.1 Geological map of the Paraguay Belt on the southeastern part of the Amazonian Craton (modified from Schobbenhaus et al., 1981; Litherland et al., 1986; Boggiani, 1998).

Sedimentological and facies studies of biostratigraphy, palaeomagnetism and isotopic data of the Neoproterozoic rocks at the edge of the Amazonian/Rio Apa Craton have increased over the past decade and an improved picture of the relationships between their sections is emerging in the context of post-glacial carbonate facies in the Corumbá and Araras groups (Boggiani et al., 2003; Nogueira et al., 2003, 2007a; Misi et al., 2007). Palaeontological data for the Corumbá Group include new genera of organic-walled microfossils (Gaucher et al., 2003) in addition to *Cloudina luciano*, *Bavlinella faveolata* and scyphozoan *Corumbella werner* (Hahn et al., 1982; Walde et al., 1982; Zaine and Fairchild, 1985, 1987). Various workers have presented composite $\delta^{13}\text{C}$, $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ records for the carbonate rocks of the Corumbá and Araras groups (Boggiani, 1998; Boggiani et al., 2003; Gaucher et al., 2003; Nogueira et al., 2003, 2007a; Alvarenga et al., 2004, 2008; Font et al., 2006), although

Figure 2.2 Schematic section spanning the border of the Amazonian Craton and the Paraguay Belt (from Almeida, 1964).

striking differences are observed between these two units. The first palaeomagnetic data on a cap carbonate from the northern Paraguay Belt were reported by Trindade et al. (2003).

This paper presents a review of the stratigraphic units in the northern and southern Paraguay belts in Brazil (Figure 2.1). Biostratigraphy, and isotopic, palaeomagnetic and radiometric data available in the literature are discussed in order to establish a chronostratigraphic correlation.

2.2. LITHOSTRATIGRAPHY

In central-western Brazil, Neoproterozoic strata are preserved on the margins of the Amazonian/Rio Apa Craton and in the Paraguay Belt. Neoproterozoic sedimentary rocks crop out in different areas from the Província Serrana in the northern Paraguay Belt to the Serra da Bodoquena in the southern Paraguay Belt, passing by the Corumbá region and extending northwest into the Tucavaca Belt (Figure 2.1). The stratigraphies of these main areas of exposures are summarised in this paper.

2.2.1. Northern Paraguay Belt

The northern Paraguay Belt comprises, from base to top, four major lithostratigraphic sequences that exhibit lateral facies variations, mainly between the western border and the central part of the basin (Figure 2.3). The older lithostratigraphic units represent a passive margin and the younger ones represent a foreland basin related to the Brasiliano–Pan-African Orogeny (Almeida, 1974).

2.2.1.1. Basal glacially influenced sequence

The basal glacially influenced sequence occurs in the cratonic cover (Puga Formation) as well as within the metasedimentary fold belt (FB, Cuiabá Group; Alvarenga and Trompette, 1992; Alvarenga et al., 2004). A correlation with the global Marinoan–Ghaub glaciation, dated at 635 Ma (Hoffmann et al., 2004; Condon et al., 2005), has been suggested for this glacial sequence in the northern Paraguay Belt (Nogueira et al., 2003; Alvarenga et al., 2004, 2008; Allen and Hoffman, 2005), despite the lack of geochronological data. The Puga Formation consists of diamictite associated with conglomerate, sandstone, siltstone and shale. The widespread occurrence of diamictite indicates a lateral transition from thinner coarse-grained beds close to the Amazonian Craton (Puga Formation) to thick fine-grained facies in the east (Cuiabá Group). The sedimentation model proposed for the glacially influenced unit involves three main glacial depositional systems: platform, slope and outer slope (Alvarenga and Trompette, 1992). The platform depositional systems, which cover the cratonic domain in the western inner shelf, were reworked by gravity flows on the outer shelf (Figure 2.3). The deposits in the inner shelf show alternating dominant massive diamictite, sandstones and fine-grained sediments with few dropstones. In the outer shelf, an association of massive diamictite, stratified diamictite and fine-grained sediments progressively replaces the dominant massive diamictite of the inner shelf. Glaciomarine sediments reworked by gravity flows and related to submarine fan deposits are associated with the slope depositional system. Diamictite, conglomerate and sandstone intercalations with occasional inverse and/or normal grading occur in the deeper parts of

Figure 2.3 Schematic stratigraphic cross-section of the Neoproterozoic depositional sequences along the southeastern edge of the Amazonian Craton, between 15° and 17° S latitude (modified from Alvarenga et al., 2008). A, B and C are the located sections illustrated in Figure 2.4.

the fan (Alvarenga and Trompette, 1992). Sandstone and siltstone intercalations represent inter-channel deposits formed by turbidity currents. Deposition on the outer slope system was dominated by fine-grained sediments (phyllite and metasilstone) related to low-density turbidity currents, in which a decrease of glacial influence is indicated by the presence of a few isolated clasts or dropstones (Alvarenga and Trompette, 1992). This model of basin filling suggests that the Amazonian Craton was the main source area of the sediments, which were later reworked during the glacial event, forming submarine channels and turbidites on the slope and outer slope (Alvarenga and Trompette, 1992). Palaeomagnetic studies on the cap carbonate covering the Puga Formation on the Amazonian Craton indicate glacial sedimentation at low latitudes (Trindade et al., 2003).

2.2.1.2. Araras Group

The Araras Group (carbonate sequence) on the southeastern border of the Amazonian Craton overlies the Puga Formation and reaches a thickness of about 100–150 m on the western border (inner shelf) of the basin. Towards the east, this carbonate succession can reach up to 1,300 m in thickness (Luz and Abreu Filho, 1978; Alvarenga et al., 2004) in the middle shelf domain, and continuing to deeper sequences mud-rich limestone and laminated metasilstone successions can be found towards the slope depositional system of the basin (Figures 2.3 and 2.4). The Araras Group can be subdivided into three formations.

The lowermost Mirassol d'Oeste Formation, which includes 18–30 m of the cap carbonate, overlies the Puga Formation (Nogueira et al., 2003; Alvarenga et al., 2008). This cap carbonate shows variable degrees of recrystallisation and contains stromatolites, tube-like structures, breccias, giant wave ripples (Nogueira et al., 2003; Alvarenga et al., 2004, 2008; Allen and Hoffman, 2005) and fan-like crystals that have been interpreted as aragonite pseudomorphs (Faulstich, 2005; Alvarenga et al., 2008).

The middle Guia Formation consists of laminated limestone, mud-limestone and mudstones up to 250 m thick in the middle shelf domain (Figure 2.4).

The upper part of the Araras Group is represented by the Nobres Formation, and it comprises shallow-water breccia, grainstone and packstone dolostone; towards the top of the formation sandstone intercalations about 1 m thick are common (Almeida, 1964; Alvarenga et al., 2004). The lower part of the Nobres Formation was recently subdivided into the Serra do Quilombo Formation (Nogueira et al., 2007a), but its extension was not confirmed for the whole belt and in this paper, the original denomination of the Nobres Formation (Hennies, 1966) is used.

Sedimentary rocks of these three formations reflect a shallower facies at the western edge of the basin (cratonic area) and progressive deeper water facies to the more distal platform in the east (fold belt).

2.2.1.3. Serra Azul Formation

The Serra Azul Formation includes discontinuous outcrops of diamictite and siltstone deposited on top of post-Marinoan-Ghaub carbonates of the Araras Group (Figures 2.3 and 2.4), and it represents the record of a second Neoproterozoic glaciation in the Paraguay Belt (Figueiredo et al., 2004, 2008; Alvarenga et al., 2007).

Figure 2.4 Measured stratigraphic sections of the Neoproterozoic units on the northern Paraguay Belt, between 15° and 17° S latitude. A, B and C are located in Figure 2.3.

The basal unit comprises approximately 70 m of massive reddish diamictite with an abundant clayey–silty matrix that becomes yellowish and stratified in the last 60 cm (Figueiredo et al., 2008). The clasts are composed of sandstones, arkoses, cherts, quartz, quartzites, metaconglomerates, carbonates, claystones, diabases, granites, gneisses, basalts and rhyolites. The size of these clasts ranges from granule to boulder, varying from highly angular to well rounded. Some clasts are striated, faceted and polished, while boulders are bullet shaped. Their poorly sorted characteristics associated with a large compositional variety of clasts and an abundance of striated clasts suggests a glacial influence on the deposition of the diamictites.

The diamictites are sharply overlain by a 25-m thick succession of reddish laminated siltstone with sparse granules. This unit grades into a 150-m thick rhythmite succession whose basal part is composed of a clayey–silty rhythmite intercalated with episodic sandstone layers (5–10 cm) with hummocky cross-stratification in the Serra Azul region (Alvarenga et al., 2007; Figueiredo et al., 2008). The upper part of the rhythmite has more contribution of sand, grading into a heterolithic succession that becomes coarser and thicker towards the top of the section (Figueiredo et al., 2008). A limestone layer has been found in the clayey–silty rhythmite in the Nobres region (Figueiredo et al., 2008).

This glacial unit has been related to the Gaskiers glaciation (Figueiredo et al., 2004; Figueiredo, 2006; Alvarenga et al., 2007; Figueiredo et al., 2008) with an age of ca. 580 Ma (Bowring et al., 2003b). The unit has not yet been found in the southern Paraguay Belt.

2.2.1.4. Upper siliciclastic unit

The upper siliciclastic unit, known as the Alto Paraguay Group (Almeida, 1964), consists of two siliciclastic formations represented by sandstones and pebble conglomerates at the base (Raizama Formation), and red shale, siltstone and arkoses at the top (Diamantino Formation). The contact between these two units is transitional and consists of sandstone, shale and siltstone intercalations (Alvarenga et al., 2000).

2.2.2. Southern Paraguay Belt – Corumbá region

The Neoproterozoic rocks in the Corumbá region are located at the margin of the Amazonian/Rio Apa Craton as subhorizontal layers locally affected by faults. Two major lithostratigraphic sequences have been described (Almeida, 1945; Dorr II, 1945). The lower sequence, lying over the granite basement, is composed of siliciclastic rocks and banded iron formation (BIF) ascribed to the Jacadigo Group, and the upper unit consists of the carbonate rocks of the Corumbá Group.

2.2.2.1. Jacadigo Group

The Jacadigo Group was deposited in a system of grabens related to a crustal stretching (Haralyi and Walde, 1986) interpreted as coeval to the glacially influenced sequence described in the northern Paraguay Belt (Trompette et al., 1998). The thickness of the Jacadigo Group ranges from 20 m out of the graben up to 700 m thick in the graben (Haralyi and Walde, 1986; Hoppe et al., 1987; Trompette et al., 1998). Two formations have been more commonly used to distinguish this group (Figure 2.5). The lower Urucum Formation, around 350 m thick at Santa Cruz valley (type section), is formed by coarse and conglomeratic sandstones with pebbles and some boulders of granite, dolostone, quartzite and rocks fragments sometimes cemented by carbonate (Almeida, 1945; Dorr II, 1945). The upper unit, the Santa Cruz Formation, starts with a continuous layer of red-violet ferruginous-manganese arkose sandstone with jaspelite/hematite cement along Urucum Hill (Almeida, 1945; Walde and Hagemann, 2007); this unit was named the Banda Alta Formation by Dorr II (1945). Towards the top of this formation, bedded hematite-rich and jasper BIF are more frequently intercalated with arkose sandstone and conglomerate (Figure 2.6). A manganese ore bed, up to 3 m thick has been mined at the underground Urucum Mine, and three other thinner Mn horizons (0.30–1.0 m) are intercalated with BIF and arkose sandstone for about 400 m of the Santa Cruz Formation (Almeida, 1945). In the Santa Cruz Formation, boulders to pebbles of granite, some more than 1.5 m in diameter, have been identified in thin arkose layers (< 5 cm) between two layers of BIF, creating a false appearance of boulders in the BIF (Figure 2.6). The lack of geochronological data and the limited extent of the Jacadigo Group constitute another problem when trying to confirm that these rocks correlate with the global Marinoan-Ghaub glaciation.

Figure 2.5 Stratigraphic section of the Neoproterozoic units on the southern Paraguay Belt, between 19° and 22° S latitude. (A) Corumbá area, (B) Puga Hill and (C) Serra da Bodoquena.

Figure 2.6 Boulder of weathered granite between layers of banded iron formation of the Santa Cruz Formation, Jacadigo Group.

2.2.2.2. Puga Formation

The Puga Formation was first described as 95 m of ‘tillites’ exposed at the isolated Puga Hill in the middle of Neogene sediments of Pantanal Basin (Maciel, 1959). This formation crops out south of Corumbá and east of Serra da Bodoquena, and it comprises sand diamictites with clasts of quartzites, gneisses and carbonate (Figure 2.5). This unit is related to the Marinoan-Ghaub glaciation, and at the Puga Hill, it is overlain by a cap carbonate (Boggiani et al., 2003).

The Puga Formation in the section type consists of massive grey and brick red diamictite. The diamictite at Puga Hill exhibits different kinds of angular clasts (quartzite, granite and limestone) 10–30 cm in size in a sandstone matrix. Evidence of glacial processes is suggested by limestones in sandstone. No striated clasts have yet been found at Serra da Bodoquena, although these are common in the northern Paraguay Belt (Alvarenga and Trompette, 1992).

2.2.2.3. Corumbá Group

The Corumbá Group in the undeformed domain, located on the right margin of the Paraguay River, is characterised by dolostones of the Bocaina Formation that are overlain by dark limestones of the Tamengo Formation, followed by shales of the Guaicurus Formation (Figure 2.5; Gaucher et al., 2003).

The lower unit (Bocaina Formation) is composed of grey dolostones, more than 100 m thick, usually as crystallised grainstone, packstone, breccia and stromatolites with crude stratification (Almeida, 1945; Gaucher et al., 2003). The dark organic-rich limestones and shales of the Tamengo Formation include a microfossil assemblage dominated by *B. faveolata* (Zaine and Fairchild, 1985, 1987), *C. luciano*, scyphozoan *C. werner* (Hahn et al., 1982; Walde et al., 1982) and *Soldadophycus bossii* (Gaucher et al., 2003). *C. werner* is of great significance to the discussion about the rise of metazoans and for biostratigraphic correlation (Erdtmann et al., 2005). A U–Pb zircon age of 543 ± 2 Ma (Ediacaran) has been obtained from an ash bed intercalated with *Cloudina*-bearing limestones of the Tamengo Formation (Babinski et al., 2008a). The upper part of the Corumbá Group includes fine-grained siliciclastic rocks from the Guaicurus Formation (Almeida, 1965; Boggiani, 1998; Gaucher et al., 2003). Diamictites and pelites with limestones have been described as a lens in pelites of the Guaicurus Formation at Laginha Mine, Corumbá, MS, which was interpreted as gravity flow deposits in a slope setting (Boggiani et al., 2004).

2.2.3. Southern Paraguay Belt – Serra da Bodoquena

On the eastern margin of the Rio Apa Craton, Neoproterozoic rocks of the Corumbá Group are preserved as different sedimentary facies. The exposures of the Corumbá Group in the Serra da Bodoquena region are in general undeformed over the Rio Apa Craton and deformed in the Paraguay Belt. In this region, the

predominantly siliciclastic units of the Corumbá Group (Cadieus and Cerradinho formations) lie under the Bocaina and Tamengo formations.

2.2.3.1. Corumbá Group

The Corumbá Group was first subdivided into the Cerradinho, Bocaina, Tamengo and Guaicurus formations (Almeida, 1965). Later, the Cadieus Formation was included as the lower unit of this group (Gaucher et al., 2003). The deposition of the Corumbá Group was initiated by crustal stretching (rift phase), including a mixed assemblage of clastic and carbonate rocks (Cadieus and Cerradinho formations) that unconformably overlies the Rio Apa Craton (Gaucher et al., 2003). The succeeding rocks include carbonate, marl and shale sediments of a drift sequence represented by the Bocaina, Tamengo and Guaicurus formations (Figures 2.5 and 2.7).

2.2.3.2. Cadieus and Cerradinho Formations

The Cadieus Formation is represented by conglomerate and arkose sandstone deposited in a proximal alluvial fan that grades to arkose sandstone, shale and grainstone of the Cerradinho Formation in a distal alluvial fan (Boggiani et al., 1993; Gaucher et al., 2003).

2.2.3.3. Bocaina Formation

The Bocaina Formation comprises shallow marine carbonates along hundreds of kilometres with various lithofacies, including grainstone, packstone, mudstone and stromatolites (Figure 2.8). In the Serra da Bodoquena, phosphatic dolostones were found at the top of this unit.

2.2.3.4. Tamengo Formation

The Tamengo Formation comprises up to 100 m of dark, organic-rich limestone-shale rhythmite with occasional grainstone limestone. Fine-grained facies in the Serra da Bodoquena region were deposited in deeper waters than those from the Corumbá region (Boggiani, 1998), which could explain the absence of the metazoan fossils *Cloudina* and *Corumbella*.

2.2.3.5. Guaicurus Formation

The Guaicurus Formation was first described by Almeida (1965) to name the thick shale sequence on top of the Corumbá Group that extends to the eastern part of Serra da Bodoquena and along the Miranda River valley (Figure 2.7).

The Tamengo and Guaicurus formations comprise a transgressive-highstand sequence that drowned the carbonate shelf of the Tamengo Formation; there is no equivalent of this unit in the northern Paraguay Belt.

Figure 2.7 Stratigraphic chart of the Neoproterozoic on the Serra da Bodoquena (modified from Boggiani and Alvarenga, 2004).

Figure 2.8 Stromatolites in dolostones of the Bocaina Formation, Serra da Bodoquena.

2.3. CHEMOSTRATIGRAPHY

The chemostratigraphic data available from the carbonate successions of the Araras and Corumbá groups consist of 300 carbon isotope and a few tens of Sr isotope data that were presented in recent papers (Boggiani et al., 2003; Gaucher et al., 2003; Nogueira et al., 2003, 2007a; Pinho et al., 2003; Alvarenga et al., 2004, 2008; Font et al., 2006; Riccomini et al., 2007).

2.3.1. Northern Paraguay Belt

2.3.1.1. Araras Group

The Araras Group reaches a thickness of about 100–150 m over the Amazon Craton (border of the basin) and up to 1,300 m thick in the Paraguay fold belt (outer shelf and slope). The underlying Puga Formation is composed predominantly of diamictites, which are continuous for 300 km along the northern Paraguay Belt, between Cáceres, Bauxi, Nobres and Serra Azul (Figure 2.1). A Marinoan–Ghaub age for the Puga Formation has been suggested (Nogueira et al., 2003, 2007a; Alvarenga et al., 2004, 2008) based on similar cap carbonate and isotope data from other Neoproterozoic basins worldwide (Alvarenga et al., 2004; Allen and Hoffman, 2005; Nogueira et al., 2007a).

2.3.1.2. Mirassol d'Oeste Formation

The basal Mirassol d'Oeste Formation is a 20-m thick sequence of transgressive cap dolomite deposited on top of diamictites over the Amazon Craton (Nogueira et al., 2003, 2007a; Alvarenga et al., 2004; Font et al., 2006) and in the FB (Alvarenga et al., 2008) (Figure 2.4). Cap dolomites over the Amazonian Craton show consistently negative $\delta^{13}\text{C}$ values ranging between -10.5‰ and -3.0‰ ($n = 96$), and $\delta^{18}\text{O}$ values between -8.4‰ and -1.9‰ ($n = 96$) (Figure 2.9). In the FB, the cap carbonate from a drill core shows $\delta^{13}\text{C}$ values ranging from -4.8‰ to -1.7‰ ($n = 7$) and $\delta^{18}\text{O}$ values between -8.0‰ and -5.1‰ ($n = 7$; Figure 2.9). A comparative carbon isotope stratigraphy between cap carbonate deposited over the Amazonian Craton and in the FB shows slightly higher $\delta^{13}\text{C}$ values in the FB (Alvarenga et al., 2008). The high and variable $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.70852–0.71435) of the cap carbonate associated with their low Sr content (44–96 ppm) and high Mn/Sr ratios (> 18) suggest that primary isotopic values were modified by post-depositional diagenetic alteration.

2.3.1.3. Guia Formation

Limestones and mud-limestones (Guia Formation) deposited on top of the cap carbonate overlying the Amazonian Craton show $\delta^{13}\text{C}$ values between -5.4‰ and -2.7‰ ($n = 10$; Alvarenga et al., 2004, 2008; Nogueira et al., 2007a), while in the FB (mid shelf to foreslope basin) $\delta^{13}\text{C}$ values range from -5.3‰ to -0.1‰

Figure 2.9 Stratigraphic section and carbon isotopic profiles for the Araras Group and limestone of the Serra Azul Formation on the northern Paraguay Belt, between 15° and 17° S latitude. (A) $\delta^{13}\text{C}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ from Nogueira et al. (2003, 2007a); (B) $\delta^{13}\text{C}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ from Alvarenga et al. (2004); and (C) $\delta^{13}\text{C}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ from Figueiredo (2006).

($n = 105$; Alvarenga et al., 2004, 2008; Nogueira et al., 2007a; Figueiredo, 2006; Figure 2.9). Oxygen isotope values for the same samples range between -8.3‰ and -5.8‰ ($n = 10$; Alvarenga et al., 2004, 2008; Nogueira et al., 2007a) over the Amazonian Craton, while in the FB, $\delta^{18}\text{O}$ values range from -10.6‰ to -6.3‰ ($n = 105$; Alvarenga et al., 2004, 2008; Nogueira et al., 2007a). These limestones and mud-limestones with high Sr concentrations (750–4,351 ppm) and low Mn/Sr ratios (0.01–0.20) have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between 0.70763 and 0.70780 ($n = 9$; Alvarenga et al., 2004, 2008) and probably represent the original Sr isotopic composition of the seawater. The limestone samples from the foreslope basin at Nossa Senhora da Guia Quarry show $\delta^{13}\text{C}$ values ranging from -5.3‰ to -0.8‰ and $\delta^{18}\text{O}$ values between -13.5‰ and -7.1‰ (Alvarenga et al., 2004).

2.3.1.4. Nobres Formation

The dolostones of the Nobres Formation are better exposed in the FB, and rocks of the Bauxi-Nobres region in the north and from the Cáceres area in the south of the FB present different isotopic pattern (Figure 2.9). In the Bauxi-Nobres region, $\delta^{13}\text{C}$ values range from -1.0‰ to $+9.6\text{‰}$ ($n = 45$), and $\delta^{18}\text{O}$ values range between -4.7‰ and -0.1‰ ($n = 45$; Pinho et al., 2003; Alvarenga et al., 2004; Figueiredo, 2006). In the Cáceres region, $\delta^{13}\text{C}$ values range from -2.2‰ to $+0.2\text{‰}$ ($n = 41$), while $\delta^{18}\text{O}$ values range between -6.7‰ and -3.4‰ ($n = 41$; Pinho et al., 2003; Nogueira et al., 2007a).

2.3.1.5. Serra Azul Formation

The isotope geochemistry data of the Serra Azul Formation include data in a limestone lens (about 12 m thick) in rhythmite from the upper part of the formation. $\delta^{13}\text{C}$ values range from -7.5‰ to -0.8‰ , and $\delta^{18}\text{O}$ values from -7.8‰ to -5.5‰ ($n = 20$; Figueiredo and Babinski, 2008; Figueiredo, 2006; Figure 2.9). Nine samples of these limestones with Sr concentrations between 714 and 885 ppm and Mn/Sr ratios below 0.2 yielded $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between 0.70863 and 0.70882 (Figueiredo and Babinski, 2008; Figueiredo, 2006).

2.3.2. Southern Paraguay Belt

2.3.2.1. Corumbá Group

The isotopic values for the carbonate rocks (Corumbá Group) in the southern Paraguay Belt were obtained for the cap carbonate overlying the diamictites at the Puga Hill and for the limestones of the Tamengo Formation in the Corumbá region (Boggiani, 1998; Boggiani et al., 2003; Gaucher et al., 2003).

2.3.2.2. Cap carbonate at Puga Hill

The cap carbonate described at Puga Hill consists of alternating pinkish and grey laminated limestone (rhythmite), with alternating peloidal and microcrystalline carbonate for ca. the first 12 m. Throughout this section, the values of $\delta^{13}\text{C}_{\text{PDB}}$ are homogeneous and are around -5‰ ($n = 9$) (Boggiani et al., 2003; Misi et al., 2007). The cap carbonate at Puga Hill is essentially calcium carbonate (MgO between 0.98% and 1.12%, $n = 9$), and the unique sedimentary feature of the cap carbonate is the rhythmic lamination.

2.3.2.3. Tamengo Formation

The Tamengo Formation occurs in Corumbá and along the eastern part of Serra da Bodoquena. The isotopic data for the limestones of the Tamengo Formation were obtained around Corumbá (Figure 2.1) in the Laginha, Saladeiro and Corcal quarries. Limestones of the lower part of the Tamengo Formation record a negative $\delta^{13}\text{C}$ excursion (-3.3‰ to -2.5‰), and in the direction of the top section, the $\delta^{13}\text{C}$ values vary between -1.3‰ and -0.3‰ , following a positive $\delta^{13}\text{C}$ excursion (up to 5.8‰) to the upper section of the formation (Figure 2.10; Boggiani, 1998; Boggiani et al., 2003; Gaucher et al., 2003; Misi et al., 2007). At Serra da Bodoquena, about 200 km southeast of Corumbá, the limestones correlated to the Tamengo Formation show constant $\delta^{13}\text{C}$ values around 3‰ (Boggiani, 1998).

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the carbonates of the Tamengo Formation for both of these areas are between 0.7084 and 0.7085 (Boggiani, 1998; Babinski et al., 2008a).

2.4. GEOCHRONOLOGICAL CONSTRAINTS

Radiometric ages on rocks from the Paraguay Belt and related cover over the Amazonian Craton consist of Sm-Nd and Rb-Sr data in some siliciclastic rocks from the Puga Formation, the Cuiabá Group, the Araras Group

Figure 2.10 Composite $\delta^{13}\text{C}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ profile for the Corumbá Group from Corumbá area, Puga Hill and Serra da Bodoquena.

and the Alto Paraguay Group, and Pb/Pb analysis on cap carbonates from the Mirassol d'Oeste Formation. These geochronological data were obtained in rocks from the northern Paraguay Belt and are presented here according to the lithostratigraphic units described earlier.

2.4.1. Glacially influenced sequence

The Nd isotope signature of the metasedimentary rocks of this sequence was obtained in phyllites, quartzites and diamictites by Dantas et al. (2009).

The glacially influenced units (Cuiabá Group and Puga Formation) are dominated by fine-grained sediments (phyllites), sandstones, conglomerates and diamictites that occur in the upper part of this unit. Twenty-four analyses on this unit show ε_{Nd} ($T = 600$) values ranging from -12 to -7 and Sm-Nd T_{DM} model ages (DePaolo, 1981) of 1.6 – 2.2 Ga. The phyllites are the most common lithology and have a large interval of ε_{Nd} ($T = 600$) values (between -12.0 and -7.4) as well as T_{DM} model ages (between 1.7 and 2.2 Ga). The diamictites also have variable T_{DM} model ages (1.6 – 2.1 Ga), but the ε_{Nd} ($T = 600$) values are quite homogeneous, between -7.0 and 8.5 . These T_{DM} model ages suggest that rocks from the Amazonian Craton could be the source of the sediments, mainly the Ventuari-Tapajós and Rio Negro-Juruena provinces (Tassinari and Macambira, 1999).

2.4.2. Araras Group

The only direct dating available for these rocks was obtained on cap carbonates from the Mirassol d'Oeste Formation, at the base of the Araras Group, deposited on the cratonic area. Many stable isotopic studies have been done in this classic outcrop (Nogueira et al., 2003; Alvarenga et al., 2004; Font et al., 2006), and the good preservation of the samples allowed us to obtain good results. Twelve carbonate samples, collected along a 40-m thick section, were analysed for Pb isotopes, some in duplicate. The $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ isotope ratios from 21 analyses range from 18.47 to 65.18 and 15.70 to 18.55, respectively. These ratios show good alignment on the Pb isochron diagram and yielded a Pb-Pb isochron age of 633 ± 25 Ma (95% confidence), which was interpreted as the depositional age of these rocks (Babinski et al., unpublished data), and which confirms the Marinoan-Ghaub age of this succession previously suggested based on C and Sr isotope data.

The Nd isotope signature and T_{DM} model ages have been determined in seven carbonates samples from the Araras Group. The Sm-Nd T_{DM} model ages vary from 1.7 to 2.1 Ga, and ε_{Nd} ($T = 600$) values from -8.2 to -11 (Dantas et al., 2009). Among these samples, the cap carbonates presented the less negative ε_{Nd} ($T = 600$) values, which are closer to values reported for the matrix of the diamictites. Model ages of the limestones and the intercalated carbonate siltstones of the Guia Formation are similar. The more negative ε_{Nd} ($T = 600$) values (-11.3 and -11.4) and T_{DM} model ages of 2.0 Ga agree with the stratigraphic interval of the upper part of the Nobres Formation, which is related to the incursion of clasts (quartz and feldspar) in the carbonate succession. These T_{DM} model ages suggest that the Amazonian Craton is the source area of the sediments.

2.4.3. Serra Azul Formation

Some indirect geochronological constraints are available for the Serra Azul diamictites deposited on top of the Araras Group carbonates. This stratigraphic position allows us to suggest that the diamictites of the Serra Azul Formation correspond to the Gaskiers glacial event, dated at 580 Ma (Bowring et al., 2003b). Geochronological data available for the Serra Azul Fm. include K-Ar ages of ca. 730 and 1,200 Ma obtained on volcanic clasts recovered from the diamictites, and Sm-Nd T_{DM} model ages of 2.0 and 1.6 Ga obtained on the matrix of the diamictites and pelites of the Serra Azul Formation shows variations in sedimentary provenance (Figueiredo and Babinski, 2008). The ε_{Nd} ($T = 600$) values range from -6.7 to -9.5 . These Sm-Nd T_{DM} ages suggest that the mixture of rocks from the Ventuari-Tapajós, Rio Negro-Juruena, and even the younger Rondonian-San Ignácio and Sunsás provinces (Tassinari and Macambira, 1999) could be the sources of sediments to the Serra Azul deposits. The source area of the younger volcanic clasts (ca. 730 Ma), however, has not yet been identified in the region and could be related to the rift phase of the Paraguay Belt during the earlier stages of Rodinia break-up.

2.4.4. Alto Paraguay Group

Geochronological data are available for siliciclastic rocks of the Raizama and Diamantino formations.

Shales from the lower part of the Diamantino Formation yielded an Rb-Sr isochron age of 568 ± 20 Ma, which was interpreted as the age of diagenesis of these sediments (Bonhomme et al., 1982; Cordani et al., 1985).

Sm–Nd T_{DM} model ages were also determined on seven samples of siliciclastic rocks from the upper unit of the Paraguay Belt, ranging from 1.5 to 1.8 Ga (Dantas et al., 2009). The ϵ_{Nd} ($T = 600$) values are not so negative and vary from -4.9 to -9.7 . These Sm–Nd values confirm the presence of a younger source of sediments filling the upper part of the basin, as confirmed by the 730 Ma volcanic clasts found in the diamictites of the Serra Azul Formation (Figueiredo and Babinski, 2008).

2.4.5. Corumbá Group – Tamengo Formation

Limestones in the Corcal Quarry close to Corumbá have abundant *Cloudina* and few intercalations of volcanic tuffs. Euhedral crystals of one of the ash beds yielded a $^{238}\text{U}/^{206}\text{Pb}$ average age of 543 ± 3 Ma ($n = 17$; 95% confidence) (Babinski et al., 2008a). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios cluster at 0.7084–0.7085, which agrees with the Ediacaran age.

2.5. PALAEOMAGNETIC CONSTRAINTS

The first palaeomagnetic data on the Neoproterozoic cap carbonate of the Puga Formation were obtained from 19 samples in the Terconi Quarry in Mirassol d'Oeste, MT (Trindade et al., 2003). These palaeomagnetic data indicate five polarity reversals along the first 20 m of the cap dolostones, suggesting a primary magnetisation and low palaeolatitudes ($22^\circ +6/-5$) for the cap dolostone (Trindade et al., 2003). The palaeomagnetic reversals imply slow sedimentation (millions of years), contrary to the snowball Earth hypothesis.

2.6. CONCLUSIONS AND REGIONAL ANALYSIS

The lithostratigraphy of the Neoproterozoic in the Paraguay Belt is summarised in the columnar sections from the northern and southern Paraguay belts (Figure 2.11). There is no consensus about the lithostratigraphic correlation between the north and the south sections of the Paraguay Belt. This will be one of the focal points of the following discussion, including an integrated discussion of the lithostratigraphy, glacial episodes, palaeontology, chemostratigraphy and ages of the rocks.

Glacial episodes have been found in both sectors of the Paraguay Belt, although the evidence is more convincing in the northern belt. The Neoproterozoic record of the northern Paraguay Belt contains two glacial events: an older one of Marinoan–Ghaub glacial age (630 Ma) represented by the Puga Formation and the Cuiabá Group rocks, and a younger one, probably related to the Gaskiers glaciation (580 Ma) recorded by the Serra Azul Formation (Figure 2.11A). The Puga Formation and the correlated Cuiabá Group are interpreted as being related to the Marinoan–Ghaub glaciation because of the cap-carbonate lithofacies, palaeomagnetic data, and associated $\delta^{13}\text{C}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ isotope trends (Nogueira et al., 2003; Trindade et al., 2003; Alvarenga et al., 2004, 2008; Font et al., 2006). Striated clasts and dropstones are direct evidence for the glaciation during the sedimentation of the Puga Formation and the Cuiabá Group, where the latter involves glacially influenced turbidite sedimentation

Northern Paraguay Belt				Southern Paraguay Belt			
Western Paraguay Belt		Eastern Paraguay Belt		Bodoquena Range		Corumbá area	
Almeida, 1964; Hennies, 1966; Luz et al., 1978; Nogueira et al., 2003; Alvarenga et al., 2007; Figueiredo et al., 2008		Almeida, 1964; Hennies, 1966; Luz et al., 1978		Almeida, 1965; Boggiani 1998		Almeida, 1945; Dorr II, 1945; Maciel, 1959	
Group	Formation	Group	Formation	Group	Formation	Group	Formation
Alto Paraguai	Diamantino	Araras	Guia	Corumbá	Guaicurus	Corumbá	Guaicurus
	Sepotuba				Tamengo		Tamengo
	Raizama				Bocaina		Bocaina
Araras	Serra Azul				Cerradinho		
	Nobres				Cadieux		
	Guia			Puga	Jacadigo	Urucun	Puga
	Mirassol d Oeste		Mirassol d Oeste				Santa Cruz
	Puga		Cuiabá				Urucun
			undivided				

Figure 2.11 Neoproterozoic stratigraphic nomenclature in the Paraguay Belt.

