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

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ABSTRACT

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

1. Introduction

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

In Brazil, diamictites are remarkably abundant in the western domain of the Araçuaí orogen, located to the east of the São Francisco craton (Fig. 1). They belong to the Macaúbas Group (Fig. 2), a thick volcano-sedimentary succession that represents a precursor basin of the Araçuaí orogen, and includes pre-glacial fluvial to marine sediments, mostly glacio-marine diamictites with minor rift-related volcanic rocks, and post-glacial passive margin sediments and oceanic deposits (Karfunkel and Hoppe, 1988; Pedrosa-Soares et al., 1992, 2001, 2008; Uhlein et al., 1999). In its original definition (e.g., Moraes and Guimarães, 1931), the name Macaúbas should be exclusively applied to refer to the extensive unit with diamictite formations that underwent regional deformation and metamorphism within the Araçuaí orogen. But diamictites also occur in relatively small areas on the São Francisco craton (Fig. 1), where the best evidence of glaciation has been reported from the Jequitaí and Bebedouro Formations. These include striated and grooved pavements, glacio-terrestrial sediments, faceted and striated flat-iron-shaped cobbles and pebbles, dropstones, and esker-like bodies (e.g., Figueiredo et al., 2009; Gravenor and Monteiro, 1983; Hettich and Karfunkel, 1978; Isotta et al., 1969; Karfunkel et al., 2002; Karfunkel and Hoppe 1988; Martins-Neto et al., 1999, 2001; Martins-Neto and Hercos, 2002; Montes et al., 1985; Rocha-Campos and Hasui 1981; Uhlein et al., 1999; Viveiros and Walde 1976). The Carrancas diamictite occurs in only a few outcrops and boreholes where it is covered by the Sete Lagoas Formation (Kuchenbecker et al., 2010; Schöll, 1972; Tuller, 2009). This diamictite forms the lowermost unit of the pelite–carbonate cratonic cover (Bambuí Group), which includes a cap carbonate dated at 740 ± 22 Ma (Babinski et al., 2007).

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

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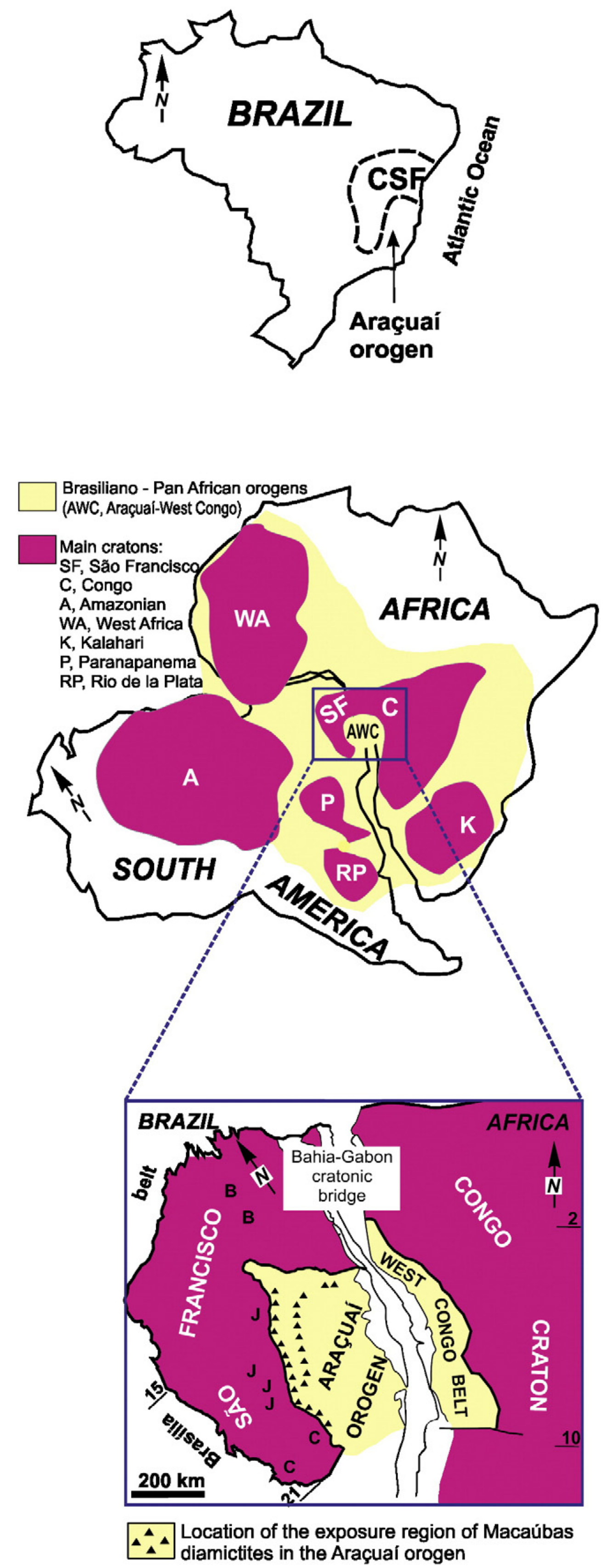

