Crustal evolution between 2.0 and 3.5 Ga in the southern Gavião block (Umburanas-Brumado-Aracatu region), São Francisco Craton, Brazil: A 3.5–3.8 Ga proto-crust in the Gavião block?

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ABSTRACT

The main evolution of the Gavião block in the Umburanas-Brumado-Aracatu region, in the state of Bahia, is defined by several sets of tonalitic-trondhjemitic and granodioritic gneisses emplaced during the Paleoproterozoic. The juvenile Bernada gneisses are emplaced at 3386 ± 9 Ma (SHRIMP zircon age). The Aracatu gneisses, probably derived from the partial melting of ca 3.4 Ga gneisses, are emplaced at 3325 ± 16 Ma. They contain inherited zircon dated at 3366 ± 15 Ma in the range of ages obtained for the juvenile Bernada gneisses. Furthermore, one core in these zircons provides an age of 3487 ± 9 Ma, which is the oldest xenocryst found in the Gavião block. A Neoproterozoic alkaline granite was emplaced at 2693 ± 5 Ma (Serra de Eixo gneiss) and corresponds to a major crustal reworking stage. All of these rocks were metamorphosed and melted at ca 2.0 Ga, as recorded by monazite ages (EPMA and LA-ICPMS) in diatexitic Archean gneisses (Aracatu) and Paleoproterozoic granites (Umburanas). The occurrence of a proto-crust ca 3.5 Ga or older in the Gavião block is discussed based on inherited zircon ages and Sm–Nd isotope signatures of the Archean gneisses.

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1. Introduction

In the São Francisco Craton, the Gavião block, which recorded a crustal history up to 3.1–3.4 Ga (references in Table 1), is one of the oldest continental segments in South America. Nevertheless, this vast area has not been subjected to extensive U–Pb in-situ zircon dating (ion probe and LA-ICPMS), with the result that its geochronological evolution is not very well constrained (Figs. 1 and 2, Table 1). The purpose of this work is to provide new in-situ U–Pb ages for the Gavião block and to discuss the petrogenetic and

geodynamic implications of these results in order to more precisely determine the crustal evolution between 1.97 and 3.49 Ga in this part of the São Francisco craton.

2. Geological setting

The basement of the São Francisco Craton, in the state of Bahia, consists of four main crustal segments: the Serrinha, Itabuna-Salvador-Curaçá, Jequié and Gavião blocks (Barbosa and Sabaté, 2002, 2004; Delgado et al., 2003 and references therein) (Fig. 1). These authors proposed that the four Archean blocks collided with each other during a Paleoproterozoic orogeny, resulting in the tectonic thrusting of the high grade Itabuna-Salvador-Curaçá block over the Jequié block, with both blocks overlapping the Gavião block (Fig. 1). This tectonic overstacking resulted in a major crustal thickening and a correlated high grade regional metamorphism between 2.0

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Table 1
Summary of the U–Pb ages (SHRIMP, TIMS and LA-ICPMS) in the Gavião block and the corresponding Nd isotopes. Methodologies: (1) SHRIMP, (2) TIMS evaporation, (3) TIMS (ID), (4) LA ICPMS, (5) EPMA. References: (1) Peucat et al. (2002), (2) Nutman and Cordani (1993), (3–4) Martin et al. (1991, 1997), (5) Marinho et al. (2008), (6) Marinho et al. (1994), (7) Nutman et al. (1994), (8) Santos Pinto et al. (1998), (9) Bastos Leal et al. (2003), (10) Bastos Leal et al. (2000), (11) Cruz et al. (2012), (12) Bastos Leal et al. (1998).

Location	Sample	Rock type	Zircon age (Ma or Ga) ± 2σ (inheritage)	Method.	Monazite age (or sphene)	Method.	Nd TDM in Ga	εt with t = zircon age	Reference
Northern Gavião block									
Mundo Novo greenst. belt	SSR680C	Metadacite	3305 ± 9	1			3.35–3.38	0	1
Mairi gneisses	SSR526A	Granodioritic gneiss	3040 ± 15	2			3.21	–2	1
W Mundo Novo gneisses	–	Tonalitic gneiss	ca 3.4 Ga	3			–	–	2
Southern Gavião block									
<i>Contendas - Mirante lineament and adjacent gneisses</i>									
Sete Voltas gneisses	AC-2B	Tonalitic gneiss	3403 ± 5 (3466 ± 8)	1			–	–	3
Sete Voltas gneisses	SV11	Old gray gneiss	3394 ± 22	2			3.7	–1 to –3	4–5
	SV11		3392 ± 21	1			–	–	4–5
Boa Vista gneisses	AC-1E	Granitic gneiss	3353 ± 5	1			–	–	3
Sete Voltas gneisses	SV20	Granodioritic gneisses	3243 ± 26	2			3.75	–3.1	4–5
	SV2	Young gray gneiss	3158 ± 6	2			3.69	–5.6	4–5
Lagoa do Morro gneisses	AC-4E	Foliated augen granodior.	3184 ± 6	1			–	–	3
Lagoa do Morro gneisses	AC-4G	Granodioritic migmatite	ca 2.85 Ga	1		1	–	–	3
Pé de Serra pluton	132	Granite-syenite	2652 ± 11	1			–	–	6
CM greenstone belt	42	Felsic sub-volcanic	3304 ± 87	3			–	–	7
CM sediment	CGM4	Upper- conglomerate	(2710–2600) (2400–2300) (2200–2150)	1			3.43–3.55	–0.4 to +0.9	8
Umburanas-Brumado-Aracatu region									
Bernada massif	BER120-1	Tonalitic gray-gneiss	3057 ± 32 to 3332 ± 4	2			3.27		9
	BER120-2		3386 ± 9	1				3.7	This work
Aracatu massif	ARA78-1	Trondjem. gray-gneiss	2284 ± 122 to 3240 ± 18	2			3.51	–	9
			3325 ± 16	1				–0.7	This work
			(3371 ± 14 and 3487 ± 9)	1				–	This work
	ARA06-2	Granodioritic gneiss	2804 ± 40 to 3230 ± 4	2		2	–	–5.5	9
			–			5			This work
			3021 ± 4 to 3259 ± 10	2			3.65	–1.1	9
Mariana	MAR12.3	Monzogranitic gneiss	2741 ± 4 to 3255 ± 14	2			3.47	0.7	9
	MAR94.1A	Monzogranitic gneiss	2794 ± 20 to 3158 ± 10	2			3.29	0.2	9
Serra de Eixo	VZP65	K rich calc-alk gneiss	2970 ± 46 to 3156 ± 14	2			–	–	9
	JC706	K rich calc-alk gneiss	2524 ± 14 to 2656 ± 10	2			3.28	–	9
	GA299	Alkaline augen gneiss	2693 ± 5	1			–	–3.1	This work
Umburanas greenst. belt	BRJ330	Quartzite	(3040 ± 12 to 3335 ± 12)	1			–	–	10
	BRJ337	Meta-andesite	2744 ± 15	2			2.80–3.46	+1.5 & –4.1	10
Caraguatãtãl massif		Syenitic augen gneisses	2696 ± 5	4			3.22–3.50	–4.0 to –5.8	11
Aracatu Paleoproterozoic granite	ARA170	Isotrope peralut. granite	1.999 ± 14 (to 2262 ± 22) 2053 ± 65 (3.25 Ga) 1	2			3.58	–15	9
			2.38 to 2.93 Ga	2			3.77	–15	This work
Umburanas granite	ARA81-4	Isotrope anatect. granite	2570 ± 20 to 2833 ± 66	2			3.30	–14	9
	ARN58-1		2848 ± 10 to 3103 ± 8	2			3.38	–14	9
	ARN60-1		3052 ± 30 to 3130 ± 14	2			3.38	–15	9
	UMB164		2791 ± 8 to 3006 ± 18	2			3.35	–15	9
	UMB165		–						This work
			–						1962 ± 13 to 2053 ± 17
			–						1971 ± 10 & 2.50 Ga
Mariana granite	MAR137	Isotrope pink granite	1852 ± 50 to 1944 ± 14	2			3.45	–10	9
Serra da Franga granite	SFR	Isotrope monzogranite	2039 ± 22	2			–	–	9
Mariana paragneiss	MAR 140	Garnet bearing micasch.	–				3.40	–	9
	MAR145	Garnet bearing micasch.	–				3.10	–	9
Espírito Santo granite	BR01	Granite	2012 ± 25	2			3.10	–12	10