(Alvarenga and Trompette, 1992). The Puga diamictites have a well-developed, post-glacial cap carbonate (Nogueira et al., 2003; Font et al., 2006), whereas the younger Serra Azul diamictite has no identified cap carbonate, and siliciclastic mudstone rock lies directly above it (Alvarenga et al., 2007; Figueiredo et al., 2008). The cap carbonate of the Araras Group, named the Mirassol d'Oeste Formation, has the same structures (tube, wave ripple and fan-like) as those found in the cap carbonate in the Otavi Group, Namibia (Hoffman and Schrag, 2002; Nogueira et al., 2003; Allen and Hoffman, 2005). The main difference between the Serra Azul and the Puga glaciogenic units is the absence of a transgressive carbonate succession overlying the diamictites of the Serra Azul Formation.

In the southern Paraguay Belt, the rocks interpreted as deposited during a glacial period are the Jacadigo Group and the Puga Formation. The presence of boulders in the iron formation of the Jacadigo Group has been used to suggest a glacial origin for these rocks (Walde et al., 1981; Hoppe et al., 1987; Urban et al., 1992). These granite pebbles and boulders generally do not cross-cut the underlying bedding, even the biggest ones, which are larger than 1 m³, and the bedding deformation around the boulders occurs associated with compaction and is more pronounced above the boulders than beneath them (Trompette et al., 1998). Trompette et al. (1998) thus argued against glacial origin of these deposits, suggesting instead reworked gravitational flows. The diamictites of the Puga Formation in the southern Paraguay Belt do not show evidence of striated clasts or dropstones, but their widespread areal distribution, underlying a carbonate succession (Araras and Corumbá groups) is strong evidence for their stratigraphic correlation to other glacial deposits of Marinoan-Ghaub age (~635 Ma) in the Paraguay Belt.

Another difference between the stratigraphy of the northern and the southern Paraguay belts is the occurrence of limestones and shales from the Tamengo and Guaicurus formations, both with microfossil contents (Gaucher et al., 2003). The Tamengo Formation is a transgressive phase at the top of the Corumbá Group, and it is not represented on the top of the Araras Group, which was covered by the younger glacial rocks of the Serra Azul Formation (Gaskiers glacial age). The depositional age of the Tamengo Formation has been considered Ediacaran (Hahn et al., 1982; Walde et al., 1982; Zaine and Fairchild, 1985; Gaucher et al., 2003), and it is now well constrained by precise SHRIMP ages on zircons extracted from volcanic ashes, yielding an age of 543 ± 3 Ma (Babinski et al., 2008a).

Regarding the composite $\delta^{13}\text{C}$ data, an important negative incursion is associated with the cap-carbonate succession in the Araras Group (Nogueira et al., 2003, 2007a; Alvarenga et al., 2004, 2008; Font et al., 2006). The negative incursion is more important in the first 20 m of the cap dolostone than in the limestone-shale rhythmite of the Guia Formation. These negative $\delta^{13}\text{C}$ excursions have also been found in the lower Corumbá Group (Puga Hill), and they demonstrate how such values are important for correlation between different sections (Boggiani et al., 2003). The positive $\delta^{13}\text{C}$ excursion in the Araras Group is seen in the Bauxi and Nobres areas, and it is probably related to the influence of high organic content.

In the northern Paraguay Belt, stratigraphic and isotopic studies have focused on the glacial and post-glacial deposits linked to the Marinoan-Ghaub glaciation, while in the southern Paraguay Belt, stratigraphic sections and isotopic profiles remain incomplete, precluding a full reconstruction of the evolution of the Paraguay Belt.

ACKNOWLEDGEMENTS

This chapter is a contribution to the International Geological Correlation Programme (IGCP) Project 478 'Neoproterozoic-Early Palaeozoic Events in SW-Gondwana'. In October 2004, a field trip was organised in the Paraguay Belt within the framework of IGCP Project 478 led by C. Gaucher and G. Germs. We thank the enthusiastic participation on this field trip from Gerard Germs, Thomas R. Fairchild, Francisco E.S. Pinho, J.C. Möller, Hans J. Hofmann, Daniel Poiré, Nicolas Beukes, Gonzalo Blanco, Jens Gutzmer, Stefan Schroeder, Andreas Braun, Ginaldo A.C. Campanha, Lucas Warren, Daniel Oliveira, Paulo Tibana, Karina Pamoukaghlian, Jane Nobre and Mario Brito. We thank Detlef Walde, Claudio Gaucher and Alcides N. Sial for constructive reviews.

CAPÍTULO 4

GEOCRONOLOGIA DOS DEPÓSITOS GLACIAIS E CARBONATOS DE CAPA DO CRÁTON DO SÃO FRANCISCO E DA FAIXA ARAÇUAÍ

Os depósitos glaciais criogenianos ocorrem tanto no Cráton do São Francisco quanto na Faixa Araçuaí que bordeja o cráton a leste. Neste capítulo serão apresentados dados geocronológicos U-Pb obtidos em zircões detríticos recuperados de diamictitos da Formação Bebedouro, base do Grupo Una, no Estado da Bahia (Misi, 1979). Estes depósitos têm sido correlacionados aos diamictitos da Formação Jequitaí, que ocorrem na base do Grupo Bambuí (Dardenne, 1978; Karfunkel & Hoppe, 1986), no cráton, e também com os depósitos glaciais descritos na Faixa Araçuaí (Uhlein et al., 1999; Pedrosa Soares et al., 2008).

As idades U-Pb obtidas nos zircões (n=92) dos diamictitos da Formação Bebedouro variam de 874 Ma a 3260 Ma, sendo que a principal área-fonte tem idade de 2,05 Ga. Populações de zircão com idades de 1,2 Ga, 1,57 Ga, 1,8 Ga, e arqueanas também foram observadas (**Figueiredo et al., 2009; item 4.1**). O zircão mais jovem datado tem idade U-Pb de 874 ± 9 Ma e estabelece a idade máxima de deposição dos diamictitos da Formação Bebedouro.

Na faixa marginal os diamictitos constituem as formações Serra do Catuni e Chapada Acauã, sobrepostos aos arenitos da Formação Duas Barras, Grupo Macaúbas, cujos zircões detríticos também foram estudados nesta tese (**Babinski et al., in press; item**

4.2). As idades U-Pb nos zircões ($n=131$) destas 3 unidades litoestratigráficas variaram entre 900 Ma e 2750 Ma. O zircão mais jovem, datado em 900 ± 21 Ma, proveniente dos arenitos da Formação Duas Barras, pré-glacial, define a idade máxima de deposição dos diamictitos do Grupo Macaúbas. A similaridade entre as idades máximas de sedimentação obtidas nos diamictitos depositados sobre o cráton (Formação Bebedouro) e na bacia precursora da Faixa Araçuaí reforçam a correlação proposta na literatura entre estas unidades (Uhlein et al., 1999; Pedrosa Soares et al., 2008).

Na região cratônica, os diamictitos são capeados por carbonatos, conhecidos como carbonatos de capa (Hoffman et al., 1998; Kennedy et al., 1998), os quais registram a rápida mudança do estado de *icehouse* para *greenhouse*. Como a mudança é rápida e os erros associados à idade são elevados, a datação destes carbonatos pode ser considerada como uma estimativa da idade do final do período glacial. Na parte sul do Cráton do São Francisco ocorrem carbonatos de capa muito bem preservados, pertencentes à base da Formação Sete Lagoas, Grupo Bambuí, que forneceram uma idade isocrônica Pb-Pb de 740 ± 22 Ma, interpretada como a idade de deposição dos carbonatos (Babinski et al., 2007; item 4.3).

As idades U-Pb determinadas em cristais de zircão detríticos, recuperados de diamictitos que afloram tanto na região cratônica quanto na Faixa Araçuaí, mostram idades máximas de deposição de 874 ± 9 Ma e 900 ± 21 Ma, respectivamente, as quais são similares dentro do erro analítico. Estes dados, associados à idade de 740 ± 22 Ma obtida nos carbonatos de capa da base da Formação Sete Lagoas, sugerem que os depósitos glaciais das formações Bebedouro, Jequitaí, Carrancas, Serra do Catuni e Chapada Acauã foram depositados durante o evento glacial do Criogeniano Médio, também conhecido como Sturtiano. Além disso, a similaridade das idades máximas obtidas neste trabalho com aquelas reportadas na literatura para o Grupo Vazante (900 Ma; Rodrigues, 2008; Rodrigues et al., 2008, 2011; Pimentel et al., 2011) reforçam a possibilidade que os depósitos glaciais que ocorrem na Formação Santo Antônio do Bonito (Dardenne, 2000) também sejam Sturtianos, como previamente sugerido por Rodrigues et al. (2008) e Azmy et al (2006).

4.1. GEOCRONOLOGIA U-Pb SHRIMP DOS ZIRCÕES DETRÍTICOS DOS DIAMICTITOS DA FORMAÇÃO BEBEDOURO, CRÁTON DO SÃO FRANCISCO

Uma das principais questões na geologia do neoproterozoico está relacionada à extensão e à dinâmica dos sistemas glaciais que estão registrados em todos os continentes. Neste trabalho são apresentadas evidências de transporte por curtas distâncias e fontes de sedimentos localizadas para a Formação Bebedouro, a qual constitui o registro de sedimentação glácio-marinha na parte centro leste do Cráton do São Francisco (SFC). São apresentados dados inéditos sobre a composição de clastos, baseados na contagem de pontos em lâminas delgadas, e datação U-Pb SHRIMP de zircões detríticos e de clastos ígneos. Análises de agrupamento da composição dos clastos revelaram uma forte variação espacial nos diamictitos indicando a presença de geleiras individuais alimentando a bacia. As idades de zircão detríticos mostram populações com idades arqueanas e paleoproterozoicas. O zircão mais jovem, datado em 874 ± 9 Ma, define a idade máxima de sedimentação dos diamictitos. Os resultados mostraram que a proveniência dos diamictitos glaciais é restrita às fontes presentes no cráton e sugerem que a deposição ocorreu em um ambiente similar aos de correntes de gelo de geleiras modernas.

Neoproterozoic glacial dynamics revealed by provenance of diamictites of the Bebedouro Formation, São Francisco Craton, Central Eastern Brazil

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ABSTRACT

One of the main questions on Neoproterozoic geology regards the extent and dynamics of the glacial systems that are recorded in all continents. We present evidence for short transport distances and localized sediment sources for the Bebedouro Formation, which records Neoproterozoic glaciomarine sedimentation in the central-eastern São Francisco Craton (SFC), Brazil. New data are presented on clast composition, based on point counting in thin section and SHRIMP dating of pebbles and detrital zircon. Cluster analysis of clast compositional data revealed a pronounced spatial variability of clast composition on diamictite indicating the presence of

individual glaciers or ice streams feeding the basin. Detrital zircon ages reveal distinct populations of Archean and Palaeoproterozoic age. The youngest detrital zircon dated at 874 ± 9 Ma constrains the maximum depositional age of these diamictites. We interpret the provenance of the glacial diamictites to be restricted to sources inside the SFC, suggesting deposition in an environment similar to ice streams from modern, high latitude glaciers.

Terra Nova, 21, 375–385, 2009

Introduction

Since the rebirth of the hypothesis of global-scale glaciation during the Neoproterozoic (Harland, 1964; Kirschvink, 1992; Hoffman *et al.*, 1998), several models have been proposed for the development of Neoproterozoic glacial deposits and post-glacial carbonate successions that are found on all continents (e.g., Hyde *et al.*, 2000; Hoffman and Schrag, 2002; Eyles and Januszcak, 2004). In the first decade of intense research, most of the field evidence came from the post-glacial carbonates, with vast documentation of anomalous sedimentary structures and isotopic signatures from the cap-carbonates of several successions worldwide (e.g., Allen and Hoffman, 2005; Corsetti and Grotzinger, 2005; Halverson *et al.*, 2005; Kaufman *et al.*, 2006). More recently, some ques-

tions have been raised on whether all the Neoproterozoic diamictite-bearing successions are indeed the record of glaciation (e.g., Eyles and Januszcak, 2004). As a consequence, there is now a renewed effort to document the detailed sedimentology of those successions (e.g., Arnaud and Eyles, 2006; Shields *et al.*, 2007; Williams *et al.*, 2008) and to establish a correlation of glacial deposits of different regions based on geochronological data (e.g., Fanning and Link, 2004; Hoffmann *et al.*, 2004; Condon *et al.*, 2005; Babinski *et al.*, 2007; Bowring *et al.*, 2007).

One of the most important and still poorly investigated aspects of Neoproterozoic diamictite sedimentology is its provenance variability (Allen *et al.*, 2004; Frimmel *et al.*, 2006; Corkoran, 2008). The hypothesis of global ice cover (e.g., Williams *et al.*, 1998; Sohl *et al.*, 1999) and its derivations (e.g. Hyde *et al.*, 2000) would imply great ice sheets covering the continents and ice shelves over the contiguous oceans. These ice sheets would promote large-scale erosion and the transport of sediment over great distances, particularly in ice

streams (e.g., Dredge, 2000). On the other hand, an alternative, tectonic model emphasizes the more localized influence of uplifted rift margin topography in giving rise to glaciers that fed the Neoproterozoic basins (e.g., Prave, 1999; Labotka *et al.*, 2000; Eyles and Januszcak, 2007). If that is the case, one could expect distinct bedrock lithologies in each of the several individual valley glaciers that fed the basin. Thus, the provenance of the glacial and glacially influenced sediments would indicate a strong variation of sources from area to area within the same basin. Analysis of the clastic provenance of Neoproterozoic diamictites can be used to test the models on glacier type and size, and glacier ice dynamics.

This work presents the results of analysis of clastic sediment provenance for the Neoproterozoic diamictites of the Bebedouro Formation of central-eastern Brazil (Fig. 1), based on the recognition of the areal variability in lithology and frequency of pebble- and cobble-sized clasts and point-counting of sandstones in thin sections. Samples were collected on the same stratigraphic level over an

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Fig. 1 Geological sketch map of the four studied areas: North (1) and South Salitre (2), Irecê (3) and Una-Utinga (4) and their location on regional geological sketch of the São Francisco Craton (at the top left), marked by a rectangle at the central-eastern portion of the Craton. (Modified after Alkmin *et al.*, 1993 and Inda and Barbosa, 1978). Outcrops A and B mark the place where the samples BA01-04, BA01-05 and BA01-09 were collected.

area of more than 16 000 km². The results are integrated with new U-Pb SHRIMP zircon geochronology data for pebbles and detrital zircon to interpret the proximity and variability of the sources of glacial sediment and to reconstruct the patterns of the ice flow.

Geological framework

The São Francisco Craton, in central-eastern Brazil, is a stable region within a collage of Neoproterozoic Brasiliano belts formed during the assemblage of the Gondwana Supercontinent. The craton preserves a vast area of Neoproterozoic sedimentary cover with incipient metamorphism and very weak deformation (Almeida, 1967). Different stratigraphic designations have been assigned to this sedimentary cover in the central-eastern and southern portions of the craton, but the simplified stratigraphic column (Fig. 2) is uniformly recognized as consisting of a lowermost glacial influenced succession

(named Jeiquitaí or Bebedouro Formation) overlain by a carbonate-dominated succession (Bambuí Group or Salitre Formation). Evidence for the glacial origin of the lowermost unit is abundant, the most striking of which is a glacially striated surface (e.g., Isotta *et al.*, 1969). Occurrences of glacial deposits are found over an area of more than 900 × 400 km, including thick deposits of the Macaúbas Group in the Araçuaí Fold Belt (Cukrov *et al.*, 2005; Pedrosa-Soares *et al.*, 2008). Regarding their tectonic setting, these successions have been interpreted as the record of a rift phase (Uhlein *et al.*, 1998; Martins-Neto *et al.*, 1999), as intracratonic deposits of unclear tectonic setting (Inda and Barbosa, 1978; Dominguez, 1996), or as the base of a foreland basin (Martins-Neto *et al.*, 2001) of the continental collision recorded in the Brasília Fold Belt to the west.

The study region, in the central-eastern São Francisco Craton, is represented by the Una Group, com-

posed of a lower clastic unit (Bebedouro Formation) and an upper carbonate unit (Salitre Formation), both cropping out in three discontinuous areas (Guimarães, 1996), related to the inner parts of large-scale synclinal folds with north–south oriented axis (Misi, 1979). These three areas are named Salitre (the northernmost one), Irecê (central) and Una-Utinga (southernmost) (Fig. 1).

The Bebedouro Formation includes diamictites, sandstones and mudstones with limestones clasts, which are often interpreted as the record of a glaciomarine environment (e.g., Söfner, 1973; Karfunkel and Hoppe, 1988; Guimarães, 1996). Previous qualitative observations on clast composition revealed a polymict character (Montes *et al.*, 1985) and corroborate the thin section descriptions presented by Kegel (1959), Söfner (1973), Inda and Barbosa (1978) and Misi (1979).

Available geochronological constraints on the Bebedouro Formation include Rb-Sr and K-Ar ages ranging from 900 to 570 Ma, determined on fine fractions separated from the siliciclastic rocks (Macedo and Bonhomme, 1981, 1984). In addition to this, Toulkeridis *et al.* (1999) presented Rb-Sr ages of 700 Ma for the overlying Salitre Formation, which suggests that the Bebedouro Formation was probably deposited during the Sturtian glacial event. The lateral correlation of the dated unit mentioned above with the Sete Lagoas Formation was interpreted as reflecting the widespread deposits of same Sturtian glacial event throughout the southern portion of the craton (Babinski *et al.*, 2007).

Depositional environments

The Bebedouro Formation is composed of two facies associations: melt-out from the sole of the ice-sheet association and a platform marine association (Table 1).

The first facies association is composed mainly of massive and crudely stratified diamictites with silt to silt-sand matrix, with rare, massive, planar-stratified and laminated sandstone intercalations, as well as siltstones with limestones (Fig. 3). The diamictites occur as tabular layers, which are laterally continuous for hundred of metres, comprising a

Fig. 2 Schematic stratigraphic columns of the Proterozoic sedimentary cover of the São Francisco Craton, showing the different designations for the glacial deposits in the southern (to the right) and central-eastern portions (to the left), after Dominguez (1996).

Table 1 Summary of the sedimentary facies, facies associations and the main depositional processes of the Bebedouro Formation.

Code	Facies	Description	Depositional processes	Facies association	Description	Interpretation
Dmm	Massive diamictite	Massive, oligomitic to polimictic, clast-rich diamictites, generally disposed in tabular beds	Melt-out deposits beneath ice-sheet or icebergs	Melt-out from the sole of the ice-sheet	Up to 20 m thick, laterally continuous succession dominated by diamictites (Dmm and Dml), with minor intercalations of sandstones (Sm, Sh, St, Shf and Scr) and mudstones (P-d) in isolated beds or organized in fining-upward metric successions.	Melt-out glaciomarine deposits from the sole of the ice-sheet, dominated by simultaneous settling of fine-grained particles and deposition of coarse particles from melting ice. Occasional density currents, probably originated in ice-contact fans, caused deposition of sandstones and mudstones.
Dms	Stratified diamictite	Crudely stratified, clast-poor, polimictic diamictites, disposed in tabular beds	Fine sediment decantation with local contribution of melt-out			
Sm	Massive sandstone	Fine-to-coarse massive sandstones with sparse outsized clasts, locally in fining-upward beds	Mass-flow deposits, possibly related to density currents originated in ice-contact fans			
Sp	Plan-parallel laminated coarse sandstones	Coarse-grained sandstones with horizontal lamination, locally in fining-upward beds	Sub-aqueous traction currents, possibly related to density currents originated in ice-contact fans			
St	Trough cross-stratified sandstones	Medium-to-coarse-grained sandstones with cross stratification, locally with mudstone intradasts and sparse pebbles	Migration of sub-aqueous dunes with sinuose crests			
Shf	Horizontally laminated fine sandstones	Fine-grained sandstones with horizontal lamination, with rare low angle cross-lamination	Upper-flow regime traction currents, possibly related to density currents originated in ice-contact fans			
Scr	Sandstones with cross-lamination	Fine-grained sandstones with climbing ripples	Simultaneous deposition from traction and suspension from decelerating currents			
P-d	Mudstones and siltstones with limestones	Horizontally laminated mudstones and siltstones with sparse pebbles and granules, locally deforming the lamination by differential compaction	Fine sediment decantation with local contribution of melt-out			
Scr	Sandstones with cross-lamination	Fine-grained sandstones with climbing ripples	Simultaneous deposition from traction and suspension from decelerating currents	Platformal marine	10 to 50 m thick succession dominated by mudstones and siltstones (Pl and Pm) with sandstone intercalations.	Marine platformal deposits dominated by settling of fine-grained particles and occasional sand input from density currents.
Sm	Massive sandstone	Fine-to-coarse massive sandstones with sparse outsized clasts, locally in fining-upward beds	Mass-flow deposits, possibly related to density currents originated in ice-contact fans		Siltstones are often laminated, associated with sand-mud rhythmites and tabular to lenticular-shaped intercalations of fine sandstones (Scr and Sm) up to few centimetres thick and few metres wide (1.5–2 m). Fining-upward beds are common.	
Pm	Massive mudstones and siltstones	Massive, with rare load structures	Settling from suspension, with sporadic fluidized flows			
Pl	Laminated mudstones and siltstones	Horizontal lamination	Settling from suspension

homogeneous succession less than 20 m thick. These diamictites are interpreted as glaciomarine deposits formed under ice shelves, in areas far basinward from the grounding line of the feeding glaciers where melt-out of glacial debris processes dominated.

Fig. 3 Stratigraphic column of the Bebedouro Formation with the two facies associations that crop out in the Una-Utinga area (4). Facies codes: (Dmm) massive diamictite; (Sh) planar-stratified and laminated sandstones; (Sm) massive sandstones with scarce lonestone clasts; (P-d) mudstones and siltstones with lonestones clasts; (Scr) Cross-laminated sandstones; (Pm) massive siltstones; (Pl) laminated mudstones and siltstones. Granulationscale: (c) claystone; (s) siltstone; (SF) fine sandstone; (SM) medium sandstone; (SC) coarse sandstone; (G) granule; (P) pebble; (Gr) gravel; (B) boulder.

The second facies association is characterized by fine-grained deposits with sandstone intercalations and scarce lonestone clasts (Fig. 3). Siltstones are often laminated, associated with sand–mud rhythmites and tabular to lenticular-shaped beds of fine sandstones up to a few centimetres thick and a few metres wide (1.5–2 m). These successions are interpreted as deep (below storm wave base) marine deposits with restricted contribution from melting icebergs. The sandstone intercalations are interpreted as the product of density currents. The whole association is seldom thicker than 10 m, reaching up to 50 m in the Una-Utinga area. The transition between the two facies associations

marks a retrogradational stacking pattern.

Diamictite and sandstone provenance

Provenance analysis of clasts in diamictites was carried out in all three areas of exposition (Table S1). More than 3000 clasts were counted and classified at 31 different locations, enabling statistical analysis and comparison of the provenance of each area. The relative variation in the frequency of lithotypes reveals that the clast composition is very heterogeneous (Figs 4 and 6). Phyllites and schists are almost exclusive to the North Salitre area, related to a local source: the Jacobina Metamorphic Complex, located less than 10 km to the east (Fig. 1), thus corroborating the conclusions of Inda and Barbosa (1978) and Montes *et al.* (1985). In the Irecê area, the provenance revealed the contribution of oolitic carbonates and calcilitites, probably derived from the Mesoproterozoic Espinhaço Supergroup, which is unconformably overlain by the Bebedouro Formation. In the southernmost area (Una-Utinga), clast composition reveals mainly granitoid sources (syeno to monzogranites and charno-enderbitic rocks), with rare amphibole-biotite schist clasts (Fig. 6).

Fig. 4 Dendrograms (Ward's method of linkage and squared euclidean distances) showing the low grade of similarity versus lithotype frequency among the studied areas marked in Fig. 1 as: 1 – North Salitre, 2 – South Salitre, 3 – Irecê, and 4 – Una-Utinga. Pebble lithology data (3000 lithoclasts counted on diamictite facies) were used as variables.

Fig. 5 Conventional Wetherill U-Pb concordia plot of SHRIMP data for zircon grains from pebbles of diamictites from the Bebebour Formation in different areas. Irecê – Area 3 (Sample FIR 32 F – Enderbite), Una-Utinga – Area 4 (Samples FUU 20 B – Granodiorite, FUU 20 E – Granodiorite, 20 A – Monzogranite, 21 F – Monzogranite). The sampling sites are located as outcrops C (Irecê) and D (Una-Utinga) in Fig. 1. Each sample mentioned above refers to a population of igneous zircons extracted from a single pebble.

Fig. 6 Sketch map with the main geotectonic units and ages of the basement of the São Francisco Craton (Gavião Block in yellow, Jequié Block in purple, Itabuna-Salvador-Curaçá in pink and Serrinha Block in light brown). The frequency of the clast compositional data for each site of study is shown in the table to the left. For reasons of map scale, a total of 33 sites of data collection were grouped in 10 representative points (red circles). Geotectonic base map modified after Barbosa and Sabaté (2004).

Cluster analysis of clast lithology data, performed after centred log-ratio transformation to avoid closure effects (Aitchison, 1982), provided a clear picture of the spatial distribution of provenance (Fig. 4), with areas separated by tens of kilometres showing distinctive sources. Indeed, all samples of each of the three studied areas are grouped in separate clusters. Radiogenic isotope dating of zircon crystals (SHRIMP U-Pb technique, Compston *et al.*, 1982) from five representative clasts from the Una-Utinga and Irecê areas reveals only Archean sources (Fig. 5, Table S2). The dating of two monzogranite and two granodiorite clasts sampled from Una-Utinga and an enderbite clast from the Irecê area indicate sources with ages in the 3.3–3.5 Ga range (correlative to the nearby basement of the craton, in the Gavião Block) and one sample of 2627 ± 20 Ma,

which can be attributed to the Jequié Block or to the Itabuna-Salvador-Curaçá Belt (Barbosa and Sabaté, 2004), both at least 150 km to the east of the deposits (Fig. 6).

Point counting of the sand-sized clasts (minerals and lithoclasts) of diamictites and the sandstones were performed in 13 samples from the three studied areas, Table S3). At least 500 points were counted in each sample, following the Gazzi-Dickinson method (Gazzi, 1966; Dickinson, 1970). The composition of the sand-sized fraction of diamictites revealed some distinctive variation from area to area, attributable to a variation in sources. The abundance of carbonate fragments in the samples from the Irecê area contrasts with the scarcity of such lithotype in the other areas. Unfortunately, the fragmentation effect masked the differences of samples with different proportions of

various types of granite, migmatite and gneiss, making the samples from Una-Utinga and Salitre very similar.

The U-Pb SHRIMP analysis of detrital zircon of the matrix of diamictites from the Salitre area (Fig. 1) shows a marked spatial variation in sources, with distinct zircon populations in three samples collected from two outcrops. A main source of *c.* 2050 Ma is represented in all samples, and a source of *c.* 1800 Ma is prevalent in two samples collected at the same outcrop A (BA01-04 and BA01-05, Table S4), but scarce in sample BA01-09, which was collected at outcrop B, 20 km towards north (Fig. 7). Several small zircon populations older than 2050 Ma are found in all samples, but the main peak ages do not coincide. Zircon crystals younger than 1800 Ma are absent in sample BA01-09, Table S5), and compose several populations in samples BA01-04 and BA01-05, which present peaks at 1.57 Ga and 1.21 Ga (Fig. 7). With the exception of mafic dikes of 1000–1100 Ma (D'Agrella-Filho *et al.*, 1990) and gabbroic sills and dikes of 1500 Ma (Babinski *et al.*, 1999; Battilani *et al.*, 2005; Guimarães *et al.*, 2005), Mesoproterozoic sources are not known in the São Francisco Craton. The identification of a zircon grain dated at 874 ± 9 Ma constrains the maximum age of the deposit (Fig. 7).

Discussion and conclusions

Provenance analysis of diamictites of the Neoproterozoic Bebedouro Formation in the central-eastern São Francisco Craton revealed that the main source areas for the sedimentary rocks are inside the Brazilian part of the craton, which is attested by the composition and zircon ages of pebbles.

Diamictites from the same stratigraphic level in three different areas revealed clear differences in pebble provenance, indicating the prevalence of source areas located less than 200 km away from the deposits. This areal heterogeneity of sources is observed in samples distributed in a N–S direction, parallel to the distribution of the basement geological units. An approximately north–south contact separates Palaeoproterozoic to Archean high-grade and granitic rocks (with minor schists and phyllites) to the east

Fig. 7 U-Pb zircon probability age distributions in Ma and Wetherill Concordia diagrams of SHRIMP data for all concordant analysis of detrital zircons collected from diamictites. (Outcrop A) Histogram for $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 41 detrital zircon grains (only ages less than 15% discordant) from samples BA01-04 and BA01-05. (Outcrop B) Histogram for $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 43 detrital zircon grains (only ages less than 20% discordant) from sample BA01-09.

from Mesoproterozoic low-grade quartzites, meta-pelites and minor meta-limestones of the Espinhaço Supergroup to the west (Fig. 1). The correlation of clast lithologies with mapped geological units of the São Francisco Craton corroborates the model of sediment transport from the east to the west proposed by Kegel (1959), Söfner (1973) and Montes *et al.* (1985), as clast lithology frequency is strongly controlled by the lithologies cropping out east of each occurrence. Quartzites, metapelites and limestones of the underlying Espinhaço Supergroup crop out east and west as well of the Salitre and Irecê areas, where the diamictites present abundant quartzite and limestone clasts. On the other hand, gneisses and granites crop out east of the Una-Utinga area and directly underlie the

diamictites, which are dominated by granitic clasts. We therefore interpret an approximately N–S ice grounding line, placed between the glaciomarine deposits and their sources to the east.

The great variation in provenance in short distances could be explained by individual valley or outlet glaciers feeding the basin or by one unique ice sheet with separated fast-flowing ice streams forming distinct E–W clast trains or plumes. Valley glaciers nucleated at high altitude and outlet glaciers present a high proportion of clasts from proximal sources (e.g., Brecke and Goode, 2007), providing a signature of provenance in marine ice-shelves (Anderson and Ashley, 1991). Nevertheless, the great dimension of the basin, the lateral continuity of sedimentary facies for hundreds of kilometres and the absence of clear

tectonic controls on basin evolution do not support the presence of elevated areas at the basin margins, contradicting the model of rift-shoulder glaciers (e.g., Eyles and Janaszczak, 2004). We therefore propose an alternative model of one ice sheet feeding the marine basin with different clast trains formed by individual fast-flowing ice streams. In some situations, such trains can be several hundreds of kilometres wide (e.g., Dredge, 2000), although ice-stream as narrow as a few kilometres have been described (e.g., Ottesen *et al.*, 2005).

Given the glaciomarine origin of the deposits of the Bebedouro Formation, additional mixing of sediment would be expected in the moving ice shelf (e.g., Talarico *et al.*, 2007) and by the action of marine currents on icebergs and sediment plumes (e.g., Diekmann and Kuhn, 1999; Phillips and Grantz, 2001). Sediment plumes transporting particles in suspension could explain in part the difficulty in recognizing the same heterogeneity of source region for the sand-sized particles that is so evident in the pebble- to cobble-sized clasts.

Populations of Archean detrital zircons (Fig. 7) ranging from 2580 to 3348 Ma corroborate the pebble dating data (2627 to 3451 Ma), pointing to sources from the basement of the São Francisco Craton (Barbosa and Sabaté, 2004). Populations in the interval of 2400 and 1800 Ma (Fig. 7) can be attributed to source rocks located in the southern portion of the São Francisco Craton (e.g., Teixeira and Figueiredo, 1991), possibly reworked from the sandstones of the Espinhaço Supergroup. The source of the 1500 Ma zircon grains (Fig. 7) may be the gabbroic sills and dikes that cross cut the sedimentary rocks of the Espinhaço Supergroup. The 1000 to 1200 Ma sources are still unknown in the São Francisco and in the Congo Craton as well (Frimmel *et al.*, 2006).

The youngest zircon dated at 874 ± 9 Ma sets the maximum depositional age of the diamictites from the Bebedouro Formation, reinforcing the Sturtian age of these glacial deposits. This source could be the recently identified 875 Ma anorogenic granite (Silva *et al.*, 2008) related to the continental rift stage of the precursor

basin of the Araçuaí fold belt, which is located *c.* 600 km to the south, at the margin of the São Francisco Craton (Pedrosa-Soares *et al.*, 2001). Another possible source would be the rhyolites (879 Ma) and granites (877 Ma) of Nchanga, in the Katangan belt, located on the eastern margin of the Congo Craton (Batumike *et al.*, 2009), which would require sediment transport over great distances, presumably crossing the Araçuaí fold belt that lay east of the basin. The great proportion of clasts and detrital zircon from sources within the São Francisco craton favours the first hypothesis. Zircons with ages in the 1.0 to 1.2 Ga interval compose the only population that cannot be identified, either in the São Francisco or the Congo cratons. One possible source of these Grenvillian-aged detrital material is the SW Amazon craton (Tohver *et al.*, 2005), a scenario we consider improbable, given the presumed width of the Clymene Ocean at the time of deposition of the Bebedouro Fm. (Trindade *et al.*, 2006). We therefore conclude that the sources of the Bebedouro diamictites are mostly restricted to the immediate basement of the glacial deposits. This scenario confirms the current palaeogeographical models that place the São Francisco and the Congo cratons in an isolated plate from the Palaeoproterozoic to the latest Neoproterozoic (e.g., Tohver *et al.*, 2006), peripheral to the main continental mass of Rodinia.

