First precise U-Pb ages of mafic dykes from the São Francisco Craton: initial barcoding of the São Francisco-Congo Craton

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Contents

Cover Picture: My co-supervisor Elson Paiva de Oliveira resting on a Uaua dyke

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EDUARDO MIRANDA DA SILVEIRA

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Abstract:

The São Francisco Craton (SFC) is an Archaean craton that hosts a significant number of mafic intrusions. Previous attempts to date the igneous emplacement of these rocks are limited to a few dates of poor precision or comprise isotopic techniques known to be sensitive to alteration and low-grade metamorphism. Here, dykes from the Uauá Block and the Curaçá belt, in the northeastern portion of the craton, as well as from the Chapada Diamantina region in the central part, were dated using U-Pb on baddeleyite (ID-TIMS) and zircon (SIMS). The older ages of 2726.2 ± 3.2 and 2623.8 ± 7.0 Ma date dykes within the Uauá Block of noritic and tholeiitic compositions, respectively, and agree within error to previously reported, but less precise, Sm-Nd ages. The dykes from Curaçá and Chapada Diamantina yielded similar baddeleyite U-Pb ages of 1506.7 ± 6.9 Ma and 1501.0 ± 9.1 Ma, respectively, with converging trends indicating a possible magmatic centre located near the western margin of the São Francisco Craton. The new 1506.7 ± 6.9 Ma age for the Curaçá dyke is significantly older than previous published age estimates for this swarm. Current geochronological and palaeomagnetic data suggest a coherent São Francisco - Congo block from at least the late Mesoproterozoic until the opening of the South Atlantic (ca. 130 Ma ago). The discovery of ca. 1500 Ma intrusions in the São Francisco Craton is a major step forward to a more complete barcode record for the SFC-Congo Craton, which can be used in future palaeoreconstructions.

Keywords: São Francisco Craton, Congo Craton, mafic intrusions, U-Pb baddeleyite, geochronology

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De första exakta U-Pb dateringarna av diabasgångar från São Francisco Craton (SFC), med implikationer för palaeorekonstruktioner

EDUARDO MIRANDA DA SILVEIRA

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Sammanfattning: São Francisco kratonen i Sydamerika är en mycket gammal landmassa som bär spår av upprepade extensionella händelser då magmor från manteln transporterades uppåt och kristalliserade som diabas. Tidigare försök att åldersbestämma dessa gångbergarter är få och begränsade till radiometriska metoder som är känsliga för metamorfos och vittring. I denna studie har diabaser från Uauá blocket och Curaçá bältet i nordöstra delen av kratonen, samt från Chapada Diamantina regionen i centrala delen, daterats med U-Pb analys av zirkon och baddeleyit. Åldrar på 2726.2 ± 3.2 and 2623.8 ± 7.0 Ma (Ma = miljoner år) erhölls för diabasgångar från Uauá blocket med noritisk respektive tholeitisk sammansättning. Dessa resultat överensstämmer med tidigare publicerade, men mindre exakta, Sm-Nd åldrar. Diabasgångar från Curaçá and Chapada Diamantina daterades med hjälp av mineralet baddeleyit till 1506.7 ± 6.9 Ma respektive 1501.0 ± 9.1 Ma. Dessa likåldriga gångsvärmar har olika riktning som konvergerar mot ett centra i norra gränsen av São Francisco Craton. Detta kan tolkas som att magmorna till gångsvärmarna matades från en gemensam magmakälla, möjligen en s.k. mantelplym. Den nya åldern för Curaçá svärmen är betydligt äldre än vad tidigare dateringsresultat indikerat. Åldersdateringar och resultat av palaeomagnetiska studier indikerar att São Francisco och Kongo kratonerna satt ihop i en gemensam landmassa från åtminstone 1500 Ma tills Atlanten öppnades för ca. 130 Ma sedan. Förekomsten av ca. 1500 Ma magmatism i São Francisco kratonen kommer att hjälpa oss att finna andra kratoner som kan ha suttit ihop med denna landmassa under Mesoproterozoikum.

Nyckelord: São Francisco kraton, Congo kraton, diabas, U-Pb, baddeleyit, geokronologi

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1 Introduction

Earth's approximately 35 main pieces of Archaean crust are today dispersed around the globe, but there are reasons to believe that all these pieces were originally amalgamated into a single supercontinent (Superia), or into a few (Superia, Valbara and Slave) large continents, 'supercratons' (Bleeker, 2003, 2004). A key question in our planet's history concerns the reconstruction of continents and, in particular, during supercontinents times. Whereas our latest supercontinent Pangea was relatively easy to reconstruct, i.e. by backtracking continents from their present positions using the sea-floor spreading record, combined with robust well-dated palaeomagnetism, the configuration of pre-Pangean supercontinents has proved much more challenging to resolve. A recent approach to obtain robust reconstructions focuses on obtaining precise U-Pb ages on intrusive and extrusive rocks of Large Igneous Provinces (LIPs), and especially their extensive dyke swarms from all cratons (e.g., Bleeker and Ernst, 2006; Ernst and Bleeker 2010; www.supercontinent.org). Specifically, cratons that share a number of coeval LIP events can be argued to have been "nearest neighbours" whereas the lack of such matches suggests they were far away from each other during this time period. Also the neighbouring cratons can be oriented with respect to each other by aligning the trends of coeval dykes on different cratons into primary radiating or linear dyke swarm patterns. Finally with widespread precise U-Pb dating the potential of robust palaeomagnetic constraints can be realized.

It is generally assumed that South America and Africa, especially the Congo Craton (CC), were amalgamated at about 2.05 Ga until the opening of the South Atlantic at 130 Ma (D'Agrella Filho et al., 1996; Feybesse et al., 1998; Pedrosa-Soares et al., 2001; Janasi, et al., 2011; Deckart, 1998; Correa-Gomes & Oliveira, 2000), yet reliable geochronological evidences are sparse. Therefore any precise ages obtained on potential LIP units from the SFC (or the CC) during this time interval would therefore contribute to building the magmatic barcode for the combined SFC-CC craton. Prior to this study there were few precise and reliable ages for these mafic intrusions within the SFC and few for its counterpart in the African side (e.g., Ernst et al., 2008; Oliveira et al., 2010). This study aims to obtain the first precise emplacement ages for key swarms in the Uauá block, Curaçá belt and Chapada Diamantina regions, using U-Pb geochronology on baddeleyite (ZrO_2) and zircon ($ZrSiO_4$) to provide the first step towards a complete barcode record for SFC to enable robust paleoreconstructions.

This study is funded by, and integrated into, an industrial-academic project that was launched 2009: "*Reconstruction of Supercontinents Back to 2.7 Ga Using the Large Igneous Province (LIP) Record: with Implications for Mineral Deposit Targeting, Hydrocarbon Resource Exploration and Earth System Evolution*" (see www.supercontinent.org). This international project seeks to reconstruct ancient supercontinents by utilizing the distribution of Large Igneous Provinces, and their dyke swarms in particular. This ambitious international project has the potential to localize the continuation of metallogenic deposits from one craton to another.

2 Geological Setting

2.1 Regional Geology

The São Francisco Craton (SFC) is located in the northeastern portion of Brazil (Fig. 1). The craton is surrounded by Neoproterozoic mobile belts (Araçuaí, Brasília, Riacho do Pontal, Rio Preto, and Sergipano).

The basement rocks of SFC are dominated by Archaean to Palaeoproterozoic high-grade migmatite and granulite gneisses, and granite-greenstones (Teixeira et al., 2000; Barbosa et al., 2003; Barbosa and Sabaté, 2004). These rocks are covered by Meso- to Neoproterozoic sedimentary rocks of the Espinhaço Supergroup and the São Francisco Supergroup (Alkmim et al. 1993; Barbosa et al., 2003; Danderfer et al. 2009). The terrains of SFC aggregated during the Archaean and Proterozoic, mainly during the time interval 2.1-1.8 Ga (Teixeira and Figueiredo, 1991; Teixeira et al., 2000), when 35% of rocks building up the Brazilian shield originated. The SFC is made up of a number of blocks (e.g. the Gavião, Jequié, Itabuna-Salvador-Curaçá orogen, and Serrinha) and greenstone belts such as the Rio Capim, Rio Itapicuru, Mundo Novo, Contendas-Mirante, Umburanas, and Rio das Velhas. Relevant here are the Serrinha and Uauá blocks, the Itabuna-Salvador-Curaçá orogen, and Mesoproterozoic sedimentary covers in the Archaean Gavião Block where the dated mafic dykes occur.