Brumado-Ibitira-Caculé region	Lagoa da Macambira Massif	BRJC02	Migmatitic gneiss	2912 ± 10 & 3300 ± 34	2	3.46	–2	12
		BRJC178	Gneiss	3202 ± 14	2	3.62	–3.5	12
		BRJC16	Granite	3146 ± 24	2	3.34	–1.5	12
		BRJC 237	Granite	2019 ± 32	2	2.70	–8	10

and 2.1 Ga (Barbosa and Sabaté, 2002, 2004; Delgado et al., 2003; Peucat et al., 2011).

The Gavião block (GB) mainly consists of Archean orthogneisses including TTG suites with U–Pb zircon ages ranging from 2.8 to 3.4 Ga (Table 1). They were metamorphosed and partially recycled into migmatites and granites at ca 2.6–2.7 Ga (Marinho, 1991; Martin et al., 1991; Nutman and Cordani, 1993; Santos Pinto et al., 1998; Cruz et al., 2012 and references therein). Disrupted supra-crustal belts, such as the Contendas-Mirante volcano-sedimentary sequence and the Umburanas and Mundo Novo greenstone belts, gave Archean zircon ages ranging between ca 2.6 and 3.3 Ga (Marinho, 1991; Cunha and Fróes, 1994; Mascarenhas and Silva, 1994; Peucat et al., 2002). The U–Pb available ages, Sm–Nd data and related references are reported in Table 1. Pioneer work (Rb–Sr and K–Ar ages) is reported in Table 1 of Cruz et al. (2012).

Paleoarchean terrains (ca 3.2–3.4 Ga, Table 1) were first recognized in the Sete Voltas and Boa-Vista/Mata Verde massifs, in the Contendas-Mirante region, which are all located on the south-eastern margin of the Gavião block (Martin et al., 1991, 1997; Nutman and Cordani, 1993). Felsic volcanics of the lower sequence of the Contendas-Mirante belt provided a TIMS zircon age of ca 3.30 Ga (Marinho, 1991, 1994). In gneisses from the western side of the Mundo-Novo belt (Fig. 1) in northern Gavião, Mougeot (1996) also obtained TIMS zircon ages ca 3.4 Ga. In a recent work, Dantas et al., 2010, reported a zircon age ca 3.5 Ga for a gneiss of the northern side of the São Francisco craton, close to Petrolina (Pernambuco).

In the studied Umburanas-Brumado-Aracatu region (UBA) (Fig. 2), several orthogneissic units crop out (Bernada, Aracatu, Eixo and Mariana massifs, Fig. 2) which have various sources and contrasted petrogenesis (Santos Pinto, 1996), their zircon evaporation ages range from the Paleo- to the Mesoarchean (Table 1). They are in tectonic contact with the Neoarchean Umburanas greenstone belt (Cunha and Fróes, 1994; Bastos Leal et al., 2003). Locally, the gneisses appear to be strongly metamorphosed during the Paleo-proterozoic event at ca 2.0 Ga. A short description of the main gneissic units is presented here and their geochemistry is summarized in Table 2. More details (petrography and geochemistry) are available in Santos-Pinto (1996) and Santos Pinto et al. (1998).

The Bernada gneissic Massif is a composite body made up of three main granodioritic to tonalitic Paleoproterozoic orthogneisses (Tables 1 and 2). The main mineralogy is composed of plagioclase An_{24–32} (45–50%), quartz (30–35%), K-feldspar (7–20%) and biotite (1–9%); the accessory phases include apatite, opaque minerals, sphene and zircon. The secondary minerals are epidote, sericite and carbonate. The granodiorites systematically have a K₂O/Na₂O ratio > 1, such that they belong to a high-K calc-alkaline suite and have a similar composition to the granite-granodiorite limit, whereas the tonalites are characterized by K₂O/Na₂O = 0.4. There are no available REE data on this suite.

The Aracatu Massif consists of two main Paleoproterozoic lithologies: granodioritic to trondhjemitic gray gneisses and granitic gneisses that display a strong foliation. Locally, these gneisses are intensively migmatized, such that they can appear as relicts or boudins in diatexitic migmatites to the SW of the town of Aracatu (Fig. 3). The trondhjemitic gneisses contain plagioclase An_{23–34} (45–55%), quartz (30–35%), K-feldspar (1–6%) and biotite (5–15%); the accessory phases are apatite, opaque minerals, sphene and zircon, while the secondary phases are epidote, sericite and carbonate. They possess typical TTG fractionated REE patterns: high light REE contents (La_N = 80–170) and very low heavy REE contents (Yb_N = 1–4), with slightly negative Eu anomalies (Eu*/Eu = 0.57–0.70). Quantitative geochemical modeling indicated that the trondhjemitic could result from 65 to 80% degree of partial melting of the old Sete Voltas-like gray gneisses (Martin et al., 1991) or alternatively, that they would be derived from the partial melting of

Fig. 1. Schematic geological map showing the limits, marginal fold belts and major structural units of the São Francisco Craton. GB – Gavião Block; JB – Jequié Block; SB – Serrinha Block; ISCB – Itabuna-Salvador-Curaçá Block. The rectangle indicates the studied area (adapted from Alkmim et al., 1993).

an Enriched Archean Tholeiite (EAT; *Condie, 1981*) that was previously contaminated by huge amounts of the old Sete Voltas gneiss-type. Located south of Aracatu (*Fig. 2*), the Mariana gneisses have a granitic K-rich calc-alkaline composition with predominant K_2O/Na_2O ratios ≥ 1 . They have the same zircon ages as the Aracatu trondhjemitic gneisses (up to 3.26 Ga; *Table 1*), with Nd model ages up to 3.6 Ga, which were interpreted as indicative of general crustal reworking processes in this area (*Santos Pinto et al., 1998*).