Our data from the São Francisco craton, together with other records of glacial deposits of Sturtian age in the Congo Craton (e.g., Frimmel *et al.*, 2006; Poidevin, 2007), suggest that almost all the São Francisco – Congo continent was covered by an ice sheet in some period of the Neoproterozoic. This ice sheet expanded towards the west, reaching the São Francisco Craton, where fast-flowing ice streams eroded bedrock and fed a marine ice shelf with calving icebergs. This scenario indicates ice sheet dynamics very similar to that of modern, high latitude glaciers, implying an active hydrological cycle, contrary to some models of extreme global glaciation. The method applied here may be used to interpret aspects of ice dynamics in other Neoproterozoic glacial successions, enabling more precise models

for the intriguing glacial events of the period.

Acknowledgements

We thank FAPESP for research grant (06/61433-8), which supported this work. We thank the graduate students Cauê Cioffi, Bernardo Freitas and Thiago Piacentini for the clast counting in the field. Earlier versions of this paper were improved by discussions with Liliane Janikian, Ricardo Trindade and André Sawakuchi. We appreciate the insightful reviews of Johan Petter Nystuen, Michael Houmark-Nielsen, as well as the editorial handling of Max Coleman.

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Received 19 November 2008; revised version accepted 11 June 2009

Supporting Information

Additional Supporting Information may be found in the online version of this article:

Table S1 Pebble compositional data for 33 counting sites of diamictites of the Bebedouro Formation.

Table S2 Summary of SHRIMP U-Pb zircon results for five lithoclasts sampled from diamictite facies at Irecê and Una-Utinga areas.

Table S3 Compositional data of thin sections (500 counting points per thin section) of diamictites and sandstones of the Bebedouro Formation. The components presented here are only the lithic fragments. For cluster analysis, the matrix as well as the accessory minerals values were not considered.

Table S4 Summary of SHRIMP U-Pb zircon results for samples BA01-04 and BA01-05 (Outcrop A in Figure 1).

Table S5 Summary of SHRIMP U-Pb zircon results for sample BA01-09 (Outcrop B in Figure 1).

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4.2. GEOCRONOLOGIA U-Pb SHRIMP DOS ZIRCÕES DETRÍTICOS DOS DIAMICTITOS DO GRUPO MACAÚBAS, FAIXA ARAÇUAÍ

Diamictitos glaciais do Grupo Macaúbas ocorrem no domínio ocidental do orógeno Araçuaí, a leste do Cráton do São Francisco. Os diamictitos também ocorrem neste cráton e na contraparte africana do orógeno Araçuaí, a Faixa Oeste Congo. Grãos de zircão detríticos recuperados da matriz dos diamictitos e arenitos do Grupo Macaúbas foram datados pelo método U-Pb SHRIMP. O estudo geocronológico estabelece a idade máxima de deposição dos diamictitos glaciais há 900 Ma e indica que múltiplas fontes forneceram sedimentos para a bacia Macaúbas, com idades variando de 900 a 2800 Ma. Idades modelo T_{DM} Sm-Nd determinadas em amostras de rocha total variam de 1,8 a 2,5 Ga e se tornam mais antigas em direção ao topo da seção. Comparação destes dados com aqueles obtidos na área cratônica sugerem que estes depósitos glaciais podem ser correlacionadas com diamictitos das formações Jequitaí e Carrancas no Cráton do São Francisco e aos da Formação *Lower Mixtite* do Grupo Oeste Congo, expostos na África. A fonte de 900 – 1000 Ma é mais provavelmente representada pelas rochas vulcânicas e granitóides dos grupos *Zadinian* – *Mayumbian* do Grupo Oeste Congo. Contudo, uma das mais importantes fontes, com idades entre 1,1 – 1,3 Ga, não foi identificada no Cráton São Francisco – Congo. Possíveis fontes destes grãos podem estar localizadas na África ou na Faixa Brasília, a oeste do Cráton do São Francisco.

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Neoproterozoic glacial deposits from the Araçuaí orogen, Brazil: Age, provenance and correlations with the São Francisco craton and West Congo belt

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ARTICLE INFO

Article history:

Received 23 November 2010

Received in revised form 15 March 2011

Accepted 5 April 2011

Available online xxxx

Keywords:

U–Pb SHRIMP geochronology

Neoproterozoic glaciations

Macaúbas Group

Brazil

Araçuaí–West Congo orogen

ABSTRACT

Glacigenic diamictite successions of the Macaúbas Group are widespread in the western domain of the Araçuaí orogen, east of the São Francisco craton (Brazil). Diamictites also occur on this craton and in the African counterpart of the Araçuaí orogen, the West Congo belt. Detrital zircon grains from the matrix of diamictites and sandstones from the Macaúbas Group were dated by the U–Pb SHRIMP technique. The geochronological study sets the maximum depositional age of the glacial diamictites at 900 Ma, and indicates multiple sources for the Macaúbas basin with ages ranging from 900 to 2800 Ma. Sm–Nd T_{DM} model ages, determined on whole rock samples, range from 1.8 Ga to 2.5 Ga and get older up-section. Comparison of our data with those from the cratonic area suggest that these glacial deposits can be correlated to the Jeiquitaí and Carrancas diamictites in the São Francisco craton, and to the Lower Mixtite Formation of the West Congolian Group, exposed in Africa. The 900–1000 Ma source is most probably represented by the Zadinian–Mayumbian volcanic rocks and related granites from the West Congo belt. However, one of the most voluminous sources, with ages in the 1.1–1.3 Ga interval, has not been detected in the São Francisco–Congo craton. Possible sources for these grains could occur elsewhere in Africa, or possibly from within the Brasília Belt in western central Brazil.

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1. Introduction

Neoproterozoic glacial deposits are found in almost all continental shields (Chumakov and Elston, 1989), thus leading some authors to propose episodes of a panglacial world (Hoffman, 2009). But the existence of such extreme glacial scenarios still requires a better knowledge of the depositional age and provenance of glacigenic deposits which are mostly represented by diamictites.

In Brazil, diamictites are remarkably abundant in the western domain of the Araçuaí orogen, located to the east of the São Francisco craton (Fig. 1). They belong to the Macaúbas Group (Fig. 2), a thick volcano-sedimentary succession that represents a precursor basin of the Araçuaí orogen, and includes pre-glacial fluvial to marine sediments, mostly glacio-marine diamictites with minor rift-related volcanic rocks, and post-glacial passive margin sediments and oceanic deposits (Karfunkel and Hoppe, 1988; Pedrosa-Soares et al., 1992,

2001, 2008; Uhlein et al., 1999). In its original definition (e.g., Moraes and Guimarães, 1931), the name Macaúbas should be exclusively applied to refer to the extensive unit with diamictite formations that underwent regional deformation and metamorphism within the Araçuaí orogen. But diamictites also occur in relatively small areas on the São Francisco craton (Fig. 1), where the best evidence of glaciation has been reported from the Jeiquitaí and Bebedouro Formations. These include striated and grooved pavements, glacio-terrestrial sediments, faceted and striated flat-iron-shaped cobbles and pebbles, dropstones, and esker-like bodies (e.g., Figueiredo et al., 2009; Gravenor and Monteiro, 1983; Hettich and Karfunkel, 1978; Isotta et al., 1969; Karfunkel et al., 2002; Karfunkel and Hoppe 1988; Martins-Neto et al., 1999, 2001; Martins-Neto and Hercos, 2002; Montes et al., 1985; Rocha-Campos and Hasui 1981; Uhlein et al., 1999; Viveiros and Walde 1976). The Carrancas diamictite occurs in only a few outcrops and boreholes where it is covered by the Sete Lagoas Formation (Kuchenbecker et al., 2010; Schöhl, 1972; Tuller, 2009). This diamictite forms the lowermost unit of the pelite-carbonate cratonic cover (Bambu Group), which includes a cap carbonate dated at 740 ± 22 Ma (Babinski et al., 2007).

The stratigraphic correlation of the Macaúbas deposits with the glacial units found in the São Francisco craton is masked by the structural complexity of the Araçuaí belt. The Macaúbas Group was thrust over the pelite-carbonate cover of the São Francisco craton and

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Fig. 1. Location of the Araçuaí orogen in Brazil (CSF, São Francisco craton) and in Western Gondwana together with its African counterpart, the West Congo belt. The exposure region of Macaúbas diamicrites (triangles) is shown in the palaeotectonic fit of the Araçuaí–West Congo orogen and related São Francisco–Congo craton. B, C and J indicate locations of Bebedouro, Carrancas and Jequitai diamicrites on the São Francisco craton.

does not show any direct field relation with the Jequitai and Carrancas diamicrites (Figs. 1 and 3). The same complexities also appear when trying to correlate these units with diamicrites found in the West Congo belt (Fig. 1), the counterpart of the Araçuaí orogen located in central west Africa (Pedrosa-Soares et al., 2008; Trompette, 1994).

The age and provenance of these diamicrites in both continents is still poorly constrained. In a regional scale, these data would allow a more refined correlation between the diamicrite successions of the Araçuaí orogen, São Francisco craton and West Congo belt. In global scale, it would enable the correlation to the Neoproterozoic glacial events of one of the largest and thickest diamicrite successions of this period (e.g. Hoffman and Li, 2009). Here we present new U–Pb SHRIMP ages obtained on detrital zircon grains recovered from pre-glacial sandstones and glaciogenic diamicrites of the Macaúbas Group. In addition, an attempt to date mafic volcanics interlayered with diamicrites of the Macaúbas Group was done. The aims of our work are (i) defining the maximum depositional age of glacial rocks and tentatively correlate them to the global Neoproterozoic glacial periods/events (e.g. Kaigas, Sturtian, Marinoan, and Gaskiers); and (ii) obtain information about the provenance of sediments deposited in the Macaúbas basin.

2. Geotectonic setting

The Araçuaí–West Congo orogen comprises the Neoproterozoic orogenic domains located to the southeast of the São Francisco craton in Brazil, and to the southwest of the Congo craton in Africa (Fig. 1). Prior to the opening of the South Atlantic Ocean, the São Francisco and Congo cratons were connected by means of a continental bridge, the Bahia–Gabon cratonic bridge (Alkmim et al., 2006; Cordani et al., 2003; Pedrosa-Soares et al., 2001, 2008; Porada, 1989; Trompette, 1994). Considering that the youngest orogenic event in the cratonic bridge occurred around 2.0 Ga, the continental link between the São Francisco and Congo cratons must have been formed during the Palaeoproterozoic and remained until the onset of the Atlantic opening. To the south of the Bahia–Gabon cratonic bridge, the Araçuaí–West Congo orogen evolved inside an embayment (an inland-sea basin partially floored by oceanic crust) into the São Francisco–Congo palaeocontinent. The evolution of this confined orogen lasted from the opening of the precursor basin at the beginning of the Neoproterozoic until late orogenic processes near the Cambrian–Ordovician boundary (Alkmim et al., 2006; Pedrosa-Soares et al., 2001, 2008).

During the Cretaceous opening of the South Atlantic Ocean, the Araçuaí–West Congo orogen was split up into two quite different but complementary counterparts: the Araçuaí orogen and the West Congo belt (Pedrosa-Soares et al., 2008). However, both counterparts inherited diamicrite successions that together with the confined nature of the precursor basin system suggest a similar paleoenvironmental evolution.

2.1. Regional geology

The Macaúbas Group represents a very extensive and thick volcano-sedimentary pile (probably thicker than 12 km; Uhlein et al., 1999) that filled in the precursor basin of the Araçuaí orogen (Fig. 2). Proximal units of the Macaúbas Group and late Tonian anorogenic intrusions record the continental rift stage of the Macaúbas basin (Pedrosa-Soares et al., 2001, 2008; Silva et al., 2008). In the southern part of this basin (south of latitude 17°S), the rift evolved into a narrow oceanic basin, represented by the distal Macaúbas Group and associated mafic–ultramafic ophiolite slivers (Pedrosa-Soares et al., 1992, 2001, 2008; Queiroga et al., 2007).

Based on the occurrence of glaciogenic diamicrites, the Macaúbas Group is subdivided, from base to top and west to east (Figs. 2 and 3), into a pre-glacial diamicrite-free succession (Matão, Duas Barras and Rio Peixe Bravo formations), a glaciogenic succession (Serra do Catani, Nova Aurora and Lower Chapada Acauã formations), and a post-glacial succession (Upper Chapada Acauã and Ribeirão da Folha formations).

Fig. 2. Geological sketch map showing the distribution of the most part of the Macaúbas Group in the Araçuaí orogen (modified from Pedrosa-Soares et al., 2007). A-B, location of section shown in Fig. 3.

The basal deposits of the Macaúbas Group are represented by the Matão, Duas Barras and Rio Peixe Bravo formations (Fig. 2). The Matão Formation consists of breccia and conglomerate, rich in sandstone pebbles and cobbles, covered by sandstone with conglomeratic lenses. Erosive unconformity and normal faults outline the contact between the basement (locally represented by eolian sandstone of the Espinhaço Supergroup) and the Matão Formation, which records sedimentation under unstable tectonic conditions related to the early rift stage of the Macaúbas basin and contains detrital zircons with ages ranging from 1.16 Ga to 2.80 Ga (Martins, 2006; Martins et al., 2008). The youngest detrital zircon dated at 1.16 Ga defines the maximum depositional age of this formation (Martins et al., 2008). The Duas Barras Formation consists of sandstone and conglomeratic sandstone with variable contents of mica, feldspar, iron oxide and/or lithic fragments, quartz

sandstone, conglomerate and rare pelite. It shows fluvial and shallow marine sedimentary facies and bimodal (NW–SE and SE–NW) paleo-current sets (Grossi-Sad et al., 1997; Martins, 2006; Martins et al., 2008; Noce, 1997). The Rio Peixe Bravo Formation consists of micaceous, ferruginous and/or feldspathic sandstone, pelite locally rich in hematite and/or graphite, and rare clast-supported conglomerate (Grossi-Sad et al., 1997; Viveiros et al., 1979). These three units show no evidence of glaciation and represent fluvial to marine sedimentation during the continental rift stage of the Macaúbas basin (Martins et al., 2008; Noce et al., 1997; Pedrosa-Soares et al., 2008).

A regional discontinuity occurs between the diamictite-free Duas Barras Formation and the overlying Serra do Catuni Formation, the most proximal glaciogenic unit of the Macaúbas Group (Karfunkel and Hoppe, 1988; Uhlein et al., 1999). Locally, the Serra do Catuni diamictite shows

Fig. 3. Geological sketch section showing stratigraphic relations in the Macaúbas Group, and the thrust front boundary between the Araçuai orogen and São Francisco craton (A–B location in Fig. 2).

an erosional channelized contact with the Duas Barras Formation, but normal and tectonic contacts are also observed (Grossi-Sad et al., 1997; Noce, 1997).

The Serra do Catuni Formation is a very extensive and homogeneous unit that persistently occurs for more than 400 km in the N–S direction, along the western border of the Araçuai orogen (Fig. 2). This formation is very rich in massive diamictite with a poorly sorted, muddy-sandstone matrix composed of detrital quartz, K-feldspar and carbonate. The metamorphic foliation is marked mainly by fine-grained muscovite and stretched quartz. The clasts, ranging in size from granules to boulders, are also poorly sorted in texture and composition (milky quartz, sandstone, granitoid, carbonate and mafic rock). Faceted and striated flat-iron-shaped cobbles and pebbles can also be found in the Serra do Catuni diamictite. Lenses of massive sandstone up to 2 m thick appear mainly in the upper part of the diamictite pile. This sandstone is poorly sorted and consists of quartz with minor K-feldspar, carbonate and iron oxides (Grossi-Sad et al., 1997; Martins, 2006).

The Nova Aurora Formation overlies the pre-glacial Rio Peixe Bravo Formation and is a distal correlative of the Serra do Catuni Formation (Fig. 2). This unit comprises diamictite and thick layers of diamictitic iron formation, with minor graded sandstone and rare pelite (Grossi-Sad et al., 1997; Pedrosa-Soares and Oliveira, 1997; Uhlein et al., 1999; Vilela, 2010; Viveiros et al. 1979).

The Chapada Acauã Formation includes a lower diamictite unit and an upper diamictite-free unit (Fig. 2). The Lower Chapada Acauã Formation consists of stratified diamictite, graded sandstone, pelite and mafic volcanic rocks (Gradim et al., 2005; Grossi-Sad et al., 1997; Pedrosa-Soares et al., 1992). A few carbonate lenses occur at the top of this formation (Grossi-Sad et al., 1997; Pedrosa-Soares and Oliveira, 1997). The mafic volcanic rocks, metamorphosed to greenschists, show pillow structures and other features of subaquatic flows (Gradim et al., 2005; Schrank et al., 1978; Uhlein et al., 1998). They have tholeiitic basalt protoliths with a dominant within-plate geochemical signature (Gradim et al., 2005), Sm–Nd T_{DM} model ages of c. 1.5 Ga, and contain detrital and xenocrystic zircons with ages ranging from Archean to Mesoproterozoic (Babinski et al., 2005). The youngest dated zircon sets the maximum age of the volcanism at 1.16 Ga. It is interpreted as a transitional mafic magma that migrated through thinned continental crust in an extensional marine basin during the transitional phase from late rift to the early passive margin of the Macaúbas basin (Pedrosa-Soares et al., 2008).

The Lower Chapada Acauã Formation gradually passes upwards to the diamictite-free package of the Upper Chapada Acauã Formation, which is formed by a sandstone–pelite succession (Fig. 2). It is interpreted as a post-glacial unit deposited in a shelf environment during the passive margin stage of the Macaúbas basin (Grossi-Sad et al., 1997; Martins-Neto et al., 2001 and Noce et al., 1997; Pedrosa-Soares and Oliveira, 1997). To the east, this unit gives way to the fine-grained

turbidites of the Ribeirão da Folha Formation (Pedrosa-Soares et al., 1992, 2008). The Ribeirão da Folha Formation is a diamictite-free unit and includes distal passive margin and ocean floor deposits (Fig. 2). The age of deposition of these sediments is constrained by the U–Pb age of 660 ± 29 Ma, obtained from euhedral crystals within a meta-plagiogranite associated with meta-mafic rocks, and interpreted as the time of magmatic crystallization of an ophiolite sliver (Queiroga et al., 2007).

The whole Macaúbas Group was affected by Ediacaran–Cambrian orogenesis and the oldest ages of c. 580 Ma from syn-collisional metamorphic–anatectic rocks set the upper limit of sedimentation for the succession (Pedrosa-Soares et al., 2008 and references therein).

3. Analytical procedures

Sm and Nd isotopic analyses were carried out at the Geochronological Research Center of the São Paulo University. Whole rock samples were powdered and dissolved in HF, HNO₃, and HCl media. Separation and purification of Sm and Nd were carried out by standard ion exchange procedures using RE and LN Eichron resins. Sm and Nd isotope data were obtained on a multi-collector MAT262 mass spectrometer. The measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to 0.7219. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of La Jolla and BCR-1, measured in the laboratory, are 0.511849 ± 0.000025 (1σ) and 0.512662 ± 0.000027 (1σ), respectively. Total blanks for Sm and Nd are 100 and 120 pg, respectively. The Sm–Nd model ages reported here are based on a depleted mantle model (DePaolo, 1981) and are interpreted as mean crustal residence ages.

Zircon grains were separated using standard heavy liquid and magnetic techniques at the Geochronological Research Center, University of São Paulo. The U–Pb analyses were carried out on the SHRIMP II ion microprobe at the Beijing SHRIMP Centre, Institute of Geology, Chinese Academy of Geological Sciences. Grains were mounted in an epoxy disk with the Temora zircon standard (417 Ma; Black et al., 2003) and polished to expose their centers. Internal structures of zircon grains were revealed by cathodoluminescence (CL) images obtained at the Institute of Mineral Resources, Chinese Academy of Geological Sciences. The analytical procedures were similar to those described by Williams (1998). Four or five scans through the mass stations were made for each age determination. U abundance was calibrated using the standard SL13 (U = 238 ppm, Williams, 1998) and $^{206}\text{Pb}/^{238}\text{U}$ ratio was calibrated using the standard Temora ($^{206}\text{Pb}/^{238}\text{U}$ age = 417 Ma; Black et al., 2003). Decay constants used for age calculation are those recommended by Steiger and Jäger (1977). Measured ^{204}Pb was applied for the common lead correction, and data processing was carried out using the Squid and Isoplot programs (Ludwig, 2001). The uncertainties for the measured ratios are given at the one sigma level. All obtained data were plotted on the concordia diagram, but only zircon ages more concordant

Table 1

Sm–Nd isotopic data from whole-rock samples from the Macaúbas Group.

Sample ID	Rock	Geological unit	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	Error	$^{143}\text{Nd}/^{144}\text{Nd}$	Error 2 σ	Age T_{DM} (Ga)	ϵ_{Nd}
MG05-04	Sandstone	Duas Barras	0.710	4.219	0.1017	0.0006	0.511645	0.000010	1.9	−19.4
MG08-06	Quartz-sandstone	Duas Barras	0.878	5.550	0.0956	0.0006	0.511618	0.000009	1.8	−19.9
MG05-01	Diamictite	Serra do Catuni	2.281	11.915	0.1158	0.0004	0.511599	0.000009	2.3	−20.3
MG05-02	Diamictite	Serra do Catuni	2.140	10.963	0.1181	0.0004	0.511613	0.000008	2.3	−20.0
MG08-09	Diamictite	Chapada Acauã Inf.	2.904	15.970	0.1100	0.0007	0.511531	0.000011	2.2	−21.6
MG08-11	Mica-rich wacke	Chapada Acauã Inf.	1.544	8.204	0.1138	0.0007	0.511573	0.000007	2.3	−20.8
MG08-12	Wacke	Chapada Acauã Sup.	2.022	10.772	0.1135	0.0007	0.511383	0.000008	2.5	−24.5
MG05-03	Basalt	Chapada Acauã Inf.	3.411	13.906	0.1483	0.0005	0.512302	0.000012	1.7	−6.6
MG05-23	Clast of basalt	Chapada Acauã Inf.	4.778	19.413	0.1488	0.0005	0.512329	0.000010	1.7	−6.0
R5.2 ^a	Basalt	Chapada Acauã Inf.	1.119	4.751	0.1424	0.0005	0.512330	0.000009	1.5	−6.0

^a Data from Babinski et al., 2005.

than 85% were used in the histogram plots. Since all U–Pb ages are older than 800 Ma, they are all mentioned as $^{207}\text{Pb}/^{206}\text{Pb}$ ages, unless referred otherwise, in order to avoid misunderstanding related to discordant ages.

4. Results

4.1. Sm–Nd data

Sm–Nd isotopic data were obtained on samples for all units of the Macaúbas Group, except the Matão Formation (Table 1, Fig. 4). Sandstones of the pre-glacial units show T_{DM} ages of 1.8–1.9 Ga, and the diamictites and sandstones from the Serra do Catuni and Lower Chapada Acauã Formations yield older T_{DM} ages of 2.2 Ga to 2.3 Ga; all samples show very negative (−20 to −24) $\epsilon_{\text{Nd}(0)}$ values, consistent with the predominance of Palaeoproterozoic rocks as the source of the sediments. These results suggest that similar source areas supplied sediments for the whole diamictite pile, but younger sediments could have predominated in the lowermost non-glacial units thus lowering slightly their T_{DM} model ages. In contrast, the T_{DM} ages determined on mafic volcanics with tholeiitic affinity from the Lower Chapada Acauã Formation are between 1.7 Ga and 1.5 Ga, and have less negative (−6) $\epsilon_{\text{Nd}(0)}$ values. These data are in agreement with those previously obtained by Babinski et al. (2005) and reinforce the idea that the mafic magma was contaminated by the thinned continental crust of the Macaúbas rift system during its ascent.

4.2. U–Pb SHRIMP data

Zircon grains separated from four clastic sedimentary rocks and from a mafic volcanic rock were dated. Given that our main objective was to

determine the maximum depositional age of glacial deposits, we selected for analyses: (i) the sandstones of the Duas Barras and Rio Peixe Bravo Formations, two of the lowermost units of the Macaúbas Group, (ii) the overlying glacial diamictites of the Serra do Catuni Formation, and (iii) a mafic volcanic rock with pillow structures from the Lower Chapada Acauã Formation (Fig. 2).

4.2.1. Sandstone from the Duas Barras Formation (MG05-04)

The detrital zircons separated from this quartzite are normally elongated and rounded, and show oscillatory zoning on the CL images (Fig. 5). Their sizes range from 80 to 300 μm . The 23 detrital zircon grains separated from this sandstone yielded U–Pb ages ranging from 2550 to 900 Ma (Table 2; Fig. 6), but most of them fall into two age intervals of 1.0 to 1.25 Ga (40% of the population), and 1.8 to 2.1 Ga (40% of the population), suggesting that those were the ages of the most important sources of the sediments (Fig. 7). The youngest zircon showed a $^{238}\text{U}/^{206}\text{Pb}$ age of 900 ± 21 Ma which is considered the maximum depositional age of the pre-glacial sandstones from the Duas Barras Formation.

4.2.2. Sandstone from the Rio Peixe Bravo Formation (OPU 2654)

Most zircon grains separated from this sample are subhedral to rounded, and are less than 200 μm long. The CL images show oscillatory zoning in most of the grains; few zircons have homogeneous internal structures (Fig. 5). Eighteen grains were analyzed from this sandstone and fourteen grains cluster at ca. 2.1 Ga yielding a $^{207}\text{Pb}/^{206}\text{Pb}$ mean age of 2097 ± 44 Ma; 3 grains have younger ages and one grain gave an older age (Table 3, Fig. 6). In the frequency histogram the 2.1 Ga peak is prominent (Fig. 7), and the distribution of ages found in these distal deposits contrasts with that observed in the more proximal units. The

Fig. 4. Nd evolution diagram for different lithostratigraphic units of the Macaúbas Group.

Fig. 5. Cathodoluminescence images of analyzed detrital zircon crystals.

Table 2

U–Pb SHRIMP data from detrital zircons of sample MG05-04 (Dua's Barras Formation).

Grain spot	U (ppm)	Th (ppm)	²³² Th/ ²³⁸ U	²⁰⁶ Pb* (ppm)	²⁰⁶ Pbc%	Radiogenic ratios						Age (Ma)				
						²⁰⁶ Pb/ ²³⁸ U	±%	²⁰⁷ Pb/ ²³⁵ U	±%	²⁰⁷ Pb/ ²⁰⁶ Pb	±	ρ	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	± % Disc.
1.1	215	193	0.93	71.4	0.09	0.385	2.10	6.58	2.2	0.1239	0.76	0.938	2101	37	2013	14 –4
2.1	190	152	0.83	50.5	0.47	0.309	2.10	5.68	2.6	0.1336	1.62	0.790	1733	32	2146	28 19
3.1	245	260	1.10	75.1	0.08	0.356	2.10	5.61	2.2	0.1142	0.81	0.931	1964	35	1867	15 –5
4.1	102	55	0.55	34.2	0.21	0.387	2.20	6.67	2.5	0.1250	1.21	0.873	2110	39	2029	21 –4
5.1	100	79	0.82	33.1	0.14	0.386	2.20	6.68	2.6	0.1256	1.46	0.829	2103	39	2037	26 –3
6.1	122	103	0.87	21.5	0.12	0.205	2.30	2.34	2.8	0.0825	1.72	0.797	1204	25	1257	34 4
7.2	110	42	0.40	16.9	0.71	0.178	0.86	1.83	3.7	0.0744	3.63	0.229	1057	8	1052	73 0
8.1	80	143	1.86	32.6	0.25	0.475	2.20	11.10	2.5	0.1694	1.20	0.880	2506	46	2552	20 2
9.1	238	263	1.14	73.8	0.09	0.361	2.10	6.08	2.3	0.1223	0.83	0.931	1986	36	1990	15 0
10.1	176	152	0.90	29.4	0.38	0.194	2.10	2.04	3.0	0.0763	2.06	0.717	1142	22	1104	41 –4
11.1	130	155	1.23	23.9	0.27	0.213	2.20	2.36	2.8	0.0802	1.83	0.762	1246	24	1201	36 –4
12.1	97	45	0.47	16.5	0.73	0.196	2.20	2.07	3.9	0.0768	3.19	0.572	1151	23	1117	64 –3
12.2	93	47	0.52	19.4	0.45	0.242	0.83	2.95	2.5	0.0883	2.40	0.326	1399	10	1390	46 –1
13.1	200	174	0.90	49.8	0.18	0.290	2.10	3.94	2.4	0.0988	1.16	0.873	1639	30	1601	22 –2
14.1	255	170	0.69	72.7	0.18	0.332	2.10	5.52	2.3	0.1207	0.84	0.928	1846	34	1967	15 6
15.1*	55	63	1.20	15.3	0.41	0.324	2.30	5.58	11	0.1248	10.75	0.213	1811	37	2025	190 11
16.1	177	138	0.81	26.8	0.19	0.176	2.10	1.76	2.7	0.0723	1.62	0.795	1046	20	995	33 –5
16.2	172	116	0.69	27.4	0.78	0.183	0.76	1.88	2.4	0.0743	2.25	0.319	1086	8	1050	45 –3
17.1	70	179	2.64	9.02	0.44	0.150	2.50	1.45	4.6	0.0701	3.86	0.548	900	21	930	79 3
18.1	62	119	1.97	9.26	2.28	0.169	1.60	1.72	9.8	0.0739	9.67	0.161	1005	15	1038	200 3
19.1	78	83	1.11	22.8	0.49	0.339	1.10	5.1	2.3	0.1090	1.99	0.484	1883	18	1784	36 –6
20.1	79	84	1.09	30	0.70	0.437	1.10	8.27	2.5	0.1371	2.27	0.423	2339	21	2191	39 –7
21.1	140	115	0.85	48.2	0.09	0.402	0.81	6.86	1.3	0.1239	0.97	0.638	2176	15	2014	17 –8
22.1	108	41	0.40	27.2	0.66	0.291	0.97	3.86	2.6	0.0961	2.37	0.379	1647	14	1549	44 –6
23.1	171	95	0.57	33.1	0.29	0.224	1.10	2.47	2.3	0.0801	2.02	0.461	1303	12	1198	40 –9

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.

Error in Standard calibration was 0.50 and 0.36% (not included in above errors).

(1) Common Pb corrected using measured ²⁰⁴Pb/²⁰⁶Pb ratio.* Analysis 15.1 has large error on the ²⁰⁷Pb/²³⁵U ratio and was not plotted on the U–Pb concordia diagram.

proximity of these deposits to the Palaeoproterozoic Porteira block suggests that it was likely the source of the sediments (Fig. 2).

4.2.3. Diamictite from the Serra do Catuni Formation (MG05-01 and MG05-02)

The detrital zircon grains separated from the matrix of diamictite MG05-01 sample are elongated to rounded, and their size range from 140 to 260 μm. The CL images show clear oscillatory zoning in most of the crystals (Fig. 5). Most of the grains are rounded indicating that they have been transported for long distances. However, 2 grains (spot 6.1 = 1.36 Ga, and spot 14.1 = 1.59 Ga; Fig. 5) show slightly euhedral shapes suggesting that their source rocks were not far from the depositional area. Forty three U–Pb ages were determined on thirty nine zircon grains (Table 4, Fig. 6). The ²⁰⁷Pb/²⁰⁶Pb ages ranged from 885 ± 70 Ma to 2740 ± 22 Ma, and despite the strong variation in ages, an expressive number of grains concentrated into two main intervals of 1.15 Ga to 1.4 Ga, and 1.8 Ga to 2.2 Ga (Fig. 7). The youngest age was determined on grain 20 where two analyses were carried out (spots 20.2 and 20.3; Table 4). Both of them are discordant and show ²⁰⁷Pb/²⁰⁶Pb ages of 885 ± 70 Ma (5% reverse discordant) and 1021 ± 51 Ma (9% discordant). However, their ²³⁸U/²⁰⁶Pb ages are in agreement at 933 ± 8 Ma, and we assume that this is the best estimate for the crystallization age of the zircons and, consequently, for the maximum depositional age of this glacial diamictite.

Zircon grains separated from the matrix of the glacial diamictite MG05-02 sample are subhedral to rounded, but few fragments of euhedral crystals are present. Their sizes range from 190 to 290 μm. Oscillatory zoning is viewed by the CL images; no metamorphic rims were observed (Fig. 5). Twenty one zircon grains were dated and the ages range from 1.1 to 2.2 Ga (Table 5; Fig. 6). One grain yielded an older ²⁰⁷Pb/²⁰⁶Pb age but is highly discordant (19%). Three main clusters are observed at 1.88 Ga, 2.09 Ga, and 2.16 Ga (Fig. 7). No Archean zircons were found in this sample. Two crystals with ages of 1.54 Ga (spot 6.1) and 1.8 Ga (spot 4.1) present euhedral shape and could indicate short transport from the source to the deposition site.

4.2.4. Mafic volcanic rock from the Lower Chapada Acauã Formation (MG05-03)

Most of the zircon grains display a rounded shape typical of detrital minerals. Their size range from 120 to 360 μm, but most of them are c. 200 μm long. Oscillatory zoning is observed on the CL images (Fig. 5). Twenty three zircon grains from this sample were dated and the ages range from 960 Ma to 2750 Ma (Table 6; Fig. 6). Considering all data, three main clusters are observed at c. 1.4 Ga, 1.7 Ga, and 2.15 Ga (Fig. 7). It is worth mentioning that the youngest ages were obtained on grain 11 (spots 11.1 and 11.2; Table 6), and both ages are discordant (19 and –11%). However, the U concentration on spot 11.1 is very low (22 ppm), and hence we prefer the ²³⁸U/²⁰⁶Pb age (1067 ± 19 Ma) obtained on spot 11.2 (U concentration of 59 ppm) as the best estimation for the crystallization age of this grain. This youngest grain (c. 1.1 Ga) has an age similar to detrital zircons from other mafic volcanic rock of the Lower Chapada Acauã Formation, which shows the same pillow structures (Babinski et al., 2005; Gradim et al., 2005).