The Serrinha Block is composed of the Mesoarchaean Santa Luz migmatite-granite basement complex, the allochtonous Archaean Uauá Block (Oliveira et al. 2010), and the Rio Capim and Rio Itapicuru greenstone belts (Fig. 2). The latter two were formed during the same period at approximately 2.1 Ga (Oliveira et al, 2010, 2011), and they are overlain to the east by sedimentary rocks of the Neoproterozoic Sergipano belt, or younger (Fig. 2). To the west, the Rio Itapicuru and the Rio Capim greenstone belts rest in tectonic contact respectively with high-grade rocks of the Serrinha Block basement complex and the Uauá Block. Mafic dykes are not found within the above greenstone belts.

The Uauá Block comprises different varieties of Archaean gneisses, intruded by tonalite-granodiorite bodies, mafic-ultramafic complexes, and layered anorthosites (Oliveira et al., 2010) with an imprint of granulite facies metamorphism followed by retrograde metamorphism under amphibolite facies conditions (Oliveira et al., 1999). The anorthosites yielded emplacement ages of ca. 3161 Ma (Paixão & Oliveira,

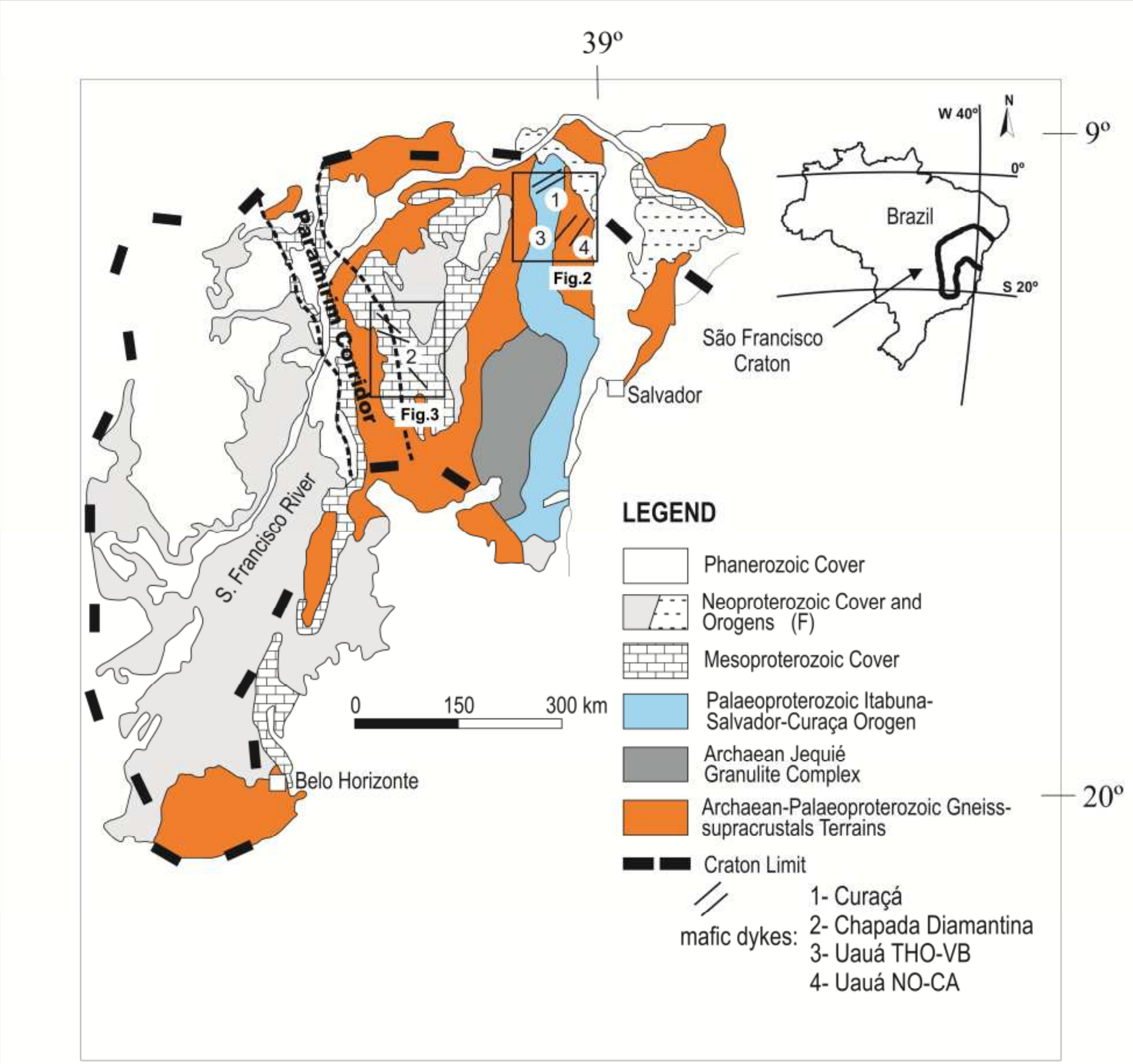

Fig. 1. Simplified geological map of the São Francisco Craton showing the location and trends of dykes investigated in this study (no. 1, 2, 3 and 4). Frames depict areas magnified in Figure 2 and 3. Modified after Oliveira et al. (2004a).

1998), whereas zircon in orthogranulites yielded a ca. 3000 Ma age (Oliveira et al, 2002), and which was interpreted to date an event of granulite facies metamorphism. Tonalites nearby the Rio Capim greenstone belt were dated at 3050 Ma and 3130 Ma (Cordani et al., 1999). The Uauá block is bordered, to the east, by the Rio Capim greenstone belt and by Neoproterozoic sedimentary rocks of continental shelf provenance incorporated into the Sergipano orogenic belt (Bueno et al., 2009; Fig. 2), that formed during the Brasiliano/ Pan-African orogeny when the Pernambuco-Alagoas Massif in the north collided with the Congo-São Francisco Craton in the south (Brito Neves et al., 1977). To the west, the Uauá Block is bounded by Archaean-Palaeoproterozoic rocks of the Caldeirão Belt along a mylonitic, up to 10 km wide, subvertical shear zone (Oliveira et al., 2002, 2010). The Caldeirão Belt rocks, consisting of quartzites, metapelites, orthogneisses, migmatites and mafic rocks have undergone an amphibolite facies metamorphism, and are in contact with the Sergipano Belt in the north. Metamorphism in the Caldeirão Belt was dated at ca. 2076 Ma ($^{207}Pb/^{206}Pb$ SHRIMP in zircon overgrowth rims), whereas deposition of sediments (e.g. quartzites) is bracketed between 2076 and 2687 Ma (Oliveira et al., 2002). 3152 ± 5 Ma old orthogneisses within this belt are among the oldest rocks in the Serrinha Block, and depleted mantle model ages indicate that even older rocks may exist within this block.

In contact with the western side of the Serrinha Block, rocks of the Palaeoproterozoic Salvador-Curaçá Belt dominate. These rocks may belong to a ca. 2.6 Ga continental arc (Oliveira et al., 2010) and were reworked by a complex collision of a number of