The Serra do Eixo Massif, to the north of Aracatu, is composed of two main gneissic units:

(1) - a K-rich calc-alkaline Mesoarchean augen-gneiss of monzogranitic composition, K_2O/Na_2O ratio ≤ 1 , zircon evaporation ages ranging from 2.80 to 3.16 Ga with Nd model ages ca 3.3 Ga, (*Tables 1 and 2*). (2) - an alkaline Neoarchean granitic gneiss, which also appears as a K-feldspar augen-gneiss. It is composed of K-feldspar (by up to 43%), plagioclase (18–23%) and biotite (<4%). The alkaline affinity is also shown by a K_2O/Na_2O ratio > 1 and $(Al_2O_3 + CaO)/(FeO_t + Na_2O + K_2O) < 1.4$ (*Sylvester, 1989*). They are very rich in LREE ($La_N = 114–263$) and have high HREE contents (Yb_N up to 25), which results in poorly fractionated patterns with $La/Yb_N = 10$. In addition, they display pronounced negative Eu anomalies ($Eu/Eu^* = 0.40–0.55$). This composition is similar to that of the ALK-3 type of Archean alkaline granites described in *Sylvester (1994)*.

They are interpreted as having been generated by 60% melting of the surrounding Mesoarchean high-K calc-alkaline augen-gneisses.

Two Paleoproterozoic granite massifs in the Umburanas-Brumado-Aracatu region (*Fig. 2, Table 1*) have been investigated. (1) The Umburanas Massif is a peraluminous biotite-muscovite granite with monazite, normative corundum $> 1\%$ and zircon with a morphology that is typical of S-type granites (*Pupin, 1980*). It shows low REE content with $La_N = 69–74$ and $Yb_N = 2.3–3$, with a small negative Eu anomaly ($Eu/Eu^* = 0.70$). The TIMS evaporation zircon ages range from 2.57 to 3.13 Ga, while the monazite ages range between 2002 ± 26 and 2049 ± 12 Ma. Geochemical modeling shows that this granite can be generated by 25% melting of the surrounding gneisses. (2) The Aracatu calc-alkaline granite (too small to be reported in *Fig. 2*) is a syeno- to monzogranite composed of plagioclase $An_{25–29}$ (15–37%), quartz (37–40%), microcline (23–48%) and biotite (<2.5%). The accessory phases are opaque minerals, sphene, apatite, zircon. Locally, this granite contains migmatitic and gneissic enclaves and schlierens. The REE patterns show $La_N = 47–61$, $Yb_N = 2.3–10$, $(La/Yb)_N = 4.7–21.7$ and negative Eu anomalies ($Eu/Eu^* = 0.64–0.73$). The single zircon ages lie between 1.95 and 2.26 Ga. Geochemical modeling suggests that it could be generated by 50% melting of the surrounding Aracatu Archean gray gneisses.

Fig. 2. Simplified geological map of the studied area (modified after Santos Pinto, 1996). White zones correspond to Proterozoic and Cenozoic covers and all the green areas to various greenstone belts. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Neoproterozoic Brazilian events are also recorded in the Gavião block with the Rb–Sr biotite-whole rock isochrons giving a 533 ± 11 Ma cooling age in the Serra do Eixo Massif and 508 ± 10 Ma in the Aracatu Massif (Santos Pinto et al., 1998). Similarly, Paleozoic ages were obtained by K–Ar dating in the Paleoproterozoic Rio do Paulo (507 ± 6 Ma), Espírito Santo (490 ± 12 Ma) and Igatemi (483 ± 5 Ma) granitic massifs (Bastos Leal et al., 2000).

3. New U–Pb results on zircon and monazite

During this work, we obtained four new SHRIMP U–Pb zircon dates. The ages are $^{207}\text{Pb}/^{206}\text{Pb}$ weighted averages of the concordant to sub-concordant analyses. They are given at $\pm 2\sigma$, using the Isoplot program (Ludwig, 2003, Table 3). In addition, we performed three in-situ U–Pb–Th analyses on the monazites (Aracatu gray-gneiss and Umburanas granite) using the EPMA and/or LA-ICPMS methods in order to better constrain the Paleoproterozoic evolution.

3.1. Zircon from a tonalitic gneiss of the Bernada Massif (sample BER-120.1)

The zircon grains are euhedral and belong to the S25 high temperature type described in Pupin (1980). They are strongly zoned as is the growth of zircon in a magmatic environment (Fig. 4). Seven analyses (out of twelve) are concordant to sub-concordant and define a weighted $^{207}\text{Pb}/^{206}\text{Pb}$ average of 3386 ± 9 Ma (MSWD = 2.5); this is interpreted as the age of the tonalitic precursor.

3.2. Zircon and monazite from trondhjemitic gray gneisses of the Aracatu Massif (samples ARA-78.1 and ARA-06.2)

The zircon grains are euhedral and belong to the S9, S13 and S18 crustal types described by Pupin (1980) (Fig. 5; Table 3). The images show complex structures with cores and overgrowths, both of which are strongly zoned. The youngest set of concordant to slightly discordant zoned overgrowths (i.e. gr. 8; Fig. 5) define a $^{207}\text{Pb}/^{206}\text{Pb}$ average age of 3325 ± 16 Ma (MSWD = 2.6) which is interpreted as the age of the magmatic precursor of the gneiss. A set of five cores (i.e. gr. 10; Fig. 4) exhibits a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3366 ± 15 Ma (MSWD = 2.9) and one other concordant core defines an age of 3487 ± 9 Ma (Fig. 5). These ages are interpreted as evidence of an inherited older crustal component in the granitic melt that emplaced at 3325 ± 16 Ma.

A Paleoproterozoic event is recorded in a second sample of these trondhjemitic gneisses (ARA-06.2, Table 1). One monazite that grew around an apatite crystal has been dated using the EPMA method. Based on a set of 200 analyses, a large range of ages is observed between 800 Ma and 2100 Ma, indicating significant lead loss in this grain (the figure is not included here). A set of 43 analyses provides an age of 1922 ± 25 Ma. The high MSWD (3.4) shows that this dataset cannot be considered as a single age population (Fig. 6a). Finally, a limited domain exhibits a population of the nine oldest analyses producing a homogeneous result of 2063 ± 33 Ma (Fig. 6b). In Fig. 6b, the error ellipses define the best-fit line (regression line), which is quite close to the theoretical

Table 2
Representative analyses and main magmatic criteria for the Archean gneisses and the Paleoproterozoic granites of the Umburanas-Brumado-Aracatu area, after Santos Pinto (1996).