Fig. 1. Location of the Araçuaí orogen in Brazil (CSF, São Francisco craton) and in Western Gondwana together with its African counterpart, the West Congo belt. The exposure region of Macaúbas diamictites (triangles) is shown in the palaeotectonic fit of the Araçuaí–West Congo orogen and related São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

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

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

2. Geotectonic setting

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

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

2.1. Regional geology

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

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

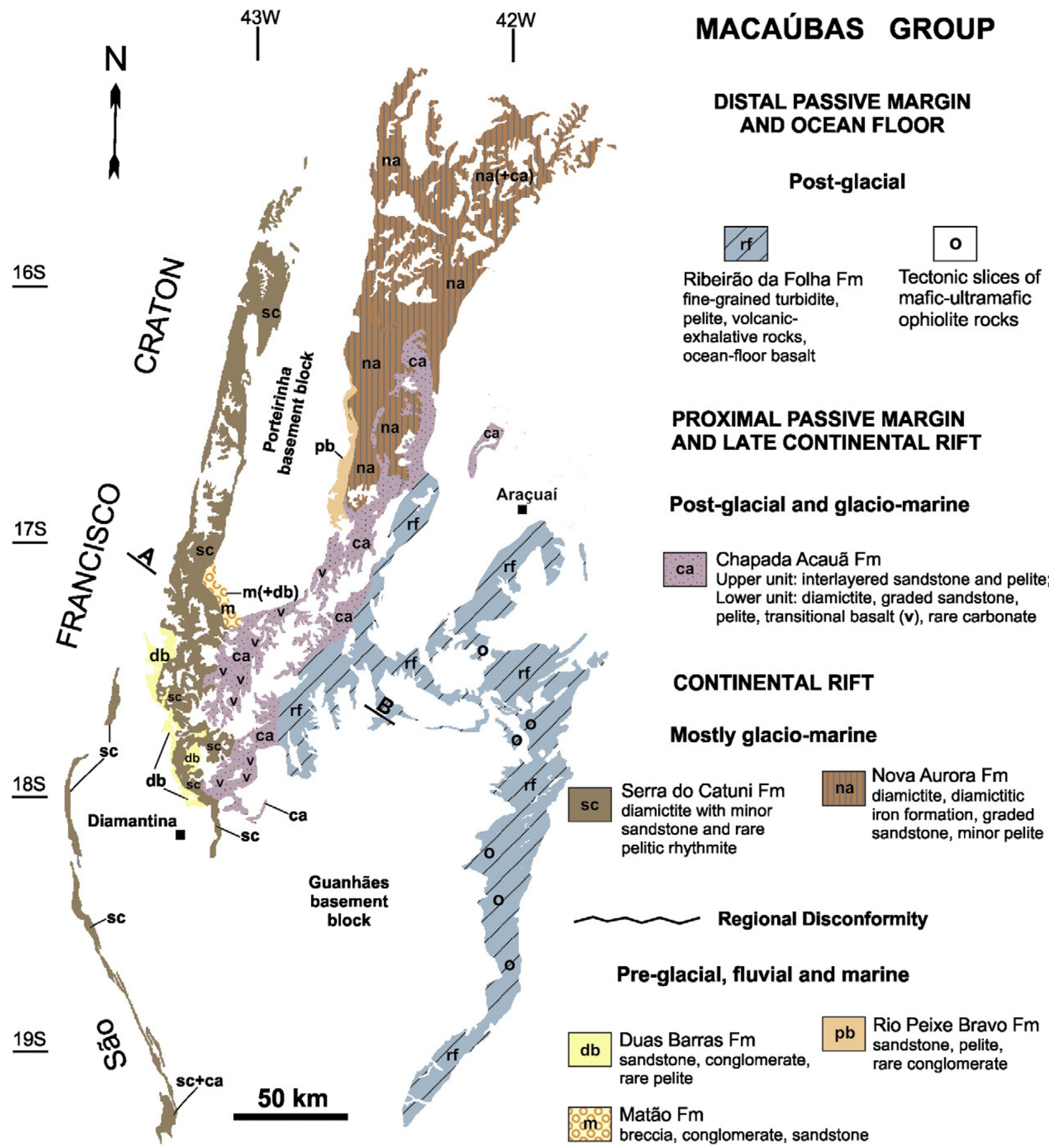

Fig. 2. Geological sketch map showing the distribution of the most part of the Macaúbas Group in the Araçuaí orogen (modified from Pedrosa-Soares et al., 2007). A-B, location of section shown in Fig. 3.