5. Discussion and correlations

The evolution of the Macaúbas basin, precursor of the Araçuaí orogen, has to be interpreted together with its African counterpart, so that we briefly synthesize the stratigraphy of the West Congo belt. The West Congo Supergroup represents the precursor basin of the West Congo belt and includes, from base to top, the Tonian Zadinian and Mayumbian groups, a very thick succession of rift-related bimodal volcanic rocks and clastic sediments, and the West Congolian Group (Fig. 8) that comprises a basal rift-related siliciclastic succession (the Sansikwa Subgroup), followed by a glacio-marine diamictite with intercalations of transitional basalts (the Lower Mixtite Formation), a post-glacial carbonate-rich succession (the Haut Shiloango Subgroup), a second diamictite unit (the Upper Mixtite Formation), a post-glacial pelite–carbonate succession (the Schisto-Calcaire Subgroup) and molasse deposits (Frimmel et al., 2006; Pedrosa-Soares et al., 2008; Tack et al., 2001). The Lower and Upper Mixtite

Fig. 6. U–Pb concordia diagrams for the five samples. Sample MG05-04: Quartzite from the Duas Barras Fm.; Sample OPU-2654: Quartzite from the Rio Peixe Bravo Fm.; Sample MG05-01: Diamictite from the Serra do Catuni Fm.; Sample MG05-02: Diamictite from the Serra do Catuni Fm.; and Sample MG05-03: Mafic volcanics.

Formations have been correlated with the old Cryogenian and young Cryogenian glacial events, respectively (Frimmel et al., 2006).

5.1. Age, provenance and correlations of Macaúbas pre-glacial and glaciogenic deposits

The oldest pre-glacial units of the Macaúbas Group, named Matão, Duas Barras and Rio Peixe Bravo Formations (Fig. 2), comprise rift-related fluvial to shallow marine sediments, deposited unconformably on the Archean–Palaeoproterozoic basement or atop sandstones of the Statherian–Mesoproterozoic Espinhaço Supergroup (Martins et al., 2008; Pedrosa-Soares et al., 2008). The maximum depositional age of the Duas Barras sandstones is 900 ± 21 Ma, constrained by the youngest detrital zircons of sample MG05-04. The zircon age spectrum determined from this sample is large (900 to 2550 Ma; Figs. 6 and 7), with main peaks at 900–1250 Ma and 1.85–2.1 Ga. This large age distribution is similar to that of the Matão Formation, from which the youngest detrital zircon was dated at 1177 ± 30 Ma (Martins et al., 2008). In contrast, the Rio Peixe Bravo Formation shows an almost single source of 2.1 Ga (Figs. 6 and 7). Such differences in age spectra for deposits close to each other suggest basin feeding from several sources and selective controls of erosion and sediment transport during the continental rift stage. Hence, though several sources provided sediments to the Duas Barras and Matão Formations, it seems that mostly Late Rhyacian rocks of the uplifted Porteirinha block fed the nearby Rio Peixe Bravo Formation (Fig. 2). The youngest detrital zircon of the Duas Barras sandstones indicates that the deposition of the Macaúbas Group took place after 900 Ma ago.

The pre-glacial deposits of the Macaúbas Group have been tentatively correlated to the Sansikwa Subgroup of the West Congolian belt (Pedrosa-Soares et al., 2008), which is also a pre-glacial rift-related unit with a maximum depositional age of 923 ± 43 Ma, and contains zircon grains with ages mainly between 900 Ma and 1200 Ma (Frimmel et al., 2006). However, in contrast with the Duas Barras age spectrum, only a few zircon grains of Archean and Palaeoproterozoic ages were observed in the Sansikwa Subgroup, indicating that in the African side Tonian to Stenian sources preferentially fed the basin. Many sources for the Tonian zircons are available in the West Congo belt including the thick Mayumbian rhyolitic lavas dated at 920 ± 8 Ma at the base and 912 ± 7 Ma at the top, the associated Mativa (924 ± 25 Ma) and Bata Kimenga (917 ± 14 Ma) subvolcanic granites, and the 999 ± 7 Ma Noqui-type granites intruded in the Palaeoproterozoic Kimezian base-

ment and in the lowermost beds of the Zadinian Group (Tack et al., 2001). In contrast, Tonian igneous sources are not so abundant in the São Francisco craton and Araçuaí orogen, being limited to mafic dikes of 1000–900 Ma (D'Agrella Filho et al., 1990; Machado et al., 1989) and a few A-type granites dated at 875 ± 9 Ma (Silva et al., 2008). This can be explained by the model of an asymmetric continental rift with the thermal axis of the Tonian basin system located in the African side, as suggested by Pedrosa-Soares et al. (2008).

Igneous sources for the zircons with ages between 1.0 Ga and 1.25 Ga have not been found either in the São Francisco craton or in the West Congo belt. Some authors have advocated that zircons with these ages may derive from the Statherian–Mesoproterozoic Espinhaço–Chapada Diamantina basin system that crops out in the São Francisco craton and Araçuaí orogen. In fact, a careful survey on the available geochronological data for the Espinhaço–Chapada Diamantina system shows a great prevalence of U–Pb zircon ages between 1.75 Ga and 1.5 Ga (Babinski et al., 1994, 1999; Battilani et al., 2005; Danderfer et al., 2009; Guimarães et al., 2005; Schobbenhaus et al., 1994). All the remaining geochronological data suggesting younger ages for deposition, except for those of Chemale et al. (2010), were obtained either on clay minerals or whole-rock samples by Rb–Sr, K–Ar and Pb–Pb radiometric methods (Távola et al., 1967; Jardim de Sá et al., 1976; Brito Neves et al., 1979; Macedo and Bonhomme, 1984; Babinski et al., 1993). These isotopic systems may have suffered partial resetting during the Brasiliano orogeny that affected several cratonic covers, yielding younger and/or geologically meaningless ages and should be interpreted with caution. The lack of source rocks with ages between 1.0 and 1.2 Ga has also been mentioned on the African side (Batumike et al., 2009; Cox et al., 2004; Frimmel et al., 2006), except in the Namaqua–Natal Province where ages of 1.1 to 1.3 Ga are reported (Evans et al., 2007; Pettersson et al., 2007). However, according to most recent palaeogeographic reconstructions this lies on the Kalahari craton and was not attached to the Congo craton in the Tonian (e.g. Collins and Pisarevsky, 2005). According to available U–Pb data we can argue that the source of the large population of 1.0–1.25 Ga zircons has to be found elsewhere.

Detrital zircons with ages in the range 1.25–1.45 Ga, also present in the pre-glacial sediments of the Macaúbas Group, could have been transported from the Kibaran belt, located on the eastern margin of the Congo craton, in Central Africa (e.g., Tack et al., 1994; Trompette, 1994). The Irumide belt, in the southeast of the Congo craton, also has protoliths of 1.36–0.95 Ga (De Waele et al. 2006) and is feasibly a source of these zircons. Nevertheless, considering the distance between these regions and the study area in a Meso–Neoproterozoic palaeogeographic fit, it is hard to believe that they could be sources of detrital zircons for the Macaúbas Group. On the other hand, U–Pb data for the Vazante (Rodrigues, 2008) and Andrelândia (Valeriano et al., 2004; Valladares et al., 2004) Groups in the Brasília belt, located west of the São Francisco craton (Fig. 1), also show detrital zircons with ages ranging from 1.0 Ga to 1.6 Ga, but mainly in the 1.2–1.3 Ga interval. Granitoids from the Brasília belt have been dated at c. 1.2 Ga (Klein, 2008) and could represent one of the sources. However, the occurrence of these zircons in both sides of the São Francisco craton strongly suggests that Mesoproterozoic sources are in the São Francisco craton, and probably hidden by Neoproterozoic and Phanerozoic covers.

The massive diamictites of the Serra do Catuni Formation represent the lowermost glaciogenic deposits of the Macaúbas Group (Fig. 2), deposited in proximal glacio-marine environments. They grade upwards and laterally to the stratified diamictites of the Lower Chapada Acauã Formation (Karfunkel and Hoppe 1988; Martins et al., 2008; Pedrosa-Soares et al., 1992, 2008). The maximum depositional age of the Serra do Catuni diamictites was given by the youngest zircon dated at 933 ± 8 Ma (Table 4; Figs. 6 and 7). The source of these zircons is considered to be the volcanic pile of the Zadinian and Mayumbian groups, and related granites (Tack et al., 2001). The main rock sources of these diamictites (samples MG05-01 and MG05-02) are in the age interval of 0.9–1.3 Ga and 1.8–

Fig. 7. Frequency histogram and probability curves of zircon Pb/Pb ages (<15% discordant) for samples of the Macaúbas Group.

2.2 Ga, with peaks at 1.3 Ga, 1.6 Ga, 1.87 Ga, 2.0 Ga, 2.1 Ga and 2.6 Ga (Figs. 6 and 7). Here again the 1.0 to 1.3 Ga sources are hard to identify, but the older ones are common in the Espinhaço Supergroup (1.5 to 1.75 Ga), and in the basement rocks of the São Francisco–Congo craton, and in the Porteirinha and Guanhanes blocks (Fig. 2).

The correlation of the Jequitai and Serra do Catuni formations, as a west to east lateral succession from glacio-terrestrial to glacio-marine

facies, is rather a consensus in the literature; the first unit including most of the glacio-terrestrial deposits left by the glaciers on the São Francisco craton (e.g., Karfunkel and Hoppe, 1988; Martins-Neto et al., 2001; Pedrosa-Soares et al., 2008; Uhlein et al., 1999). This interpretation is also supported by the age spectra of detrital zircons from both formations, as shown in this paper, and by Buchwaldt et al. (1999) and Rodrigues (2008).

Table 3

U–Pb SHRIMP data from detrital zircons of sample OPU 2654 (Rio Peixe Bravo Formation).

Grain. spot	U (ppm)	Th (ppm)	²³² Th/ ²³⁸ U	²⁰⁶ Pb* (ppm)	²⁰⁶ PbC% (ppm)	Radiogenic ratios						ρ	Age (Ma)				
						²⁰⁶ Pb/ ²³⁸ U	$\pm\%$	²⁰⁷ Pb/ ²³⁵ U	$\pm\%$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm$		²⁰⁶ Pb/ ²³⁸ U	$\pm$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm$	% Disc.
1.1	104	65	0.64	28.3	2.29	0.309	2.9	5.45	5.2	0.128	4.3	0.556	1735	44	2069	76	16
2.1	49	19	0.40	18.4	0.75	0.430	3.2	7.57	4.7	0.128	3.4	0.682	2306	62	2067	60	–12
3.1	58	36	0.64	21.8	0.56	0.431	3.0	7.69	4.3	0.129	3.1	0.703	2311	59	2090	54	–11
4.1	137	131	0.98	50.0	0.72	0.420	2.9	7.43	3.6	0.128	2.2	0.797	2262	55	2074	38	–9
5.1	71	92	1.35	23.2	0.39	0.381	3.1	6.62	4.0	0.126	2.4	0.790	2079	56	2044	43	–2
6.1	41	19	0.49	14.4	0.20	0.406	3.2	6.83	4.2	0.122	2.7	0.758	2197	59	1986	49	–11
7.1	65	45	0.71	24.3	0.11	0.433	2.9	8.12	3.7	0.136	2.2	0.800	2318	57	2178	38	–6
8.1	105	13	0.12	39.0	0.30	0.430	2.8	8.21	3.3	0.139	1.7	0.850	2305	54	2210	30	–4
9.1	54	63	1.20	14.4	0.82	0.309	3.1	4.09	5.7	0.096	4.7	0.549	1737	47	1548	89	–12
10.1	45	42	0.96	12.5	5.32	0.307	3.6	4.27	14.0	0.101	14.0	0.252	1725	55	1640	260	–5
11.1	49	24	0.51	17.9	2.93	0.418	3.5	7.60	7.2	0.132	6.3	0.487	2250	67	2125	110	–6
12.1	77	42	0.57	28.3	2.41	0.419	2.9	7.78	5.2	0.135	4.3	0.563	2258	56	2157	75	–5
13.1	47	27	0.59	18.3	1.00	0.446	3.1	7.99	5.0	0.130	3.9	0.628	2378	63	2096	68	–13
14.1	56	19	0.36	19.5	0.32	0.406	3.2	7.42	4.1	0.132	2.6	0.775	2199	59	2130	46	–3
15.1	120	35	0.30	45.0	0.35	0.434	2.7	7.92	3.2	0.132	1.7	0.854	2324	53	2130	29	–9
16.1	62	26	0.44	24.7	0.00	0.467	3.2	8.33	3.8	0.129	2.0	0.850	2470	66	2089	35	–18
17.1	158	127	0.83	52.2	0.32	0.382	2.7	6.43	3.1	0.122	1.6	0.862	2087	48	1985	28	–5
18.1	235	85	0.37	91.9	0.14	0.455	2.6	9.54	2.8	0.152	1.1	0.928	2417	53	2370	18	–2

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.

Error in Standard calibration was 0.58% (not included in above errors).

1) Common Pb corrected using measured ²⁰⁴Pb/²⁰⁶Pb ratio.

The Lower Chapada Acauá Formation, a thick glacio-marine pile of debris flow deposits (the stratified diamictites) interbedded with sand–pelite turbidites, covers the proximal Serra do Catuni diamictites but also represents a distal equivalent of them, depicting a continuous record of the same glacial event (Karfunkel and Hoppe, 1988; Pedrosa-Soares et al., 1992, 1998, 2008; Martins-Neto et al., 2001; Uhlein et al., 2007; Martins et al., 2008). Rift-related, mafic volcanic rocks with pillow structures and transitional geochemical signature occur interbedded in the Lower Chapada Acauá Formation (Gradim et al., 2005). Attempts to date them were not successful due to the lack of original igneous zircon crystals (Babinski et al., 2005; this work). The youngest detrital grain dated here at 1067 Ma sets the maximum age of the volcanism. The presence of detrital zircons ranging from 1.07 Ga to 2.75 Ga represents the variety of rocks assimilated by the mafic magma and, therefore, available in the sediment pile crossed by it. The Sm–Nd T_{DM} model age of 1.7 Ga also suggests that the magma was contaminated by the continental crust. These tholeiitic basalts are present in other parts of the Lower Chapada Acauá Formation and show similar Sm–Nd T_{DM} model ages and detrital zircons ranging from 1.16 to 2.67 Ga (Babinski et al., 2005; Gradim et al., 2005). In the West Congo belt, basalts with pillow structures interlayered with diamictites of the Lower Mixtite Formation, as well as associated feeder dykes and sills, also show a rift-related, transitional geochemical signature (De Paepe et al., 1975; Kampunzu et al., 1991), but they are undated (Frimmel et al., 2006). Based on the striking similarity of this volcanism and stratigraphic relation in both Macaúbas and West Congolian basins, Pedrosa-Soares et al. (2008) suggested the correlation between the Lower Chapada Acauá and Lower Mixtite formations.

Most of the diamictite–turbidite package of the Lower Chapada Acauá Formation seems to represent resedimented mass flows supplied by the glaciated onshore region (Gradim et al., 2005; Martins et al., 2008; Martins-Neto et al., 2001; Uhlein et al., 1999), so that it can be related to the deglaciation phase of the Macaúbas basin. This succession is transgressive over the Serra do Catuni Formation, contains a few carbonate lenses at its top, and passes upward and laterally to successions completely free of diamictites—the Upper Chapada Acauá and Ribeirão da Folha Formations. This suggests a correlation of the Lower Chapada Acauá Formation with the Carrancas diamictite (Pedrosa-Soares et al., 2008), which is covered by the unit that includes the c. 740 Ma cap carbonate on the São Francisco craton (Babinski et al., 2007).

It is worth mentioning that the Sm–Nd model ages show older ages towards the top of the sedimentary succession (Table 1). The lowermost unit, the Duas Barras Formation, yields the younger T_{DM} ages of 1.8–1.9 Ga. Up-section the T_{DM} ages become older, ranging from 2.2 to 2.3 Ga in the diamictites of the Serra do Catuni and Lower Chapada Acauá Formations. The older T_{DM} age (2.5 Ga) was determined on a sandstone of the diamictite-free Upper Chapada Acauá Formation. These Sm–Nd model ages suggest that sources with older ages (Archean protoliths) had a higher contribution in the upper units of the Macaúbas basin. This behavior could be explained by the evolution of the rift. In the early rift stages, the main sources of sediments were the 900–1000 Ma volcanic pile of the Zadinian and Mayumbian groups and related anorogenic granites eroded from uplifted shoulders and internal horsts of the rift. These are common sources in the basal units of the Macaúbas and West Congolian groups. As the basin evolved, these Neoproterozoic sources were replaced by older sources—mainly Palaeoproterozoic and Archean ones. This is observed in the Serra do Catuni and Lower Chapada Acauá Formations, which have older T_{DM} ages compared to the pre-glacial units of the Macaúbas Group. In the West Congo Belt, the 1.8–1.9 Ga T_{DM} ages are constant along the sedimentary pile, but the presence of Archean and Palaeoproterozoic zircons, although not very common, significantly increases in the Haut Shiloango Subgroup (40%) compared to the Sansikwa Subgroup (15%), (Frimmel et al., 2006). Because T_{DM} ages on samples from the West Congo belt are all around 1.8–1.9 Ga, and the Archean and Palaeoproterozoic sources are much less important compared to the Tonian ones, we speculate that the older components have come mostly from the basement of the São Francisco craton.

5.2. Possible correlations of the Macaúbas diamictites with Neoproterozoic glacial events

The U–Pb ages obtained on the lower formations of the Macaúbas Group indicate that their deposition took place after 900 Ma ago, but do not give any direct constraint regarding to which Neoproterozoic glaciation these deposits are related. Also, the Jequitaiá, Carrancas and Bebedouro diamictites, deposited on the São Francisco craton (Fig. 1), only contain zircons older than 875 Ma (Buchwaldt et al., 1999; Figueiredo et al., 2009; Rodrigues, 2008). Since the Carrancas and Jequitaiá diamictites underlie the pelite–carbonate cratonic cover that includes a basal cap carbonate dated at 740 ± 22 Ma (Babinski et al., 2007), we suggest that the correlative Macaúbas diamictites were deposited during the Early Cryogenian (known as old Cryogenian) glacial event.

Table 4

U–Pb SHRIMP data from detrital zircons of sample MG05-01 (Serra do Catuni Formation).

Grain. spot	U (ppm)	Th (ppm)	²³² Th/ ²³⁸ U	Radiogenic ratios								ρ	Age (Ma)				% Disc.
				²⁰⁶ Pb* (ppm)	²⁰⁶ Pbc %	²⁰⁶ Pb/ ²³⁸ U	±%	²⁰⁷ Pb/ ²³⁵ U	±%	²⁰⁷ Pb/ ²⁰⁶ Pb	±		²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	
1.1	104	58	0.57	36.4	0.09	0.405	2.5	7.190	2.8	0.129	1.3	0.886	2190	46	2084	23	5
2.1	215	111	0.54	66	0.47	0.357	2.4	5.650	2.6	0.115	1	0.921	1969	41	1875	18	−5
3.1	112	141	1.30	31.8	0.08	0.330	0.99	5.083	1.8	0.112	1.5	0.549	1837	16	1829	27	0
4.1	149	73	0.51	26	0.21	0.204	0.92	2.353	2.0	0.084	1.8	0.447	1196	10	1286	36	7
5.1	102	56	0.56	32.6	0.14	0.370	1.0	6.300	1.9	0.124	1.5	0.551	2027	18	2010	27	−1
6.1	103	43	0.43	18.9	0.12	0.211	1.3	2.535	3.5	0.087	3.2	0.376	1235	15	1363	63	9
7.1	114	186	1.69	36.5	0.71	0.372	1.2	6.400	1.7	0.125	1.2	0.692	2040	21	2025	22	−1
8.1	84	44	0.55	29.2	0.25	0.405	1.1	7.420	2.1	0.133	1.8	0.526	2190	21	2138	32	−2
9.1	158	79	0.52	51.7	0.09	0.380	0.82	7.135	1.3	0.136	0.96	0.649	2077	15	2179	17	5
10.1	139	76	0.56	34.9	0.38	0.291	0.91	3.870	1.8	0.096	1.5	0.519	1647	13	1555	28	−6
10.2	82	46	0.58	20.6	0.27	0.291	1.2	3.920	2.6	0.098	2.3	0.473	1645	18	1583	43	−4
11.1	250	241	0.99	92.8	0.73	0.431	0.67	11.010	0.9	0.185	0.64	0.724	2310	13	2700	11	14
12.1	141	410	3.00	24.6	0.45	0.202	1.2	2.290	2.5	0.082	2.3	0.455	1185	12	1253	44	5
13.1	98	38	0.40	20.1	0.18	0.237	1.1	3.000	4.4	0.092	4.3	0.242	1370	13	1465	82	6
14.1	316	122	0.40	75.5	0.18	0.278	0.64	3.752	1.1	0.098	0.91	0.576	1580	9	1586	17	0
15.1	52	21	0.41	18.6	0.41	0.415	1.4	8.150	2.2	0.143	1.8	0.609	2237	26	2257	31	1
16.1	101	140	1.43	32.4	0.19	0.372	1.0	6.440	1.7	0.125	1.3	0.607	2041	18	2035	24	0
17.1	105	159	1.56	32.4	0.78	0.359	0.99	6.140	1.8	0.124	1.5	0.556	1976	17	2017	26	2
18.1	159	64	0.41	29.9	0.44	0.219	0.87	2.541	1.7	0.084	1.5	0.509	1276	10	1298	29	2
19.1	65	24	0.38	22.8	2.28	0.409	1.2	7.800	1.8	0.139	1.4	0.648	2208	22	2209	24	0
20.2	83	221	2.75	11.2	0.49	0.156	0.96	1.472	3.6	0.069	3.4	0.269	933	8	885	71	−5
20.3	57	169	3.07	7.61	0.70	0.156	1.1	1.573	2.7	0.073	2.5	0.410	933	10	1021	51	9
21.1	82	56	0.71	28.9	0.09	0.411	1.1	7.620	1.7	0.135	1.3	0.620	2217	20	2158	23	−3
22.1	80	49	0.64	25.1	0.66	0.365	1.1	6.280	2.0	0.125	1.7	0.543	2004	19	2027	30	1
23.1	48	48	1.03	21.7	0.29	0.520	1.7	13.610	2.1	0.190	1.4	0.771	2699	36	2741	22	2
24.1	103	131	1.32	30.4	0.14	0.344	1.2	5.920	1.8	0.125	1.4	0.664	1904	20	2029	24	6
25.1	97	90	0.96	15.5	0.35	0.186	1.4	2.149	3.4	0.084	3.2	0.395	1100	14	1288	62	15
26.1	63	51	0.83	20.9	0.19	0.386	1.2	7.400	2.0	0.139	1.6	0.616	2103	22	2216	27	5
27.1	83	36	0.45	23.5	0.20	0.330	1.1	5.310	2.1	0.117	1.8	0.513	1837	17	1909	32	4
28.1	251	74	0.30	51.7	0.16	0.239	2.3	2.858	2.6	0.087	1.1	0.902	1380	29	1357	21	−2
29.1	53	33	0.64	15.4	1.24	0.332	2.7	5.270	5.0	0.115	4.2	0.546	1849	44	1882	75	2
30.1	103	87	0.87	33.6	0.37	0.378	2.4	6.270	2.7	0.120	1.2	0.896	2068	43	1960	21	−6
31.1	83	20	0.26	13.8	0.80	0.193	2.5	2.043	4.4	0.077	3.6	0.568	1137	26	1117	73	−2
32.1	264	71	0.28	47.6	0.32	0.209	2.4	2.277	2.9	0.079	1.7	0.805	1225	26	1170	34	−5
33.1	70	48	0.71	24.2	0.95	0.397	2.6	7.250	3.2	0.132	1.9	0.811	2157	47	2130	32	−1
34.1	43	22	0.53	18.4	0.26	0.493	2.7	12.210	3.0	0.180	1.3	0.894	2583	57	2650	22	3
34.2	66	43	0.67	29.6	0.36	0.518	2.5	12.660	2.8	0.177	1.3	0.883	2692	55	2627	22	−2
35.1	83	57	0.70	27.7	0.45	0.384	4.5	6.660	4.8	0.126	1.5	0.946	2096	81	2039	27	−3
36.1	137	58	0.43	23	1.02	0.193	2.4	2.142	4.3	0.080	3.5	0.566	1140	25	1204	70	5
37.1	145	66	0.47	58.7	0.35	0.468	2.4	10.980	2.6	0.170	0.85	0.943	2475	50	2559	14	3
37.2	203	150	0.76	85.3	0.21	0.489	2.3	11.860	2.4	0.176	0.66	0.963	2565	49	2616	11	2
38.1	101	70	0.72	27.6	0.38	0.316	2.5	4.890	2.9	0.112	1.6	0.847	1768	39	1838	28	4
39.1	96	54	0.58	36.6	0.43	0.441	2.5	8.050	2.7	0.132	1.2	0.897	2355	49	2130	21	−11

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.

Error in Standard calibration was 0.52, 0.30 and 0.90% (not included in above errors).

(1) Common Pb corrected using measured ²⁰⁴Pb/²⁰⁶Pb ratio.

Based on recent ⁸⁷Sr/⁸⁶Sr data (Frimmel et al., 2006) and previous data (Frimmel et al., 2002) from carbonates, the Haut Shiloango Subgroup has been interpreted as a post-Sturtian open marine carbonate platform. The carbonates of the Schisto-Calcaire unit are distinguished by negative $\delta^{13}\text{C}$ values (–1 to –3‰), which are interpreted as a signature of post-glacial cap carbonates above the Upper Mixtite Formation. Although no radiometric age control exists for these rocks, they must be older than 558 ± 29 Ma, which is the maximum depositional age for the overlying Inkisi Subgroup. An upper age limit is also given by the metamorphic event that affected the whole West Congolian Group at c. 566 Ma (Frimmel et al., 2006).

If the U–Pb SHRIMP ages on detrital zircons recovered from the Haut Shiloango Subgroup (Frimmel et al., 2006) are considered, another scenario can be suggested. The youngest detrital zircon dated from the Haut Shiloango Group is 547 ± 45 Ma (5% discordant). Thus, considering the error, the maximum depositional age of the Haut Shiloango would be 592 Ma. If this is correct, the Upper Mixtite Formation which is in higher stratigraphic position would be correlated to the Gaskiers glaciation event (Bowring et al., 2003). Some samples of the Haut Shiloango Subgroup present ⁸⁷Sr/⁸⁶Sr ratios of 0.7071 that could be

correlated to post-old Cryogenian values. However, the $\delta^{13}\text{C}$ values are very positive (+3 to +8‰) and could not represent those of post-old Cryogenian cap carbonates (which are typified by negative $\delta^{13}\text{C}$ values). This would imply in a hiatus in the middle of the Haut Shiloango Subgroup, the lowest part being Early Cryogenian and the upper part being younger than 590 Ma.

The same scenario has been recently identified in the Bambuí Group, a pelite–carbonate cover of the São Francisco craton. Cap carbonates of the Sete Lagoas Formation, the lowermost unit of the Bambuí Group, yielded a Pb–Pb isochron age of 740 ± 22 Ma (Babinski et al., 2007), supporting a old Cryogenian-age for the underlying glacial deposits. At the middle of the Sete Lagoas Formation a sharp shift on $\delta^{13}\text{C}$ values is observed, increasing from 0 to +12‰. This shift is accompanied by a marked change in sedimentary facies, with increase in detrital input (Kuchenbecker et al., 2010; Vieira et al., 2007a). Detrital zircons (n = 63) recovered from pelites above the C-isotopic shift yielded U–Pb ages ranging from 610 Ma to 850 Ma (Rodrigues, 2008). These data suggest that the lowermost part of the Sete Lagoas Formation represents a post-Sturtian cap carbonate, and a hiatus of c. 130 m.y. exists between the lower and the middle–upper part of the Sete Lagoas Formation. A

Table 5

U–Pb SHRIMP data from detrital zircons of sample MG05-02 (Serra do Catuni Formation).

Grain. spot	U (ppm)	Th (ppm)	²³² Th ²³⁸ U	²⁰⁶ Pb* (ppm)	²⁰⁶ Pbc %	Radiogenic ratios						ρ	Age (Ma)				% Disc.
						²⁰⁶ Pb/ ²³⁸ U	$\pm$ %	²⁰⁷ Pb/ ²³⁵ U	$\pm$ %	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm$		²⁰⁶ Pb/ ²³⁸ U	$\pm$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm$	
1.1	77	47	0.62	26.9	0.17	0.406	2.1	7.20	2.3	0.129	0.94	0.914	2197	39	2079	16	–6
2.1	66	44	0.70	23.1	0.09	0.409	2.3	7.56	2.6	0.134	1.3	0.858	2211	42	2150	24	–3
3.1	121	130	1.11	39	0.12	0.375	2.1	6.70	2.4	0.130	1.1	0.893	2052	38	2093	19	2
4.1	129	77	0.62	38.2	0.31	0.345	2.2	5.28	2.6	0.111	1.3	0.864	1909	37	1817	23	–5
5.1	180	129	0.74	50.9	0.12	0.328	2.1	4.93	2.3	0.109	1.0	0.897	1829	33	1784	19	–2
6.1	149	40	0.28	33.8	0.05	0.265	2.1	3.49	2.5	0.096	1.3	0.853	1513	29	1541	24	2
7.1	115	68	0.61	20.9	0.20	0.210	2.2	2.30	3.1	0.079	2.2	0.711	1231	24	1183	43	–4
8.1	162	94	0.60	28.3	0.20	0.203	2.1	2.19	2.8	0.078	1.8	0.756	1190	23	1151	37	–3
9.1	144	90	0.65	41.5	0.14	0.336	2.1	5.28	2.4	0.114	1.1	0.881	1866	34	1866	21	0
10.1	183	222	1.25	49.3	0.32	0.313	2.2	5.53	2.4	0.128	1.1	0.892	1755	33	2072	19	15
11.1	169	178	1.09	33.7	0.14	0.232	2.2	2.81	2.7	0.088	1.6	0.806	1347	27	1376	31	2
12.1	52	64	1.26	17.5	0.15	0.389	2.3	7.12	2.8	0.133	1.6	0.832	2116	42	2136	27	1
13.1	101	35	0.36	35.7	0.24	0.409	2.2	7.77	2.5	0.138	1.2	0.879	2211	41	2198	20	–1
14.1	341	302	0.92	105	0.85	0.356	2.0	7.67	2.3	0.156	1.2	0.870	1962	35	2417	20	19
15.1	99	172	1.80	32.7	0.15	0.385	2.2	6.83	2.5	0.129	1.2	0.872	2101	39	2078	22	–1
16.1	125	95	0.78	37.6	0.34	0.348	2.1	5.84	2.5	0.122	1.3	0.862	1925	36	1982	22	3
17.1	46	61	1.39	15	0.26	0.382	2.4	7.08	2.9	0.134	1.6	0.823	2087	42	2155	29	3
18.1	72	74	1.07	21.1	0.32	0.340	2.4	5.34	2.9	0.114	1.6	0.829	1887	39	1863	29	–1
19.1	129	84	0.67	38.2	0.04	0.344	2.1	5.46	2.4	0.115	1.1	0.887	1905	35	1882	20	–1
20.1	60	36	0.62	22	0.27	0.423	2.3	7.93	2.8	0.136	1.6	0.826	2274	44	2176	27	–4
21.1	82	79	1.00	29.7	0.17	0.423	2.3	7.62	2.6	0.131	1.2	0.882	2272	44	2108	22	–8

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.

Error in Standard calibration was 0.50% (not included in above errors but required when comparing data from different mounts).

(1) Common Pb corrected using measured ²⁰⁴Pb/²⁰⁶Pb ratio.

detailed sedimentological study has established a disconformity just where the $\delta^{13}\text{C}$ data increase to positive values (Vieira et al., 2007a, b). In addition, seismic data also support this disconformity at the lower part of the Sete Lagoas Formation (Zalán and Romeiro-Silva, 2007), indicating that the Carrancas and Jeiquai diamictites together with the c. 740 Ma cap carbonate would be an older succession compared to the rest of the Bambuí Group which is younger than 610 Ma (Rodrigues, 2008).

6. Conclusions

The U–Pb and Sm–Nd results presented here were interpreted in terms of the joint evolution of the Macaúbas and West Congolian basins (Fig. 8). This evolution should be envisaged in the scenario of a mantle-activated asymmetric rift, in which the magma-rich zone was inherited by the West Congo belt after the opening of the South Atlantic Ocean. This rift evolved to a confined basin partially floored

Table 6

U–Pb SHRIMP data from detrital zircons of sample MG05-03 (Lower Chapada Acauá Formation).