Sample	Bernada massif		Aracatu/Mariana massifs		Serra de Eixo massif		Umburanas massif	
	BER120-1	BER120-2	ARA 78-1	MAR94-1	ARA 170	VZP65	GA 299	UMB 165
Rock type	Tonalitic gneiss	Granodioritic to granitic gneiss	Trondhjemitic gneiss	Monzogranite gneiss	Monzogranite	Monzogranitic gneiss	Alkaline granitic gneiss	Granodiorite to biot-musc granite
Longitude (°W)	41 14 36	41 14 36	41 27 32	41 23 33	41 34 41	41 21 10	41 25 21	41 26 23
Latitude (°S)	14 16 37	14 16 37	14 25 44	14 34 13	14 17 15	14 16 30	14 17 10	14 07 11
SiO ₂ (wt%)	74.2	74.4	72.6	70.0	74.4	72.1	74.0	71.4
TiO ₂	0.39	0.22	0.2	0.29	0.13	0.22	0.39	0.24
Al ₂ O ₃	13.0	13.9	15.4	15.7	14.4	15.6	11.9	15.8
Fe ₂ O ₃	3.4	1.7	1.7	2.1	1.4	2.0	4.7	2.1
MnO	0.07	0.04	0.03	0.04	0.06	0.03	0.05	0.07
MgO	0.6	0.4	0.4	0.8	0.6	0.5	0.3	0.9
CaO	2.2	1.4	2.1	2.2	0.6	2.2	1.1	1.9
Na ₂ O	4.4	4.9	4.9	4.7	3.2	5.2	2.4	4.4
K ₂ O	1.7	2.13	2.74	4.02	5.2	2.1	5.1	3.1
P ₂ O ₅	0.06	0.05	0.5	0.16	0.06	0.06	0.07	0.06
LOI	0.76	0.55	0.46	0.42	0.45	0.70	0.57	0.42
Total	101	101	101	100	100	101	100	100
La (ppm)	—	—	24	65	19	70	115	16.0
Ce	—	—	56	138	41	136	255	34.3
Pr	—	—	—	16	—	14	—	—
Nd	—	—	25	61	18	49	100	14.2
Sm	—	—	2.9	14	1.8	6.2	18	1.9
Eu	—	—	0.6	1.4	0.3	1.5	2.3	0.6
Gd	—	—	2.1	13	1.7	4.7	16	1.6
Tb	—	—	—	2.5	—	0.5	—	—
Dy	—	—	1.2	17	1.2	2.2	18	1.1
Ho	—	—	0.2	3.8	0.2	0.4	3.5	0.2
Er	—	—	0.5	11	0.6	0.9	9.3	0.5
Tm	—	—	—	1.8	—	0.1	—	—
Yb	—	—	0.4	12	0.6	0.8	7.4	0.5
Lu	—	—	0.1	1.7	0.1	0.1	0.9	0.1
Eu/Eu*	—	—	0.7	0.3	0.6	0.82	0.4	0.99
LaN/YbN	—	—	39.1	3.7	22.2	61.9	10.4	22.5
K ₂ O/Na ₂ O	0.38	1.06	0.56	0.86	1.6	0.40	2.12	0.7
A/CNK	0.99	1.04	1.03	0.98	1.21	1.05	1.04	1.12
A/CNK-A/NK	Metaluminous	Peraluminous	Peraluminous	Metaluminous	Peraluminous	Peraluminous	Peraluminous	Peraluminous
Magmatic serie	Calc-alkaline	Calc-alkaline	Trondhjemitic	High-K calc-alkaline to alkaline	High-K calc-alkaline	Calc-alkaline	Alkaline	High-K calc-alkaline
Epsilon Nd (t)	—	(+3.7)	(-1)	(+1 to -1)	(-15)	0	(-3)	(-15)
Inherited zircons	—	No	3.37 & 3.5 Ga	—	Up to 3.25 Ga	—	No	Up to 3.0 Ga
Possible source	Partial melting of a deep mafic source	Crustal reworking or contamination of a juvenile source	Crustal reworking or contamination of a juvenile source	Crustal reworking	Crustal reworking	Crustal reworking or contamination of a juvenile source	Crustal reworking	Crustal reworking
Geochemical modeling	—	Partial melting of old Sete Voltas gneiss or of a EAT amphibolite with high rate of crustal contamination	Partial melting of juvenile Archean crust	Partial melting of juvenile Archean crust	Partial melting of Aracatu gray gneiss	Partial melting of grt amphibolite with crust. contamination or Aracatu gray gneiss melting	Partial melting of S. Eixo granitic gneiss type	Partial melting of Archean tonalitic gneisses

Fig. 3. Relics of Archean boudined gneisses within diatexitic Paleoproterozoic migmatites, close to the town of Aracatu.

isochron, and makes the mean age of 2063 Ma, calculated at the centroid of the population significant (Cocherie and Albarède, 2001). These data demonstrate the Paleoproterozoic age for the crystallization of monazite, disrupted by the partial opening of the U–Th–Pb system up to 800 Ma.

3.3. Zircon from the alkaline granitic gneiss of the Serra de Eixo (sample GA-299)

Zircon grains are euhedral, strongly zoned magmatic grains and have a typical high temperature alkaline morphology (types S24 and S25 in Pupin, 1980, Fig. 7 and Table 3). The major set (13/15) of data is concordant to sub-concordant and defines a weighted average of $^{207}\text{Pb}/^{206}\text{Pb}$ ages at 2693 ± 5 Ma (MSWD = 0.98), which is interpreted as the age of the alkaline magmatic protolith.

3.4. Zircon from the Aracatu granite (sample ARA-170)

Zircon grains are euhedral, brown, with a high U content and are partially or totally metamict; nevertheless, some cores and overgrowths were preserved (Fig. 8; Table 3). They belong to the S4, S5 and S9 types described in Pupin (1980), which this author considered as characteristic of anatectic granites. Despite a careful selection of the analysis sites, several grains have a high common lead content and are strongly discordant, resulting in high errors on the ages (spots 1.2 and 3.1; Table 2). The core of grain 1 is slightly discordant and gives a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3245 ± 25 Ma (spot 1.1), which fixes a minimum age for the inheritance from the Archean gneissic source. Grain 2 is an overgrowth with a relatively low common lead content (Table 3); it is discordant (13%) and has a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2036 ± 19 Ma (Table 1). The entire dataset for the overgrowths defines an inaccurate discordia intercept at 2058 ± 89 Ma which nevertheless records the Paleoproterozoic history of this granite.