The basal deposits of the Macaúbas Group are represented by the Matão, Duas Barras and Rio Peixe Bravo formations (Fig. 2). The Matão Formation consists of breccia and conglomerate, rich in sandstone pebbles and cobbles, covered by sandstone with conglomeratic lenses. Erosive unconformity and normal faults outline the contact between the basement (locally represented by eolian sandstone of the Espinhaço Supergroup) and the Matão Formation, which records sedimentation under unstable tectonic conditions related to the early rift stage of the Macaúbas basin and contains detrital zircons with ages ranging from 1.16 Ga to 2.80 Ga (Martins, 2006; Martins et al., 2008). The youngest detrital zircon dated at 1.16 Ga defines the maximum depositional age of this formation (Martins et al., 2008). The Duas Barras Formation consists of sandstone and conglomeratic sandstone with variable contents of mica, feldspar, iron oxide and/or lithic fragments, quartz sandstone, conglomerate and rare pelite. It shows fluvial and shallow marine sedimentary facies and bimodal (NW–SE and SE–NW) paleocurrent sets (Grossi-Sad et al., 1997; Martins, 2006; Martins et al., 2008; Noce, 1997). The Rio Peixe Bravo Formation consists of micaceous, ferruginous and/or feldspathic sandstone, pelite locally rich in hematite and/or graphite, and rare clast-supported conglomerate (Grossi-Sad et al., 1997; Viveiros et al., 1979). These three units show no evidence of glaciation and represent fluvial to marine sedimentation during the continental rift stage of the Macaúbas basin (Martins et al., 2008; Noce et al., 1997; Pedrosa-Soares et al., 2008).

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

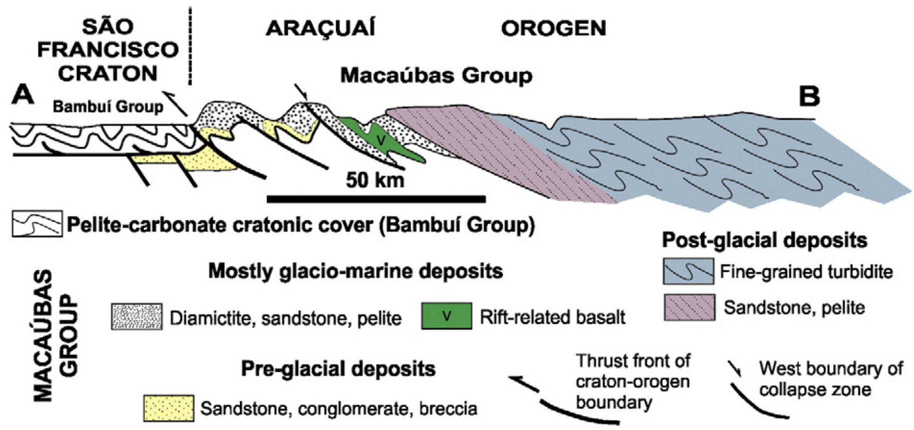

Fig. 3. Geological sketch section showing stratigraphic relations in the Macaúbas Group, and the thrust front boundary between the Araçuaí orogen and São Francisco craton (A-B location in Fig. 2).

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

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

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

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

The Lower Chapada Acauã Formation gradually passes upwards to the diamictite-free package of the Upper Chapada Acauã Formation, which is formed by a sandstone–pelite succession (Fig. 2). It is interpreted as a post-glacial unit deposited in a shelf environment during the passive margin stage of the Macaúbas basin (Grossi-Sad et al., 1997; Martins-Neto et al., 2001 and Noce et al., 1997; Pedrosa-Soares and Oliveira, 1997). To the east, this unit gives way to the fine-grained turbidites of the Ribeirão da Folha Formation (Pedrosa-Soares et al., 1992, 2008). The Ribeirão da Folha Formation is a diamictite-free unit and includes distal passive margin and ocean floor deposits (Fig. 2). The age of deposition of these sediments is constrained by the U–Pb age of 660 ± 29 Ma, obtained from euhedral crystals within a meta-plagiogranite associated with meta-mafic rocks, and interpreted as the time of magmatic crystallization of an ophiolite sliver (Queiroga et al., 2007).