Grain. spot	U (ppm)	Th (ppm)	²³² Th ²³⁸ U	²⁰⁶ Pb* (ppm)	²⁰⁶ Pbc %	Radiogenic ratios						ρ	Age (Ma)				% Disc.
						²⁰⁶ Pb/ ²³⁸ U	$\pm$ %	²⁰⁷ Pb/ ²³⁵ U	$\pm$ %	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm$		²⁰⁶ Pb/ ²³⁸ U	$\pm$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm$	
1.1	83	44	0.55	27.9	0.21	0.391	1.5	6.810	2.2	0.127	1.5	0.706	2126	28	2049	27	–4
2.1	93	47	0.53	32.5	0.15	0.405	1.6	7.430	2	0.133	1.1	0.829	2194	30	2137	19	–3
3.1	94	78	0.85	31.5	0.35	0.388	1.8	6.870	2.2	0.129	1.3	0.816	2114	32	2077	22	–2
4.1	103	90	0.90	31.1	0.38	0.351	1.8	5.290	2.5	0.110	1.7	0.715	1937	30	1791	31	–8
5.1	94	53	0.58	34.6	0.19	0.428	1.8	7.770	2.1	0.132	1.2	0.838	2296	35	2121	20	–8
6.1	160	65	0.42	46.6	0.26	0.338	1.5	4.869	1.9	0.105	1.1	0.805	1876	25	1707	21	–10
7.1	260	79	0.31	47.9	0.21	0.214	1.5	2.384	2	0.081	1.4	0.737	1249	17	1219	27	–3
8.1	241	315	1.35	66	2.88	0.310	1.6	5.470	3.9	0.128	3.6	0.401	1739	24	2073	63	16
9.1	57	27	0.48	26	0.25	0.529	1.8	13.500	2.1	0.185	1.1	0.852	2738	39	2698	18	–1
10.1	88	78	0.92	30.6	0.23	0.402	1.8	7.430	2.1	0.134	1.2	0.838	2180	33	2151	20	–1
11.1	22	23	1.05	3.12	1.64	0.160	2.5	1.760	10	0.079	9.7	0.254	959	23	1183	190	19
11.2	59	62	1.10	9.19	0.94	0.180	1.9	1.764	5.1	0.071	4.7	0.373	1067	19	959	97	–11
12.1	43	39	0.94	14.5	0.61	0.388	1.9	6.740	2.9	0.126	2.2	0.656	2114	34	2043	39	–3
13.1	48	12	0.25	16.8	0.47	0.403	2.1	7.180	2.8	0.129	1.9	0.735	2182	38	2089	33	–4
14.1	67	65	1.00	18.4	0.51	0.319	1.8	4.560	3.1	0.104	2.6	0.563	1787	27	1691	47	–6
15.1	126	74	0.60	42	0.18	0.388	1.6	7.280	1.9	0.136	10	0.845	2115	28	2177	17	3
16.1	136	59	0.44	64.6	0.06	0.552	1.6	14.510	1.7	0.191	0.67	0.919	2833	36	2748	11	–3
17.1	144	61	0.43	56.7	0.11	0.457	1.5	9.920	1.9	0.158	1	0.832	2425	31	2429	17	0
18.1	162	224	1.43	51.2	0.11	0.367	1.5	6.180	1.8	0.122	0.89	0.864	2015	27	1987	16	–1
19.1	216	100	0.48	53.4	0.34	0.286	1.5	4.053	1.9	0.103	1.1	0.805	1623	22	1673	20	3
20.1	210	200	0.99	59.3	0.10	0.329	1.6	5.199	1.8	0.115	0.92	0.863	1834	25	1873	17	2
21.1	198	234	1.22	45	0.13	0.264	1.5	3.312	1.9	0.091	1.1	0.797	1512	20	1444	22	–5
22.1	157	147	0.96	32	0.19	0.236	1.6	2.764	2.1	0.085	1.4	0.738	1367	19	1314	27	–4
23.1	243	83	0.35	53.9	0.15	0.258	1.5	3.188	1.8	0.090	1.1	0.814	1477	20	1421	20	–4

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.

Error in Standard calibration was 0.63% (not included in above errors).

(1) Common Pb corrected using measured ²⁰⁴Pb/²⁰⁶Pb ratio.

Fig. 8. Correlation stratigraphic sketch for the Proximal Macaúbas Group (Araçuaí orogen), West Congolian Group (West Congo belt) and covers of the São Francisco craton. 1, rudites; 2, sandstones; 3, pelites; 4, diamictites; 5, mafic volcanic rocks with pillow structure and transitional geochemical signature; 6, carbonate rocks; 7, outsized clasts (dropstones); 8, mafic dykes; 9, anorogenic granites; and 10, disconformity or erosional unconformity. Compiled ages are from: *, Frimmel et al. (2006); **, Silva et al. (2008); *** Machado et al. (1989).

by oceanic crust (Alkmim et al., 2006; Pedrosa-Soares et al. 2001, 2008).

The basin began to open in the Stenian–Tonian boundary, when it started to be filled by the thick Zadinian–Mayumbian volcano-sedimentary succession (1000–900 Ma), that together with the related intrusions represent the thermal–magmatic axis of the asymmetric rift. No correlative of the Zadinian and Mayumbian groups is known in the Araçuaí orogen. However, the presence of the c. 875 Ma A-type granites in the northeast tip of this orogen suggests that the thermal axis of the rift migrated to the west (in relation to the present-day geography) in the Late Tonian (Silva et al., 2008), so that the basin widening shifted to the Brazilian side. At that time, after 900 Ma ago, deposition of the lower diamictite-free sedimentary succession of the Macaúbas basin began, probably coeval with the Sansikwa Subgroup of the West Congolian basin. At least part of the Macaúbas pre-glacial deposits are interpreted to be contemporaneous with the magmatic episode that generated the c. 875 Ma granites (Fig. 8).

After this Late Tonian sedimentation, a climatic change to glacial conditions affected the São Francisco–Congo continent and neighboring basins, so that a huge amount of diamictites started to fill the Macaúbas basin. The correlations between the Macaúbas diamictites and those covered by the c. 740 Ma cap carbonate on the São Francisco craton suggest that this climatic change can be correlated to the Early Cryogenian (known as Old Cryogenian) glacial event. The Serra do Catuni Formation is considered to be an equivalent of the Jeiquitá diamictite, depicting a lateral succession of glacial environments from glacio-terrestrial, on the São Francisco craton, to proximal glacio-marine along the western region of

the Macaúbas basin. The transitional mafic volcanism of the Lower Chapada Acauã Formation signals the end of the continental rift stage, which seems to be coeval with a climatic change to non-glacial conditions indicated by the absence of diamictites in the upper units of the Chapada Acauã Formation. This implies that at least part of the Chapada Acauã diamictites record the deglaciation process in the region. The Chapada Acauã diamictite succession is considered to be equivalent to the Carrancas diamictite, placed on the southern São Francisco craton, and to the Lower Mixtite Formation of the West Congo belt (Fig. 8). The Upper Mixtite Formation of the West Congo belt would have no equivalent in the Macaúbas basin, and might represent a local glaciation in the African side, coeval or younger than the upper part of the pelite–carbonate cover of the São Francisco craton, which maximum age is c. 590 Ma.

The continental rift to passive margin transition is related to the generation of oceanic crust in the distal Macaúbas basin, where the Ribeirão da Folha Formation was deposited (Figs. 2 and 3). The U–Pb dating of zircon crystals from plagiogranites of an ophiolite hosted by the Ribeirão da Folha Formation resulted in an age around 660 Ma (Queiroga et al., 2007). Considering that both the diamictite-free Ribeirão da Folha and Upper Chapada Acauã formations are lateral equivalents, the age of the ophiolite provides further evidence that the Macaúbas diamictites might record a glaciation older than 660 Ma. In fact, the deglaciation process can explain the rise of the sea level followed by the deposition of the c. 740 Ma cap carbonates over diamictites on the São Francisco craton, and the diamictite-free Upper Chapada Acauã Formation can be considered a correlative of the lower part of the pelite–carbonate cratonic cover.

Acknowledgments

The authors acknowledge financial support provided by FAPESP (Fundação de Amparo à Pesquisa do Estado de São Paulo), CNPq (Conselho Nacional de Desenvolvimento Científico e Tecnológico), FINEP (Financiadora de Estudos e Projetos), and FAPEMIG (Fundação de Amparo à Pesquisa de Minas Gerais). Fernando Alkmim is thanked for showing us some of the outcrops studied here. Alan Collins kindly reviewed a previous version of the manuscript. Constructive reviews by Daniel Le Heron and Stephen Marshak contribute to improve the original manuscript. We dedicate this work to our beloved friend and coauthor – the late Carlos Maurício Noce – who contributed significantly for the understanding of the São Francisco Craton and Araçuai Fold Belt evolution.

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4.3. GEOCRONOLOGIA Pb-Pb DOS CARBONATOS DE CAPA DA FORMAÇÃO SETE LAGOAS, GRUPO BAMBUÍ, CRÁTON DO SÃO FRANCISCO

O final do Neoproterozoico é marcado por depósitos glaciais mas a cronologia destes depósitos é prejudicada pela escassez de dados geocronológicos. Neste trabalho são apresentados novos dados radiométricos obtidos em carbonatos de capa da base da Formação Sete Lagoas que estão depositados sobre unidades glaciais no Cráton do São Francisco. Seis amostras de uma fácies rica em cristais de pseudomorfos de aragonita, mostrando texturas originais e razões $^{87}\text{Sr}/^{86}\text{Sr}$ constantes e iguais a 0,7075, forneceram uma idade isocrônica Pb-Pb de 740 ± 22 Ma, a qual é interpretada como a idade de deposição destas rochas, devido à excelente preservação. Esta idade pode ser usada para inferir uma paleolatidade baixa a moderada de $20^\circ - 30^\circ$ para a deposição dos carbonatos e dos depósitos glaciais. Além disso, como esta idade se superpõe com idades obtidas para outras sucessões neoproterozoicas, os resultados deste trabalho reforça a ideia da existência de um amplo intervalo para o evento Sturtiano, sugerindo que este evento representa diferentes glaciações discretas ou um evento longo que pode ser durado cerca de 80 Ma.

Direct dating of the Sete Lagoas cap carbonate (Bambu Group, Brazil) and implications for the Neoproterozoic glacial events

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ABSTRACT

The end of the Neoproterozoic is punctuated by glacial deposition, but the chronology of these deposits is hindered presently by the paucity of geochronological data. Here, we present new radiometric dating for the basal Sete Lagoas cap carbonate deposits that overlie glacial units in the So Francisco craton. Six samples from aragonite-pseudomorph crystal-rich facies, showing pristine textures and constant ⁸⁷Sr/⁸⁶Sr ratios around 0.7075, yielded a Pb–Pb isochron age of 740 ± 22 Ma, which is interpreted as the depositional age

for these remarkably preserved rocks. This age can be used to infer a low-to-moderate palaeolatitude of 20–30 for carbonate (and glacial) deposition. In addition, as it overlaps the ages obtained for the oldest Neoproterozoic glacial successions, our result reinforces the idea of a long-standing ‘Sturtian’ interval, suggesting that this event represents either different discrete glaciations or a protracted event encompassing almost 80 Ma.

Terra Nova, 19, 401–406, 2007

Introduction

The end of the Neoproterozoic was marked by important changes in Earth’s climate. Clues to these environmental perturbations are recorded in glacial deposits, found to be deposited in equatorial latitudes (Sohl *et al.*, 1999), which are frequently overlain by dolomites and limestones, known as ‘cap carbonates’ (e.g. Kennedy *et al.*, 1998). These rocks show enigmatic sedimentary structures and strong temporal anomalies in δ¹³C and ⁸⁷Sr/⁸⁶Sr (Kaufman *et al.*, 1997). Such strong climatic variations seem to have occurred at least three times between 750 and 580 Ma, commonly termed the ‘Sturtian’, ‘Marinoan’ and ‘Gaskiers’ glacial events (Halverson *et al.*, 2005).

Different models have been developed to explain the Neoproterozoic ice ages along with cap carbonates and associated biogeochemical anomalies (Hoffman and Schrag, 2002). But some topics remain controversial, including the chronology

(and synchrony) of the glacial events and their impact on the biosphere (Corsetti *et al.*, 2006; Fairchild and Kennedy, 2007). As the geochronological database grows, accumulated ages for the two younger events cluster at 635 and 580 Ma (Bowring *et al.*, 2003; Schaefer and Burgess, 2003; Calver *et al.*, 2004; Hoffmann *et al.*, 2004; Condon *et al.*, 2005). In contrast, Sturtian successions show a wide spread of ages varying from 740 to 650 Ma (e.g. Lund *et al.*, 2003; Fanning and Link, 2004, 2006). Palaeontological information from before, during and after Neoproterozoic glacial events is still very scarce (e.g. Corsetti *et al.*, 2006). Recently, biomarkers were used to infer a complex and productive microbial ecosystem, including phototrophic bacteria and eukaryotes, associated with glacial deposits of the So Francisco craton, Brazil (Olcott *et al.*, 2005). These results imply that sunlight was available during glacial deposition and the glacial event had little impact on the ecosystem. But again, the age of these deposits is poorly constrained and their correlation with the aforementioned glacial events is still very speculative. Here, we present direct dating of the basal cap carbonate unit (Sete Lagoas Formation) of the Bambu Group and discuss its implications the age of glaciation in the

So Francisco craton and the timing of Neoproterozoic glacial events.

So Francisco Basin

The So Francisco basin covers more than 300 000 km² and contains an extensive sedimentary cover of Neoproterozoic clastic and carbonate rocks. The basal Macabas Group includes glacial diamictites (Karfunkel and Hoppe, 1988). These deposits are overlain by the Bambu Group, which comprises a lower interval of marine strata and an upper succession of shallow-water to alluvial strata. The lower marine succession (Sete Lagoas Formation) is the target of this study.

Few age constraints are available for the Macabas and Bambu Groups. Pb–Pb and U–Pb SHRIMP ages from detrital zircons suggest a maximum age for the Macabas diamictites of *c.* 900 Ma (Pedrosa-Soares *et al.*, 2000), and *c.* 875 Ma for the correlative Bebedouro Formation (Babinski *et al.*, 2004). For the Bambu Group, previous Pb–Pb isochron ages obtained on the Sete Lagoas Formation suggested a minimum depositional age of 686 ± 69 Ma (Babinski *et al.*, 1999). However, most ages on the carbonate rocks are younger than 550 Ma for revealing a widespread isotopic resetting in the region (D’Agrella-Filho *et al.*, 2000; Trindade *et al.*, 2004).

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Sete Lagoas Formation

The Sete Lagoas Formation is more than 200 m thick in the studied region (Fig. 1). It comprises two shallowing-upward megacycles, recording two transgressive events (Vieira *et al.*, 2007).

The first megacycle started with sea-floor precipitates, including crystal fans and crusts. These deposits are characterized by negative $\delta^{13}\text{C}$ values, which increase towards 0‰ accompanying the vanishing of sea-floor precipitates upward (Figs 2 and 3). This initial deposition marks a pronounced transgression over the São Francisco craton, putting carbonate deposits over Archean to Palaeoproterozoic basement rocks. The deep platform deposits are overlain by crystalline limestone with structures related to storm-wave action, including swaley and hummocky cross-stratification. Values of $\delta^{13}\text{C}$ for these rocks are limited to a narrow range around 0‰ (Fig. 2).

The second megacycle starts with deposits of mixed deep platform covering a wide surface. The abundance of pelites and the lack of wave structures in the basal unit suggest deposition in the deepest offshore zone. The cycle continues with deposits of black

Fig. 2 Sections of Sambra and Tatiana, with carbon-isotopic curves and $^{87}\text{Sr}/^{86}\text{Sr}$ values (within rectangles). Arrows indicate geochronological samples.

crystalline limestone with wave structures and microbial boundstone, deposited in coastal environments. This transgression contrasts sharply

with the first one by the lack of oversaturation events and the absence of crystal fans. They systematically show positive $\delta^{13}\text{C}$ values reaching up to +14‰, which are interpreted as related to anomalously high burial rates of organic matter with negative carbon rectope values in the upper black limestones (see discussion in Iyer *et al.*, 1995 and Vieira *et al.*, 2007). It is worth mentioning that such positive values are characteristic of post-Sturtian but pre-Marinoan carbonates (Kaufman *et al.*, 1997). This megacycle ends with a rapid transgression marked by the deposition of siltstones of the Serra de Santa Helena Formation.

Methods

Analyses were carried out at the Center of Geochronological Research, University of São Paulo. Pb isotope analyses followed the leaching technique described in Babinski *et al.* (1999), and only the second leachate technique described in (L2) was analysed. Residue fractions were determined on three samples from Sambra carbonates. Isotopic ratios were corrected for a fractionation factor of

Fig. 1 Geological map of the southern border of the São Francisco craton (SFC) with locations of the studied sections (modified from Vieira *et al.*, 2007). SA: Sambra quarry; TA: Tatiana quarry.

Fig. 3 Crystal-bearing facies in Sambra. In (a) and (b) centimetre scale cycles of alternating crystal-rich layers and (darkgrey) micrite. (c) Photomicrograph of aragonite-pseudomorph crystal (left) with square termination replaced by neomorphic sparry calcite.

0.12‰ amu^{-1} based on analyses of NBS 981 standard. Blanks were 30–50 pg and had a negligible effect on the measured isotopic compositions. For isochron calculation, uncertainties of 0.071‰ and 0.074‰ (2σ) were

assumed, respectively, for $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ ratios, on the basis of NBS 981 standard reproducibility. The analytical errors indicated in Table 1 were used only for samples with errors higher than those of the

standard. Isochron ages were determined using Model 1 of Isoplot software (Ludwig, 1999); errors are reported as 95% confidence level.

Sr isotope analyses were conducted on the second leachate obtained with 1.0 N HCl; the first leachate was obtained with 0.1 N HCl and the supernatant was discarded. Solutions were converted to 2 M HNO_3 and loaded through Sr-SPEC resin. Sr was eluted with 0.05 M HNO_3 , loaded onto a Ta filament and analysed in a VG-354 mass spectrometer.

Carbon and oxygen isotope analyses were performed after CO_2 gas extraction from powdered carbonates in a high vacuum line after reaction with 100% H_3PO_4 at room temperature for 24 h. Following cryogenic cleaning, the released CO_2 was analysed in a EUROPA GEO 20–20 mass spectrometer, using IAEA standards as well as a secondary standard. Results are reported in the conventional notation in per mil (‰) relative to the PDB (Pee Dee Belemnite) standard. Uncertainties are 0.1‰ for both carbon and oxygen isotope results.

New age constraints

Samples for the Pb and Sr isotopic study were collected at the Sambra and Tatiana sections (Figs 1 and 2), which are located in the stable area of the basin. Samples are rich in calcite crystal fans, interpreted as pseudomorphed aragonite. These deposits form tabular layers up to 30 m thick, organized in centimetre-scale cycles of lime mudstone and crystals (Fig. 3a). Pseudomorphed aragonite crystals are black to dark grey with acicular morphology, laterally connected to thin, millimetre-thick cement crusts (Fig. 3b). Under the microscope, the crystals show squared terminations with a marked contrast to the micrite matrix (Fig. 3c). Detailed petrographic studies on Sambra were performed by Peryt *et al.* (1990).

Six carbonate samples were selected for isotopic dating in Sambra quarry: four from the base of the quarry, corresponding to the crystal-rich sediments, and two from storm-influenced crystalline limestones (Fig. 2). The four lowermost samples contained both precipitate and microspar components; these were analysed

Table 1 Pb isotopic compositions obtained from carbonates of the Sete Lagoas Formation, Bambuí Group, Brazil.

Sample	$^{206}\text{Pb}/^{204}\text{Pb}$	Error	$^{207}\text{Pb}/^{204}\text{Pb}$	Error	$^{208}\text{Pb}/^{204}\text{Pb}$	Error
Sambra Quarry						
MG01-34.00a	21.997	0.056	15.860	0.058	38.892	0.060
MG01-34.00b	32.518	0.092	16.535	0.086	38.372	0.094
MG01-34.00b (R)	33.570	0.042	16.566	0.038	36.785	0.027
MG01-34.00b1	32.833	0.066	16.559	0.066	38.390	0.068
MG01-34.1.0a	21.024	0.032	15.809	0.032	38.388	0.032
MG01-34.1.0b	28.773	0.040	16.295	0.040	38.129	0.040
MG01-34.1.0b (R)	29.570	0.060	16.333	0.058	38.259	0.033
MG01-34.2.0a	22.662	0.030	15.910	0.030	38.866	0.030
MG01-34.2.0b	27.545	0.040	16.227	0.038	38.893	0.038
MG01-34.2.0b (R)	28.635	0.068	16.265	0.068	38.905	0.035
MG01-34.4.0a	21.611	0.022	15.839	0.021	38.575	0.022
MG01-34.4.0b	29.100	0.029	16.315	0.030	38.165	0.028
MG01-34.07	19.517	0.048	15.703	0.048	39.029	0.052
MG01-34.09	19.139	0.012	15.681	0.011	38.812	0.013
Tatiana Quarry						
MG03-26.1a	19.413	0.048	15.761	0.048	40.193	0.048
MG03-26.1b	28.136	0.060	16.307	0.060	39.917	0.060
MG03-26.2a	19.609	0.052	15.776	0.052	40.453	0.052
MG03-26.2b	21.206	0.108	15.876	0.116	40.343	0.120
MG03-26.3a	20.152	0.016	15.816	0.016	40.944	0.020
MG03-26.3b	21.435	0.052	15.893	0.052	40.499	0.056

Isotopic compositions are corrected for mass fractionation. Errors are 2σ (‰). a, micrite; b, pseudomorphed aragonite; R, residue.

separately. They yielded variably radiogenic Pb isotopic compositions (Table 1). Fibrous precipitates systematically show more radiogenic Pb ratios probably due to the higher U/Pb initial ratios in the aragonite precursors (Table 1, Fig. 4a). The Pb isotope ratios defined a straight line, which yielded a Pb–Pb isochron age of 740 ± 22 Ma (Fig. 4a). Samples from the lowermost part of the quarry showed $^{87}\text{Sr}/^{86}\text{Sr}$ ratios around 0.7074, increasing upward to 0.7081 (Fig. 2). These Sr isotopic results are consistent with those presented by Misi *et al.* (2007) for the same outcrop.

Three carbonate samples from Tatiana quarry were selected for the geochronology study, corresponding to the upper part of the cement-crust facies (Fig. 2). Unlike the Sambra samples, only ghosts of the original precipitate and microspar components of the cement-crust facies were preserved in Tatiana quarry. The fibrous precipitates also showed more radiogenic Pb ratios (Table 1, Fig. 4b). The Pb data defined a straight line, which yielded a Pb–Pb isochron age of 681 ± 50 Ma (Fig. 4b). $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in this quarry are similar to

those obtained from the upper part of Sambra, varying from 0.7079 to 0.7082 (Fig. 2).

Discussion and conclusions

Given the widespread presence of carbonate rocks atop Neoproterozoic glacial successions, the Pb–Pb isochron method would be a suitable technique for directly assessing the age of these deposits. The errors in Pb–Pb ages are typically large ($> \pm 20$ Ma) owing to the typically low concentrations of ^{235}U (half-life 704 Ma) in young carbonates as well as the low spread in U/Pb ratios in carbonate rocks, usually related to post-depositional interaction with continental water (Jahn and Cuvellier, 1994). Hence, only ancient carbonate rocks with a simple geological history are suitable for Pb–Pb dating. Despite its low precision, this method can still discriminate different Neoproterozoic glacial events.

The Pb isotopic compositions obtained on carbonates from the lower section of the Sambra quarry presented a large spread of Pb and a tight alignment, thus yielding a good-quality Pb–Pb isochron age of 740 ± 22 Ma, which is interpreted as the depositional age of these rocks. The excellent petrographical and geochemical preservation of the carbonates, with no substantial recrystallization, support this interpretation. Their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are constant and low (~ 0.7074) at the base of the quarry, where strontium concentrations are high, probably retaining the original sea-water signature. These values mimic results obtained from other well-preserved sections of the Bambuí Group (e.g. Misi *et al.*, 2007), and compare well with those obtained for other Neoproterozoic successions. Notwithstanding this, they give an ambiguous age estimation varying from 750 to 600 Ma according to the available master curves (e.g. Jacobsen and Kaufman, 1999; Halverson *et al.*, 2007). An additional argument indicating that these samples retain a pristine isotopic signal is the strong control of microfacies on the initial U/Pb ratios; crystal fans presenting more radiogenic Pb ratios than the micrite (Table 1, Fig. 4a). It is worth mentioning that analytical results obtained on the residue of Sambra

quarry samples fit on the same 740 Ma isochron (Table 1, Fig. 4a). These results exclude contamination from an older, detrital component that could yield an apparently older isochron age.

In contrast to the well-preserved carbonate samples from Sambra Quarry, Tatiana carbonates are strongly recrystallized and their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios reach values of 0.7082. The 681 ± 50 Ma isochron age obtained for these samples is comparable to the oldest Pb–Pb ages of 686 ± 69 Ma previously obtained on the same unit (Babinski *et al.*, 1999). These ages are interpreted as minimum depositional ages owing to textural and Sr isotopic evidence of post-depositional fluid percolation. They could have been affected by the widespread fluid circulation event recorded in the isotopic and magnetic signals of carbonate rocks throughout the São Francisco craton (D'Agrella-Filho *et al.*, 2000; Trindade *et al.*, 2004). At any rate, these ages present large analytical errors that overlap the more precise age determined on the well-preserved Sambra carbonates.

The older Neoproterozoic glacial deposits, assembled under the 'Sturtian' event, present the largest number of published ages (Halverson, 2006). They also show the largest spread when compared with the younger 'Marinoan' and 'Gaskiers' events. An age of 712 Ma was obtained on tuffs interbedded in glacial sediments of the Gubrah Formation, Oman (Allen *et al.*, 2002). A similar age of 709 ± 5 Ma was obtained immediately below the glaciogenic Scout Mountain Member diamictite, Pocatello Formation, Idaho, whereas a tuff collected 20 m above the diamictite yielded an age of 667 ± 5 Ma (Fanning and Link, 2004). Comparatively younger ages were also found for volcanic rocks intercalated with Rapitan-correlative sediments of the Edwardsburg Formation (Lund *et al.*, 2003). Recent Re–Os and U–Pb SHRIMP data from the actual Sturtian succession in Australia provide even younger ages, varying from 658 to 643 ± 2.4 Ma (Fanning and Link, 2006; Kendall *et al.*, 2006). On the other hand, significantly older ages were obtained for the Rosh Pinah volcanics (741 ± 6 Ma) and volcanic breccias from the Kundelunga

Fig. 4 Pb–Pb isochron diagrams for (a) Sambra quarry and (b) Tatiana quarry. Black squares: crystal fans; grey squares: micrite; empty diamonds: residue. Results from residues were not used in the linear regressions.

Group (735 ± 5 Ma) interpreted as minimum ages for the glacial event in southern Namibia and Zambia, respectively (Frimmel *et al.*, 1996; Key *et al.*, 2001). The Sete Lagoas age of 740 ± 22 Ma agrees well with such oldest 'Sturtian' ages and does not overlap the younger 'Sturtian' ages obtained in Oman and Idaho even considering the large errors inherent in the Pb–Pb isochron system. All these results suggest that either the Sturtian glaciation is a protracted event encompassing almost 80 Ma or it represents different discrete glacial events. It is also interesting to note that Sete Lagoas cap carbonates show sedimentary structures, for instance aragonite pseudomorph crystal fans, considered to be typical of younger 'Marinoan' cap carbonates (e.g. Kennedy *et al.*, 1998; see also discussion in Corsetti and Lorentz, 2006). This indicates that caution must be taken in making correlations based solely on cap carbonate facies.

Finally, it is worth mentioning that good-quality palaeomagnetic poles with ages overlapping those of the Sete Lagoas carbonates are available for the Congo–São Francisco craton (Tohver *et al.*, 2006). Poles for the c. 743 Ma Mbozi complex (Meert *et al.*, 1995) and the c. 744 Ma Nosib Group (McWilliams and Kröner, 1981; age constraints given by Hoffman *et al.*, 1996) place the southern São Francisco craton at a low-to-intermediate latitude (20 – 30°) by ~ 740 Ma. Our geochronological results and these palaeolatitude estimates are useful constraints for the palaeoenvironmental scenario proposed by Olcott *et al.* (2005). They imply that the complex ecosystem interpreted for glacial units in the western part of the São Francisco craton was indeed developed at low-to-intermediate latitudes during a glacial event within the time frame of the 'Sturtian' ice age.

Acknowledgements

This research was partially funded by CNPq and FAPESP (03/08716-3 and 05/5321-1). We appreciate the insightful reviews by Frank Corsetti, Thomas Nägler and Philip Allen, as well as the editorial handling of Max Coleman. We have benefited from discussions on the IGCP-512 forum (<http://www.IGCP512.org>).

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Received 16 May 2007; revised version accepted 6 July 2007

CAPÍTULO 5

GEOCRONOLOGIA DOS DEPÓSITOS GLACIAIS E CARBONATOS DE CAPA DO CRÁTON AMAZÔNICO E DA FAIXA PARAGUAI

Os depósitos glaciais criogenianos da Formação Puga ocorrem tanto sobre o Cráton Amazônico quanto na Faixa Paraguai, na borda leste do cráton, e são recobertos por carbonatos.

Neste capítulo serão apresentados dados geocronológicos U-Pb obtidos em zircões detríticos recuperados de diamictitos da Formação Puga (Maciel, 1959), coletados na porção sul da Faixa Paraguai. Estes depósitos têm sido correlacionados com os diamictitos que ocorrem na porção norte da faixa (Alvarenga et al., 2009, in press; Boggiani, 1998), e em exposições sobre o cráton (Alvarenga et al., 1992).

As idades U-Pb obtidas em cristais de zircão ($n = 340$) separados da matriz dos diamictitos da Formação Puga coletados em 2 áreas variam de 706 Ma a 2100 Ma (apenas 2 cristais arqueanos foram encontrados) (**Babinski et al., submetido; item 5.1**). Diamictitos provenientes do Morro do Puga, área tipo da formação, forneceram zircões com idades entre 970 e 2100 Ma, enquanto que os diamictitos coletados na região da Serra da Bodoquena possuem zircões com idades entre 706 e 1990 Ma, sendo que a principal fonte (mais de 70% dos grãos) tem idade de 1760 Ma. Esses dados permitem sugerir que as rochas coletadas no Morro do Puga têm sua proveniência ligada às rochas do Cráton Amazônico, enquanto que as da Serra da Bodoquena tiveram suas áreas-fontes no Bloco

Rio Apa. Quanto à origem da fonte mais jovem, de 700 Ma, não há evidências desta ser local, podendo ter sido transportada a partir do arco magmático de Mara Rosa, na Faixa Brasília, ou de rochas do paleocontinente Laurentia; outras possibilidades, menos prováveis, são que ela tenha sido totalmente erodida ou encontra-se recoberta por sedimentos fanerozoicos. Os grãos de zircão mais jovens delimitam a idade de deposição dos diamictitos da Formação Puga em 706 ± 9 Ma.

Na borda sudoeste do Cráton Amazônico afloram carbonatos de capa depositados diretamente sobre os diamictitos da Formação Puga. Estes carbonatos pertencem à Formação Mirassol do Oeste (Nogueira et al., 2003) e foram datados pelo método isocrônico Pb-Pb. A idade obtida foi de 633 ± 25 Ma, a qual é interpretada como da época de deposição destas rochas (**item 5.2**). A idade obtida nos carbonatos de capa da Formação Mirassol do Oeste, associada às idades máximas de deposição dos diamictitos da Formação Puga, reforçam a ideia que estas rochas tenham sido depositadas durante o Evento Glacial Marinoano, ocorrido há 636 Ma, no Criogeniano Superior.

5.1. GEOCRONOLOGIA U-Pb SHRIMP DOS ZIRCÕES DETRÍTICOS DOS DIAMICTITOS DA FORMAÇÃO PUGA

A Faixa Paraguai é constituída por uma espessa sucessão sedimentar depositada na borda sudoeste do Cráton Amazônico e do Bloco Rio Apa. Na parte sul da faixa, a base da pilha sedimentar é marcada por um nível de rochas glaciais da Formação Puga associado a formações ferríferas bandadas (BIFs), cuja idade é assumida como Marinoana (ca. 636 Ma), apesar da falta de dados geocronológicos. Neste trabalho são apresentadas idades U-Pb SHRIMP de zircões detríticos separados de 6 amostras de diamictitos de duas localidades (Morro do Puga e região da Serra da Bodoquena, MS). As idades U-Pb (ca. 130 grãos) determinadas em duas amostras do Morro do Puga variam entre 970 e 2100 Ma. Rochas com estas idades ocorrem no Cráton Amazônico sugerindo que esta é a mais provável fonte dos sedimentos. Zircões detríticos (ca. 230 grãos) de diamictitos de Bodoquena (200 km a sul do Morro do Puga) possuem idades entre 706 e 1990 Ma. Uma fonte de 1760 Ma é significativamente mais importante, perfazendo mais de 70% dos grãos

analisados, e pode estar relacionada às rochas do Bloco Rio Apa. O zircão mais jovem tem 706 ± 9 Ma e define a idade máxima de deposição da Formação Puga. Possíveis fontes desta população mais jovem podem ser as rochas do arco magmático Mara Rosa, Faixa Brasília, ou rochas de faixas móveis do continente Laurentia que, em reconstruções paleogeográficas, é posicionado a oeste da área estudada. A idade máxima de deposição dos diamictitos (e BIFs associados), vinculada aos dados isotópicos de C e Sr ($\delta^{13}\text{C} = -5.0$ e $^{87}\text{Sr}/^{86}\text{Sr} = 0.7077$) determinados em carbonatos de capa do Morro do Puga, sugerem que estes foram depositados após 700 Ma e podem representar um registro do evento glacial Marinoano (ca. 636 Ma).

1 **Geochronological constraints on Neoproterozoic Puga diamictites and associated**
2 **BIFs in the Southern Paraguay Belt, Brazil**

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Submitted to publication at Gondwana Research on September 7th, 2011

Abstract

The Paraguay belt comprises a thick sedimentary succession deposited on the southwestern border of the Amazonian craton and the Rio Apa Block. The base of the succession in the southern Paraguay belt is marked by a level of glacially derived deposits from the Puga Formation associated to banded iron formations, which are assumed to be Marinoan in age (635 Ma) in spite of the lack of geochronological data. Here we present the first U-Pb SHRIMP ages on detrital zircon grains separated from the matrix of six samples of these diamictites from two different localities (Puga Hill and Bodoquena area). U-Pb ages determined from two samples (ca. 130 grains) of Puga Hill show a large variation between 970 Ma and 2100 Ma. Rocks with these ages can be found in the Amazonian craton suggesting that it is the most probable source of the sediments. Detrital zircons (ca. 230 grains) from the Bodoquena area (about 200 km south of Puga Hill) range from 706 to 1990 Ma. The 1760 Ma source is significantly more important in these samples, comprising more than 70% of analysed grains, and corresponds to the adjacent Rio Apa Block. The youngest zircon was dated at 706 ± 9 Ma, thus constraining the maximum depositional age for the Puga Formation. Possible sources for this younger population could be either the juvenile Mara Rosa magmatic arc in the Brasilia belt, or the rocks from the Laurentian external fold belts located to the west of the sampled area in Neoproterozoic paleogeographic reconstructions. The maximum depositional age of the diamictites (and associated BIFs), together with carbonate carbon and strontium isotope data ($\delta^{13}\text{C} = -5.0$ and $^{87}\text{Sr}/^{86}\text{Sr} = 0.7077$) in Puga Hill, indicate that they were deposited after 700 Ma, suggesting that they may represent the Marinoan event.