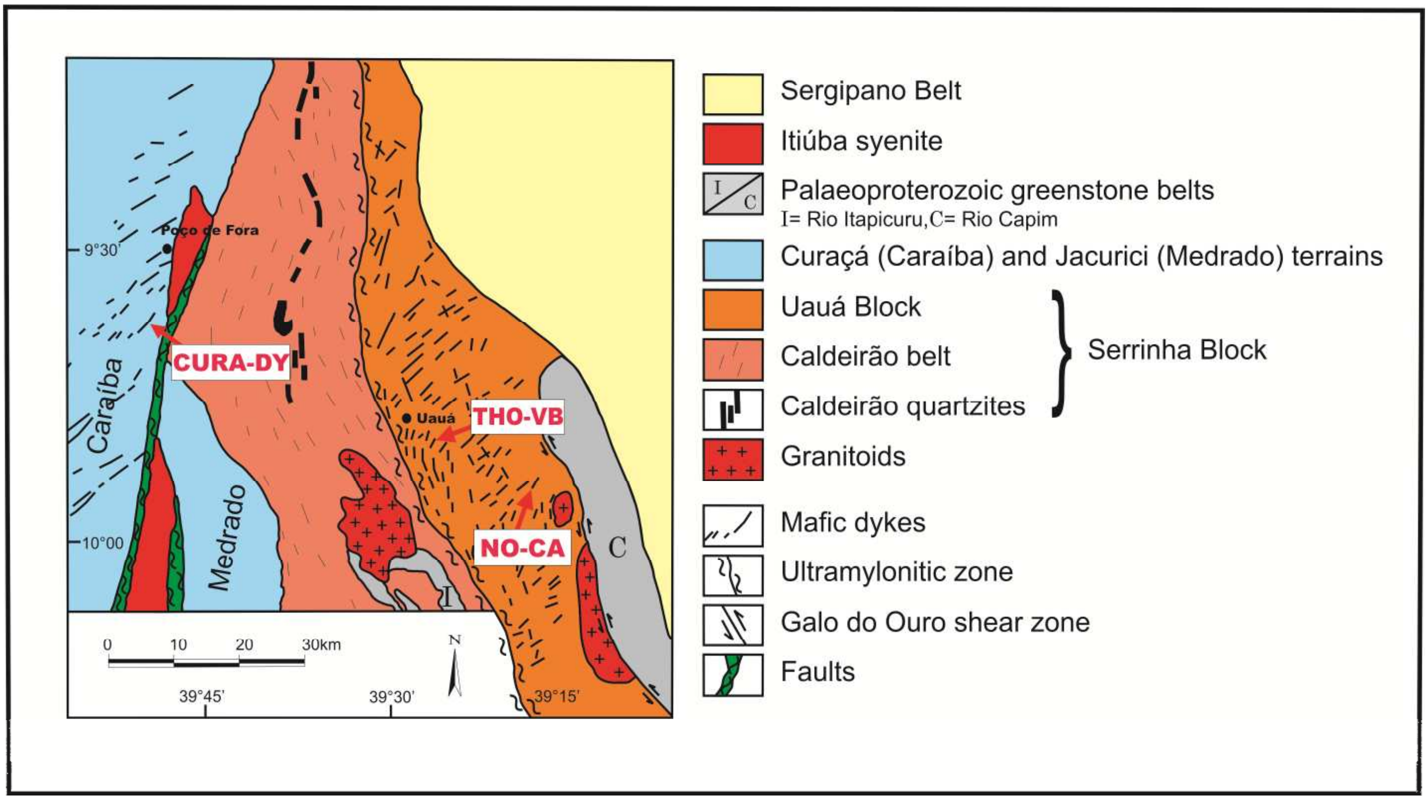

Fig. 2. Overview geological map of the Curaçá and Uauá dykes. Modified after Menezes Leal et al. (1995); Oliveira & Tarney (1995); Oliveira et al. (2002). THO-VB = Tholeiitic, NO-CA = Noritic. Note the gradual change in trend directions of THO-VB and NO-CA dykes when approaching the ultramylonitic zone that separates the Caldeirão belt and the Uauá block, indicative of sinistral shearing.

blocks (e.g. Gavião, Jequié and Serrinha blocks) around 2.1 Ga (Barbosa & Sabaté, 2002, 2004; Oliveira et al., 2002, 2010). The Salvador-Curaçá Belt is the northern segment of the Itabuna-Salvador-Curaçá orogen, which has a length of 800 km. The southern segment has been called either the Atlantic coast granulite belt (Mascarenhas, 1979 cited in Oliveira et al., 2010) or the Itabuna complex (Teixeira & Figueiredo, 1991).

The Curaçá terrain, as part of the Salvador-Curaçá Belt, is well-known by its Cu-deposits (Caraíba and Surubim, e.g. Oliveira and Tarney 2004; D'el Rey Silva et al., 2007) (Fig. 2). It is separated from the similar composition Jacurici terrain, rich in Cr, (Medrado complex, e.g. Oliveira et al., 2004a, see Fig. 2) in the Serrinha Block by the 2084 Ma Itiúba syenite (Oliveira et al, 2004a). Both terranes are mainly composed of tonalite-granodiorite plutons with basic-ultrabasic enclaves, and supracrustal rocks that have undergone granulite facies metamorphism (Barbosa & Sabaté, 2004; Oliveira et al., 2010). The ore-bearing igneous protolith (norites) in the Curaçá and the Jacurici terranes were dated at 2580 Ma and 2085 Ma, respectively (Oliveira et al., 2004a).

The Gavião block is mainly composed of 2.8–2.9 Ga amphibolite facies tonalite–granodiorite orthogneisses, gneiss–amphibolite associations, and greenstone belts (Barbosa and Sabaté, 2004, and references therein). An old nucleus of trondhjemite–tonalite–granodiorite (TTG) includes some of the oldest rocks in South America, with ages ranging from 3.4 to 3.2 Ga (e.g. Martin et al., 1997). Basement rocks of the block are covered by the Espinhaço Supergroup.

The Espinhaço Supergroup is an intracratonic basin composed of quartzites, volcanic rocks, carbonates and mafic dykes, and has undergone low-grade metamorphism under greenschist facies conditions. It is divided into the Rio dos Remédios, Paraguaçu and Chapada Diamantina groups (Teixeira et al., 2010) in the Chapada Diamantina region and into Borda Leste and Serra Geral groups in the northern portion of Serra do Espinhaço (Barbosa & Dominguez, 1996). Magmatic zircons of the Rio dos Remédios Group dated at 1748 ± 1 Ma (Babinsky et al., 1994) were interpreted to date the beginning of the Espinhaço Supergroup sedimentation. The Espinhaço Supergroup is generally interpreted as a rift sequence (Martins-Neto et al. 2000), probably an aulacogen (e.g. Alkmim et al. 2006), and is broadly separated into two exposure areas by the Paramirim Corridor (Figure 1).

The Paramirim Corridor (Alkmim et al., 1993) is a deformed zone that crosses the SFC with a NNW-SSE trend and is temporally linked to the Brasiliano cycle (650-550 Ma). Teixeira and Figueiredo (1991) described this structure in terms of extensional grabens and intracratonic basins aged between 1.8 and 1.2 Ga within the Espinhaço Belt. It is also identified by the intrusion of alkaline complexes and A-granites into an extensional setting (Teixeira et al., 2010, and references therein).

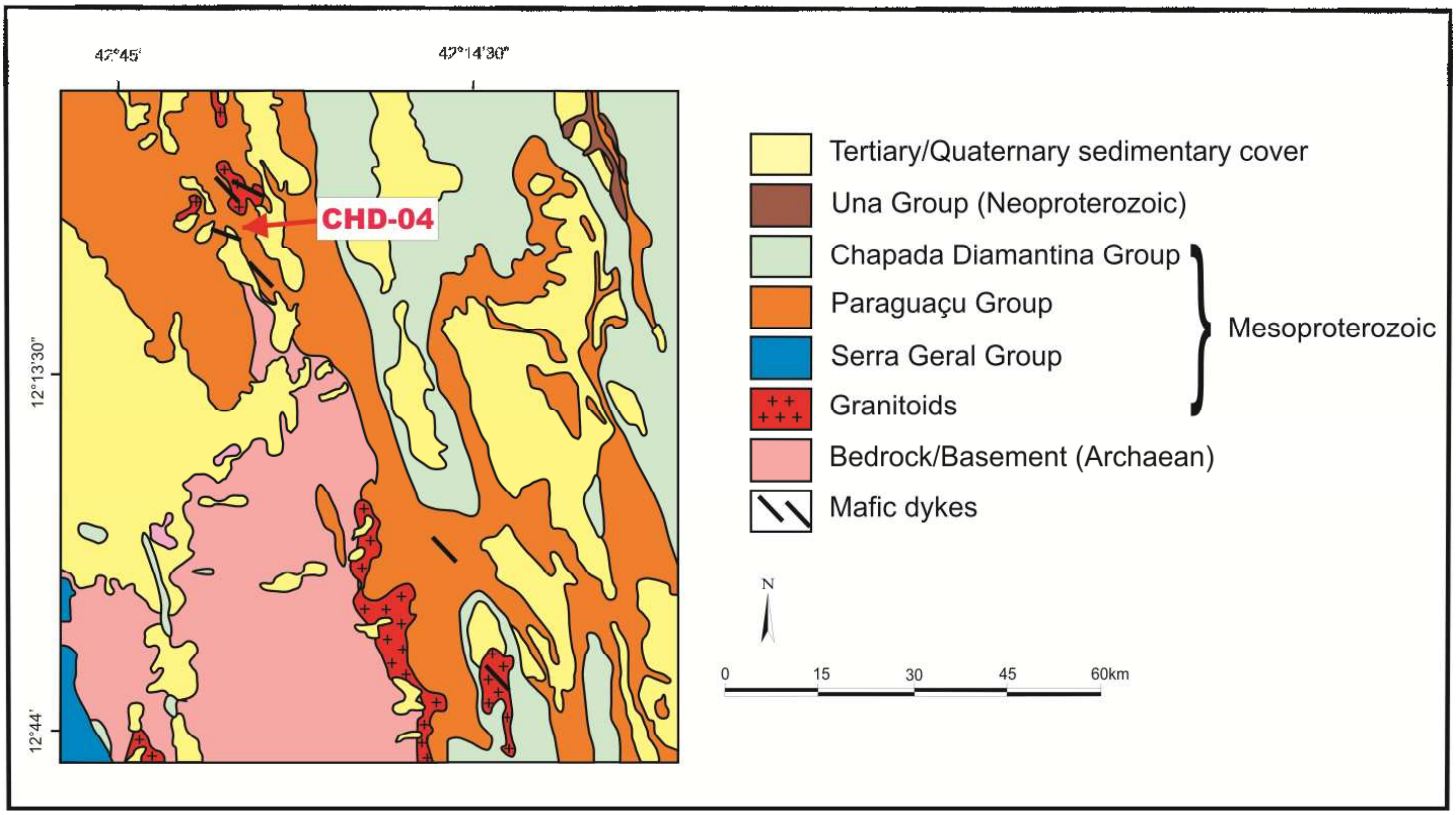

Fig. 3. Overview geological map of Chapada Diamantina dykes. Modified after Bandeira (2010).