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

3. Analytical procedures

Sm and Nd isotopic analyses were carried out at the Geochronological Research Center of the São Paulo University. Whole rock samples were powdered and dissolved in HF, HNO_3, and HCl media. Separation and purification of Sm and Nd were carried out by standard ion exchange procedures using RE and LN Eichron resins. Sm and Nd isotope data were obtained on a multi-collector MAT262 mass spectrometer. The measured $^{143}Nd/^{144}Nd$ ratios were normalized to 0.7219. The $^{143}Nd/^{144}Nd$ ratios of La Jolla and BCR-1, measured in the laboratory, are 0.511849 ± 0.000025 (1σ) and 0.512662 ± 0.000027 (1σ), respectively. Total blanks for Sm and Nd are 100 and 120 pg, respectively. The Sm–Nd model ages reported here are based on a depleted mantle model (DePaolo, 1981) and are interpreted as mean crustal residence ages.

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

Table 1
Sm–Nd isotopic data from whole-rock samples from the Macaúbas Group.

Sample ID	Rock	Geological unit	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	Error	$^{143}Nd/^{144}Nd$	Error 2σ	Age T_{DM} (Ga)	$\varepsilon_{(0)}$
MG05-04	Sandstone	Duas Barras	0.710	4.219	0.1017	0.0006	0.511645	0.000010	**1.9**	−19.4
MG08-06	Quartz-sandstone	Duas Barras	0.878	5.550	0.0956	0.0006	0.511618	0.000009	**1.8**	−19.9
MG05-01	Diamictite	Serra do Catuni	2.281	11.915	0.1158	0.0004	0.511599	0.000009	**2.3**	−20.3
MG05-02	Diamictite	Serra do Catuni	2.140	10.963	0.1181	0.0004	0.511613	0.000008	**2.3**	−20.0
MG08-09	Diamictite	Chapada Acauã Inf.	2.904	15.970	0.1100	0.0007	0.511531	0.000011	**2.2**	−21.6
MG08-11	Mica-rich wacke	Chapada Acauã Inf.	1.544	8.204	0.1138	0.0007	0.511573	0.000007	**2.3**	−20.8
MG08-12	Wacke	Chapada Acauã Sup.	2.022	10.772	0.1135	0.0007	0.511383	0.000008	**2.5**	−24.5
MG05-03	Basalt	Chapada Acauã Inf.	3.411	13.906	0.1483	0.0005	0.512302	0.000012	**1.7**	−6.6
MG05-23	Clast of basalt	Chapada Acauã Inf.	4.778	19.413	0.1488	0.0005	0.512329	0.000010	**1.7**	−6.0
R5.2[a]	Basalt	Chapada Acauã Inf.	1.119	4.751	0.1424	0.0005	0.512330	0.000009	**1.5**	−6.0

[a] Data from Babinski et al., 2005.

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

4. Results

4.1. Sm–Nd data

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

4.2. U–Pb SHRIMP data

Zircon grains separated from four clastic sedimentary rocks and from a mafic volcanic rock were dated. Given that our main objective was to determine the maximum depositional age of glacial deposits, we selected for analyses: (i) the sandstones of the Duas Barras and Rio Peixe Bravo Formations, two of the lowermost units of the Macaúbas Group, (ii) the overlying glacial diamictites of the Serra do Catuni Formation, and (iii) a mafic volcanic rock with pillow structures from the Lower Chapada Acauã Formation (Fig. 2).

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

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

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

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

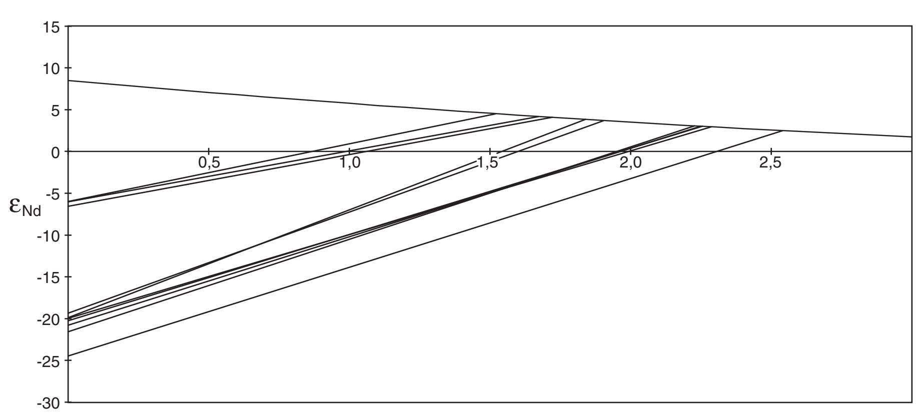

Fig. 4. Nd evolution diagram for different lithostratiraphic units of the Macaúbas Group.