Key-words: U-Pb SHRIMP geochronology, Southern Paraguay belt, Puga diamictites, Glaciation, Neoproterozoic, Sediment provenance.

1. Introduction

The Neoproterozoic encompasses significant paleoenvironmental changes, including the oxygenation of the atmosphere and the deep-ocean and events of extreme glaciations where ice caps reached tropical latitudes at least three times (Hoffman, 2009) in the Sturtian (~720 Ma), Marinoan (~635 Ma) and Gaskiers (~580 Ma) events. This period is also marked by the renaissance of banded iron formations after more than one billion years (Klein and Beukes, 1993). The Snowball Earth hypothesis postulates that widespread anoxia due to the ice-covered ocean enabled iron and manganese to accumulate in the seawater to be precipitated later in large amounts with the ventilation of the oceans in the glacial aftermath (Kirschvink, 1992). The BIF-diamictite association is thus one of the principal evidences of the Snowball Earth hypothesis (Hoffman et al., 1998, Hoffman and Schrag, 2002). One of the most important Neoproterozoic ore deposits of iron and manganese occurs in central Brazil, within the glacial Puga Formation (Trompette et al., 1998; Klein and Ladeira, 2004; Piacentini et al., 2008).

In central Brazil, glacial deposits in the São Francisco and Amazonian craton, and their respective marginal fold belts have been correlated to these three intervals. The older glacial event occurs exclusively in the São Francisco Craton and the Araçuaí fold belt, comprising diamictites and related facies from the Jequitaí and Bebedouro formations and the Macaúbas Group (Karfunkel and Hoppe, 1988; Uhlein et al., 1999; Figueiredo et al., 2009; Babinski et al., 2011). These glacial rocks are covered by limestones from the Sete Lagoas Formation, for which a Pb-Pb isochron age of 740 ± 22 Ma recently was obtained (Babinski et al., 2007). The ages of the other two glacial episodes in Central Brazil are much less well constrained. Both of them occur in the Amazonian Craton and the adjacent Paraguay belt, comprising the thick glaciomarine

deposits of the Puga Formation that occur along the Paraguay belt in its northern and southern branches (e.g., Alvarenga et al., in press), and the younger diamictites of the Serra Azul Formation restricted to the northern Paraguay belt (Alvarenga et al., 2007; Figueiredo et al., 2008). In this paper, we report new U-Pb SHRIMP ages on detrital zircons from Puga diamictites cropping out in the southern branch of the Paraguay belt. These data allow us to better constrain the detritus source areas and the age of deposition of both the glacial sediments and the iron-ores hosted by them.

2. Geological Setting

The Neoproterozoic Paraguay belt comprises a thick sedimentary succession metamorphosed at low-grade, comprises the Paraguay belt, which borders the southeastern margin of the Amazonian craton and the Rio Apa Block (Figure 1). The belt is divided in northern and southern branches, and their connection is hidden by Cenozoic deposits of the Pantanal basin. Both branches present a basal diamictite, the Puga Formation, covered by a thick carbonate succession, Araras Group in the north and Corumbá Group in the south. This lithostratigraphic similarity is the basis for regional correlation schemes between the two branches, but differences in their stratigraphy and evolution suggest otherwise (e.g., Alvarenga et al., 2009). But these models are difficult to test due to the scarcity of geochronological data in both parts of the Paraguay belt, particularly in its southern branch.

The Puga Formation consists of diamictite associated with conglomerate, sandstone, siltstone and shale. The widespread occurrence of diamictite indicates lateral transition from thinner coarse-grained beds close to the Amazonian Craton (Puga Formation) to thick fine-grained facies in the east (Cuiabá Group) (Alvarenga et al., in press). The sedimentation model proposed for the glacially influenced unit involves

three main glacial depositional systems: platformal, slope and outer slope (Alvarenga and Trompette, 1992). Discontinuous outcrops of the Puga Formation extend for more than 1000 km along the Paraguay Belt and over the Amazonian Craton and the Rio Apa Block (Fig. 1). The glacial diamictite succession was first described from the isolated Puga hill by Maciel (1959), who proposed a glaciogenic origin for this unit.

The southern part of the Paraguay belt extends southwards from the Corumbá to the Bodoquena Range (Figure 2). It extends westwards from Corumbá in western Brazil, into Bolivia as a large syncline known as the Tucavaca Belt (Litherland et al., 1986). In the Corumbá area, two lithostratigraphic units can be distinguished from base to top (Almeida, 1945; Dorr II, 1945): the Jacadigo and Corumbá groups. The Jacadigo Group is made up of Urucum and Santa Cruz formations. The basal Urucum Formation is composed of siliciclastic rocks, including arkosic sandstones and conglomeratic sandstones, unconformably overlying the basement of the Amazonian Craton. The overlying Santa Cruz Formation comprises jaspers capped by banded hematite-rich jaspilites with arkoses and manganese ore intercalations. According to Haralyi and Walde (1986) e Alvarenga et al. (in press) the Jacadigo Group was deposited in a system of grabens related to a process of crustal stretching. Basement boulders as big as 1 m³ were described in arkosic intercalations which cap the manganese beds. A glacially-influenced environment was invoked to explain the “dropped boulders” (Barbosa, 1949), although ice-rafted dropstones have not been found (Alvarenga and Trompette, 1992). Although geochronological data are not available for this unit, it has been interpreted as coeval to the glacially influenced Puga Formation described in the both parts of the belt. This idea has been also influenced by the Snowball Earth Hypothesis (Kirschvink, 1992; Hofmann et al., 1998), which predicts that large Fe deposits would form as a result of ocean anoxia during a long-lived and global-scale

glaciation. But this interpretation is contentious. Recently this unit was reinterpreted as continental siliciclastic deposits formed in alluvial fans, lacustrine and fluvial braided systems during rift initiation (Freitas et al., 2011), being in this way better correlated to the deposits of the Chiquitos-Tucavaca Belt, to the north of the area. The Corumbá Group, by its turn, is made up of dolostones at the base (Bocaina Formation) and limestones containing Ediacaran fauna at the top (Tamengo Formation). Fossils include *Corumbella weneri* (Hahn et al., 1982), acritarchs such as *Balvinella faveolata*, and *Cloudina luciano* (Zaine and Fairchild, 1987). This fauna can be correlated to that described by Germs (1972) in the lower part of the Nama Group, in Namibia (Trompette et al., 1998).

The relationship between the Jacadigo and Corumbá groups is still a matter of debate, mainly because both are affected by tectonism, which have obliterated the original contact relationships between them (Trompette et al., 1998). Observations from outcrops in Bolivia support the proposition of Haralyi (1972), that the Jacadigo Group is older and overlain by the rocks of the Corumbá Group.

2.1. Study area and sampling description

The main goal of our study was to determine the maximum depositional age of the glacial diamictites, and obtain information about the provenance of the sediments. Two areas ca. 200 km were selected which represent different sectors of the belt: (i) in the border of the Amazonian craton, the Puga Hill area represents the type section of the diamictites of the Puga Formation, and (ii) the Bodoquena Range is located to the east of the Rio Apa Block (Figure 1). Two samples were collected at the Puga Hill and two other samples were collected at the Bodoquena area, corresponding to the north and central sectors of the southern Paraguay belt, respectively.

Puga Hill: It is an isolated hill in the Paraguay river; the access is limited since it can only be reached by boat. This type section is made up by a massive diamictite with a sandy matrix, which is about 95 m thick; few dropstones are observed (Figure 3A, 3B). These diamictites are capped by pink limestone (Figure 3C) of the Bocaina Formation interpreted as a cap carbonate by Boggiani et al. (2003). Two samples (04-PUGA-2A and -2B) were collected at this site.

Bodoquena Area: It is located about 200 km south of the Puga Hill. At this site sediments are interpreted as glacio-marine. The diamictites show sandy and/or silty matrix and occur along the axis of folded rocks, being overlain by carbonates of the Corumbá Group. The clasts show a large range in size, form and lithology. At some sites (Santa Terezinha Farm; sample MS06-03, Fig. 3D), blocks of gneiss in the diamictite can reach 1.5 meters and some granitic clasts are as large as 40 cm. The size of the clasts decrease towards the northeast where clasts are usually about 5 cm or smaller (samples MS06-05 and -06, Figure 3E). About 10 km northeast, at the São Manoel Farm (sample MS06-07), a banded iron formation occurs as a 2 m thick bed, with a NS strike and dipping 45° to east, confined into a massive foliated diamictite with ferruginous matrix (Figures 3F-3H) affected by low grade metamorphism (Piacentini et al., 2008). Geological maps identify these metasediments as belonging to the Cuiabá Group (Corrêa et al., 1979, Nogueira et al., 1978, Godói, 2001) but they were redefined by Lacerda Filho (2006) as belonging to the Puga Formation.

3. Analytical Procedures

Zircon grains were separated using standard heavy liquid and magnetic techniques at the Geochronological Research Center, University of São Paulo. The U-Pb analyses were carried out on the SHRIMP I (04-PUGA-2A, MS06-03, MS06-07) and SHRIMP RG (04-PUGA-2B, MS06-05, MS06-06) ion microprobe at the Australian

National University. Grains were mounted in an epoxy disk with the Temora zircon standard (417 Ma; Black et al., 2003) and polished to expose their centers. Internal structures of zircon grains were revealed by cathodoluminescence (CL) image. The analytical procedures were similar to those described by Williams (1998). Five scans through the mass stations were made for each age determination. U abundance was calibrated using the standard SL13 (U = 238 ppm, Williams, 1998) and $^{206}\text{Pb}/^{238}\text{U}$ ratio was calibrated using the standard Temora (Black et al., 2003). Decay constants used for age calculation are those recommended by Steiger and Jäger (1977). Measured ^{204}Pb was applied for the common lead correction, and data processing was carried out using the Squid and Isoplot programs (Ludwig, 2001). The uncertainties for the measured ratios are given at the one sigma level. All obtained data were plotted on the concordia diagram, but only zircon ages with concordance degree > 85% were plotted on the age probability diagrams. During data discussion, U-Pb ages older than 800 Ma are mentioned as $^{207}\text{Pb}/^{206}\text{Pb}$ ages, and younger than 800 Ma are referred as $^{238}\text{U}/^{206}\text{Pb}$ ages.

4. Results

Zircon grains separated from six clastic sedimentary rocks were dated by U-Pb Sensitive High Resolution Ion Microprobe (SHRIMP) technique. Two samples were collected at the Puga Hill, and six samples are located on the Bodoquena area, and about 350 zircon grains were analyzed.

Puga Hill Diamictites:

Sample 04-PUGA-2A: The detrital zircon crystals separated from the matrix of this diamictite are mostly round or slightly rounded; some are broken grains with sharp edges. On the cathodoluminescent (CL) images oscillatory zoning can be observed. Few grains present core and rims, but only the later phases were dated. Their size

range from 60 to 180 μm , but most of them are ca. 100 μm (Figure 4). Sixty-five grains were dated and the age of those which are more than 85% concordant range from 970 to 2045 Ma (Table 1). Six grains were very discordant (more than 20%) and have high common Pb content, so these ages will not be considered to constrain the maximum depositional age of this rock (Table 1). All the data are plotted on the U-Pb concordia diagram (Figure 5), where clusters are observed between 1200-1400 Ma and 1700-1900 Ma, and the youngest concordant zircon has an age of 1000 ± 41 Ma (grain 54.1 in Table 1). On the probability histogram the main zircon populations show age peaks at 1224, 1736, and 1843 Ma (Figure 6).

Sample 04-PUGA-2B: Most of the zircon crystals from this sample are subhedral, elongated and slightly rounded, and show oscillatory zoning on the CL images (Figure 4). Their size range widely from 70 to 400 μm , but most of them are ca. 150 μm . Sixty eight grains yielded ages ranging from 990 to 2120 Ma (Table 2), and the main peak ages are at 995, 1208, 1537, 1782, 1858, and 2118 Ma (Figures 5, 6). Only one grain has an Archean age (2708 ± 20 Ma; not shown in the diagrams) indicating that this is not an important source of the sediments. The youngest concordant zircon has an age of 988 ± 17 Ma (grain 22.1 in Table 2), which sets the maximum depositional age of the Puga Formation at the Puga Hill.

Bodoquena Area Diamictites:

Sample MS06-03: It is a diamictite with sandy matrix and contains clasts reaching up to 30 cm in diameter, mostly from granites, gneisses and sandstones. The zircon crystals recovered from the matrix are elongated, slightly rounded and some are broken. Their size is between 50 and 180 μm , and display oscillatory zoning on CL images (Figure 4). Fifty nine grains were dated and the ages range from 1500 to 2000 Ma; some grains have high common Pb and present large errors on their ages (Table 3; Figure 5).

On the probability diagram, an expressive peak at 1728 Ma indicates that this is the predominant source area of the sediments (Figure 6). Only one grain has a Neoproterozoic age (729 ± 10 Ma). Some grains (about 12%) are discordant but are adjusted to a discordia with an upper intercept at 1700 Ma suggesting that they belong to the same population and have suffered recent Pb loss.

Sample MS06-05: It is a siltstone with small clasts of gneisses, granites and sandstones, generally smaller than 5 cm. The zircons are mostly subhedral, some are prismatic and show rounded edges; broken grains are common. CL images show oscillatory zoning and overgrowths are not observed (Figure 4). Their size range from 80 to 330 μm , but the majority are between 100 to 150 μm . The U-Pb ages determined on 62 grains range from 706 to 1830 Ma (Table 4; Figure 5), but two main populations with cluster ages at 726 and 1744 Ma predominate (Figure 6). The 700-720 Ma zircon population represents 8% of the sample; the youngest zircon has a U-Pb age of 706 ± 9 Ma, and establishes the maximum depositional age of the Puga diamictite at the Bodoquena area.

Sample MS06-06: This diamictite has the same matrix but less clasts than the sample MS06-05. The zircons are also subhedral, slightly rounded and their size is mostly ca. 120 μm . Fifty seven grains were dated and the ages are concentrated between 1200 and 1860 Ma (Table 5); it was found one Neoproterozoic grain (727 ± 9 Ma) and one Archean crystal (3003 ± 9 Ma; not shown in diagrams). Zircons from this sample show Pb loss and about 20% of the dated grains are more than 15% discordant (Figure 5). On the probability diagram, the most expressive peaks are at 1262 and 1766 Ma indicating that these are the main source of the detritus (Figure 6).

Sample MS06-07: Zircon grains were recovered from the matrix of a diamictite with banded iron intercalations. The zircons are euhedral to subhedral, pyramidal, some are

slightly rounded, and display igneous oscillatory zoning on CL images (Figure 4). Their size is mostly between 80 and 190 μm long. Fifty two grains were dated and except one Neoproterozoic grain (714 ± 13 Ma), all the others apparently come from the same source of ca. 1759 Ma (Table 6, Figures 5, 6)

5. Discussion and Conclusions

U-Pb data obtained on diamictites from the Puga Formation combined with field evidence allow us to discuss the provenance of the sediments and the maximum depositional age of the Puga diamictites in the southern branch of the Paraguay belt.

Constrasting U-Pb data determined on detrital zircons from diamictites collected at the Puga Hill, located close to the Amazonian Craton, and at the Bodoquena area, on the eastern side of the Rio Apa Block, about 200 km apart, indicate different source regions for provenance of the sediments in these areas (Figure 7). Diamictites from the Puga Hill show a variety of source-rocks compared to those from the Bodoquena area where sources of c. 710-730 Ma and 1730-1770 Ma have contributed to the initial sedimentation of the precursor basin of the Paraguay belt.

U-Pb ages (133 grains; Tables 1, 2) from the Puga Hill range from 970 Ma to 2120 Ma; only one Archean grain (2.7 Ga) was dated. Sources with ages of 990, 1200, 1400, 1540, 1750, 1780, and 1840 Ma predominate (Fig. 7), and all of them can be found in the nearby Amazonian craton, which has been divided into 6 geochronological provinces by Tassinari and Macambira (1999) (Figure 1): Central Amazonian Province (> 2.3 Ga), Maroni - Itacaiúnas Province (2.2 – 1.95 Ga), Ventuari – Tapajós (1.95 – 1.8 Ga), Rio Negro - Juruena Province (1.8 – 1.55 Ga), Rondonian – San Ignácio (1.55 – 1.3 Ga), and Sunsás (1.25 – 1.0 Ga).

According to this simple geochronological scheme, we suggest that the oldest provinces, the Central Amazonian Province (> 2.3 Ga) and the Maroni - Itacaiúnas Province (2.2 to 1.95 Ga) have not furnished sediments to the Puga diamictites. Rocks younger than 1.9 Ga represent the most important detritus source to the diamictites of the Puga Hill (Fig. 7). Sources of the 1840 Ma zircons could come from rocks of the Ventuari – Tapajós Province (1.95 – 1.8 Ga), which includes gneisses with U/Pb ages of 1.85-1.83 Ga at the Tapajós area (about 1500 km to the north). In addition, alkaline and calc-alkaline acid to intermediate volcanic rocks (Irirí Formation) and associated Maloquinha granitoids, with ages between 1.89 to 1.84 Ga have been described at the southern part of the Ventuari – Tapajós Province (Santos, 2003). Another possibility is that the 1.84 Ga zircons represent reworked grains from the metasediments of the Aguapeí Belt (Geraldes et al., 2001) located at the southern Amazonian craton, just to the north of the study area (Sunsás Province in Figure 1). This belt has detrital zircons with ages ($n=270$) ranging from 1.9 to 1.3 Ga (Vargas-Mattos et al., 2007). Moreover, the proximity of the Aguapeí belt to the study area favors this interpretation.

The source area of the 1780 Ma and 1540 Ma sediments is the Rio Negro - Juruena Province (1.8 – 1.55 Ga), most probably the Alto Jauru Terrane, including Alto Jauru suite (1.79 – 1.74 Ga) and Cachoeirinha suite (1.58 – 1.52 Ga), and other rocks dated in the southern part of the craton (Geraldes et al., 2001, Neder et al., 2002, Payolla et al., 2002, Pinho et al., 2003, Ruiz et al., 2004). The provenance of detrital zircons with ages of 1400 Ma is probable the Santa Helena arc (1.45 – 1.42 Ga; Geraldes et al., 2001) and Rio Branco anorogenic plutonic suite formed behind the arc (1.47 – 1.42 Ga; Geraldes et al., 2004). The 1200 Ma detrital zircons are found in the Aguapeí e Nova Brasilândia sequences in the Sunsás Province (Santos et al., 2000), which could have been eroded and acted as source for the Puga diamictites, as already suggested for the

1840 Ma zircons. The magmatic source of the 970-1000 Ma ages can be related to the widespread anorogenic granites and mafic complexes intruding the rocks in the southern part of the Craton (Litherland et al., 1986; Bettencourt et al., 1999, Rizzotto et al., 1999; Geraldès et al., 2001; Sparrenberger et al., 2002).

A very distinct scenario was observed from the detrital zircons from Puga diamictites cropping out at the Bodoquena area, east of the Rio Apa Block (Figures 1, 2). The U-Pb ages (227 grains) range from 706 Ma to 1990 Ma (except one Archean grain of 3.0 Ga) with a major cluster at 1758 Ma and two smaller peaks at 729 Ma and 1223 Ma (Fig. 7). The most obvious source of the 1.76 Ga zircons is the adjacent Rio Apa Block, where rocks with ages between 1.95 and 1.7 Ga are common (Cordani et al., 2010). Zircons with ages of 1.2 Ga have not been identified in this block yet. Similarly, rocks of 700 - 730 Ma have not been found in the area nor in the nearby cratons. However, it is worth mention that clasts of felsic volcanics with K-Ar ages of 730 Ma and 1100 -1200 Ma have been found in the Serra Azul Formation in the Northern Paraguay belt (Figueiredo, 2006, Alvarenga et al., 2007, Figueiredo et al., 2008). Also, volcanic rocks from Nova Xavantina region (Northern Paraguay Belt) yielded zircons exhibiting complex internal pattern, and suggest a U-Pb SHRIMP crystallization age of ca. 820 Ma (Dantas et al., 2007). In summary, although some hints suggest the 700 – 800 Ma volcanic event, no rocks were found yet in the area and the source of these zircons remain unknown and could have come either from other cratonic units of the supercontinent Rodinia beyond the Amazonian craton or from the Goiás Magmatic Arc (Pimentel and Fuck, 1992) which is located east of the Paraguay Belt.

The provenance study also shows that the sediments from the ferruginous diamictite (sample MS06-07 in Figures 5, 6) were sourced entirely from the Rio Apa Block, showing that glacial deposits contain mostly local or regional sources and do not

represent long-travelled sediments. The same glacial dynamics were observed on diamictites from the São Francisco craton where all sources of the sediments were identified as the basement or supracrustal rocks from the surrounding uplifted areas (Figueiredo et al., 2009). If it is correct to assume that most of the sources are not long-travelled sediments, the 700 Ma young source which is not found in Amazonian craton or the Paraguay belt basement must have been covered by sediments or was mostly eroded. Otherwise, if a distant source is to be considered, the best candidates are igneous sources from the eastern margin of Laurentia, which were likely juxtaposed to the Rio Apa and the Amazonian craton at Neoproterozoic times (McCausland et al., 2007, 2011).

The youngest zircon dated from diamictites collected at the Puga Hill, the type area of the glacial diamictites, is 970 Ma (Table 1). Unfortunately, this age does not help to constrain the depositional age of these sediments. However, U-Pb ages determined on diamictites collected at the Bodoquena area, yielded younger zircons and the age of 706 ± 9 Ma determined on the youngest dated zircon constrains the maximum depositional age of this diamictites. Considering the triad of glacial events of Neoproterozoic age (Sturtian: 720 Ma, Marinoan: 635 Ma, and Gaskiers: 580 Ma; Hoffman and Li, 2009), the Puga diamictites are better correlated to the Marinoan event given the carbon and strontium isotope signals of the Puga Hill cap carbonates displaying $\delta^{13}\text{C}$ of -5.0 ‰ (Boggiani et al., 2003) and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7077 (Sr concentration of 400 ppm) which are typical Sr isotopic ratios of Marinoan carbonates (Halverson et al., 2007; Melezhik et al., 2008).

Acknowledgments

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333 The authors acknowledge financial support provided by FAPESP (Fundação de
334 Amparo à Pesquisa do Estado de São Paulo) and CNPq (Conselho Nacional de
335 Desenvolvimento Científico e Tecnológico).
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544 **Table Captions**

545 **Table 1.** U-Pb SHRIMP data from detrital zircons of sample 04-PUGA-2A

546 **Table 2.** U-Pb SHRIMP data from detrital zircons of sample 04-PUGA-2B

547 **Table 3.** U-Pb SHRIMP data from detrital zircons of sample MS06-03

548 **Table 4.** U-Pb SHRIMP data from detrital zircons of sample MS06-05

549 **Table 5.** U-Pb SHRIMP data from detrital zircons of sample MS06-06

550 **Table 6.** U-Pb SHRIMP data from detrital zircons of sample MS06-07

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Figure Captions

Figure 1. Geological map of the Amazonian Craton showing the geochronological provinces defined by Tassinari and Macambira (1999), the Rio Apa Craton and Paraguay belt (modified from Cordani et al, 2009).

Figure 2. Detailed geological map of the Paraguay belt on the southeastern part of the Amazonian Craton and Rio Apa Block (modified from Alvarenga et al., 2000 and Litherland et al., 1986), showing the location of the study areas (Puga Hill and Bodoquena).

Figure 3. Puga Formation and cap carbonate. **A, B:** dropstones of sandstone(A) and carbonate (B) diamictite at the Puga Hill; **C:** pink dolostone interpreted as cap carbonate at the Puga Hill; **D:** diamictite with clasts as large as 40 cm from the Bodoquena area; **E:** diamictite with clasts smaller than 5 cm from the Bodoquena area; **G, H, I:** Outcrops of the ferruginous diamictite from the Bodoquena area.

Figure 4. Cathodoluminescence images of analyzed detrital zircon crystals.

Figure 5. U-Pb Concordia diagrams for the six diamictite samples.

Figure 6. Frequency histogram and probability curves of zircon Pb/Pb ages (< 15% discordant) for samples from the Bodoquena (A to D) and the Puga Hill (E-F) areas.

Figure 7. Frequency histogram and probability curves of zircon Pb/Pb ages (< 15% discordant) for samples of the Puga Formation collected at the Puga Hill and in the Bodoquena area.

Figure 1

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Figure 7

Table 1

Table 1. Summary of SHRIMP U-Pb zircon results for sample 04-PUGA-2A

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Total Ratios				Radiogenic Ratios						Age (Ma)				% Disc	
							²³⁸ U/ ²⁰⁶ Pb		²⁰⁷ Pb/ ²⁰⁶ Pb		²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²³⁵ U		²⁰⁷ Pb/ ²⁰⁶ Pb		²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²⁰⁶ Pb			
							±	±	±	±	±	±	±	±	±	±	±	±	±	±		
							±	±	±	±	±	±	±	±	±	±	±	±	±	±		
1.1	38	22	0.58	10.6	0.001508	2.39	3.078	0.053	0.1221	0.0013	0.3171	0.0058	4.434	0.260	0.1014	0.0057	0.311	1775	28	1650	103	-8
2.1	222	119	0.54	56.6	0.000422	0.65	3.377	0.044	0.1174	0.0006	0.2942	0.0038	4.530	0.078	0.1117	0.0012	0.760	1662	19	1827	20	9
3.1	394	217	0.55	63.4	0.000950	1.58	5.344	0.070	0.0958	0.0006	0.1842	0.0024	2.094	0.052	0.0824	0.0017	0.535	1090	13	1256	41	13
4.1	116	88	0.76	31.0	0.000178	0.28	3.216	0.044	0.1089	0.0014	0.3101	0.0043	4.552	0.093	0.1065	0.0016	0.674	1741	21	1740	28	0
5.1	210	195	0.93	59.2	0.000209	0.32	3.042	0.039	0.1154	0.0005	0.3277	0.0042	5.087	0.074	0.1126	0.0008	0.888	1827	21	1842	12	1
6.1	833	50	0.06	118.2	0.002321	3.79	6.053	0.074	0.1218	0.0004	0.1589	0.0020	1.961	0.059	0.0895	0.0024	0.415	951	11	1414	52	33
7.1	315	203	0.64	46.7	0.000341	0.58	5.799	0.074	0.0761	0.0005	0.1714	0.0022	1.684	0.032	0.0712	0.0010	0.671	1020	12	964	29	-6
8.1	470	388	0.83	61.5	0.002307	3.64	6.571	0.083	0.1358	0.0012	0.1467	0.0019	2.108	0.065	0.1042	0.0029	0.419	882	11	1701	52	48
9.1	147	66	0.45	31.0	0.000257	0.42	4.061	0.057	0.0913	0.0006	0.2452	0.0034	2.964	0.057	0.0877	0.0011	0.731	1414	18	1375	25	-3
10.1	476	209	0.44	122.1	0.000260	0.40	3.350	0.042	0.1142	0.0004	0.2973	0.0037	4.536	0.062	0.1107	0.0006	0.911	1678	18	1810	10	7
11.1	159	91	0.57	28.5	0.000486	0.82	4.783	0.063	0.0848	0.0006	0.2073	0.0028	2.228	0.048	0.0779	0.0013	0.613	1215	15	1145	34	-6
12.1	675	382	0.57	110.0	0.002378	3.67	5.277	0.064	0.1460	0.0004	0.1825	0.0023	2.867	0.072	0.1139	0.0025	0.496	1081	12	1863	40	42
13.1	225	137	0.61	58.7	0.000494	0.76	3.286	0.042	0.1249	0.0006	0.3020	0.0039	4.924	0.078	0.1182	0.0011	0.815	1701	19	1930	16	12
14.1	122	109	0.89	33.1	0.000345	0.53	3.169	0.043	0.1174	0.0007	0.3138	0.0042	4.877	0.081	0.1127	0.0011	0.813	1760	21	1844	17	5
15.1	75	54	0.72	19.2	0.000087	0.14	3.351	0.048	0.1109	0.0009	0.2980	0.0043	4.510	0.084	0.1098	0.0013	0.768	1682	21	1796	22	6
16.1	658	247	0.38	73.5	0.001567	4.33	7.687	0.094	0.1001	0.0006	0.1267	0.0016	1.377	0.035	0.0789	0.0018	0.482	756	9	1145	48	34
17.1	97	38	0.39	19.4	0.000300	0.50	4.290	0.060	0.0875	0.0008	0.2320	0.0033	2.663	0.061	0.0833	0.0015	0.612	1345	17	1275	36	-5
18.1	334	189	0.57	61.6	0.000096	0.16	4.649	0.059	0.0814	0.0005	0.2148	0.0027	2.372	0.036	0.0801	0.0007	0.829	1254	14	1199	17	-5
19.1	104	80	0.77	29.8	0.000424	0.66	2.993	0.041	0.1164	0.0007	0.3319	0.0046	5.063	0.104	0.1106	0.0017	0.674	1847	22	1810	27	-2
20.1	362	321	0.89	98.7	0.000872	1.35	3.147	0.039	0.1241	0.0006	0.3135	0.0039	4.852	0.083	0.1123	0.0013	0.728	1758	19	1836	21	4
21.1	198	133	0.67	51.8	0.000319	0.50	3.287	0.042	0.1114	0.0005	0.3027	0.0039	4.469	0.068	0.1071	0.0009	0.851	1705	19	1750	15	3
22.1	608	708	1.16	104.4	0.002169	3.42	5.000	0.061	0.1337	0.0004	0.1932	0.0024	2.771	0.072	0.1040	0.0024	0.480	1139	13	1697	42	33
23.1	141	55	0.39	24.8	0.000395	0.66	4.891	0.065	0.0835	0.0006	0.2031	0.0027	2.181	0.051	0.0779	0.0015	0.570	1192	15	1144	39	-4
24.1	286	230	0.80	65.2	0.000671	1.05	3.766	0.047	0.1148	0.0005	0.2628	0.0033	3.828	0.062	0.1057	0.0011	0.782	1504	17	1726	18	13
25.1	54	41	0.76	16.5	0.000698	1.09	2.825	0.043	0.1195	0.0010	0.3501	0.0054	5.312	0.143	0.1101	0.0024	0.571	1935	26	1800	40	-7
26.1	125	109	0.87	36.1	0.000390	0.60	2.964	0.040	0.1178	0.0007	0.3353	0.0045	5.203	0.087	0.1125	0.0011	0.798	1864	22	1841	18	-1
27.1	108	40	0.37	19.7	0.000191	0.32	4.707	0.065	0.0836	0.0007	0.2118	0.0029	2.363	0.043	0.0809	0.0010	0.760	1238	16	1220	23	-2
28.1	206	83	0.40	36.4	0.000209	0.35	4.877	0.063	0.0851	0.0005	0.2043	0.0026	2.315	0.046	0.0822	0.0012	0.653	1199	14	1249	29	4
29.1	141	52	0.37	25.6	0.000326	0.55	4.728	0.063	0.0834	0.0006	0.2104	0.0028	2.286	0.048	0.0788	0.0013	0.641	1231	15	1167	32	-5
30.1	127	98	0.77	32.6	0.000354	0.55	3.339	0.044	0.1190	0.0008	0.2979	0.0040	4.694	0.085	0.1143	0.0014	0.734	1681	20	1869	22	10
31.1	48	43	0.90	14.3	0.000301	0.46	2.915	0.045	0.1204	0.0010	0.3414	0.0053	5.476	0.119	0.1163	0.0018	0.710	1894	25	1900	28	0
32.1	150	137	0.91	39.8	0.000241	0.38	3.232	0.042	0.1080	0.0005	0.3082	0.0040	4.451	0.070	0.1048	0.0009	0.836	1732	20	1710	16	-1
33.1	85	51	0.60	21.4	0.000460	0.74	3.397	0.048	0.1034	0.0008	0.2922	0.0042	3.911	0.086	0.0971	0.0016	0.650	1653	21	1569	31	-5
34.1	164	101	0.62	29.1	0.000861	1.45	4.820	0.064	0.0901	0.0006	0.2045	0.0027	2.195	0.068	0.0779	0.0022	0.434	1199	15	1143	55	-5
35.1	181	88	0.49	33.4	0.000188	0.31	4.655	0.061	0.0842	0.0005	0.2141	0.0028	2.407	0.040	0.0815	0.0008	0.791	1251	15	1234	20	-1
36.1	524	369	0.70	99.1	0.003773	6.09	4.547	0.056	0.1463	0.0004	0.2065	0.0027	2.678	0.117	0.0940	0.0039	0.298	1210	14	1509	79	20
37.1	188	137	0.73	46.9	0.000452	0.70	3.439	0.044	0.1199	0.0006	0.2887	0.0037	4.529	0.073	0.1138	0.0011	0.801	1635	19	1860	18	12
38.1	86	50	0.58	14.3	0.000751	1.27	5.167	0.074	0.0860	0.0008	0.1911	0.0028	1.984	0.068	0.0753</							

