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Archean crustal evolution in the Southern São Francisco craton, Brazil: constraints from U-Pb, Lu-Hf and O isotope analyses

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Abstract

In this study we present U–Pb and Hf isotope data combined with O isotopes in zircon from Neoarchean granitoids and gneisses of the southern São Francisco craton in Brazil. The basement rocks record three distinct magmatic events: Rio das Velhas I (2920–2850 Ma), Rio das Velhas II (2800–2760 Ma) and Mamona (2750–2680 Ma).

The three sampled metamorphic complexes (Bação, Bonfim and Belo Horizonte) have distinct ϵ_{Hf} vs. time arrays, indicating that they grew as separate terranes. Paleoarchean crust is identified as a source which has been incorporated into younger magmatic rocks via melting and mixing with younger juvenile material, assimilation and/or source contamination processes. The continental crust in the southern São Francisco craton underwent a change in magmatic composition from medium- to high-K granitoids in the latest stages, indicating a progressive HFSE enrichment of the sources that underwent anatexis in the different stages and possibly shallowing of the melting depth. Oxygen isotope data shows a secular trend towards high $\delta^{18}\text{O}$ (up to 7.79‰) indicating the involvement of metasediments in the petrogenesis of the high potassium granitoids during the Mamona event. In addition, low $\delta^{18}\text{O}$ values (down to 2.50‰) throughout the Meso- and Neoarchean emphasize the importance of meteoritic fluids in intra-crustal magmatism.

We used hafnium isotope modelling from a compilation of detrital zircon compositions to constrain crustal growth rates and geodynamics from 3.50 to 2.65 Ga. The modelling points to a change in geodynamic process in the southern São Francisco craton at 2.9 Ga, from a regime dominated by net crustal growth in the Paleoarchean to a Neoarchean regime marked by crustal reworking. The reworking pro-

cesses account for the wide variety of granitoid magmatism and is attributed to the onset of continental collision.

Keywords: crustal evolution; zircon; U–Pb dating; Lu–Hf isotopes; O isotopes; São Francisco craton

1. Introduction

Zircon has been extensively used for geochronological and geochemical studies for the past decades. This is primarily due to its abundance in crustal rocks, its resistance to weathering, its ability to retain complex growth zoning and its ability to be precisely dated. Combined U–Pb, Lu–Hf and O isotope studies on zircons have been widely used to trace the evolution of the continental crust, particularly in the Archean (e.g. Dhuime et al., 2012; Kemp et al., 2009a; Naeraa et al., 2012; Pietranik et al., 2008; Zeh et al., 2009, 2014).

The southern São Francisco craton (SSFC) encompasses a typical segment of Archean/Paleoproterozoic continental crust. The basement consists of granitoid-gneiss complexes exposed within domelike structures, in contact with Neoarchean and Paleoproterozoic supracrustal sequences. The geology of the SSFC has been investigated for over three centuries, primarily for its significant economic resources, and is a key location for studying Archean crustal evolution in South America (e.g. Carneiro et al., 1997; Farina et al., 2015a, 2015b; Hartmann et al., 2006; Koglin et al., 2014; Lana et al., 2013; Machado and Carneiro, 1992; Machado et al., 1992, 1996; Moreira et al., 2016; Noce et al., 1998, 2005; Romano et al., 2013; Teixeira

and Figueiredo, 1991; Teixeira et al., 1996). Although a substantial number of these studies have aimed to unravel the Archean magmatic history of the SSFC, researchers are only recently starting to use a more regional approach. Geochronological studies have shown that several Meso- and Neoarchean magmatic pulses have led to the construction of this crustal segment, followed by an episode of crustal reworking leading to the emplacement of large K-rich granitoids and the subsequent stabilization of the SSFC (Lana et al., 2013; Romano et al., 2013). Farina et al. (2015a) assembled a large geochemical database on rocks from the basement and showed that in the Neoarchean there was a pronounced change in the composition of the crust, with a transition from medium-K to high-K magmatism. These authors proposed that this transition reflected the onset of basin deposition followed by the reworking of these rocks in the lower crust. Previous studies, however, lack the support of isotopic data. In order to place better constraints on the late-Archean evolution of the SSFC, we present the first set of U-Pb, Lu-Hf and O isotope data on single zircon grains from the basement rocks. This type of combined dataset has proven a powerful tool for deciphering geodynamic processes, particularly from times and places from which other geological records are scarce or absent.

In this paper we present a comprehensive dataset of zircon Lu-Hf and O isotope analyses for 30 gneisses, granitoids and amphibolitic dikes from the SSFC. This allows us to address several questions with implications for the Neoarchean crustal evolution of the southern São Francisco craton:

- (i) What were the mechanisms of crust formation?
- (ii) What was the nature and importance of crustal recycling/reworking and crust-mantle interactions?