3.5. Monazite from the Umburanas biotite-rich granite

Monazite ages were obtained using the EPMA and LA-ICPMS methods. The monazite grains are anhedral and more or less round, with limited zoning when observed in back-scattered (BS) images (Fig. 10). Some inclusions, probably of thorite, are visible (white zones in grain 1). The whole set of EPMA ages (170 analyses) obtained from nine monazite grains provides a weighted average of

1962 ± 13 Ma (MSWD = 1.5) (Fig. 9a). A set of 33 analyses, that is significantly older than the previous set, defines an isochron age of 2053 ± 17 Ma (MSWD = 1.2; Fig. 9b). This age is similar to the oldest one obtained by TIMS from the same sample (2049 ± 12 Ma; Table 1). In order to try to understand the range of ages observed (ca 90 Ma) in this sample, four of the monazite crystals were analyzed by the LA-ICPMS isotopic method (Table 4). A set of 30 analyses is concordant to slightly discordant and provides a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1971 ± 10 Ma (MSWD = 0.98). Two analyses are discordant and significantly older: spot (9.8) with a $^{207}\text{Pb}/^{206}\text{Pb}$ minimum age of 2200 ± 50 Ma and spot (5.7) at 2492 ± 70 Ma. They indicate the presence of inherited zones in the monazite grains, even if no specific structure was visible on the BS images (Fig. 10). These inherited zones could also account for some older ages (ca 2.05 Ga) obtained by the EPMA & TIMS evaporation methods. Consequently, the set of youngest ages obtained by (1) TIMS at 2002 ± 26 Ma, (2) the whole set of EPMA ages at 1962 ± 13 Ma, and (3) the LA-ICPMS method at 1971 ± 10 Ma, are probably the most significant; the latter age is retained as the age of the anatectic Umburanas granite.

4. Discussion

As expected, in terrains that underwent multiple tectono-metamorphic events, the SHRIMP zircon ages are significantly older than the TIMS evaporation single zircon ages. During this work, we observed that the range of TIMS ages is broadly underestimated by approximately 1.0–2.8% compared to the SHRIMP ages. On the other hand, this work allowed to define three major Paleo- and Neoproterozoic magmatic events in the gneisses of the UBA region at ca 3.4, 3.3 and 2.7 Ga and to recognize the existence of a proto-crust up to 3.5 Ga.

4.1. The Paleoproterozoic events

These events might be correlated with those evidenced in the Contendas-Mirante gneisses (Table 1). The 3386 ± 9 Ma plutonic age of the Bernada gray gneiss is the same as that of the Sete Voltas “old gneisses” dated by SHRIMP at 3403 ± 5 and 3392 ± 21 Ma (Table 1 and references therein). Both series have been interpreted as TTG suites derived from the partial melting of a mafic source leaving a garnet-bearing residue, which accounts for the low HREE content and the strong fractionated REE patterns (Martin et al., 1997). Consequently, both the Bernada gray gneisses and Sete Voltas old gneisses are considered as a juvenile input to the Archean continental crust that took place ca 3.4 Ga ago. The positive ϵ_{Nd} value (+3.7 at 3386 Ma; Tables 1 and 2 and No. 2 in Fig. 11) for the Bernada gneisses is in good agreement with a juvenile origin from a depleted mantle at ca 3.4 Ga. This is the most positive ϵ_{Nd} value ever obtained for gneisses of the Gavião block. It constitutes important evidence of the existence of a depleted mantle in this region during the Paleoproterozoic. By contrast, the old Sete Voltas gneisses exhibit negative ϵ_{Nd} values (between –1 and –3 at 3.4 Ga; Martin et al., 1997– No. 1 in Fig. 11), suggesting that these rocks contain an older crustal component. This is constrained by the presence of a concordant inherited zircon core dated at 3466 ± 8 Ma by Nutman and Cordani (1993).

The Aracatu gray gneisses give a significantly younger age at 3325 ± 16 Ma (No. 3 in Fig. 11), which could probably be correlated with the 3353 ± 5 Ma age of the Boa Vista/Mata Verde gneisses (Nutman and Cordani, 1993). These ages are also in the range of the ca 3.30 Ga age of the Lagoa da Macambira gneisses dated by Bastos Leal et al. (1998) in the adjacent Brumado-Caculé-Caetité region. The Aracatu gray gneisses have a negative ϵ_{Nd} value (–1 at 3325 Ma) and contain inherited zircon cores dated at 3371 ± 14 Ma,

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Radiogenic ratios			ρ	Age (Ma)	% Conc						
							²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb			²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²⁰⁶ Pb					
Sample BER120-1																		
1.1	499	104	0.21	110	0.002463	3.19	0.2478	0.0044	7.61	0.21	0.2227	0.0047	0.64	1427	23	3000	34	48
2.1	369	244	0.66	220	0.000023	0.03	0.6954	0.0110	27.26	0.44	0.2843	0.0009	0.98	3403	42	3387	5	100
3.1	154	41	0.26	41	0.001140	1.40	0.3042	0.0053	11.12	0.23	0.2651	0.0031	0.83	1712	26	3277	18	52
4.1	192	98	0.51	110	0.000010	0.01	0.6670	0.0087	26.19	0.36	0.2848	0.0014	0.94	3294	34	3389	7	97
5.1	295	133	0.45	173	0.000057	0.07	0.6824	0.0084	26.64	0.34	0.2832	0.0011	0.96	3354	32	3381	6	99
6.1	105	50	0.48	60	—	<0.01	0.6635	0.0098	26.29	0.42	0.2873	0.0017	0.93	3281	38	3403	9	96
7.1	193	100	0.52	109	0.000017	0.02	0.6558	0.0086	25.48	0.35	0.2818	0.0013	0.95	3251	33	3373	7	96
8.1	304	148	0.49	137	0.000057	0.07	0.5220	0.0068	19.31	0.27	0.2683	0.0014	0.93	2708	29	3296	8	82
8.2	347	195	0.56	178	0.000005	0.01	0.5970	0.0237	22.35	0.89	0.2715	0.0009	1.00	3018	96	3315	5	91
9.1	265	184	0.69	155	0.000019	0.02	0.6818	0.0087	26.94	0.36	0.2866	0.0011	0.96	3351	34	3399	6	99
10.1	179	73	0.41	96	0.000126	0.15	0.6226	0.0091	23.79	0.37	0.2771	0.0015	0.93	3120	36	3347	9	93
11.1	313	157	0.50	183	0.000006	0.01	0.6810	0.0084	26.54	0.34	0.2826	0.0010	0.96	3348	32	3378	6	99
Sample ARA78																		
1.1	273	47	0.17	150	0.000029	0.03	0.6389	0.0080	23.99	0.33	0.2723	0.0014	0.93	3185	32	3319	8	96
1.2	550	394	0.72	235	0.000362	0.44	0.4956	0.0058	18.60	0.24	0.2722	0.0015	0.90	2595	25	3319	9	78
2.1	335	67	0.20	162	0.000389	0.48	0.5608	0.0094	20.53	0.36	0.2655	0.0015	0.95	2870	39	3279	9	88
2.2	281	86	0.31	144	0.000241	0.29	0.5947	0.0085	23.70	0.37	0.2890	0.0017	0.92	3009	34	3412	9	88
3.1	115	49	0.43	65	0.000110	0.13	0.6564	0.0097	24.41	0.39	0.2697	0.0016	0.93	3253	38	3304	9	98
4.1	442	50	0.11	213	0.000026	0.03	0.5603	0.0067	17.87	0.22	0.2313	0.0008	0.96	2868	28	3061	6	94
4.2	219	82	0.37	135	0.000001	<0.01	0.7160	0.0127	29.94	0.56	0.3033	0.0019	0.95	3481	48	3487	9	100
5.1	261	69	0.26	134	0.000294	0.36	0.5961	0.0164	22.51	0.63	0.2738	0.0014	0.98	3014	66	3328	8	91
6.1	267	131	0.49	151	—	<0.01	0.6581	0.0083	25.23	0.33	0.2780	0.0011	0.95	3260	32	3352	6	97
7.1	107	30	0.28	65	—	<0.01	0.7064	0.010