Table 2. Summary of SHRIMP U-Pb zircon results for sample 04-PUGA-2B

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Total Ratios				Radiogenic Ratios						Age (Ma)						% Disc
							²³⁸ U/ ²⁰⁶ Pb		²⁰⁷ Pb/ ²⁰⁶ Pb		²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²³⁵ U		²⁰⁷ Pb/ ²⁰⁶ Pb		²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²⁰⁶ Pb				
							±	±	±	±	±	±	±	±	±	±	±	±	±	±			
1.1	197	146	0.74	55.8	0.000096	0.15	3.041	0.039	0.1145	0.0008	0.3283	0.0042	5.124	0.077	0.1132	0.0009	0.852	1830	20	1851	14	1	
2.1	115	104	0.90	18.7	0.000112	0.19	5.288	0.076	0.0782	0.0010	0.1887	0.0027	1.994	0.042	0.0766	0.0012	0.682	1115	15	1111	31	0	
3.1	51	15	0.29	16.3	0.000154	0.23	2.717	0.051	0.1343	0.0018	0.3672	0.0068	6.696	0.163	0.1322	0.0021	0.767	2016	32	2128	27	5	
4.1	158	55	0.35	34.8	0.000095	0.15	3.894	0.052	0.0966	0.0011	0.2564	0.0034	3.369	0.063	0.0953	0.0012	0.722	1471	18	1534	24	4	
5.1	95	46	0.48	16.5	0.000089	0.15	4.956	0.074	0.0821	0.0011	0.2015	0.0030	2.246	0.050	0.0809	0.0013	0.675	1183	16	1218	32	3	
6.1	517	138	0.27	100.7	-	<0.01	4.411	0.052	0.0860	0.0005	0.2267	0.0027	2.692	0.036	0.0861	0.0005	0.888	1317	14	1340	12	2	
7.1	43	15	0.35	6.9	0.000158	0.27	5.397	0.103	0.0753	0.0017	0.1848	0.0036	1.860	0.076	0.0730	0.0026	0.471	1093	19	1014	73	3	
8.1	133	80	0.61	34.4	0.000544	0.85	3.306	0.045	0.1138	0.0010	0.2999	0.0041	4.398	0.095	0.1064	0.0018	0.633	1691	20	1738	31	3	
9.1	54	15	0.28	9.2	-	<0.01	4.972	0.087	0.0835	0.0015	0.2015	0.0035	2.362	0.059	0.0850	0.0015	0.697	1183	19	1316	35	10	
10.1	76	46	0.60	16.2	0.000131	0.21	4.018	0.063	0.0928	0.0012	0.2483	0.0039	3.113	0.072	0.0909	0.0015	0.679	1430	20	1445	32	1	
11.1	87	53	0.61	15.7	0.000240	0.40	4.747	0.073	0.0832	0.0012	0.2098	0.0032	2.308	0.056	0.0798	0.0015	0.627	1228	17	1192	38	-2	
12.1	466	111	0.24	109.1	0.000009	0.02	3.665	0.043	0.0951	0.0007	0.2728	0.0032	3.573	0.050	0.0950	0.0007	0.834	1555	16	1528	15	-3	
13.1	131	61	0.46	26.1	-	<0.01	4.314	0.060	0.0868	0.0009	0.2322	0.0032	2.825	0.054	0.0882	0.0011	0.731	1346	17	1388	25	3	
14.1	264	85	0.32	61.6	0.000026	0.04	3.688	0.048	0.0980	0.0012	0.2710	0.0035	3.649	0.067	0.0976	0.0013	0.706	1546	18	1580	24	2	
15.1	688	220	0.32	107.0	0.000763	1.27	5.523	0.073	0.0912	0.0005	0.1787	0.0024	1.982	0.042	0.0804	0.0013	0.628	1060	13	1208	32	12	
16.1	136	40	0.29	31.6	0.000558	0.90	3.700	0.055	0.1024	0.0009	0.2679	0.0040	3.497	0.104	0.0947	0.0024	0.503	1530	20	1522	49	-1	
17.1	251	53	0.21	58.6	0.000080	0.13	3.681	0.051	0.0954	0.0006	0.2713	0.0038	3.528	0.056	0.0943	0.0007	0.867	1548	19	1514	15	-2	
18.1	165	90	0.55	50.9	0.000038	0.06	2.785	0.040	0.1200	0.0008	0.3589	0.0052	5.914	0.094	0.1195	0.0008	0.908	1977	24	1949	12	-1	
20.1	79	39	0.49	15.8	-	<0.01	4.315	0.070	0.0886	0.0012	0.2321	0.0038	2.883	0.061	0.0901	0.0012	0.776	1346	20	1427	25	6	
21.1	183	89	0.49	47.5	0.000014	0.02	3.309	0.047	0.1094	0.0008	0.3021	0.0043	4.549	0.073	0.1092	0.0008	0.892	1702	21	1786	13	5	
22.1	371	20	0.05	53.0	0.000038	0.06	6.020	0.082	0.0726	0.0006	0.1660	0.0023	1.649	0.026	0.0721	0.0006	0.850	990	12	988	17	0	
23.1	165	48	0.29	30.1	0.000045	0.08	4.709	0.069	0.0807	0.0008	0.2122	0.0031	2.342	0.045	0.0801	0.0010	0.763	1240	17	1199	24	-3	
24.1	422	212	0.50	85.2	0.000034	0.06	4.260	0.057	0.0866	0.0005	0.2346	0.0031	2.787	0.041	0.0861	0.0005	0.914	1359	16	1341	11	-1	
25.1	103	62	0.61	30.1	0.000091	0.14	2.932	0.047	0.1162	0.0010	0.3405	0.0055	5.397	0.099	0.1149	0.0010	0.873	1889	26	1879	16	-1	
26.1	375	64	0.17	101.3	0.003634	5.66	3.175	0.043	0.1595	0.0007	0.2971	0.0042	4.515	0.181	0.1102	0.0041	0.353	1677	21	1803	68	7	
27.1	63	19	0.30	10.7	0.000112	0.19	5.084	0.087	0.0810	0.0013	0.1963	0.0034	2.150	0.055	0.0794	0.0015	0.667	1156	18	1183	38	2	
28.1	47	38	0.80	13.8	0.000187	0.29	2.923	0.053	0.1146	0.0015	0.3411	0.0062	5.270	0.133	0.1121	0.0020	0.720	1892	30	1833	32	-3	
29.1	97	63	0.65	26.3	0.000048	0.07	3.161	0.052	0.1092	0.0010	0.3161	0.0052	4.731	0.090	0.1085	0.0011	0.855	1771	25	1775	18	0	
30.1	243	107	0.44	74.3	0.000549	0.83	2.816	0.039	0.1326	0.0007	0.3522	0.0049	6.082	0.103	0.1253	0.0012	0.819	1945	23	2033	17	4	
31.1	305	251	0.82	86.3	0.000018	0.03	3.036	0.042	0.1091	0.0006	0.3293	0.0045	4.944	0.074	0.1089	0.0006	0.919	1835	22	1781	11	-3	
32.1	559	139	0.25	116.1	0.001739	2.81	4.139	0.055	0.1180	0.0007	0.2348	0.0032	3.038	0.084	0.0938	0.0023	0.490	1360	17	1505	45	10	
33.1	63	54	0.86	16.9	0.000494	0.77	3.186	0.055	0.1153	0.0014	0.3115	0.0054	4.663	0.132	0.1086	0.0024	0.611	1748	26	1776	41	2	
34.1	113	83	0.73	31.8	-	<0.01	3.064	0.047	0.1070	0.0010	0.3268	0.0050	4.875	0.090	0.1082	0.0011	0.826	1823	24	1769	19	-3	
35.1	50	13	0.26	26.7	0.000300	0.41	1.619	0.032	0.1898	0.0021	0.6150	0.0121	15.782	0.364	0.1861	0.0023	0.850	3090	48	2708	20	-14	
36.1	140	133	0.94	38.6	0.000034	0.05	3.127	0.046	0.1136	0.0009	0.3196	0.0047	4.986	0.083	0.1131	0.0009	0.884	1788	23	1851	14	3	
37.1	139	122	0.88	32.4	0.000085	0.14	3.676	0.055	0.0976	0.0009	0.2717	0.0041	3.613	0.065	0.0965	0.0010	0.824	1549	21	1557	19	0	
38.1	141	51	0.36	25.3	0.000036	0.06	4.784	0.072	0.0823	0.0011	0.2089	0.0031	2.357	0.049	0.0818	0.0012	0.720	1223	17	1242	28	2	
39.1	123	41	0.34	23.7	0.000029	0.05	4.440	0.068	0.0882	0.0010	0.2251	0.0034	2.725	0.052	0.0878	0.0010	0.793	1309	18	1378	23	5	
40.1	215	75	0.35	40.4	-	<0.01	4.578	0.067	0.0817	0.0007	0.2186	0.0032	2.484	0.043	0.0824	0.0007	0.855	1274	17	1255	17	-2	
41.1	166	107	0.64	46.8	0.000037	0.06	3.051	0.044	0.1068	0.0008	0.3275	0.0047	4.800	0.079	0.1063	0.0008	0.879	1826	23	1737	14	-5	
42.1	189	78	0.41	44.8	0.000015	0.02	3.627	0.052	0.0958	0.0007	0.2757	0.0039	3.635	0.059	0.0956	0.0008	0.875	1569	20	1541	15	-2	
43.1	160	2	0.01	46.1	0.000000	0.00	2.977	0.043	0.1109	0.0008	0.3359	0.0049	5.135	0.084	0.1109	0.0008	0.892	1867	24	1814	13	-3	
44.1	157	63	0.40	22.2	0.000115	0.20	6.070	0.091	0.0702	0.0009	0.1644	0.0025	1.555	0.033	0.0686	0.0010	0.712	981	14	886	31	-11	
45.1	15	16	1.06	4.6	0.000172	0.26	2.824	0.076	0.1205	0.0027	0.3531	0.0095	5.756	0.219	0.1182	0.0032	0.710	1950	45	1929	48	-1	
46.1	180	100	0.55	40.4	0.000040	0.07	3.833	0.056	0.0952	0.0008	0.2607	0.0038	3.404	0.058	0.0947	0.0009	0.848	1494	19	1522	17	2	
47.1	101	50	0.50	14.5	0.000247	0.42	5.993	0.095	0.0735	0.0011	0.1662	0.0027	1.604	0.052	0.0700	0.0020	0.497	991	15	929	57	-7	
48.1	57	25	0.44	11.7	0.000088	0.14	4.187	0.074	0.0913	0.0017	0.2385	0.0042	2.963	0.084	0.0901	0.0020	0.622	1379	22	1428	42	3	
49.1	269	184	0.68	65.4	0.000107	0.17	3.533	0.050	0.1002	0.0007	0.2826	0.0040	3.848	0.064	0.0988	0.0009							

Table 3

Table 3. Summary of SHRIMP U-Pb zircon results for sample MS06-03

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	Total Ratios				Radiogenic Ratios						Age (Ma)				% Disc		
						f206 %	²³⁸ U/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	±	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	ρ	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²⁰⁶ Pb	±			
						±	±	±	±	±	±	±	±	±	±	±	±	±				
1.1	102	79	0.78	26.9	0.000564	0.89	3.250	0.044	0.1088	0.0009	0.3050	0.0041	4.248	0.107	0.1010	0.0021	0.541	1716	20	1643	39	-4
2.1	69	47	0.69	18.1	0.000508	0.80	3.255	0.049	0.1106	0.0013	0.3048	0.0046	4.357	0.120	0.1037	0.0024	0.545	1715	23	1691	43	-1
3.1	129	128	0.99	33.4	0.000756	1.19	3.314	0.043	0.1127	0.0009	0.2982	0.0039	4.209	0.101	0.1024	0.0021	0.548	1682	19	1668	37	-1
4.1	109	49	0.45	29.8	0.000577	0.90	3.126	0.042	0.1096	0.0009	0.3170	0.0043	4.443	0.105	0.1017	0.0020	0.578	1775	21	1655	36	-7
5.1	292	368	1.26	66.0	0.000496	0.80	3.801	0.043	0.1112	0.0006	0.2610	0.0030	3.757	0.061	0.1044	0.0012	0.710	1495	15	1704	21	12
6.1	133	130	0.98	35.3	0.000143	0.22	3.232	0.042	0.1076	0.0008	0.3087	0.0040	4.497	0.079	0.1056	0.0013	0.737	1734	20	1726	22	-1
7.1	163	155	0.95	46.1	0.000248	0.38	3.043	0.038	0.1104	0.0009	0.3273	0.0041	4.829	0.086	0.1070	0.0014	0.705	1825	20	1749	23	-4
8.1	79	61	0.77	21.3	0.000633	0.99	3.186	0.046	0.1106	0.0011	0.3108	0.0045	4.366	0.116	0.1019	0.0023	0.549	1745	22	1659	41	-5
9.1	206	116	0.56	40.5	0.000572	0.94	4.373	0.052	0.1027	0.0007	0.2265	0.0027	2.960	0.064	0.0948	0.0017	0.554	1316	14	1523	34	14
10.1	151	110	0.73	39.3	0.000526	0.83	3.305	0.041	0.1102	0.0008	0.3001	0.0037	4.263	0.079	0.1030	0.0014	0.672	1692	18	1679	25	-1
11.1	90	42	0.46	21.3	0.000531	0.85	3.606	0.051	0.1023	0.0010	0.2749	0.0039	3.601	0.092	0.0950	0.0020	0.555	1566	20	1528	40	-2
12.1	128	94	0.73	34.3	0.000575	0.90	3.221	0.041	0.1095	0.0008	0.3077	0.0040	4.312	0.101	0.1016	0.0020	0.555	1729	20	1654	36	-5
13.1	350	392	1.12	82.6	0.000264	0.42	3.636	0.041	0.1084	0.0005	0.2738	0.0031	3.957	0.054	0.1048	0.0008	0.814	1560	15	1711	15	9
14.1	53	45	0.85	14.1	0.001467	2.31	3.220	0.052	0.1122	0.0013	0.3034	0.0051	3.840	0.184	0.0918	0.0041	0.351	1708	25	1463	85	-17
15.1	222	353	1.59	53.8	0.001202	1.92	3.539	0.042	0.1216	0.0007	0.2771	0.0033	4.017	0.097	0.1051	0.0022	0.497	1577	17	1717	39	8
16.1	191	215	1.12	55.5	0.000131	0.20	2.956	0.038	0.1118	0.0007	0.3376	0.0043	5.122	0.081	0.1101	0.0010	0.803	1875	21	1800	17	-4
17.1	114	37	0.32	27.7	0.000774	1.24	3.530	0.048	0.0979	0.0011	0.2798	0.0039	3.360	0.132	0.0871	0.0032	0.357	1590	20	1363	71	-17
18.1	89	72	0.82	24.0	0.000361	0.57	3.177	0.044	0.1102	0.0010	0.3130	0.0044	4.544	0.097	0.1053	0.0017	0.656	1755	21	1720	29	-2
19.1	181	157	0.86	49.3	0.000269	0.42	3.155	0.039	0.1103	0.0007	0.3157	0.0039	4.639	0.077	0.1066	0.0012	0.737	1769	19	1742	21	-2
20.1	74	54	0.73	20.7	0.000441	0.69	3.090	0.045	0.1135	0.0011	0.3214	0.0047	4.763	0.117	0.1075	0.0021	0.598	1797	23	1757	36	-2
21.1	194	142	0.73	50.6	0.000282	0.44	3.297	0.040	0.1080	0.0007	0.3019	0.0036	4.334	0.066	0.1041	0.0010	0.787	1701	18	1699	17	0
22.1	260	177	0.68	70.4	0.000338	0.53	3.180	0.038	0.1094	0.0008	0.3128	0.0038	4.520	0.085	0.1048	0.0015	0.641	1755	18	1711	26	-3
23.1	165	162	0.98	46.4	0.000351	0.54	3.054	0.038	0.1101	0.0007	0.3257	0.0041	4.731	0.078	0.1054	0.0011	0.758	1817	20	1721	20	-6
24.1	309	478	1.54	65.5	0.000501	0.82	4.060	0.046	0.1111	0.0006	0.2443	0.0028	3.511	0.057	0.1042	0.0012	0.701	1409	14	1701	21	17
25.1	189	168	0.89	51.5	0.000115	0.18	3.164	0.038	0.1073	0.0007	0.3155	0.0038	4.600	0.067	0.1058	0.0008	0.834	1768	19	1727	15	-2
26.1	428	218	0.51	79.4	0.001049	1.74	4.635	0.051	0.1132	0.0010	0.2120	0.0024	2.886	0.060	0.0987	0.0018	0.530	1239	13	1600	33	23
27.1	204	131	0.64	55.6	0.000315	0.49	3.150	0.038	0.1097	0.0007	0.3159	0.0038	4.590	0.076	0.1054	0.0012	0.728	1770	19	1721	21	-3
28.1	186	136	0.73	42.1	0.000275	0.44	3.804	0.046	0.0972	0.0007	0.2617	0.0032	3.369	0.063	0.0934	0.0013	0.659	1499	16	1495	27	0
29.1	97	83	0.86	27.7	0.000579	0.90	3.004	0.041	0.1119	0.0013	0.3299	0.0046	4.733	0.131	0.1040	0.0025	0.506	1838	22	1697	44	-8
30.1	131	89	0.68	34.0	0.000382	0.60	3.305	0.042	0.1133	0.0008	0.3008	0.0039	4.481	0.090	0.1081	0.0017	0.640	1695	19	1767	28	4
31.1	231	116	0.50	46.8	0.001997	3.29	4.248	0.056	0.1219	0.0023	0.2277	0.0031	2.955	0.126	0.0941	0.0038	0.318	1322	16	1511	76	12
32.1	40	28	0.71	10.9	0.000904	1.41	3.109	0.056	0.1155	0.0015	0.3171	0.0058	4.508	0.178	0.1031	0.0036	0.463	1776	28	1681	65	-6
33.1	245	151	0.62	67.4	0.000476	0.74	3.123	0.036	0.1097	0.0006	0.3178	0.0037	4.522	0.088	0.1032	0.0016	0.603	1779	18	1682	29	-6
34.1	510	429	0.84	73.5	0.002077	3.56	5.960	0.067	0.1269	0.0009	0.1618	0.0019	2.191	0.065	0.0982	0.0027	0.392	967	11	1590	51	39
35.1	216	224	1.03	60.5	0.000345	0.54	3.076	0.036	0.1100	0.0006	0.3233	0.0038	4.695	0.073	0.1053	0.0011	0.761	1806	19	1720	19	-5
36.1	217	186	0.85	67.7	0.000142	0.21	2.757	0.033	0.1218	0.0006	0.3619	0.0043	5.985	0.089	0.1199	0.0011	0.798	1991	20	1955	16	-2
37.1	256	134	0.52	64.9	0.000323	0.51	3.395	0.039	0.1099	0.0008	0.2931	0.0034	4.264	0.066	0.1055	0.0011	0.744	1657	17	1723	19	4
38.1	323	186	0.58	75.2	0.000210	0.34	3.690	0.042	0.0953	0.0005	0.2701	0.0031	3.440	0.051	0.0924	0.0009	0.767	1541	16	1475	18	-5
39.1	220	149	0.68	58.6	0.000674	1.06	3.220	0.038	0.1137	0.0007	0.3073	0.0037	4.426	0.093	0.1044	0.0018	0.577	1727	18	1705	31	-1
40.1	279	218	0.78	75.2	0.000247	0.39	3.188	0.037	0.1099	0.0006	0.3125	0.0036	4.592	0.066	0.1066	0.0009	0.799	1753	18	1742	16	-1
41.1	273	160	0.58	59.9	0.000818	1.33	3.923	0.047	0.1136	0.0007	0.2515	0.0031	3.551	0.091	0.1024	0.0023	0.478	1446	16	1668	41	13
42.1	143	81	0.57	39.3	0.000300	0.47	3.135	0.040	0.1121	0.0011	0.3175	0.0041	4.730	0.093	0.1081	0.0016	0.652	1778	20	1767	27	-1
43.1	204	114	0.56	55.7	0.000152	0.24	3.149	0.038	0.1103	0.0007	0.3168	0.0038	4.730	0.070	0.1083	0.0009	0.809	1774	19	1771	16	0
44.1	143	99	0.69	14.8	0.001207	0.61	8.301	0.113	0.0687	0.0015	0.1179	0.0017	0.829	0.066	0.0510	0.0040	0.179	729	10			
45.1	59	49	0.83	15.3	0.002378	3.77	3.295	0.053	0.1335	0.0055	0.2921	0.0058	4.062	0.472	0.1009	0.0116	0.171	1652	29	1640	213	-1
46.1	167	143	0.86	45.0	0.000231	0.36	3.186	0.040														

Table 4

Table 4. Summary of SHRIMP U-Pb zircon results for sample MS06-05

Grain spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f206 %	Total Ratios				Radiogenic Ratios						ρ	Age (Ma)				
							²³⁸ U/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²⁰⁶ Pb		%				
							±	±	±	±	±	±	±	±	±	±		Disc				
1.1	55	32	0.57	15.4	0.000150	0.23	3.073	0.045	0.108	0.001	0.325	0.005	4.756	0.094	0.106	0.001	0.748	1813	23	1736	24	-4
2.1	247	380	1.54	65.2	0.000089	0.14	3.251	0.038	0.108	0.001	0.307	0.004	4.503	0.062	0.106	0.001	0.844	1727	18	1737	14	1
3.1	212	283	1.34	56.7	0.000177	0.28	3.206	0.037	0.109	0.001	0.311	0.004	4.590	0.067	0.107	0.001	0.804	1746	18	1749	16	0
4.1	119	107	0.90	33.6	0.000000	0.00	3.032	0.038	0.107	0.001	0.330	0.004	4.868	0.074	0.107	0.001	0.833	1838	20	1749	15	-5
5.1	98	81	0.83	26.7	0.000000	0.00	3.146	0.041	0.108	0.001	0.318	0.004	4.721	0.079	0.108	0.001	0.780	1779	20	1761	19	-1
6.1	153	84	0.55	28.1	0.000045	0.08	4.671	0.058	0.080	0.001	0.214	0.003	2.347	0.038	0.080	0.001	0.761	1250	14	1186	21	-5
7.1	181	130	0.72	18.9	0.000184	0.02	8.210	0.103	0.064	0.001	0.122	0.002	1.056	0.022	0.063	0.001	0.606	741	9			
8.1	158	124	0.78	43.5	0.000000	0.00	3.125	0.050	0.108	0.001	0.320	0.005	4.758	0.085	0.108	0.001	0.900	1790	25	1763	14	-1
9.1	85	103	1.20	23.5	-	<0.01	3.116	0.043	0.110	0.001	0.321	0.004	4.877	0.083	0.110	0.001	0.806	1795	21	1802	18	0
10.1	248	206	0.83	67.5	0.000036	0.06	3.150	0.037	0.107	0.001	0.317	0.004	4.652	0.062	0.106	0.001	0.876	1777	18	1737	12	-2
11.1	101	46	0.45	27.6	0.000093	0.15	3.163	0.042	0.107	0.001	0.316	0.004	4.604	0.080	0.106	0.001	0.761	1769	20	1728	21	-2
12.1	94	68	0.73	9.6	0.000182	<0.01	8.377	0.121	0.063	0.001	0.119	0.002	1.037	0.026	0.063	0.001	0.578	728	10			
13.1	94	97	1.03	26.2	0.000091	0.14	3.092	0.041	0.111	0.001	0.323	0.004	4.900	0.086	0.110	0.001	0.768	1804	21	1800	20	0
14.1	100	93	0.93	28.1	0.000057	0.09	3.060	0.043	0.107	0.001	0.327	0.005	4.791	0.084	0.106	0.001	0.806	1821	22	1739	19	-5
15.1	237	270	1.14	67.0	0.000197	0.31	3.035	0.038	0.113	0.001	0.328	0.004	4.979	0.079	0.110	0.001	0.797	1831	20	1799	17	-2
16.1	338	306	0.91	91.9	0.000031	0.05	3.160	0.036	0.108	0.001	0.316	0.004	4.704	0.060	0.108	0.001	0.905	1772	18	1764	10	0
17.1	178	160	0.90	47.9	0.000042	0.07	3.182	0.039	0.107	0.001	0.314	0.004	4.615	0.066	0.107	0.001	0.852	1761	19	1742	14	-1
18.1	187	125	0.67	47.0	0.000039	0.06	3.414	0.041	0.103	0.001	0.293	0.004	4.153	0.060	0.103	0.001	0.846	1655	18	1677	14	1
19.1	194	154	0.79	50.1	0.000594	0.93	3.327	0.040	0.115	0.001	0.298	0.004	4.410	0.117	0.107	0.003	0.468	1680	18	1756	43	4
20.1	283	34	0.12	59.1	0.000037	0.06	4.111	0.048	0.089	0.001	0.243	0.003	2.973	0.043	0.089	0.001	0.802	1403	15	1398	17	0
21.1	100	58	0.58	18.3	0.000157	0.26	4.703	0.065	0.082	0.001	0.212	0.003	2.330	0.051	0.080	0.001	0.637	1240	16	1189	33	-4
22.1	150	73	0.49	41.4	0.000084	0.13	3.110	0.039	0.109	0.001	0.321	0.004	4.779	0.075	0.108	0.001	0.803	1795	20	1765	17	-2
23.1	99	72	0.73	26.8	0.000107	0.17	3.180	0.043	0.109	0.001	0.314	0.004	4.659	0.082	0.108	0.001	0.775	1760	21	1759	20	0
24.1	108	105	0.98	29.1	-	<0.01	3.183	0.043	0.108	0.001	0.314	0.004	4.699	0.078	0.108	0.001	0.810	1762	21	1773	18	1
25.1	157	239	1.52	42.7	-	<0.01	3.158	0.081	0.106	0.001	0.317	0.008	4.633	0.129	0.106	0.001	0.921	1775	40	1732	20	-2
26.1	113	93	0.83	30.6	0.000114	0.18	3.158	0.059	0.108	0.001	0.316	0.006	4.657	0.109	0.107	0.001	0.799	1770	29	1747	26	-1
27.1	100	87	0.87	26.9	0.000058	0.09	3.189	0.043	0.108	0.001	0.313	0.004	4.640	0.082	0.107	0.001	0.769	1757	21	1756	21	0
28.1	202	101	0.50	54.7	0.000000	0.00	3.170	0.056	0.109	0.001	0.315	0.006	4.737	0.090	0.109	0.001	0.926	1768	27	1781	13	1
29.1	202	131	0.65	54.0	0.000046	0.07	3.207	0.039	0.112	0.001	0.312	0.004	4.780	0.068	0.111	0.001	0.861	1749	19	1820	13	4
30.1	91	60	0.66	20.0	0.000226	0.37	3.922	0.055	0.094	0.001	0.254	0.004	3.187	0.074	0.091	0.002	0.605	1459	18	1447	35	-1
31.1	131	60	0.46	30.4	0.000141	0.23	3.689	0.049	0.100	0.001	0.270	0.004	3.670	0.065	0.098	0.001	0.744	1543	18	1595	22	3
32.1	270	121	0.45	49.0	0.000042	0.07	4.733	0.057	0.081	0.001	0.211	0.003	2.352	0.036	0.081	0.001	0.789	1235	13	1216	18	-2
33.1	129	99	0.76	34.7	0.000076	0.12	3.202	0.042	0.107	0.001	0.312	0.004	4.536	0.074	0.105	0.001	0.799	1750	20	1722	18	-2
34.1	77	52	0.67	21.8	-	<0.01	3.031	0.045	0.107	0.001	0.330	0.005	4.908	0.092	0.108	0.001	0.781	1840	24	1762	21	-4
35.1	82	46	0.56	18.6	0.000041	0.07	3.809	0.056	0.098	0.001	0.262	0.004	3.523	0.071	0.097	0.001	0.730	1502	20	1575	26	5
36.1	117	71	0.61	31.0	0.000068	0.11	3.249	0.045	0.108	0.001	0.307	0.004	4.550	0.093	0.107	0.002	0.671	1728	21	1755	28	2
37.1	182	80	0.44	18.6	0.000083	0.06	8.404	0.109	0.064	0.001	0.119	0.002	1.059	0.022	0.065	0.001	0.634	724	9			
38.1	298	156	0.52	81.4	0.000044	0.07	3.144	0.037	0.107	0.001	0.318	0.004	4.650	0.063	0.106	0.001	0.876	1779	18	1734	12	-3
39.1	161	122	0.75	42.1	0.000034	0.05	3.294	0.048	0.108	0.001	0.303	0.004	4.509	0.077	0.108	0.001	0.852	1708	22	1763	16	3
40.1	99	80	0.81	27.2	0.000069	0.11	3.111	0.043	0.108	0.001	0.321	0.004	4.747	0.084	0.107	0.001	0.783	1795	22	1753	20	-2
41.1	164	159	0.97	44.1	0.000047	0.07	3.189	0.048	0.107	0.001	0.313	0.005	4.578	0.083	0.106	0.001	0.827	1757	23	1731	19	-2
42.1	282	134	0.48	57.2	0.001177	1.87	4.225	0.052	0.117	0.001	0.232	0.003	3.218	0.088	0.101	0.002	0.458	1346	15	1633	45	18
43.1	107	84	0.78	27.7	0.000086	0.14	3.319	0.049	0.106	0.002	0.301	0.004	4.333	0.096	0.104	0.002	0.665	1696	22	1705	30	1
44.1	147	134	0.91	40.0	0.000027	0.04	3.163	0.041	0.107	0.002	0.316	0.004	4.644	0.094	0.107	0.002	0.650	1771	20	1742	28	-2
45.1	103	75	0.73	28.0	0.000092	0.14	3.150	0.044	0.108	0.001	0.317	0.004	4.646	0.084	0.106	0.001	0.781	1775	22	1737	21	-2
46.1	188	129	0.69	49.4	-	<0.01	3.259	0.041	0.106	0.001	0.307	0.004	4.502	0.067	0.106	0.001	0.843	1725	19	1739	15	1
47.1	298	73	0.24	64.3	0.000026	0.04	3.986	0.048	0.089	0.001	0.251	0.003	3.055	0.044	0.088	0.001	0.836	1443	15	1390	15	-4
48.1																						