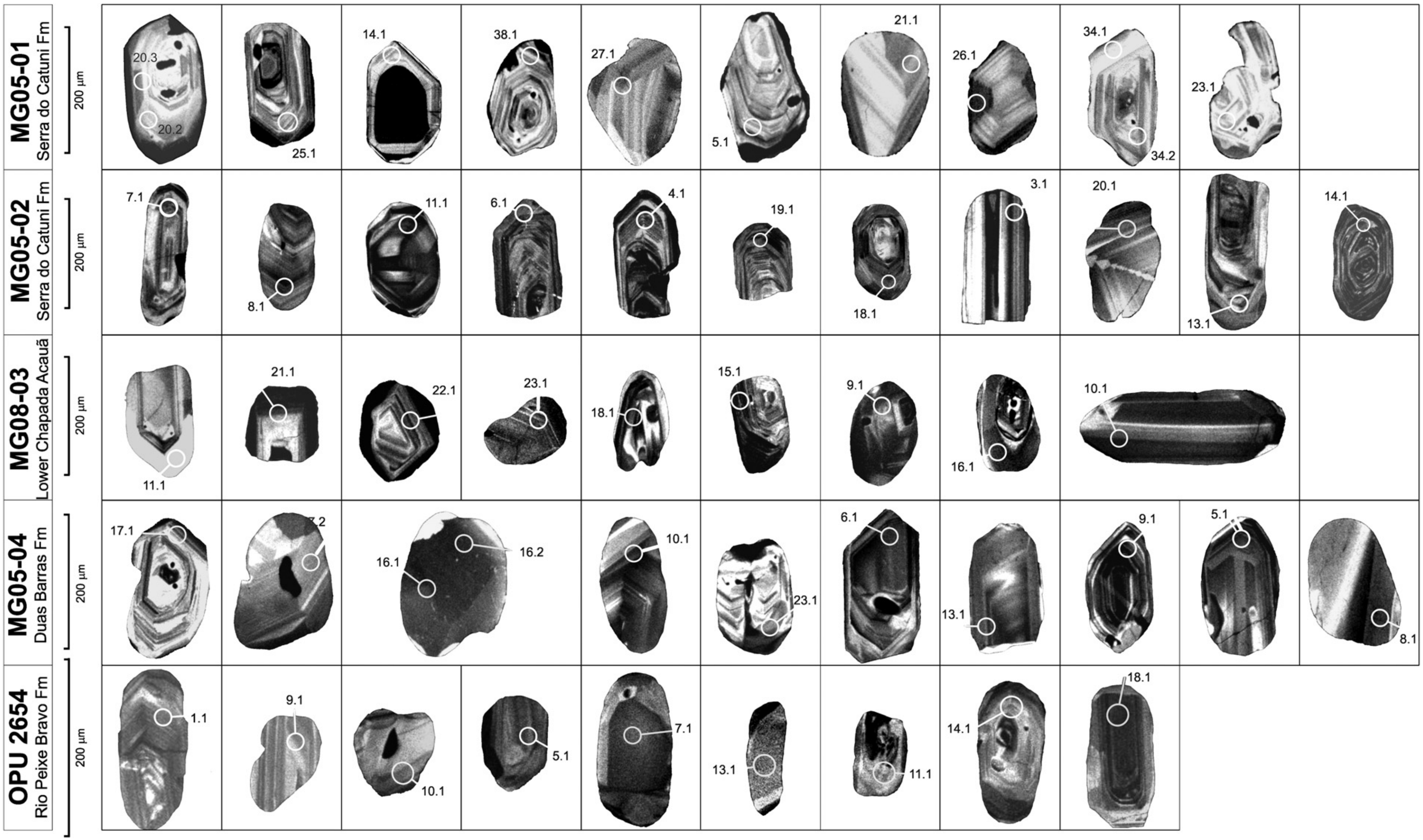

Fig. 5. Cathodoluminescence images of analyzed detrital zircon crystals.

Table 2
U–Pb SHRIMP data from detrital zircons of sample MG05-04 (Duas Barras Formation).

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

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.50 and 0.36% (not included in above errors).
(1) Common Pb corrected using measured $^{204}Pb/^{206}Pb$ ratio.
* Analysis 15.1 has large error on the 207Pb/235U ratio and was not plotted on the U–Pb concordia diagram.