2.2 Mafic dykes in the São Francisco Craton

2.2.1. Uauá dykes

The Uauá Block has been intruded by a set of dykes commonly referred to as the Uauá dyke swarm, and representing one of the most prominent swarms in the São Francisco Craton (Oliveira et al., 1999 (Fig. 2). The dykes comprise both tholeiitic and noritic compositions, the former structurally younger than the latter (Oliveira et al., 2010; Oliveira, 2011). The dykes gradually change their trend from NE-SW to NW-SE (Fig. 2) as they approach the Caldeirão Belt, where they have been stretched and boudinaged by deformation along the Caldeirão Shear Belt. Samples of both the younger and older dykes in this swarm have been previously dated at 2586 ± 66 Ma and 2744 ± 65 Ma, respectively (Sm-Nd; Oliveira et al., 1999). Oliveira et al. (2002) related the deformation of dykes near the contact with the Caldeirão Belt to the displacement of the Uauá Block at some time between 2039 and 2077 Ma.

2.2.2. Curaçá dyke swarm

The Curaçá dyke swarm (Fig. 2) is situated in the Curaçá river valley, intruding the Palaeoproterozoic metamorphic rocks of the Caraíba complex and is overlain by the Neoproterozoic rocks of the Canudos Group (Bastos Leal et al., 1995). It is one of the most alkaline swarms in the Precambrian of the Brazilian territory (Oliveira & Tarney, 1995). The same authors proposed that the Curaçá dykes intruded the crust between 650 and 700 Ma (Rb/Sr ages by Bastos Leal, 1992) as a consequence of an extension created by the collision of the Pernambuco-Alagoas massif with the São Francisco Craton. Prior to the present study there had been no attempts to date this dyke swarm.

2.2.3. Chapada Diamantina dykes and sills

The dykes of Chapada Diamantina (Fig. 3) intruded metasedimentary rocks of the Espinhaço Supergroup (e.g. Chapada Diamantina, Paraguaçu groups) and Archaean basement rocks (Menezes Leal et al., 2010). Some geochronology on Chapada Diamantina dykes and sills has been previously attempted (e.g. Babinsky et al. 1999; Guimarães et al. 2005; Battilani et al. 2005). In the Brotas de Macaúbas region of Chapada Diamantina, the sandstones of the Mangabeira Formation (Paraguaçu Group) were intruded by the Lagoa de Dentro gabbro sill (at least 30 m wide). U-Pb zircon ages for this gabbro, yielded an upper intercept age of 1514 ± 22 Ma (Babinsky et al., 1999). Guimarães et al. (2005) also dated gabbros and obtained U-Pb zircon analyses ages of 1496 ± 3.2 Ma for the mafic dykes in the Lagoa do Dionísio area which cross-cut the metaconglomerates, and conglomerate sandstones of the Morro do Chapéu Formation (the uppermost unit of the Chapada Diamantina Group). Muscovite-martite (iron-oxide) dykes, with NNW-SSE trend, cut the subhorizontal sandstones of the lowermost unit of the Chapada Diamantina Group (Tombador Formation) and were dated by the Ar-Ar method and yielded ages

Table 1. U-Pb baddeleyite TIMS data

Sample	Fraction	Number of grains	Pbc/ Pbtot[1)]	^{206}Pb/ ^{204}Pb	^{207}Pb/ ^{235}U	± 2s % err	^{206}Pb/ ^{238}U	± 2s % err	^{207}Pb/ ^{235}U	^{206}Pb/ ^{238}U	^{207}Pb/ ^{206}Pb	± 2s % err	Concord- ance
				raw[2)]	[corr][3)]				[age, Ma]				
NO-CA	a	5	0,009	3217	13,1613	0,23	0,51170	0,22	2691,3	2663,9	2711,9	1,2	0,977
	b	8	0,003	3990	13,4370	0,26	0,51993	0,26	2710,9	2698,9	2719,8	1,5	0,991
	c	6	0,005	2957	12,8056	0,31	0,50129	0,29	2665,5	2619,3	2700,7	1,5	0,963
Curaçá	a	11	0,016	1317	3,2856	0,60	0,25447	0,59	1477,6	1461,5	1500,8	2,1	0,969
	b	22	0,017	1661	3,3441	0,43	0,25857	0,41	1491,4	1482,5	1504,0	1,6	0,982
	c	25	0,017	1564	3,2756	0,47	0,25361	0,46	1475,3	1457,1	1501,5	1,6	0,965
CHD-04	a	7	0,019	1191	3,1914	0,70	0,24847	0,69	1455,1	1430,6	1491,0	2,0	0,954
	b	6	0,013	1496	3,1314	0,55	0,24405	0,52	1440,4	1407,7	1489,0	3,9	0,938
	c	6	0,096	419	2,9387	0,88	0,23042	0,84	1391,9	1336,7	1477,7	6,4	0,875
THO-VB	a	5	0,049	986	6,1342	0,73	0,29514	0,48	1995,1	1667,1	2354,4	8,5	0,708

1) Pbc = common Pb; Pbtot = total Pb (radiogenic + blank + initial).

2) measured ratio, corrected for fractionation and spike.

3) isotopic ratios corrected for fractionation (0.1% per amu for Pb), spike contribution, blank (2 pg Pb and <1 pg U), and initial common Pb. Initial common Pb corrected with isotopic compositions from the model of Stacey and Kramers (1975) at the age of the sample.

of 1512 ± 6 Ma and 1514 ± 5 Ma (Batillani et al, 2005, 2007). These ages are essentially the same as those obtained using the U-Pb method and are thought to date the intrusion of muscovite-martite dykes.

3 Geochronology

Samples for geochronology were collected by Elson Paiva de Oliveira (State University of Campinas, Brazil) and Angela Beatriz de Menezes Leal (Federal University of Bahia, Brazil), and prepared at the Department of Geology at Lund University. All mass spectrometric analyses were performed at the Laboratory of Isotope Geology (LIG) and the NORDSIM laboratory at the Natural History Museum of Stockholm.

3.1. Isotopic dilution TIMS geochronology

Samples were crushed to small cm-size chips using a sledge hammer and subsequently turned into powder using a mill tray. The recovered powder was carefully mixed with water before being loaded in portions (ca. 50 g per portion) on a Wilfley table (for details see Söderlund & Johansson, 2002). Immediately after loading, the finest material is washed off whereas the remaining material will start to move slowly across the table. Blades and needles of baddeleyite grains have high surface/volume ratios and almost "float" on top of the table deck, thus moving at low speed across the table. After 40-60 s, only the smallest and densest mineral grains (including baddeleyite) remain on the table, and these are rinsed off into a beaker using a squirt bottle. The procedure was repeated 3-4 times for each sample. The grains were then transferred to a petri dish where the magnetic minerals were removed using a pencil magnet wrapped into a thin plastic film (replaced with a new film each time to avoid cross contamination between samples). The baddeleyite grains (and zircon grains for sample THO-VB) were handpicked under a microscope.

For each of samples, CURA-DY, CHD-04 and

Table 2. U-Pb zircon SIMS data

Sample	Th/U calc	Th/U meas	^{207}Pb/ ^{235}U	± s (%)	^{206}Pb/ ^{238}U	± s (%)	^{207}Pb/ ^{235}U	± s	^{206}Pb/ ^{238}U	± s	^{207}Pb/ ^{206}Pb	± s
									[age, Ma]			
THO-VB												
n3425-01a	0,89	0,91	11,440	1,14	0,4767	0,98	2559,7	10,7	2512,9	20,4	2597,1	9,7
n3425-02a	0,83	0,82	13,227	1,16	0,5342	1,09	2696,0	11,0	2759,0	24,5	2649,1	6,6
n3425-03a	0,95	0,90	13,522	1,07	0,5385	1,03	2716,8	10,2	2777,0	23,2	2672,4	5,0
n3425-04a	0,56	0,57	12,174	1,00	0,5002	0,93	2617,9	9,4	2614,6	19,9	2620,5	6,3
n3425-05a	0,45	0,54	9,200	1,10	0,3992	0,93	2358,1	10,2	2165,3	17,0	2529,3	10,1
n3425-06a	0,49	0,48	12,779	1,10	0,5209	0,95	2663,5	10,4	2703,1	21,0	2633,5	9,3
n3425-07a	0,61	0,61	12,123	1,00	0,4986	0,93	2614,0	9,4	2607,6	19,9	2618,9	6,0
n3425-07b	0,71	0,72	13,057	1,02	0,5294	0,92	2683,8	9,7	2739,0	20,6	2642,4	7,4
n3425-08a	0,55	0,72	8,595	1,08	0,3741	1,00	2296,0	9,9	2048,6	17,6	2524,2	6,8
n3425-08b	0,55	0,63	8,955	1,24	0,4091	1,06	2333,4	11,4	2210,8	19,9	2442,4	10,9

NO-CA, the baddeleyite grains (4 to 25) considered to be of the best quality were transferred to Teflon dissolution capsules and washed repeatedly in MQ-$H_2$0. After that, 12 drops of MQ water and 12 drops of nitric acid (HNO_3) were added before putting the capsules on a hot plate at 100º C for about 1 hour. Thereafter the sample was successively washed in MQ water and spiked with a small portion (ca. 10 mg) ^{205}Pb, $^{233\text{-}236}$U isotopic tracer solution.