Fig. 4. Concordia diagram with cathodo-luminescence (CL) and transmitted light (TL) images for magmatic zircon grains of a trondhjemitic gneiss from the Bernada Massif (sample BER120-1). SHRIMP analyses and error ellipses are reported at 1σ . The spot size reported is around 30 μm .

which are within the time range obtained for the emplacement of the Bernada gneiss (3386 ± 9 Ma). Moreover, these results are in agreement with the geochemical modeling carried out by Santos Pinto (1996) which showed that the Aracatu trondhjemite could have been generated by a high degree of melting of the old Sete Voltas gray gneiss-type (Table 2). The 3487 ± 9 Ma age measured on one zircon core from the Aracatu gneisses is one of the oldest age reported in the Gavião block so far; it is a new evidence of the occurrence of a ~ 3.5 Ga old proto-crust in this region.

4.2. The existence of a Neoproterozoic magmatism in the UBA region

A Neoproterozoic magmatic event is confirmed by the occurrence of the alkaline granitic gneisses of Serra do Eixo that emplaced 2693 ± 5 Ma ago. This age is similar to that recently obtained for the syenites and syeno-granites of the Caraguatá Massif (Table 1; Cruz

Fig. 6. a) Weighted average age calculated using the EPMA technique applied to one single monazite grain from the ARA 06-2 gneiss sample. b) isochron plot of Th/Pb vs. U/Pb for the oldest domain of the monazite grain from the ARA6-2 gneiss sample.

Fig. 5. Concordia diagram with cathodo-luminescence (CL) images for magmatic zircons of a trondhjemitic gneiss from the Aracatu Massif (sample ARA78-1). SHRIMP analyses and error ellipses are reported at 1σ . The spot size reported is around 30 μm .

et al., 2012), 60 km to the north of the UBA region, in the Chapada Diamantina (Fig. 2). It is also in the same age range as that obtained for the Pé de Serra granitic to syenitic pluton (U–Pb SHRIMP age of 2652 ± 11 Ma; Marinho et al., 2008) in the eastern part of the Contendas-Mirante belt, (Fig. 2). At 2.7 Ga, the Eixo gneisses had a ε_{Nd} value ca -3 , whereas it ranges between -4 and -6 in the Caraguatá syenite (Cruz et al., 2012). These negative ε_{Nd} values and the corresponding model age (3.28 Ga, Table 1) indicate that the ca 2.7 Ga alkaline magmatism is primarily the result of crustal reworking processes from the surrounding Paleoproterozoic gneisses. Furthermore, the felsic volcanics that emplaced 2.74 Ga ago in the Umburanas greenstone belt (Table 1) are characterized by $\varepsilon_{\text{Nd}} = -4$, which is in agreement with a crustal input. However, another sample from the same volcanics has a ε_{Nd} value equal to $+1.5$, which also suggests some juvenile input from a depleted mantle ca 2.7 Ga (Bastos Leal et al., 2003).

4.3. What is the age of the proto-crust in the Gavião block?

The occurrence of inherited zircon grains with ages up to 3.5 Ga indicates the existence of an old continental crust that was

Fig. 7. Concordia diagram with cathodo-luminescence (CL) and transmitted light (TL) images for magmatic zircons of an alkaline gneiss from the Serra de Eixo Massif (sample GA 299). SHRIMP analyses and error ellipses are reported at 1σ . The spot size reported is around 30 μm .

probably totally destroyed and recycled during the main Paleo-archean events that took place between 3.3 and 3.4 Ga. The Nd isotopes also provide further evidence of the recycling of an older proto-crust. Fig. 11 reports the $\epsilon_{\text{Nd}t}$ values for the Archean rocks versus their zircon ages. Most of these rocks have negative $\epsilon_{\text{Nd}t}$ values, suggesting that they either had a significant crustal residence time before their formation or they incorporated an older crustal component. The Nd model ages (Table 1) range between 3.3 and 3.7 Ga in the UBA region (Santos Pinto, 1996), 3.5 and 3.7 Ga in the Sete Voltas gneisses (Martin et al., 1997) and 3.3 and 3.7 Ga in the Brumado-Caculé-Caetité region (Bastos Leal et al., 1998).

On the other hand, it should be noted that the model age (3.27 Ga; Table 1) obtained for the Bernada gneisses (No. 2 in Fig. 11) is slightly younger than its zircon age (3.39 Ga). In order to

Fig. 8. Concordia diagram with cathodo-luminescence (CL) and transmitted light (TL) images for magmatic zircon grains from a peraluminous granite of the Aracatu Massif (sample ARA 170). SHRIMP analyses and error ellipses are reported at 1σ . The spot size reported is ca 30 μm .

Fig. 9. a Weighted average age calculated using the EPMA technique applied to nine monazite grains from the Umburanas granite (sample UMB 165). b Isochron plot of Th/Pb vs. U/Pb (EPMA) for grains 7 and 9 from sample UMB 165.

Fig. 10. Concordia diagram with back-scattered images for monazite grains from the Umburanas granite (sample UMB 165). LA-ICPMS analyses and error ellipses are reported at 1σ . The spot size reported is 7 μm .

Table 4

Summary of the monazite LA-ICPMS data.