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

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

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

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

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

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

5. Discussion and correlations

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

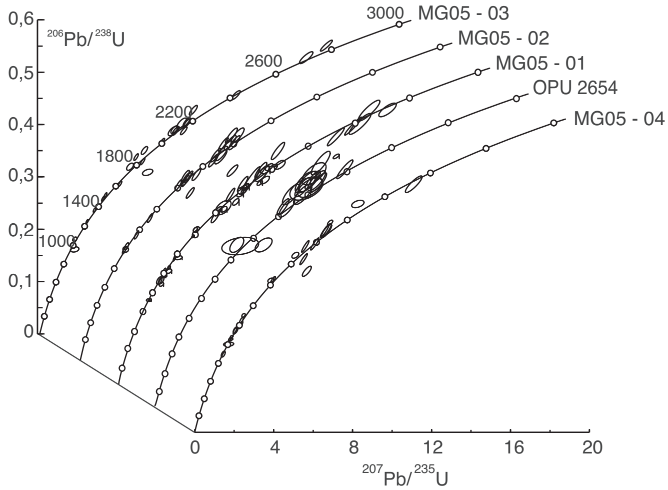

Fig. 6. U–Pb concordia diagrams for the five samples. Sample MG05-04; Quartzite from the Duas Barras Fm.; Sample OPU-2654: Quartzite from the Rio Peixe Bravo Fm.; Sample MG05-01: Diamictite from the Serra do Catuni Fm.; Sample MG05-02: Diamictite from the Serra do Catuni Fm.; and Sample MG05-03: Mafic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

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

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

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

The pre-glacial deposits of the Macaúbas Group have been tentatively correlated to the Sansikwa Subgroup of the West Congolian belt (Pedrosa-Soares et al., 2008), which is also a pre-glacial rift-related unit with a maximum depositional age of 923 ± 43 Ma, and contains zircon grains with ages mainly between 900 Ma and 1200 Ma (Frimmel et al., 2006). However, in contrast with the Duas Barras age spectrum, only a few zircon grains of Archean and Palaeoproterozoic ages were observed in the Sansikwa Subgroup, indicating that in the African side Tonian to Stenian sources preferentially fed the basin. Many sources for the Tonian zircons are available in the West Congo belt including the thick Mayumbian rhyolitic lavas dated at 920 ± 8 Ma at the base and 912 ± 7 Ma at the top, the associated Mativa (924 ± 25 Ma) and Bata Kimenga (917 ± 14 Ma) subvolcanic granites, and the 999 ± 7 Ma Noqui-type granites intruded in the Palaeoproterozic Kimezian basement and in the lowermost beds of the Zadinian Group (Tack et al., 2001). In contrast, Tonian igneous sources are not so abundant in the São Francisco craton and Araçuaí orogen, being limited to mafic dikes of 1000–900 Ma (D'Agrella Filho et al., 1990; Machado et al., 1989) and a few A-type granites dated at 875 ± 9 Ma (Silva et al., 2008). This can be explained by the model of an asymmetric continental rift with the thermal axis of the Tonian basin system located in the African side, as suggested by Pedrosa-Soares et al. (2008).