The baddeleyite grains were dissolved in 10 drops of a 10:1 HF:HNO_3 solution at 205º C for 12 hours and then evaporated on a hot plate at 100º C. Sample were dissolved in 10 drops of 3.1 N HCl and then transferred into 50 µl column filled with anion resin. To remove the Zr and Hf, 16 drops (2+2+2+10) of 3.1 N HCl were added to the columns. U and Pb were eluted from the columns by using MQ water (3+18+10+25 drops). One drop of phosphoric acid (H_3PO_4) was added directly to the capsules before samples were evaporated on the hot plate at 100º C overnight.

The samples were loaded with 2 µl of silica gel on an outgassed Re (Rhenium) filament. The silica gel was used to retard the volatilisation of Pb, so the filament could be heated at high temperatures to provide a stable Pb beam during measurement (Dickin, 2005). All the pipette tips were rinsed repeatedly in 2% HNO_3 before contact with the sample. A 1 A (Ampere) current was set through the filament when adding the sample in 2-3 portions onto the centre of the Re filament and then slowly increased until white smoke was released (at ca. 2.4 A), meaning that the phosphoric acid had burned out. After a short glowing the filament was ready to put into the mass spectrometer.

The total procedural blank for the chemical separation was less than 2 pg for Pb and 0.2 pg for U. The U and Pb were analysed in a TIMS Finnigan Triton mass spectrometer using a Secondary Electron Multiplier (SEM) equipped with a RPQ filter. During the analyses, the mass discrimination correction was 0.1% per atomic mass unit. The decay constants used are those from Jaffey et al. (1971). The initial common lead composition was based using the model of Stacey and Krames (1975), taken at the age of the sample.

3.2. SIMS geochronology

Ten zircon grains of sample THO-VB (Fig. 4), that had been separated in the same way as the baddeleyite grains (and having non-inherited characteristics and averaging 30 µm in size) were chosen for the analysis. The grains were mounted in epoxy resin and polished and checked in the SEM (scanning electronic microscope) to ensure a primary origin of the selected grains, i.e., no signs of inheritance. The zircon grains were analysed in a secondary-ion mass spectrometry (SIMS) Cameca 1280 at the NORDSIM laboratory, following standard procedures for zircon analyses (Whitehouse & Nemchin, 2009). The samples were placed in the ion source and a beam of light ions (e.g., O^-) shot and sputtered the polished surface of the zircon grains destroying an area of 20 µm and yielding a secondary beam of Pb that was analysed by mass spectrometry (Ireland & Williams, 2003; Dickin, 2005). The advantage of the Cameca 1280 is that it has a high

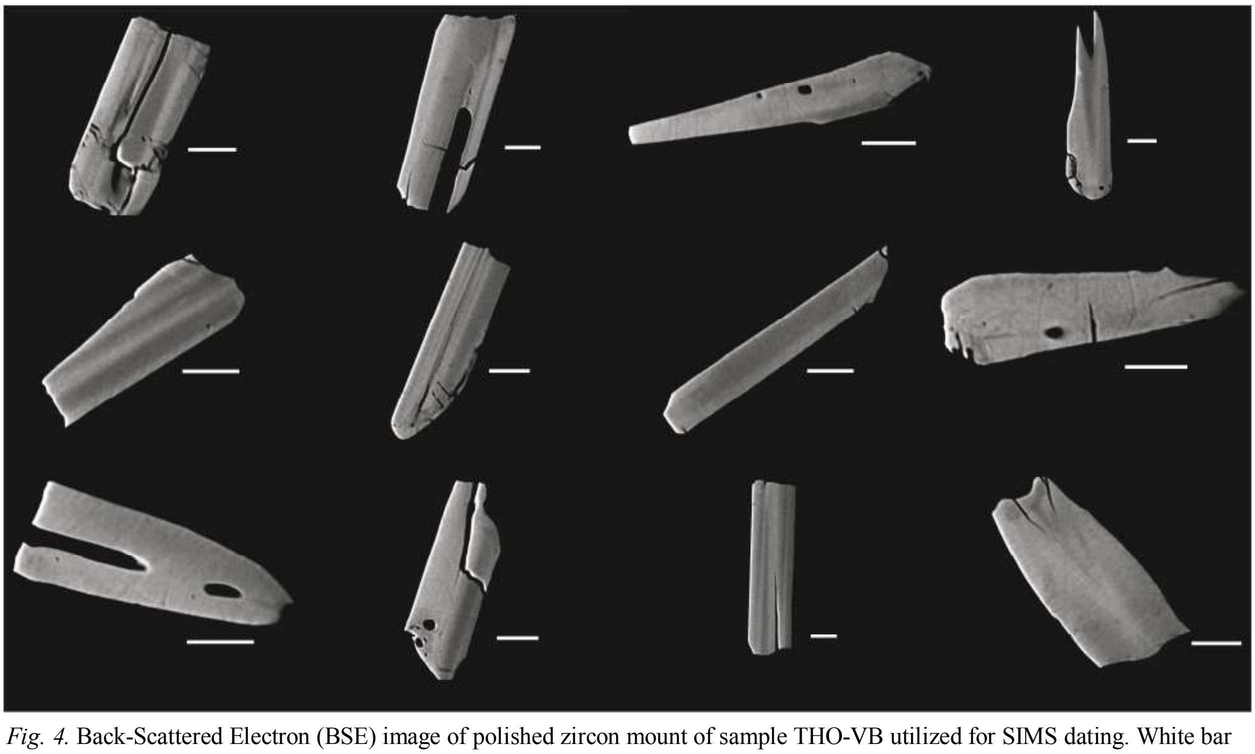

Fig. 4. Back-Scattered Electron (BSE) image of polished zircon mount of sample THO-VB utilized for SIMS dating. White bar =10 µm. Note typical magmatic features of this homogenous zircon population.

Table 3. Discriptive data of U-Pb dated samples (the datum used for coordinates is Corrego Alegre, rather close to WGS84)

Sample	Unit	Latitude	Longitude	Direction	Age, error (2σ)
NO-CA	Uauá dyke	09º55'58.0" S	39º19'08.5" W	N30°	2726.2 ± 3.2
THO-VB	Uauá dyke	09º50'43.8" S	39º28'10.8" W	N40°	2623.8 ± 7.0
Curaçá	Curaçá dyke	09º34'03.0" S	39º47'24.1" W	N70°	1506.7 ± 6.9
CHD-04	Chapada Diamantina dyke	12º01'51.3" S	42º36'52.0" W	N110°	1501.0 ± 9.1

mass resolution allowing a large separation between different masses. Otherwise, interferences by species such as HfO^{+2} and HfSi could spoil the analysis, since they have almost the same mass as the Pb isotopes. The ages and plot calculations for both baddeleyite and zircon were obtained using Isoplot (Ludwig, 2003).

4. Sample description and U-Pb results

U-Pb data is shown in Concordia diagrams in Figure 5. Table 1 reports U-Pb TIMS data of samples NO-CA, Curaçá CURA-DY and CHD-04, whereas SIMS data of THO-VB is listed in Table 2. The coordinates of the dykes were acquired with a handheld GPS, and these, in addition to trend directions and U-Pb ages, are reported in Table 3.

The NO-CA sample is from a coarse-grained, ca. 40 m thick noritic dyke of the Uauá swarm (Figs. 1 and 2). The trend of the dyke is N30º. The sample was collected 7 m away from the chilled margin. Dark brown, approximately 30 μm large baddeleyite grains, a few with faint lighter rims in other portions, were extracted from this sample. Regression of three variably discordant baddeleyite analyses yields an upper intercept age of 2726.2 ± 3.2 Ma (MSWD = 0.1).

The THO-VB sample is a medium-grained tholeiite that belongs to the Uauá dyke swarm as well. This dyke has N40º trend and it is approximately 20 m in width. This sample was also collected at a ca. 7 m distance from the contact. Ten zircon grains dated with the SIMS yielded an age of 2623.2 ± 7.0 Ma (MSWD = 0.8). Two fractions (analysis 03a and 08b, Table 2) plotted slightly off the discordia line and were excluded from the age calculation. One fraction comprising five grains was dated by the TIMS dilution technique. This single analysis plotted strongly discordant (ca. 30%) but falls along the SIMS regression line.