Grain spot	U (ppm)	Th (ppm)	U/Th	Pb (ppm)	Radiogenic ratios				ρ	Age (Ma)				Conc. %
					$^{207}\text{Pb}/^{235}\text{U}$	$\pm 1\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm 1\sigma$		$^{206}\text{Pb}/^{238}\text{U}$	$\pm 1\sigma$	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm 1\sigma$	
1.1	58294	874	67	4211	5.429	0.088	0.3236	0.0037	0.71	1807	18	1981	28	91
1.2	51689	1050	49	4995	5.835	0.089	0.3468	0.0039	0.74	1920	19	1986	26	97
1.3	67417	1188	57	6033	5.386	0.084	0.3227	0.0037	0.73	1803	18	1972	27	91
1.4	55110	1182	47	5328	5.798	0.090	0.3441	0.0039	0.73	1906	19	1989	27	96
1.5	47353	626	76	4534	5.845	0.097	0.3528	0.0041	0.70	1948	19	1959	29	99
1.6	47978	601	80	4663	5.932	0.099	0.3546	0.0041	0.69	1957	19	1977	29	99
1.7	47557	551	86	4444	5.916	0.104	0.3554	0.0042	0.67	1960	20	1968	30	100
1.8	53322	649	82	5071	5.836	0.101	0.3529	0.0041	0.67	1948	20	1956	30	100
5.1	48016	629	76	3936	5.429	0.101	0.3214	0.0038	0.64	1796	19	1994	32	90
5.2	40670	696	58	3733	5.578	0.102	0.3334	0.0039	0.64	1855	19	1976	31	94
5.3	42889	529	81	4066	5.914	0.112	0.3554	0.0042	0.63	1960	20	1967	33	100
5.4	45624	677	67	4295	5.751	0.108	0.3455	0.0041	0.63	1913	20	1967	32	97
5.5	64647	445	145	5419	5.976	0.124	0.3465	0.0043	0.59	1918	20	2030	36	94
5.6	55636	470	118	5141	6.200	0.124	0.3586	0.0043	0.60	1976	21	2035	34	97
5.7	59316	259	229	4662	9.591	0.203	0.4256	0.0053	0.59	2286	24	2492	35	92
5.8	58694	862	68	4589	5.670	0.122	0.3273	0.0041	0.58	1825	20	2038	37	90
7.1	20368	699	29	2046	5.797	0.083	0.3529	0.0040	0.78	1948	19	1944	25	100
7.2	21180	697	30	2152	5.804	0.083	0.3529	0.0040	0.78	1948	19	1946	25	100
7.3	24232	658	37	2421	5.857	0.085	0.3552	0.0040	0.77	1959	19	1951	25	100
7.4	17418	332	52	1699	6.186	0.097	0.3572	0.0041	0.74	1969	19	2038	27	97
7.5	19565	504	39	1996	5.900	0.097	0.3584	0.0042	0.71	1975	20	1948	29	101
7.6	21909	714	31	2244	5.903	0.089	0.3553	0.0040	0.75	1960	19	1964	26	100
7.7	22040	909	24	2346	5.918	0.086	0.3596	0.0040	0.77	1980	19	1947	25	102
7.8	26715	795	34	2562	5.674	0.089	0.3350	0.0038	0.73	1863	19	1999	27	93
9.1	30358	2668	11	3655	5.956	0.088	0.3611	0.0042	0.78	1987	20	1941	25	102
9.2	30844	755	41	3167	6.066	0.095	0.3601	0.0043	0.75	1983	20	1979	26	100
9.3	40588	890	46	4013	5.884	0.091	0.3558	0.0042	0.76	1962	20	1946	26	101
9.4	30839	717	43	3160	5.889	0.094	0.3586	0.0042	0.74	1976	20	1934	27	102
9.5	30675	2656	12	3730	6.022	0.089	0.3605	0.0042	0.78	1984	20	1965	25	101
9.6	36273	737	49	3606	6.144	0.099	0.3624	0.0043	0.74	1994	20	1991	27	100
9.7	36033	775	47	3534	5.893	0.098	0.3556	0.0043	0.72	1961	20	1951	29	101
9.8	30771	898	34	3347	7.417	0.113	0.3885	0.0045	0.77	2116	21	2200	25	96

obtain a model age that is equal to the zircon age, we need to use a more depleted mantle at 3.39 Ga with a present-day $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of = 0.513235 (epsilon zero value of +11.6). Using this mantle as a model for the other gneisses, the set of model ages becomes older by ca 100 Ma, ranging between 3.4 and 3.8 Ga. Consequently, the existence of a crust older than 3.4 Ga is also shown by the Nd

isotopes and we cannot rule out finding zircon xenocrysts or detrital zircon grains, or even rocks, giving ages of 3.6 Ga as in the Serrinha block (Rios et al., 2008) or up to 3.8 Ga, as recorded in detrital zircon grains from sediments of the Quadrilátero Ferrífero region, in the southern São Francisco Craton (Hartmann et al., 2006).

Fig. 11. Nd vs. time isotopic evolution diagram for the gneisses from the Gavião Block, which are dated using U–Pb zircon techniques (references in Table 1). SHRIMP (1–6) and LA-ICPMS (7) ages = large yellow circles: 1- Sete Votas 3.4 Ga gneiss, 2- Bernada 3.4 Ga gneiss, 3- Aracatu 3.3 Ga gneiss, 4- Mundo Novo 3.3 Ga meta-dacite, 5-Sete Voltas/Lagoa do Morro 3.2 Ga gneisses, 6- Serra de Eixo 2.7 Ga alkaline gneisses, 7- Caraguatá 2.7 Ga meta-syenite. TIMS evaporation (minimum) ages = small black circles: a- Mairi 3.0 Ga gneiss, b- Sete Voltas 3.2 Ga gneiss, c- Aracatu 3.2 Ga gneiss, d- Mariana 3.3 Ga gneiss, e- Serra de Eixo 3.2 Ga gneiss, f- Umburanas dacite, g- Lagoa Macambira 3.3 Ga gneiss, h- Lagoa Macambira 3.2 Ga gneiss, i- Lagoa Macambira 3.1 Ga granite. The error envelope is reported for the dated gneisses and felsic volcanics in the Gavião block (references in Table 1).

4.4. Paleoproterozoic metamorphism and the related magmatism

A ca 2.0 Ga event is recorded by the monazite (EPMA ages ranging from 1922 ± 25 to 2063 ± 33 Ma) in the 3.33 Ga Aracatu gneisses. They indicate that a high temperature metamorphism, reaching migmatization, affected the Archean rocks during the Paleoproterozoic. A related magmatic event is shown by the SHRIMP zircon age obtained for the Aracatu granite; even if it is imprecise (2053 ± 65 Ma), it is still within the range of the monazite ages. The occurrence of an Archean core (ca 3.25 Ga) in one zircon is in agreement with the Nd model ages for all the peraluminous granites (3.10–3.77 Ga; Table 1) and shows that these granites result from the partial melting of the Archean surrounding gneissic basement. The La-ICPMS monazite age obtained for the Umburanas granite at 1971 ± 10 Ma must also be related to the ca 2.0 Ga crustal melting event obtained on similar granites (Santos Pinto et al., 1998; Bastos Leal et al., 2000).

4.5. Occurrences of paleoarchean relicts in the South America cratons and in Western Africa

Before the opening of the South Atlantic Ocean, South America and Africa constituted a single continental mass. The Amazonian Craton was connected with the West Africa Craton and the São Francisco Craton was connected with the Congo Craton (ref. below).

Relicts of Paleoproterozoic terranes remain extremely scarce in these cratons, as it will be noticed and discussed here. A general figure showing the location of the main cratons in South America can be found in Hartmann et al. (2001). Cordani (2003) provides a similar figure showing the relationship with the Congo Craton.