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

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

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

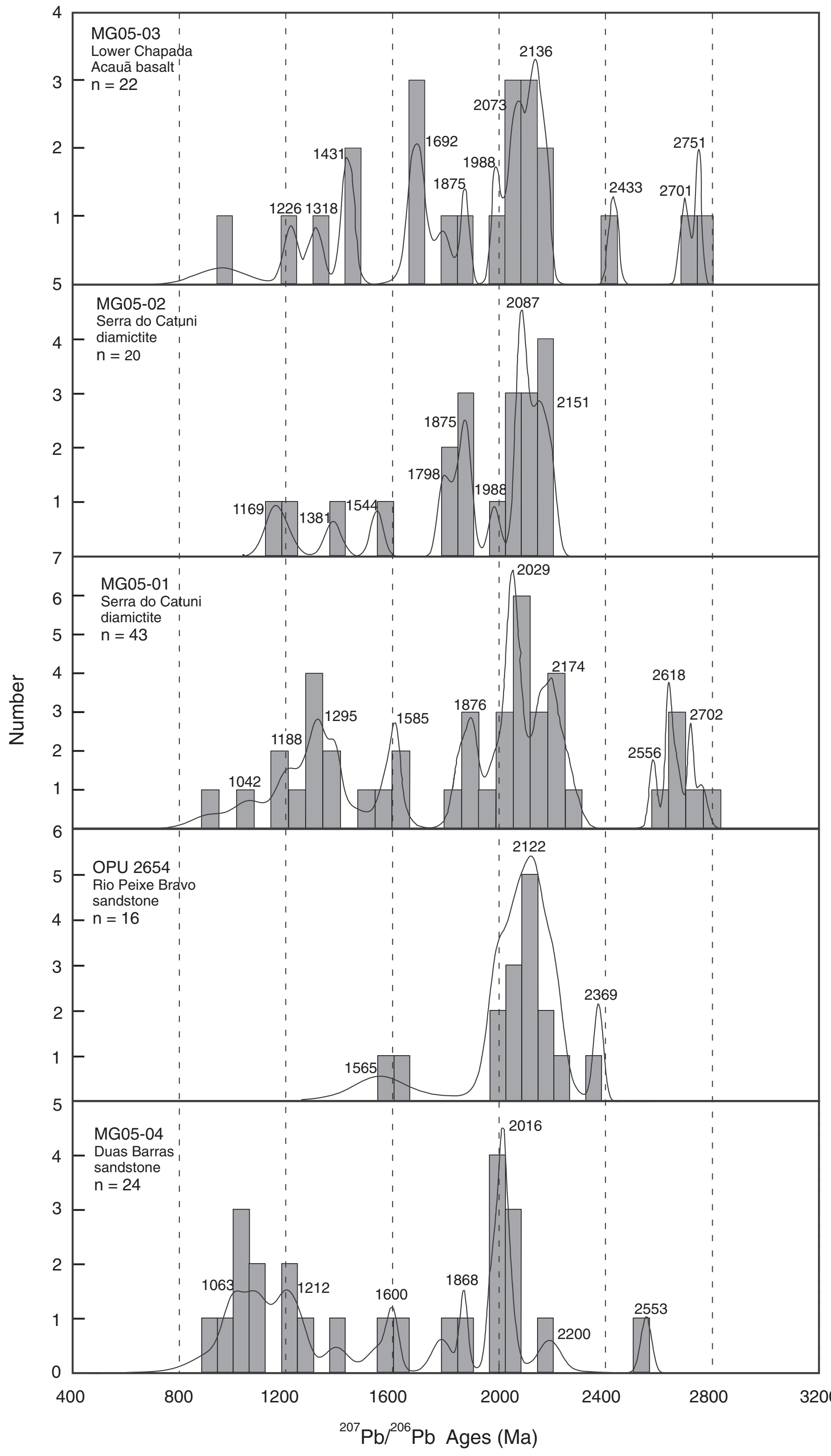

Fig. 7. Frequency histogram and probability curves of zircon Pb/Pb ages (<15% discordant) for samples of the Macaúbas Group.

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

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

Table 3
U–Pb SHRIMP data from detrital zircons of sample OPU 2654 (Rio Peixe Bravo Formation).

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

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.58% (not included in above errors).
1) Common Pb corrected using measured $^{204}Pb/^{206}Pb$ ratio.

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

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

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

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

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

Table 4
U–Pb SHRIMP data from detrital zircons of sample MG05-01 (Serra do Catuni Formation).

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

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.52, 0.30 and 0.90% (not included in above errors).
(1) Common Pb corrected using measured ^{204}Pb/^{206}Pb ratio.

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

If the U–Pb SHRIMP ages on detrital zircons recovered from the Haut Shiloango Subgroup (Frimmel et al., 2006) are considered, another scenario can be suggested. The youngest detrital zircon dated from the Haut Shiloango Group is 547±45 Ma (5% discordant). Thus, considering the error, the maximum depositional age of the Haut Shiloango would be 592 Ma. If this is correct, the Upper Mixtite Formation which is in higher stratigraphic position would be correlated to the Gaskiers glaciation event (Bowring et al., 2003). Some samples of the Haut Shiloango Subgroup present ^{87}Sr/^{86}Sr ratios of 0.7071 that could be correlated to post-old Cryogenian values. However, the δ^{13}C values are very positive (+3 to +8‰) and could not represent those of post-old Cryogenian cap carbonates (which are typified by negative δ^{13}C values). This would imply in a hiatus in the middle of the Haut Shiloango Subgroup, the lowest part being Early Cryogenian and the upper part being younger than 590 Ma.

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

Table 5
U–Pb SHRIMP data from detrital zircons of sample MG05-02 (Serra do Catuni Formation).

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

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.50% (not included in above errors but required when comparing data from different mounts).
(1) Common Pb corrected using measured ^{204}Pb/^{206}Pb ratio.

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

6. Conclusions

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

Table 6
U–Pb SHRIMP data from detrital zircons of sample MG05-03 (Lower Chapada Acauã Formation).

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

Notes: Errors are 1-sigma; Pbc and Pb* indicate the common and radiogenic portions, respectively.
Error in Standard calibration was 0.63% (not included in above errors).
(1) Common Pb corrected using measured ^{204}Pb/^{206}Pb ratio.