The Curaçá sample (CURA-DY) is from a medium-grained, more alkaline, diabase and belongs to a N70º-trending dyke swarm (Fig. 2). The width of the dyke is ca. 15 m. The sample was collected at least 3 m from the chilled margin of the dyke. Approximately 60 grains of light to dark brown baddeleyite were ex-

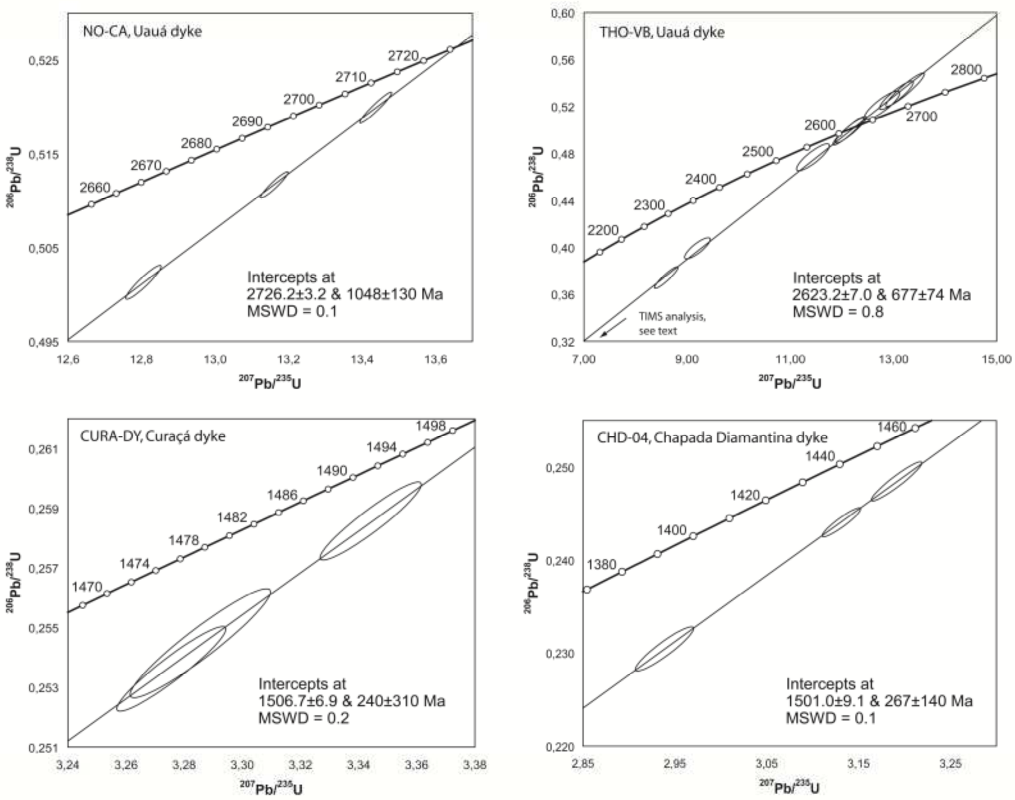

Fig. 5. U-Pb Concordia diagrams of NO-CA, THO-VB, Curaçá and CHD-04. All samples were dated using ID-TIMS on baddeleyite except the THO-VB; this sample was dated by SIMS analysis of zircons. Ellipses depict 2 sigma errors.

tracted. Analysis of three fractions yielded an upper intercept age of 1506.7 ± 6.9 Ma (MSWD = 0.2).

The CHD-04 sample is a medium-grained tholeiite with a greenish appearance due to weathering. The dyke trends N110º and has a width of ca. 40 m. It was sampled 5 m away from the contact. The baddeleyite grains chosen for isotopic analyses were light-brown coloured. Regression of three variably discordant baddeleyite analyses yields an upper intercept age of 1501.0 ± 9.1 Ma (MSWD = 0.1).

5 Discussion

5.1. U-Pb data of investigated dykes

Baddeleyite in mafic dykes can confidently be linked to crystallization of magmas since baddeleyite cannot survive melting processes or metamorphism. In fresh dolerite, baddeleyite generally yields concordant to nearly concordant U-Pb dates, and ID-TIMS is typically preferred as this technique yields U-Pb dates at high precision.

All samples dated in this study carry fresh baddeleyite except the THO-VB dyke, which instead has abundant zircon. Although the general use of zircon to date mafic dykes can be questioned, a primary origin for the zircons in THO-VB is evident. As seen in Figure 4, these grains represent a homogeneous population with typical magmatic shape (e.g., having pyramidal tips), no signs of zircon overgrowth (due to metamorphism), sharp edges between the crystal surfaces. In addition, absence of gneisses in this region (i.e. from the basement) with the same age as the zircons (e.g. Oliveira et al., 2002) enforces that these zircons were crystallized directly from the magma.

One fraction of zircon from sample THO-VB was initially analyzed using TIMS. This single fraction plots ca. 71% concordant, indicating significant Pb-loss though the zircons look pristine and crystalline. Therefore, spot dating using SIMS was chosen for constraining the age of this sample. As seen in Figure 5, the single TIMS-analysis plots along the discordia line constrained from the less discordant SIMS analyses, which demonstrates Pb-loss in zircon took place preferentially from the margins of the grains. The upper intercept age is almost unaffected when including the TIMS analysis in the regression (2624.5±7.1 Ma, MSWD=1.1), yet the less discordant SIMS data alone is preferred as the age of THO-VB, hence dated at 2623.8 ± 7.0 Ma. Notably, the SIMS and TIMS data collectively show zircon in this sample is cogenetic and of the same age.

The three baddeleyite fractions of the Curaçá dyke do not constitute a three-point regression as the two most discordant analyses plot on top of each other. Free regression yields an upper intercept of 1506.7 ± 6.9 Ma and a lower intercept of roughly 250 Ma, similar to the lower intercept in the regression of the CHD dyke. Using a forced lower intercept at 0 ± 100 Ma, for the Curaçá dyke, lowers the upper intercept to 1502.5 ± 2.3 Ma (MSWD = 1.2), and into agreement with the age of the CHD dyke (1501.0 ± 9.1 Ma). However, because both samples (CHD-04 and CURA-DY) indicate isotopic disturbance at the same time (ca. 300 Ma) the slightly older result is preferred for the Curaçá dyke.

5.2. New ages of mafic dykes in the São Francisco Craton

The 2726.2 ± 3.2 and 2623.8 ± 7.0 Ma U-Pb baddeleyite ages of the NO-CA and THO-VB dykes, respectively (Fig. 2 and 5), are consistent with the less precise 2744 ± 65 Ma and the 2586 ± 66 Ma ages for corresponding dykes (Oliveira et al., 1999), constrained from whole-rock Sm-Nd isochron geochronology of pyroxenite-norite dykes and tholeiite-norite dykes within the Uauá block. In contrast, Bastos Leal et al. (1994) reported emplacement ages of 2.4 and 2.0 Ga by Rb-Sr for these two dykes of the present study, but the mobility of both Rb and Sr during low-grade metamorphism and alteration probably explains the younger Rb-Sr dates. The U-Pb baddeleyite ages, and to some extent the Sm-Nd whole-rock results, collectively indicate a prolonged event of dyking of approximately 100 Myr for the Uauá dyke swarm. Additional precise dating of dykes in the Uauá block are needed to reveal whether dyking was continuous or took place during two or more discrete events. The new ages obtained for the Uauá dykes are supported by the cross-cutting field relationship, whereas the THO-VB (tholeiitic) cuts the NO-CA (noritic). The younger age of ca. 2624 Ma sets a maximum age of sinistral shearing between the Caldeirão belt and the Uauá block, hence corroborating with the interpretation of Oliveira et al. (2002) that dated this event of shearing between 2039 and 2077Ma. Since field relations indicate that tholeiitic dykes cut the noritic dykes, a ca 100 Myr age difference between the two composition types is plausible, although additional dating are required to establish an older noritic pulse followed by a younger tholeiitic pulse.

Bastos Leal (1992) also reported Rb-Sr ages in the range of 700-650 Ma for the intrusion of the Curaçá dyke swarm. These results led to an interpreted link with the Brasiliano/Pan-African cycle. However, the much older date of 1506.7 ± 6.9 Ma obtained for the Curaçá dyke in this study again demonstrates significant isotopic disturbance of the Rb-Sr systematics and, more importantly, precludes a link to the Brasiliano cycle for the Curaçá dyke swarm. Rather, as discussed below the 1506.7 ± 6.9 Ma age of the Curaçá dyke, together with the similar age of the Chapada Diamantina dyke sample, has the potential to provide important constraints for palaeocontinental reconstructions involving the São Francisco Craton.