In the northern part of South America, the high-grade terranes of Imataca in Venezuela recorded main crustal stages ca 2.7–3.0 Ga and up to 3.2 Ga (Tassinari et al., 2004). Until now, the pioneer work of Montgomery and Hurley (1978), with ages ca 3.4–3.7 Ga, has not been confirmed. Archean rocks are also described in the central Amazonian province, in the Carajás region and south of the Amapá region (Brazil). The main stages of crustal growth are ca 2.9–3.0 Ga and 3.2–3.3 Ga, respectively, based on zircon ages and Nd isotopes (Althoff et al., 2004; Rosa-Costa et al., 2006; Oliveira et al., 2009; Almeida et al., 2011 and references therein). Detrital zircon grains from a quartzite provided SHRIMP ages up to 3.4–3.7 Ga, which indicate the existence of Paleoproterozoic rocks in the Carajás province (Macambira et al., 1998). SHRIMP ages up to 3.4–3.5 Ga (Rino et al., 2003), obtained for zircon grains from sands collected at the mouth of the Amazon River, also constrain the occurrence of Paleoproterozoic rocks in the Amazonian Craton.

The West Africa Craton is related to the Amazonian Craton in geodynamical reconstructions (i.e. Cordani, 2003; Cordani and Teixeira, 2007 and references therein). It also contains Paleoproterozoic relicts as in the Reguibat Rise (Mauritania). SHRIMP zircon ages of orthogneisses are ca 3.5–3.6 Ga (Potrel et al., 1996) and the main crust was formed ca 2.9–3.2 Ga (Chardon, 1996; Potrel et al., 1998; Key et al., 2008). In the Man Rise (Guinea), tonalitic gneisses and granulitic gabbros ca 3.5–3.6 Ga are also described (SHRIMP zircon ages in Thiéblemont et al., 2001; Barth et al., 2002). The main crust was developed ca 2.8–3.3 Ga (Kouamelan et al., 1997) in the Ivory Coast. Between the West Africa Craton and the Congo Craton, the Benin-Nigeria shield, which corresponds to the Pan-African mobile belt, also contains relicts of Paleoproterozoic age (Kudan gneisses ca 3.57 Ga; Kröner et al., 2001).

In Brazil, Archean relicts were also recognized in the São José do Campestre massif located to the north of the São Francisco Craton, in the Borborema Province close to the city of Natal (Fig. 1). Relicts of tonalitic gneisses exhibit the oldest zircon ages between 3.4 and 3.5 Ga with Nd T_{DM} ages > 3.7 Ga (Dantas et al., 2004). Other felsic gneisses, including reworked and juvenile crusts, are 3.0–3.3 Ga old. Late Archean syeno-granites are emplaced at around 2.7 Ga. These rocks are also involved in Paleoproterozoic accretionary orogenic processes ca 2.2–2.0 Ga. This sequence of ages is rather similar to the sequence recognized in the Gavião block and our results reinforce the correlation suggested by Dantas et al. (2004) between the São José do Campestre Massif and the São Francisco Craton. In the São Francisco Craton, the oldest Paleoproterozoic gneisses and relicts are between 3.4 and 3.8 Ga, as detailed above in this article. In Uruguay, the Rio de la Plata Craton contains Paleoproterozoic relicts with a zircon age for a metatonalite ca 3.41 Ga in the Nico Perez terrane (Hartmann et al., 2001). In Argentina, a xenocryst of zircon, also dated at ca 3.41 Ga, was found in the Patagonian batholith (Rolando et al., 2002).

The Congo Craton is generally correlated to the São Francisco Craton because of: (1) the direct connection between the cratons when the South Atlantic Ocean was re-closed, and (2) similarities between their Archean and Paleoproterozoic evolutions. Several papers have already discussed this correlation (i.e. Alkmim et al., 2006; De Waele et al., 2008; Cordani et al., 2009). Juvenile magmatism occurred ca 2.9–3.0 Ga in the NW part of the Congo Craton (Toteu et al., 2001; Poulet et al., 2007; Tchameni et al., 2010). A large alkaline crustal magmatism, derived from the partial melting of surrounding TTG suites, developed ca 2.7 Ga and exhibits model ages between 2.9 and 3.4 Ga (Tchameni et al., 2000; Shang et al.,

2010); it reinforces the similarities between the two cratons. In the central part of the Congo Craton, the oldest zircon grains from sands of rivers (Batumike et al., 2009) indicate Archean ages ca 2.5–2.9–3.2 Ga, with the oldest age at 3.6 Ga. According to Barbosa and Sabaté (2004), the Archean Gavião block (Brazil) and the Chaillu Massif in the NW Congo Craton occupy symmetric and analogous positions in the accretionary build-up, both acting as uplifted crustal segments producing the Paleoproterozoic Jacobina–Contendas and Francevillian Basins, which are also in symmetric positions. Moreover, the subduction of the continental crust led to the crustal thickening of both Paleoproterozoic and Archean units ca 2.1–2.0 Ga in the São Francisco and in NW Congo cratons.

5. Conclusion

The geological history of the Gavião block started by significant juvenile magmatism emplaced ca 3.4 Ga. This magmatism could be purely juvenile (Bernada gneisses) or could incorporate variable amounts of older crustal materials (Sete Voltas old gneisses), which provides evidence for the occurrence of a proto-crust (≥ 3.5 Ga). While emplaced and stabilized, this crust began to be actively recycled either directly or through the sedimentary cycle. Such reworking events could have taken place between 3.3 and 3.2 Ga ago (Aracatu gneisses & Sete Voltas young gneisses) and during the Neoproterozoic (~ 2.7 Ga, Serra do Eixo & Caraguatá alkaline gneisses). The Paleoproterozoic melting of the Archean crust is probably related to the major crustal thickening process following the ca 2.1 Ga collision on the eastern side of the Gavião block with the Jequié granulites.

Relicts ca 3.4–3.5 Ga or older are exceptional in both the South American Cratons as well as in the Western Africa Archean Cratons. They provide evidence for proto-crusts up to 3.8 Ga old which are often totally destroyed during the main accretional processes observed in the cratons and developed between ca 3.0 and 3.4 Ga. The Gavião block in the São Francisco Craton constitutes an important region in terms of the study of early crustal evolution in South America.

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Appendix

Zircon analyses were performed using the SHRIMP II at Canberra, ANU, Australia, following the methods given in Williams (1998). Uncertainties for the individual analyses (ratios and ages) are given at the 1σ level (Table 3), but the uncertainties in calculated weighted mean ages (Ludwig, 2003) are reported at 2σ . Two samples of monazite (EPMA dating) were analyzed using a Cameca SX 50 electron microprobe (BRGM, Orléans, France). The analytical procedure is detailed in Cocherie et al. (1998) and Cocherie and Albarède (2001). U–Th–Pb geochronology of monazite was conducted by LA-ICPMS at the UMR 6524, Clermont-Ferrand (France). Data are corrected for U–Pb fractionation occurring during laser sampling and for instrumental mass discrimination (mass bias) by standard bracketing with repeated measurements of Moacyr

monazite standard (Seydoux-Guillaume et al., 2004). Ages were generated using the Isoplot/Ex software package by Ludwig (2003). The U–Th–Pb concentrations were calibrated relative to the certified contents of the Moacyr monazite standard. More details are reported in Hurai et al. (2010). The Nd model ages were calculated using a linear model of depleted mantle from 4.54 Ga with a present-day $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.51315 (epsilon zero value of +10) and a $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.2137 (DePaolo, 1988, Table 1).

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