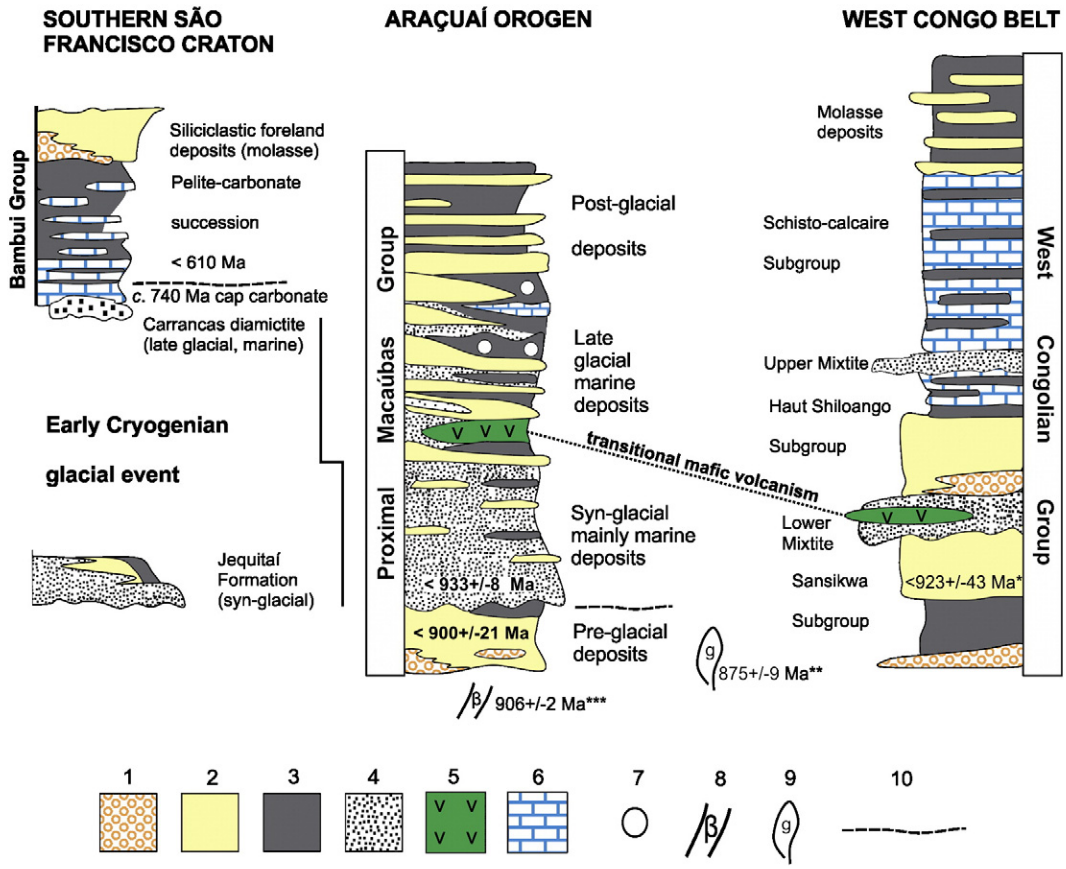

Fig. 8. Correlation stratigraphic sketch for the Proximal Macaúbas Group (Araçuaí orogen), West Congolian Group (West Congo belt) and covers of the São Francisco craton. 1, rudites; 2, sandstones; 3, pelites; 4, diamictites; 5, mafic volcanic rocks with pillow structure and transitional geochemical signature;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

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

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

After this Late Tonian sedimentation, a climatic change to glacial conditions affected the São Francisco–Congo continent and neighboring basins, so that a huge amount of diamictites started to fill the Macaúbas basin. The correlations between the Macaúbas diamictites and those covered by the *c.* 740 Ma cap carbonate on the São Francisco craton suggest that this climatic change can be correlated to the Early Cryogenian (known as Old Cryogenian) glacial event. The Serra do Catuni Formation is considered to be an equivalent of the Jequitaí diamictite, depicting a lateral succession of glacial environments from glacio-terrestrial, on the São Francisco craton, to proximal glacio-marine along the western region of the Macaúbas basin. The transitional mafic volcanism of the Lower Chapada Acauã Formation signals the end of the continental rift stage, which seems to be coeval with a climatic change to non-glacial conditions indicated by the absence of diamictites in the upper units of the Chapada Acauã Formation. This implies that at least part of the Chapada Acauã diamictites record the deglaciation process in the region. The Chapada Acauã diamictite succession is considered to be equivalent to the Carrancas diamictite, placed on the southern São Francisco craton, and to the Lower Mixtite Formation of the West Congo belt (Fig. 8). The Upper Mixtite Formation of the West Congo belt would have no equivalent in the Macaúbas basin, and might represent a local glaciation in the African side, coeval or younger than the upper part of the pelite–carbonate cover of the São Francisco craton, which maximum age is *c.* 590 Ma.

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

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