The dyke of the Chapada Diamantina swarm, here dated at 1501 ± 9.1 Ma, lies ca. 400 km southwest

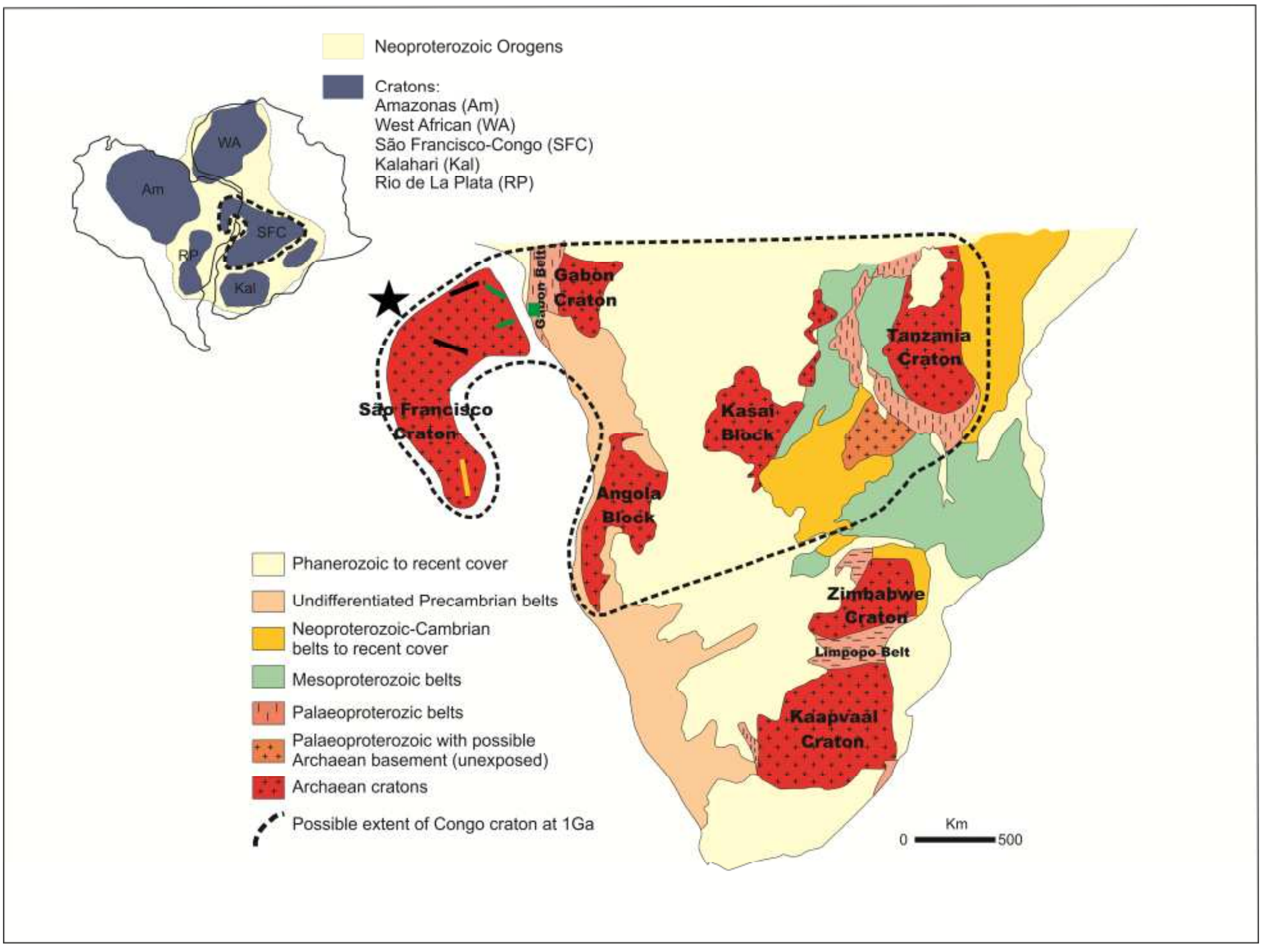

Fig. 6. 1500 Ma dykes from Curaçá and Chapada Diamantina in the São Francisco Craton (black dashes), 1100 Ma dykes from Januária swarm (in Chaves & Neves, 2005) in the São Francisco (yellow dash) and 920 Ma Salvador and Ilhéus-Olivença dykes (Heaman, 1991; Evans et al., 2010) and Gangila Basalt (Tack et al., 2001) (green square and dashes). Black star shows possible mantle plume centre. Modified after De Waele et al. (2008) and Danderfer et al. (2009).

from the Curaçá dyke (Fig. 1). Previous age determinations fall close to this result: a gabbro sill dated at 1514±22 Ma (U-Pb zircon, Babinsky et al., 1999), 1496±3 Ma intrusive gabbros in the Lagoa do Dionisio region (U-Pb zircon, Guimarães et al., 2005) and Ar-Ar muscovite in muscovite-martite fels ages of 1512±6 and 1514±5 Ma (Battilani et al., 2005). Figueiredo et al. (2009) suggested that zircons, roughly dated at 1500 Ma, in diamictites, could possibly have a provenance from gabbroic dykes and sills that cross-cut the Espinhaço Supergroup.

5.3. Palaeocontinental reconstructions in the late Archaean

The origin of noritic dykes in the Uauá block has been a matter of study for many years, but a reliable and precise age has not been available until now. The 2726 Ma age on the NO-CA norite sample cannot be easily matched with any other units in the São Francisco craton, which is not expected since the Uauá block was not amalgamated with the present-day surrounding blocks until approximately 2.0 Ga. However, the Uauá block can be temporally linked to many Late Archaean tholeiitic komatiite (LIP-style) greenstone belts elsewhere around the globe, such as the Abitibi of Superior craton which contains tholeiitic basalts (2730-2705 Ma, Dostal & Mueller, 1997; Wyman, 2003) and basaltic volcanism with minor komatiitic rocks in Kam group of the Slave craton (2730-2700 Ma, Bleeker, 2003), both in Canada. In addition, similar ages have been presented for the Kylena stage (2765 Ma) of Fortescue Group of Pilbara craton in Australia and for the Klipriviersberg Group (2714 ± 8 Ma) of Ventersdorp magmatism in the Kaapvaal craton in southern Africa (Eriksson et al., 2002, Taylor et al., 2010). In addition, Arndt et al. (1991) and Nelson (1998), cited in Eriksson et al. (2002) obtained U-Pb zircon ages for a tuff (2715 ± 6 Ma) from an unit correlated to the Fortescue Group (Tumbiana Formation) and a dacite of 2715 ± 6 Ma that overlies the Tumbiana Formation. Eriksson et al. (2002) proposed the ca. 40 Myr age span of these magmatic units to be related to mantle plume activity.

Bastos Leal et al. (2003) present $^{207}Pb/^{206}Pb$

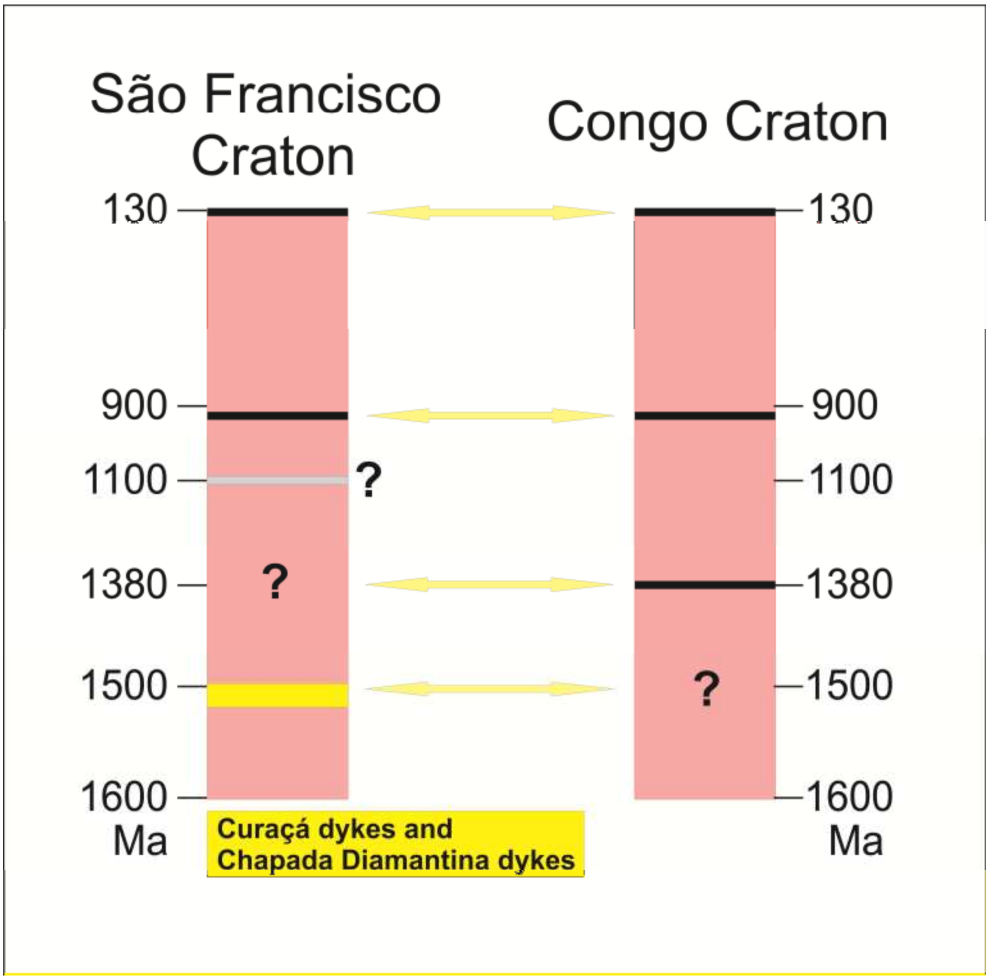

Fig. 7. 1500 Possible matches of mafic magmatism between the SFC and the Congo cratons. The thickness of the black and grey bars show the average of radiometric ages, whereas the thickness of the yellow bar depicts 2 s errors.