Table 5. Summary of SHRIMP U-Pb zircon results for sample MS06-06

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Total Ratios				Radiogenic Ratios						Age (Ma)				% Disc	
							²³⁸ U/ ²⁰⁶ Pb		±	²⁰⁷ Pb/ ²⁰⁶ Pb		±	²⁰⁶ Pb/ ²³⁸ U		±	²⁰⁷ Pb/ ²³⁵ U		±	²⁰⁶ Pb/ ²³⁸ U			±
							±	±	±	±	±	±	±	±	±	±	±	±	±	±		±
1.1	77	149	1.9	22.3	0.000070	0.11	2.970	0.042	0.1147	0.0010	0.3363	0.0048	5.277	0.089	0.1138	0.0010	0.842	1869	23	1861	16	0
2.1	93	83	0.9	26.0	0.000073	0.11	3.074	0.043	0.1095	0.0010	0.3249	0.0046	4.858	0.087	0.1085	0.0012	0.789	1814	22	1774	20	-2
3.1	146	125	0.9	40.2	0.000003	0.00	3.120	0.040	0.1159	0.0008	0.3205	0.0041	5.119	0.075	0.1158	0.0008	0.869	1792	20	1893	13	5
4.1	87	144	1.7	25.9	0.000197	0.31	2.899	0.042	0.1121	0.0013	0.3438	0.0050	5.189	0.110	0.1094	0.0017	0.687	1905	24	1790	28	-6
5.1	156	99	0.6	29.5	0.000373	0.62	4.531	0.058	0.0871	0.0008	0.2193	0.0028	2.474	0.057	0.0818	0.0015	0.567	1278	15	1241	37	-3
6.1	124	109	0.9	33.8	0.000025	0.04	3.154	0.042	0.1074	0.0008	0.3169	0.0042	4.678	0.075	0.1071	0.0009	0.833	1775	21	1750	16	-1
7.1	264	130	0.5	79.6	0.000068	0.11	2.851	0.035	0.1114	0.0007	0.3504	0.0043	5.338	0.077	0.1105	0.0008	0.852	1937	20	1807	14	-7
8.1	275	239	0.9	56.0	0.000266	0.45	4.222	0.053	0.0835	0.0007	0.2358	0.0030	2.594	0.052	0.0798	0.0012	0.631	1365	16	1192	31	-15
9.1	492	205	0.4	62.0	0.002972	4.94	6.816	0.082	0.1235	0.0010	0.1395	0.0018	1.568	0.082	0.0816	0.0041	0.249	842	10	1235	99	32
10.1	211	158	0.7	57.2	0.000013	0.02	3.161	0.039	0.1095	0.0007	0.3163	0.0039	4.768	0.066	0.1093	0.0007	0.897	1772	19	1788	11	1
11.1	141	129	0.9	66.7	0.000183	0.24	1.819	0.024	0.2252	0.0012	0.5485	0.0072	16.875	0.240	0.2231	0.0013	0.917	2819	30	3003	9	6
12.1	417	250	0.6	95.1	0.000277	0.46	3.765	0.048	0.0898	0.0007	0.2644	0.0034	3.130	0.061	0.0859	0.0012	0.660	1512	17	1335	28	-13
14.1	467	716	1.5	65.4	0.001850	2.97	6.133	0.073	0.1224	0.0008	0.1582	0.0019	2.113	0.065	0.0969	0.0027	0.397	947	11	1564	53	39
15.1	432	434	1.0	75.8	0.000681	1.08	4.899	0.057	0.1111	0.0007	0.2019	0.0024	2.834	0.050	0.1018	0.0014	0.660	1186	13	1657	25	28
16.1	339	393	1.2	92.0	0.000385	0.60	3.166	0.040	0.1120	0.0008	0.3139	0.0040	4.621	0.084	0.1068	0.0014	0.695	1760	20	1745	24	-1
17.1	99	93	0.9	32.8	0.000515	0.81	2.597	0.038	0.1126	0.0012	0.3819	0.0057	5.562	0.151	0.1056	0.0024	0.545	2085	26	1725	42	-21
18.1	162	51	0.3	36.5	0.000961	1.56	3.810	0.051	0.1057	0.0010	0.2584	0.0036	3.290	0.101	0.0923	0.0025	0.448	1482	18	1474	52	-1
19.1	59	70	1.2	13.8	0.000358	0.56	3.672	0.059	0.1165	0.0014	0.2708	0.0044	4.168	0.104	0.1116	0.0021	0.646	1545	22	1826	34	15
20.1	273	141	0.5	66.8	0.000123	0.20	3.503	0.043	0.1027	0.0007	0.2849	0.0035	3.968	0.059	0.1010	0.0008	0.826	1616	18	1643	15	2
21.1	287	326	1.1	77.4	0.000725	1.12	3.184	0.038	0.1216	0.0007	0.3106	0.0037	4.786	0.082	0.1118	0.0014	0.702	1744	18	1828	22	5
22.1	93	64	0.7	25.5	-	<0.01	3.147	0.045	0.1093	0.0010	0.3181	0.0045	4.828	0.085	0.1101	0.0011	0.809	1780	22	1801	19	1
23.1	106	52	0.5	19.8	-	<0.01	4.595	0.066	0.0829	0.0010	0.2177	0.0031	2.494	0.047	0.0831	0.0010	0.759	1270	16	1272	24	0
24.1	130	53	0.4	37.2	-	<0.01	3.008	0.040	0.1133	0.0009	0.3326	0.0044	5.207	0.080	0.1135	0.0009	0.869	1851	21	1857	14	0
25.1	370	165	0.4	65.8	0.000686	1.14	4.830	0.057	0.0907	0.0006	0.2047	0.0025	2.287	0.049	0.0810	0.0014	0.562	1200	13	1222	35	2
26.1	215	117	0.5	22.1	0.000068	<0.01	8.375	0.109	0.0634	0.0008	0.1192	0.0016	1.015	0.021	0.0618	0.0010	0.634	727	9			
27.1	343	196	0.6	94.8	0.000004	0.01	3.105	0.037	0.1087	0.0005	0.3220	0.0038	4.825	0.063	0.1087	0.0006	0.917	1799	19	1778	9	-1
28.1	80	82	1.0	19.8	0.000239	0.38	3.454	0.052	0.1098	0.0012	0.2885	0.0044	4.236	0.083	0.1065	0.0013	0.776	1634	22	1740	23	6
29.1	104	85	0.8	35.1	0.000416	0.65	2.555	0.048	0.1115	0.0018	0.3889	0.0075	5.672	0.228	0.1058	0.0037	0.478	2118	35	1728	65	-23
30.1	156	134	0.9	37.9	0.000226	0.36	3.541	0.048	0.1089	0.0011	0.2814	0.0038	4.104	0.084	0.1058	0.0016	0.665	1598	19	1728	28	8
31.1	416	350	0.8	81.7	0.019087	29.01	4.370	0.065	0.3775	0.0068	0.1624	0.0059	2.757	0.710	0.1231	0.0314	0.142	970	33	2001	453	52
32.1	74	19	0.3	19.5	0.000000	0.00	3.255	0.049	0.1067	0.0014	0.3073	0.0046	4.522	0.090	0.1067	0.0014	0.758	1727	23	1744	24	1
33.1	165	152	0.9	38.4	0.001756	2.96	3.694	0.053	0.1015	0.0012	0.2627	0.0040	2.774	0.163	0.0766	0.0044	0.255	1504	20	1110	114	-35
34.1	171	172	1.0	51.6	0.000302	0.47	2.847	0.041	0.1099	0.0012	0.3496	0.0051	5.096	0.126	0.1057	0.0021	0.590	1933	24	1727	37	-12
35.1	563	406	0.7	125.9	0.000838	1.30	3.839	0.045	0.1228	0.0008	0.2571	0.0031	3.949	0.071	0.1114	0.0015	0.658	1475	16	1823	25	19
36.1	145	153	1.1	45.0	0.000187	0.29	2.762	0.045	0.1160	0.0016	0.3610	0.0060	5.649	0.155	0.1135	0.0025	0.602	1987	28	1856	40	-7
37.1	119	91	0.8	33.8	0.000076	0.12	3.035	0.044	0.1097	0.0009	0.3291	0.0048	4.931	0.085	0.1087	0.0010	0.839	1834	23	1778	17	-3
38.1	114	77	0.7	30.1	0.000932	1.49	3.258	0.056	0.1115	0.0022	0.3024	0.0054	4.113	0.228	0.0986	0.0052	0.324	1703	27	1599	98	-7
39.1	174	162	0.9	41.6	0.000618	0.99	3.603	0.051	0.1070	0.0011	0.2748	0.0039	3.731	0.095	0.0985	0.0021	0.561	1565	20	1595	39	2
41.1	305	324	1.1	89.1	0.000059	0.09	2.942	0.035	0.1160	0.0006	0.3396	0.0041	5.395	0.072	0.1152	0.0007	0.899	1885	20	1884	11	0
42.1	255	217	0.9	70.5	0.000032	0.05	3.107	0.038	0.1082	0.0006	0.3217	0.0039	4.781	0.066	0.1078	0.0007	0.885	1798	19	1762	12	-2
43.1	189	93	0.5	55.0	0.000134	0.21	2.957	0.039	0.1062	0.0009	0.3375	0.0045	4.856	0.082	0.1044	0.0011	0.788	1875	22	1703	19	-10
44.1	306	207	0.7	77.3	0.000597	0.95	3.399	0.043	0.1058	0.0008	0.2914	0.0037	3.922	0.085	0.0976	0.0017	0.593	1648	19	1579	32	-4
45.1	156	65	0.4	41.8	0.000095	0.15	3.208	0.042	0.1087	0.0008	0.3113	0.0041	4.612	0.072	0.1075	0.0009	0.842	1747	20	1757	15	1
46.1	160	130	0.8	43.5	0.000112	0.17	3.161	0.041	0.1135	0.0008	0.3158	0.0041	4.877	0.079	0.1120	0.0011	0.812	1769	20	1832	17	3
47.1	185	176	1.0	51.8	0.000032	0.05	3.064	0.039	0.1093	0.0008	0.3262	0.0042	4.898	0.073	0.1089	0.0008	0.860	1820	20	1781	14	-2
48.1	213	175	0.8	57.0	0.000024	0.04	3.207	0.040	0.1079	0.0007	0.3117	0.0039	4.624	0.067	0.1076	0.0008	0.867	1749	19	1759	13	1
49.1	274	145	0.5	59.0	0.000077	0.13	3.986	0.050	0.0908	0.0007	0.2505	0.0031	3.100	0.052	0.0897	0.0010	0.749	1441	16	1420	21	-1
50.1	198	227	1.1	56.7	0.000630	0.98	2.998	0.042	0.1177	0.0011	0.3303	0.0047	4.972	0.135								

Table 6. Summary of SHRIMP U-Pb zircon results for sample MS06-07

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Total Ratios				Radiogenic Ratios						ρ	Age (Ma)				% Disc
							²³⁸ U/ ²⁰⁶ Pb		²⁰⁷ Pb/ ²⁰⁶ Pb		²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²³⁵ U			²⁰⁶ Pb/ ²³⁸ U		²⁰⁷ Pb/ ²⁰⁶ Pb				
							±	±	±	±	±	±	±	±	±	±		±	±			
1.1	159	99	0.62	43.3	0.000000	0.00	3.142	0.039	0.1063	0.0008	0.3182	0.0039	4.665	0.067	0.1063	0.0008	0.855	1781	19	1737	14	-3
2.1	170	165	0.97	45.2	0.000058	0.09	3.223	0.039	0.1090	0.0008	0.3100	0.0038	4.624	0.067	0.1082	0.0009	0.838	1741	19	1769	14	2
3.1	212	171	0.81	56.9	0.000009	0.01	3.196	0.038	0.1087	0.0007	0.3128	0.0037	4.681	0.064	0.1085	0.0007	0.868	1755	18	1775	12	1
4.1	175	139	0.79	46.3	0.000026	0.04	3.244	0.039	0.1074	0.0008	0.3081	0.0037	4.549	0.065	0.1071	0.0008	0.853	1731	18	1751	14	1
5.1	279	289	1.03	66.6	0.000204	0.32	3.604	0.041	0.1100	0.0007	0.2766	0.0031	4.088	0.060	0.1072	0.0010	0.765	1574	16	1752	17	10
6.1	114	81	0.71	29.6	-	<0.01	3.306	0.043	0.1073	0.0010	0.3025	0.0039	4.486	0.071	0.1075	0.0010	0.824	1704	19	1758	16	3
7.1	114	90	0.79	31.0	0.000035	0.05	3.165	0.043	0.1077	0.0010	0.3158	0.0042	4.669	0.077	0.1072	0.0010	0.813	1769	21	1753	18	-1
8.1	37	32	0.86	10.0	0.000157	0.25	3.175	0.058	0.1081	0.0017	0.3142	0.0057	4.589	0.121	0.1059	0.0020	0.692	1761	28	1730	35	-2
9.1	223	239	1.07	60.2	0.000021	0.03	3.182	0.038	0.1091	0.0007	0.3142	0.0037	4.712	0.064	0.1088	0.0007	0.871	1761	18	1779	12	1
10.1	216	186	0.86	58.3	-	<0.01	3.189	0.037	0.1082	0.0007	0.3136	0.0037	4.682	0.062	0.1083	0.0007	0.880	1758	18	1771	12	1
11.1	192	137	0.71	52.3	0.000030	0.05	3.158	0.040	0.1080	0.0007	0.3165	0.0040	4.694	0.068	0.1076	0.0008	0.872	1773	19	1759	13	-1
12.1	298	405	1.36	75.5	0.000039	0.06	3.391	0.040	0.1085	0.0006	0.2948	0.0034	4.388	0.058	0.1080	0.0006	0.890	1665	17	1765	11	6
13.1	246	307	1.25	58.4	0.000159	0.25	3.624	0.042	0.1085	0.0007	0.2753	0.0032	4.037	0.057	0.1064	0.0008	0.825	1567	16	1738	15	10
14.1	348	318	0.92	90.5	0.000150	0.23	3.305	0.037	0.1107	0.0006	0.3018	0.0034	4.523	0.058	0.1087	0.0007	0.863	1700	17	1777	12	4
15.1	145	128	0.88	38.7	0.000031	0.05	3.221	0.040	0.1082	0.0008	0.3103	0.0039	4.610	0.069	0.1077	0.0009	0.838	1742	19	1762	15	1
16.1	341	339	0.99	71.8	0.000220	0.34	4.079	0.046	0.1094	0.0006	0.2443	0.0027	3.586	0.049	0.1065	0.0008	0.817	1409	14	1740	14	19
17.1	29	10	0.34	6.1	-	<0.01	4.029	0.080	0.0935	0.0032	0.2491	0.0050	3.313	0.139	0.0965	0.0036	0.475	1434	26	1557	69	8
18.1	118	108	0.92	29.1	0.000046	0.07	3.478	0.045	0.1080	0.0010	0.2873	0.0037	4.252	0.068	0.1073	0.0010	0.808	1628	19	1754	17	7
19.1	118	60	0.51	31.0	-	<0.01	3.274	0.042	0.1088	0.0010	0.3055	0.0040	4.591	0.073	0.1090	0.0010	0.815	1718	20	1783	17	4
20.1	149	164	1.10	39.6	0.000063	0.10	3.233	0.040	0.1084	0.0008	0.3090	0.0038	4.583	0.069	0.1076	0.0009	0.830	1736	19	1759	15	1
21.1	212	267	1.26	52.9	0.000126	0.20	3.437	0.040	0.1097	0.0007	0.2904	0.0034	4.322	0.061	0.1079	0.0009	0.824	1644	17	1765	15	7
22.1	77	68	0.88	20.7	-	<0.01	3.201	0.046	0.1075	0.0012	0.3129	0.0045	4.692	0.090	0.1088	0.0014	0.756	1755	22	1779	23	1
23.1	123	62	0.51	31.7	0.000008	0.01	3.322	0.043	0.1051	0.0009	0.3010	0.0039	4.358	0.068	0.1050	0.0009	0.824	1696	19	1714	16	1
24.1	52	38	0.72	5.3	0.000597	1.05	8.452	0.155	0.0703	0.0025	0.1171	0.0022	0.995	0.077	0.0616	0.0046	0.246	714	13			
25.1	306	332	1.08	68.5	0.000098	0.15	3.834	0.043	0.1085	0.0006	0.2604	0.0029	3.847	0.051	0.1072	0.0007	0.859	1492	15	1752	12	15
26.1	253	209	0.82	66.6	0.000112	0.17	3.268	0.038	0.1103	0.0011	0.3054	0.0035	4.583	0.072	0.1088	0.0012	0.736	1718	17	1780	19	3
27.1	160	134	0.84	42.5	0.000053	0.08	3.241	0.040	0.1079	0.0008	0.3083	0.0038	4.555	0.067	0.1072	0.0009	0.834	1732	19	1752	15	1
29.1	84	53	0.63	20.5	0.000065	0.10	3.501	0.050	0.1079	0.0012	0.2853	0.0041	4.212	0.078	0.1071	0.0013	0.768	1618	20	1750	22	8
30.1	225	174	0.78	61.7	0.000086	0.14	3.127	0.037	0.1073	0.0007	0.3194	0.0038	4.675	0.066	0.1062	0.0008	0.845	1787	18	1735	14	-3
31.1	285	310	1.09	75.6	0.000050	0.08	3.234	0.037	0.1079	0.0006	0.3090	0.0035	4.567	0.059	0.1072	0.0007	0.879	1736	17	1752	11	1
32.1	165	180	1.09	43.5	0.000110	0.17	3.259	0.040	0.1075	0.0008	0.3063	0.0038	4.479	0.069	0.1060	0.0010	0.804	1723	19	1732	17	1
33.1	246	204	0.83	65.8	0.000040	0.06	3.209	0.037	0.1077	0.0007	0.3114	0.0036	4.599	0.062	0.1071	0.0007	0.865	1748	18	1751	12	0
34.1	172	107	0.62	43.4	0.000048	0.08	3.403	0.042	0.1051	0.0008	0.2936	0.0036	4.230	0.063	0.1045	0.0009	0.833	1660	18	1705	15	3
35.1	328	242	0.74	84.5	0.000042	0.07	3.334	0.037	0.1084	0.0006	0.2997	0.0034	4.456	0.056	0.1078	0.0006	0.888	1690	17	1763	11	4
36.1	200	193	0.96	51.5	0.000027	0.04	3.346	0.040	0.1080	0.0008	0.2987	0.0036	4.434	0.064	0.1076	0.0009	0.832	1685	18	1760	15	4
37.1	253	261	1.03	62.7	0.000126	0.22	3.461	0.042	0.1086	0.0007	0.2883	0.0035	4.241	0.062	0.1067	0.0009	0.836	1633	18	1744	15	6
38.1	159	150	0.94	42.2	0.000104	0.16	3.247	0.041	0.1087	0.0008	0.3075	0.0039	4.550	0.071	0.1073	0.0010	0.803	1728	19	1754	17	1
39.1	140	121	0.87	36.8	-	<0.01	3.260	0.042	0.1077	0.0009	0.3068	0.0039	4.561	0.070	0.1078	0.0009	0.836	1725	19	1763	15	2
40.1	125	96	0.76	33.1	0.000068	0.11	3.261	0.043	0.1073	0.0009	0.3063	0.0040	4.495	0.073	0.1064	0.0010	0.803	1723	20	1739	18	1
41.1	146	91	0.63	31.9	0.000250	0.40	3.917	0.084	0.1042	0.0010	0.2543	0.0055	3.534	0.093	0.1008	0.0015	0.821	1460	28	1639	28	11
42.1	182	156	0.86	47.8	0.000056	0.09	3.264	0.040	0.1088	0.0008	0.3061	0.0037	4.561	0.069	0.1081	0.0010	0.806	1722	18	1767	16	3
43.1	196	178	0.91	50.9	0.000025	0.04	3.309	0.040	0.1082	0.0008	0.3021	0.0036	4.494	0.063	0.1079	0.0008	0.855	1702	18	1764	13	4
44.1	140	117	0.83	38.1	-	<0.01	3.155	0.041	0.1087	0.0009	0.3170	0.0041	4.762	0.074	0.1089	0.0009	0.835	1775	20	1782	16	0
45.1	84	64	0.76	22.8	0.000140	0.22	3.174	0.046	0.1112	0.0012	0.3144	0.0045	4.737	0.092	0.1093	0.0014	0.741	1762	22	1787	24	1
46.1	146	123	0.85	37.6	0.000051	0.08	3.337	0.060	0.1104	0.0010	0.2995	0.0054	4.529	0.091	0.1097	0.0010	0.890	1689	27	1794	17	6
47.1	100	77	0.77	27.1	0.000110	0.17	3.156	0.044	0.1089	0.0011	0.3164	0.0044	4.683	0.085	0.1074	0.0013	0.767	1772	22	1755	21	-1
48.1	137	144	1.05	37.4	-	<0.01	3.146	0.041	0.1084	0.0010	0.3180	0.0041	4.764	0.077	0.1086	0.0010	0.801	1780	20	1777	18	0
49.1	94	77	0.82	25.0	0.000052	0.08	3.229	0.047	0.1083	0.0012	0.3094	0.0045	4.590	0.084	0.1076	0.0012	0.785	1738	22	1759	21	1
50.1	168	194	1.16	44.2	0																	

5.2. GEOCRONOLOGIA DOS CARBONATOS DE CAPA DA FORMAÇÃO MIRASSOL D'OESTE, CRÁTON AMAZÔNICO

5.2.1. Introdução

Uma ocorrência singular do par diamictito – carbonato de capa foi descrita na borda sudeste do Cráton Amazônico, próximo à cidade de Mirassol d'Oeste, MT. Neste local, os carbonatos estão depositados diretamente sobre os diamictitos da Formação Puga e estruturas preservadas indicam que o diamictito estava ainda inconsolidado quando os carbonatos foram sedimentados (Nogueira et al., 2003), demonstrando a rápida transição do estado *Icehouse* – *Greenhouse*. Considerando-se a rapidez do processo, é possível assumir que a idade determinada nos carbonatos de capa pode definir a época do final da glaciação.

Embora a metodologia isotópica Pb-Pb tenha sido criticada por alguns autores devido à abertura do sistema U-Pb por processos pós-deposicionais (Macdonald et al., 2010), outros consideram como uma possível alternativa de datação quando outros métodos não podem ser aplicados (Condon & Bowring, in press). De qualquer forma, a datação Pb-Pb em carbonatos de capa pode ser a única alternativa para obtenção de idade absoluta nesses litotipos quando não há tufos vulcânicos intercalados neles ou nos diamictitos sotopostos.

É importante lembrar que um dos maiores problemas em todas as hipóteses que tentam explicar as mudanças climáticas do Neoproterozoico é a falta de idades acuradas para as rochas que registram estes eventos. Embora os erros usualmente determinados nas idades Pb-Pb sejam altos, decorrentes da pequena quantidade de ^{235}U (o pai radioativo do ^{207}Pb radiogênico), as idades obtidas através deste método permitem discriminar entre os diferentes eventos glaciais desta era. Aqui será apresentada uma idade isocrônica Pb-Pb determinada em carbonatos da Formação Mirassol d'Oeste, que capeiam os diamictitos da Formação Puga.

5.2.2. *Geologia Regional e descrição das amostras*

A área de estudo situa-se na Pedreira Terconi, próxima à cidade de Mirassol d'Oeste, MT, na borda sudeste do Cráton Amazônico (Fig. 5.1A, B, C). Os carbonatos estudados pertencem à Formação Mirassol d'Oeste (Nogueira et al., 2003), base do Grupo Araras, e estão depositados diretamente sobre os diamictitos da Formação Puga (Fig. 5.1D).

Na área de estudo, os carbonatos ocorrem em uma frente de lavra abandonada com cerca de 45 metros de espessura. Os primeiros 20 metros são representados por dolomitos de capa (Formação Mirassol d'Oeste), enquanto que a sequência superior é constituída por carbonatos calcínicos (Formação Guia). A coluna estratigráfica com a localização das amostras analisadas é apresentada na Figura 5.2. As amostras foram coletadas e gentilmente cedidas para este estudo pelo Prof. Carlos Alvarenga (UnB).

Os carbonatos desta pedreira (Fig. 5.3A, B, C) têm sido intensivamente estudados sob vários aspectos, incluindo: sedimentologia, estratigrafia, geoquímica elementar e isotópica, biomarcadores e paleomagnetismo (Alvarenga et al., 2004, 2008; Elie et al., 2007; Font et al., 2005, 2006, 2010; Nogueira et al., 2003, 2007, Trindade et al., 2003). Estes estudos se devem à excelente preservação de estruturas típicas de carbonatos de capa, tais como: tubos de escape de gases (*tubestones*; Fig. 5.3D), marcas de ondas gigantes (*mega ripples* ou *pseudo-teepe structures*; Fig. 5.3E) e cristais de pseudomorfo de aragonita (Fig. 5.3F). Evidências geoquímicas adicionais de que os carbonatos foram depositados imediatamente após a glaciação são dadas pelos valores de $\delta^{13}\text{C}$ negativos, entre -9 e -3 ‰ (Alvarenga et al., 2004, 2008; Font et al., 2006, Nogueira et al., 2003, 2007).

Apesar da falta de dados geocronológicos, uma idade Marinoana (Criogeniano Superior) tem sido atribuída às rochas das formações Mirassol d'Oeste e Puga a partir da correlação com perfis isotópicos de outras sucessões bem datadas (e.g. Formação Maieberg, Namíbia).

Figura 5.1. Localização do Cráton Amazônico (A) e da Faixa Paraguai (B). Mapa geológico da área estudada (C) com a localização da Pedreira Terconi (TR) e coluna estratigráfica regional (D). Mapa extraído de Font et al. (2006).

Figura 5.2. Seção estratigráfica da Pedreira Terconi, mostrando a posição das amostras analisadas neste trabalho, as quais foram cedidas pelo Prof. Carlos J. Alvarenga, UnB. Seção extraída de Trindade et al. (2003).

Figura 5.3. Rochas aflorantes na Pedreira Terconi. (A) Vista geral da parte superior do afloramento; (B) Detalhe do contato entre o diamictito e o carbonato de capa; (C) Presença de óleo em porosidade secundária nos calcários da Formação Guia; (D) Tubos de escape de gases nos dolomitos; (E) Mega-ripples no topo do pacote dos dolomitos; (F) Pseudomorfos de cristais de aragonita nos calcários (foto de Elie et al., 2007). Para visualizar a posição das amostras ao longo da seção, ver Fig. 5.2.

5.2.3. Materiais e Métodos

Foram analisadas 12 amostras de carbonato, sendo 7 de dolomitos e 5 de calcáreos (Fig. 5.2). Todos os carbonatos estavam bem preservados e porções que apresentavam veios ou aparência de recristalização foram descartadas. As amostras foram quebradas em fragmentos menores que 3 mm, lavadas, secadas e selecionadas sob lupa binocular.

Os procedimentos químico-analíticos seguiram aqueles descritos por Babinski (1993). Cerca de 500 mg de amostra foram parcialmente dissolvidos com 2 mL de HBr 0,7 N por cerca de 2 horas. A solução foi descartada (L1), a amostra foi lavada com H₂O e seca. A amostra restante foi novamente atacada com 10 mL de HBr 0,7 N, tendo reagido por cerca de 12 horas. A solução foi transferida para um Savillex[®], evaporada e o resíduo foi dissolvido em 1 mL de HBr. Esta solução foi submetida à purificação através da técnica de troca iônica, onde cerca de 35 µL de resina aniônica AG1-X8, 200-400 mesh, forma Cl⁻, pré-tratada, foram adicionados em uma coluna de polietileno. A resina foi lavada, sucessivamente, com 1,0 mL de HCl 6N, 300 µL de água Milli-Q[®] e, posteriormente, condicionada com 300 µL de HBr 0,7 N. A solução de amostra foi adicionada e a resina foi lavada com 300 µL de HBr 0,7 N por três vezes consecutivas. O Pb foi eluído com 1,0 mL de HCl 6N. A amostra foi purificada duas vezes para garantir a pureza do Pb eluído. Para isto a solução obtida foi evaporada e novamente convertida para 300 µL de HBr 0,7 N e repassada na coluna seguindo as mesmas etapas descritas acima. À solução final foram adicionados 5 µL de H₃PO₄ 0,25M e evaporada à secura.

As composições isotópicas foram determinadas por Espectrometria de Massa Termo-iônica (TIMS) usando o espectrômetro de Massa Finnigan MAT, equipado com 8 coletores tipo Faraday, usando detecção simultânea, em modo estático. Os dados foram corrigidos para o fracionamento isotópico de 0,09 ‰ u.m.a.⁻¹, com base nas mais de 200 análises obtidas no padrão de Pb comum SRM 981. Os brancos analíticos de Pb foram inferiores a 50 pg. A idade Pb-Pb foi determinada através do Programa *Isoplot* (Ludwig, 2001), assumindo os erros de 0,072% e 0,078% para as razões ²⁰⁶Pb/²⁰⁴Pb e ²⁰⁷Pb/²⁰⁴Pb, respectivamente, que representam os erros (1 sigma) determinados no padrão de Pb comum SRM 981.

5.2.4. Resultados

Das 12 amostras de carbonatos analisadas, 9 foram preparadas em duplicata, resultando em 21 análises. Os dados analíticos obtidos são apresentados na Tabela 5.1. As composições isotópicas determinadas nos carbonatos mostraram grande variação, com razões $^{206}\text{Pb}/^{204}\text{Pb}$ entre 18,505 e 65,179, $^{207}\text{Pb}/^{204}\text{Pb}$ de 15,698 a 18,550 e $^{208}\text{Pb}/^{204}\text{Pb}$ entre 38,175 e 52,616. As razões mais radiogênicas foram medidas nos calcários calcíticos, onde pseudomorfos de cristais de aragonita são encontrados, e as razões menos radiogênicas foram observadas na fácies dolomítica (Fig. 5.2). Esta ampla variação nas razões permitiu a obtenção de uma isócrona Pb-Pb de 633 ± 25 Ma (1 sigma), com MSWD de 3.3 (Fig. 5.4). Em função da boa preservação das rochas analisadas esta idade é interpretada como a idade de deposição dos carbonatos de capa da Formação Mirassol d'Oeste.

Tabela 5.1. Dados isotópicos de Pb obtidos nos carbonatos da Formação Mirassol d'Oeste, Cráton Amazônico.

Amostra	$^{206}\text{Pb}/^{204}\text{Pb}$	erro (%)	$^{207}\text{Pb}/^{204}\text{Pb}$	erro (%)	$^{208}\text{Pb}/^{204}\text{Pb}$	erro (%)
Mira-1a	18,541	0,056	15,745	0,053	38,336	0,054
Mira-1b	18,505	0,017	15,698	0,018	38,175	0,018
Mira-2	18,463	0,029	15,707	0,028	38,209	0,029
Mira-3a	20,644	0,107	15,831	0,109	38,829	0,109
Mira-3b	20,734	0,050	15,833	0,048	38,849	0,049
Mira-4	19,115	0,014	15,725	0,013	38,537	0,014
Mira-5a	22,215	0,058	15,892	0,058	38,698	0,059
Mira-5b	23,126	0,023	15,985	0,023	38,854	0,023
Mira-6a	26,494	0,025	16,166	0,024	38,923	0,025
Mira-6b	26,094	0,059	16,165	0,059	38,923	0,059
Mira-8a	37,002	0,019	16,827	0,017	42,943	0,018
Mira-8b	35,750	0,023	16,786	0,023	42,815	0,023
Mira-9a	65,179	0,020	18,550	0,019	44,058	0,019
Mira-9b	65,134	0,027	18,536	0,027	43,776	0,027
Mira-10	24,502	0,011	16,089	0,011	42,070	0,011
Mira-11a	30,509	0,046	16,455	0,045	42,589	0,047
Mira-11b	36,124	0,038	16,801	0,038	42,503	0,038
Mira-13a	41,777	0,013	17,108	0,012	50,656	0,011
Mira-13b	45,358	0,024	17,299	0,023	52,616	0,024
Mira-15a	23,094	0,048	15,984	0,048	38,261	0,050
Mira-15b	23,403	0,018	15,987	0,017	38,209	0,019

Figura 5.4. Isócrona Pb-Pb dos carbonatos de capa da Formação Mirassol d'Oeste, MT. Quadrados preenchidos representam dados obtidos em amostras da fácies calcítica e quadrados vazados em amostras da fácies dolomítica.

5.2.5. Discussão e Conclusão

A idade isocrônica Pb-Pb de 633 ± 25 Ma, determinada nos carbonatos de capa da Formação Mirassol d'Oeste, demonstra que a metodologia isotópica Pb-Pb permite obter idades com boa precisão (5%) neste tipo de rocha. Adicionalmente, a idade obtida indica que o par diamictito-carbonato de capa do sudoeste do Cráton Amazônico foi depositado durante no final do evento glacial do Criogeniano Superior, comumente denominado de Glaciação Marinoana.

A idade de 633 Ma, apesar do erro, vem contribuir para o escasso banco de dados geocronológicos disponível para sucessões depositadas durante a Glaciação Marinoana. As idades U-Pb em zircão de 636 Ma obtidas em tufos vulcânicos intercalados nos

diamictitos da Namíbia e China (Hoffmann et al., 2004; Zhang et al., 2008), associadas à idade obtida neste trabalho, sugerem que este evento glacial pode ser tido abrangência global, como previamente sugerido por Hoffman (2009). Contudo, um maior número de dados geocronológicos de boa qualidade devem ser obtidos em sucessões glaciais ou em carbonatos de capa para confirmar ou não esta evidência.

CAPÍTULO 6

CONCLUSÕES

As pesquisas realizadas e resumidas neste trabalho permitem concluir que:

Dois eventos glaciais criogenianos estão registrados nas rochas do Brasil Central. Um evento glacial mais antigo do Criogeniano Médio, conhecido como Sturtiano, é representado pelas rochas depositadas sobre o Cráton do São Francisco e na Faixa Araçuaí. Um evento mais jovem do Criogeniano Superior, conhecido como Marinoano, foi identificado em rochas depositadas sobre o Cráton Amazônico e na Faixa Paraguai.

O evento glacial do Criogeniano Médio é representado pelos diamictitos da Formação Bebedouro, depositados sobre o Cráton do São Francisco, na Bahia, e são interpretados como cronocorrelatos dos diamictitos das formações Jequitaí e Carrancas. Os zircões recuperados dos diamictitos forneceram idades U-Pb SHRIMP entre 874 e 3260 Ma ($n = 92$). O zircão mais jovem datado tem idade U-Pb de 874 ± 9 Ma e estabelece a idade máxima de deposição dos diamictitos da Formação Bebedouro.

Registro deste evento glacial também ocorre em rochas da Faixa Araçuaí, onde os diamictitos constituem as formações Serra do Catuni e Chapada Acauã Inferior, sobrepostos aos arenitos da Formação Duas Barras, unidades basais do Grupo Macaúbas. As idades U-Pb nos zircões detríticos ($n = 131$) destas 3 unidades litoestratigráficas variaram entre 900 Ma e 2750 Ma. O zircão mais jovem, datado em 900 ± 21 Ma, proveniente dos arenitos da Formação Duas Barras, pré-glacial, define a idade máxima de deposição dos diamictitos do Grupo Macaúbas. A similaridade entre as idades máximas de

sedimentação obtidas nos diamictitos depositados sobre o cráton (Formação Bebedouro) e na bacia precursora da Faixa Araçuaí reforçam a correlação entre estas unidades.

Na região cratônica, os diamictitos são capeados por carbonatos capa pertencentes à base da Formação Sete Lagoas, Grupo Bambuí, que forneceram uma idade isocrônica Pb-Pb de 740 ± 22 Ma, interpretada como a idade de deposição dos carbonatos.

As idades dos zircões detríticos determinados nos diamictitos depositados sobre o cráton e na Faixa Araçuaí, associadas à idade Pb-Pb determinada no carbonato de capa da base da Formação Sete Lagoas, permitem propor que estas rochas representem o registro da Glaciação do Criogeniano Médio (Evento Sturtiano) no Brasil Central. Adicionalmente, zircões detríticos provenientes da Formação Santo Antônio do Bonito, base do Grupo Vazante, Faixa Brasília, indicam idades máximas de deposição de 900-930 Ma (Rodrigues et al., 2011), similares aos dados obtidos neste trabalho. Estas idades determinadas na Faixa Brasília sugerem a extensão do evento glacial que ocorre sobre o cráton e na Faixa Araçuaí, para os depósitos da margem ocidental do Cráton do São Francisco.

O evento glacial do Criogeniano Superior (Marinoano) é representado pelos diamictitos da Formação Puga, que ocorrem tanto no Cráton Amazônico quanto na Faixa Paraguai. A matriz dos diamictitos da Faixa Paraguai Sul forneceram cristais de zircão detríticos com idades U-Pb variando entre 706 e 1990 Ma ($n = 360$). O zircão mais jovem datado tem idade de 706 ± 9 Ma e estabelece a idade máxima de deposição dos diamictitos na Faixa Paraguai Sul.

Carbonatos de capa da Formação Mirassol d'Oeste, depositados diretamente sobre os diamictitos da Formação Puga, no Cráton Amazônico, foram datados pela metodologia isotópica Pb-Pb e forneceram uma idade isocrônica de 633 ± 25 Ma, a qual é interpretada como a época de sedimentação destes carbonatos.

As idades U-Pb obtidas em zircões detríticos da Formação Puga na Faixa Paraguai Sul, associados à idade isocrônica Pb-Pb determinada nos carbonatos de capa da Formação Mirassol d'Oeste, permitem sugerir que este par diamictito - carbonato de capa foi

depositado durante o evento glacial do Criogeniano Superior, comumente denominado de Glaciação Marinoana, ocorrida há ca. 636 Ma.

Apesar dos erros elevados determinados nas idades Pb-Pb dos carbonatos de capa, sua associação com os dados geocronológicos U-Pb em zircões detríticos obtidos neste trabalho sugere que as Glaciações do Criogeniano Médio (Sturtiana) e do Criogeniano Superior (Marinoana), no Brasil Central, ocorreram nas épocas sugeridas recentemente na literatura (há ca. 720 Ma e 636 Ma, respectivamente; Hoffman & Li, 2009). É possível que o amplo intervalo de idades proposto para a Glaciação Sturtiana, variando entre 670 e 740 Ma, seja reflexo do uso de idades máximas e mínimas de deposição como limitantes do período glacial, o que não seria correto. Se apenas as idades U-Pb determinadas em níveis vulcânicos intercalados nas rochas glaciogênicas fossem consideradas, um intervalo bem mais curto seria definido. De acordo com os poucos dados disponíveis na literatura, obtidos em sucessões sin-glaciais, um período de 712 a 716 Ma (Bowring et al., 2007; MacDonald et al., 2010) poderia ser selecionado para a Glaciação Sturtiana. Com relação aos diamictitos Marinoanos (Criogeniano Superior), apesar das poucas idades disponíveis, um intervalo bastante restrito e próximo à idade de 636 Ma parece ser aceito como a melhor época para este evento. Contudo, o limitado conjunto de dados geocronológicos dificulta a definição destes intervalos como globais. Em suma, é óbvia a necessidade de um maior número de dados radiométricos em sucessões glaciogênicas, além de relações estratigráficas claras e melhor definidas, para promover um melhor entendimento sobre o número, a duração e a sincronia (ou não) das glaciações neoproterozoicas, principalmente as criogenianas.

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