zircon evaporation age of 2744 ± 15 Ma for a meta-andesite of the intermediate unit of the Umburanas greenstone belt in the Gavião Block. Noce et al. (2005) reported SHRIMP and ID-TIMS U–Pb zircon ages for three volcaniclastic graywacke samples at 2792 ± 11, 2773 ± 7 and 2751 ± 9 Ma, indicating a duration of about 40 Myr of felsic magmatism.

The 2623.8 ± 7.0 Ma age of tholeiite sample THO-VB does not match any other LIP (R. Ernst, pers. comm., 2009), although the interval between 2650-2580 Ma is a major magmatic event in the Caraíba arc (in the SFC) (Oliveira et al., 2010) and so the possibility of linking this tholeiitic Uauá dyke to mafic-ultramafic complex in the Salvador-Curaçá orogen remains possible (e.g., Oliveira, 2011).

5.4. 1500 Ma dyke swarms in the São Francisco Craton: a possible link to a mantle plume

The new age obtained for the dykes of the Curaçá region (1506.7 ± 6.9 Ma) allows for matching the SFC with other cratons around the globe during the Mesoproterozoic and, the age match with the Chapada Diamantina dyke swarm age of 1501 ± 9.1 Ma can have implications for the setting of these two dyke swarms within the São Francisco Craton.

The 1501 ± 9.1 Ma age for the Chapada Diamantina dykes is within error the same as for the Curaçá sample, though these dykes are as much as 400 km apart (Fig. 1). Assuming the U-Pb baddeleyite ages are representative for all dykes of each swarm, then one interpretation is that they define a radiating pattern with magmas fed from a common igneous centre to the NW (Figs. 1 and 6). However, the Chapada Diamantina dykes are thoileiitic in composition whereas the Curaçá dykes are more alkaline. If magmas for the two swarms were indeed fed from a common igneous source, then there must be a time gap between the swarms to develop the recorded chemical difference between the dyke swarms.

Alternatively, the Chapada Diamantina dykes manifest a late intracratonic rifting event linked to the Paramirim Corridor. However, the onset of rifting along the Paramirim Corridor is constrained from the 1748 ± 1 Ma age of sedimentation (Babinsky et al., 1994), hence some 200 Myr earlier than the here dated

Chapada Diamantina dyke. There also seems to be a small but distinct difference in directions between the Paramirim Corridor and the dated dyke of the Chapada Diamantina dyke (sample CHD-04). The existence of a more distal dyke swarm (Curaçá dykes), perpendicular to the Paramirim Corridor, further suggests that the ca. 1500 Ma dykes in SFC are unrelated to rifting along the Paramirim Corridor.

5.5 A LIP barcode for the combined SFC – Congo Craton

The 130 Ma sill from Paraná Basin in the southern Brazilian territory is considered to be one of the largest LIPs in the world (Deckart et al., 1998). Together with its counterpart in the southwest African territory, i.e. the Etendeka magmatic province, they are among the largest continental flood basalts in the world (Janasi et al., 2011) and manifest break up of South American and African continents.

Figure 7 outlines possible age matches of mafic magmatism between the SFC and the Congo cratons using the concept of barcode matching (Bleeker & Ernst, 2006). Assuming that the SFC and the Congo Craton were amalgamated from at least 2.0 Ga until 130 Ma (see section 1), it is expected that regional events of mafic magmatism are to be found in both the SFC and Congo Craton. However, there are not many such LIP events or dyke swarms yet identified, or that have been properly dated. The only robust match during the Mesoproterozoic is the Salvador dykes in the São Francisco Craton dated at ca. 920 Ma (924±24 Ma, Heaman, 1991; 921.5 ± 4.3 and 924.2 ± 3.8, Evans et al., 2010). A similar age of 920 ± 8 Ma (Tack et al., 2001) was obtained for the Gangila basalts, which are associated with the bimodal Mayumbian volcanism in the West Congo belt. The ca. 920 Ma event is believed to manifest an attempt of break-up of the SFC-Congo Craton.

The extensive Umkundo event in southern Africa is precisely dated by U-Pb baddeleyite and ages of sills and dykes of this event cluster between 1112 and 1108 Ma (Hanson et al. 2004; U. Söderlund, unpubl. data). Mafic dykes of the Januária swarm in the southern part of the SFC may represent a further westerly extension of Umkundo intrusions though such inference relies on ca. 1.1 Ga K-Ar dates (Parenti Couto et al., 1983 cited in Chavez & Neves, 2005). Tholeiitic dykes of the same age (1.1-0.9 Ga) intrude Proterozoic terrains within the Araçuaí mobile belt (e.g. Diamantina and Conceição do Mato Dentro) also in the southern SFC.

Another possible age match is at 1380 Ma as represented by the Kunene Intrusive Complex in SW Angola (Drüppel et al., 2007; Ernst et al., 2008) and a mafic-ultramafic belt In the eastern portion of the Congo craton (Maier et al., 2007), both dated at 1380 Ma. However, corresponding coeval units are yet to be discovered in the SFC.

Fig. 8. Black dots mark 1500 Ma intrusions around the globe. The map show the position of SFC (upper curved portion of the SFC-Congo Craton) between Baltica and Siberia.

The here presented ages of 1501 ± 9.1 and 1506.7 ± 6.9 Ma for the Chapada Diamantina and the Curaçá dykes in SFC is a major step forward towards a complete "barcode" for the SFC-Congo Craton that can be used in global reconstructions. However, similarly old intrusions in the Congo Craton remains to be identified and properly dated.

5.6. 1.5 Ga dykes in a global perspective

Approximately 1.5 Ga dykes and sills are present not only in the SFC, but also in Fennoscandia (see Bingen et al., 2005; Bingen et al., 2008; Zhao et al., 2004) and northern Siberia (see Ernst et al., 2008). More specifically, the ca. 1500 Ma intrusive gabbro-dolerite-granite associations (see Table 1 in Ernst et al., 2008; see also Zhao et al., 2004) in southern Baltica (Scandinavia) and in the Rjukan Group in the Telemark Sector (Bingen et al., 2005; Bingen et al, 2008, Zhao et al., 2004) require some attention, since it is unclear if these terrains were amalgamated during the Grenvillean-Svecofannian orogeny, or earlier.

Within northern Siberia the Kuonamka dyke was dated at 1503 ± 5 Ma (U-Pb baddeleyite; Table 1 in Ernst et al., 2008; Ernst et al., 2000) whereas a large diabase sill in the Kyutingde Formation from the Olenek uplift yielded a U-Pb baddeleyite age of 1473 ± 24 Ma (Wingate et al., 2009). Collectively, these dates form suggest placing SFC-Congo Craton between Northern Siberia and Baltica in the Mesoproterozoic (Fig. 8).

6 Conclusion

- The noritic and tholeiitic dykes within the Uauá Block yielded U-Pb dates of 2726.2 ± 3.2 and 2623.8 ± 7.0 Ma, respectively, and are the oldest dykes yet dated in this crustal block. The older noritic dykes may belong to a late Archaean superplume event as supported by 2.76-2.71 Ga mafic magmatism common on many cratons worldwide.
- The new 1506.7 ± 6.9 Ma age for the Curaçá dyke is significantly older than previous published age estimated for these dykes. Collectively with the Chapada Diamantina dykes, here dated at 1501.0 ± 9.1 Ma, this result can be used to further reinforce a coherent São Francisco-Congo block during the late Mesoproterozoic.
- The trends of Chapada Diamantina and Curaçá dykes define a radiating pattern that may be linked to a possible magmatic centre at the western margin of the São Francisco Craton

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