2. Geological setting

The São Francisco craton represents one of the oldest segments of continental crust exposed in South America. It is composed of several Archean to Paleoproterozoic blocks, thought to have amalgamated during the 2.2-1.9 Ga Transamazonian orogenic event, and is bounded on all sides by Neoproterozoic orogenic belts (Almeida et al., 1981; Barbosa and Sabaté, 2004; Teixeira and Figueiredo, 1991). The southern edge of the craton exposes a section of Archean and Paleoproterozoic crust, including the Quadrilátero Ferrífero mining district that hosts world-class iron and gold deposits (Dorr, 1969; Lobato et al., 2001).

The SSFC was formed during a succession of magmatic pulses spanning from 3200 to 2600 Ma, in part concomitant with the deposition of a greenstone belt sequence, the Rio das Velhas Supergroup. The latest magmatic event saw the emergence of a stable continental platform, enabling the deposition of a thick Paleoproterozoic succession of volcanic, sedimentary and chemical strata, including the Minas Supergroup (Carneiro, 1992; Lana et al., 2013; Machado et al., 1992; Romano et al., 2013; Teixeira et al., 1996) (Fig. 1). The Archean basement in the SSFC is exposed within dome-like bodies reaching several tens of kilometres across (e.g. the Bação and Bonfim complexes), separated by elongate troughs containing polydeformed, low-grade supracrustal sequences. Together, these complexes have a typical Archean dome-and-basin geometry (Alkmim and Marshak, 1998; Marshak et al., 1992, 1997).

The Archean basement in the SSFC consists mainly of banded orthogneisses, intruded by several generations of granitoid bodies, leucogranitic sheets and dikes and late pegmatitic and aplitic veins. The gneisses display a complex pattern of am-

phibolite facies foliation, and locally exhibit stromatic migmatitic features (Lana et al., 2013). Lana et al. (2013) identified three main periods of magmatism in the SSFC, spanning between 3220 and 2770 Ma. Firstly, two gneisses cropping out in the south of the Santa Barbara complex were dated at 3212-3210 Ma, defining the eponymous Santa Barbara magmatic event. These ages are further supported by Sm-Nd T_{DM} model ages obtained by Teixeira et al. (1996) from medium- to high-grade gneisses located to the west of the SSFC (Campo Belo complex), and by a subset of detrital zircons from the Rio das Velhas and Minas Supergroups, suggesting the existence of fragments of Paleoarchean crust in the SSFC (Hartmann et al., 2006; Koglin et al., 2014; Machado et al., 1992, 1996; Moreira et al., 2016). The Rio das Velhas I and II (RVI and RVII) events are represented by gneisses and granitoids from the three main complexes (Bação, Bonfim and Belo Horizonte) yielding ages of 2920-2850 and 2800-2760 Ma respectively (Farina et al., 2015a and references therein). The widespread distribution of RVI and RVII rocks in and around the SSFC, and the presence in the detrital record of a large number of ca. 2800 and 2900 Ma zircon grains (Hartmann et al., 2006; Koglin et al., 2014, Moreira et al., 2016) emphasize the importance of the RVI and RVII as crust forming events.

The presence in the SSFC of several mostly undeformed granitoid batholiths dated at 2780-2770 Ma (namely the Samambaia and Caeté batholiths, Machado and Carneiro, 1992; Machado et al., 1992) indicates that the whole crust in the SSFC experienced regional metamorphism at the end of the RVII event. This crustal segment was later affected by one final magmatic event (namely the Mamona event; Farina et al., 2015a), responsible for the production of voluminous granitoid batholiths, smaller granitic domains and leucogranitic sheets and dikes, intruded in and around the older gneissic-greenstone crust between 2750 and 2700 Ma (Machado et al., 1992; Noce

et al., 1998; Romano et al., 2013). These large granitoid batholiths represent ~30% of the exposed surface of the SSFC (Romano et al., 2013). They are typically weakly foliated bodies, mostly encountered on the topographically higher outskirts of the Bonfim and Belo Horizonte complexes. In contrast, the older gneisses are found in the more eroded centre of the domes (Fig. 1). Noce et al. (1998) and Romano et al. (2013) also reported the occurrence in the south of the Bonfim complex of small granitoid bodies dated at 2613-2612 Ma. Romano et al. (2013) infer that these bodies account for less than 1% of the granitoid crust of the SSFC. In a recent contribution, Farina et al. (2015a) provided the first detailed database of major and trace element compositions for gneisses and granitoids of the SSFC. These authors observed that the crust in the SSFC experienced a major compositional change during the Mamona event, shifting from medium- to high-K magmatism. As opposed to what several studies had previously assumed, Farina et al. (2015a) demonstrated that the composition of the majority of the pre-Mamona intrusives does not match perfectly that of “true” Archean TTGs as described by Moyen (2011) (“true” TTGs refer to magmas that are formed via partial melting of an oceanic mafic crust, without any involvement from the continental crust). Instead, Farina et al. (2015a) highlighted the “hybrid” nature of the medium-K rocks, which display whole rock major and trace element compositions that are intermediate between those of “true” TTGs and experimental melts derived from partial melting of TTGs, leading these authors to suggest an origin via mixing between an end-member derived by partial melting of metamafic crust and a component resulting from the reworking of older TTGs. On the other hand, the major and trace element compositions of the high-K granitoids argue in favour of derivation from low-degree partial melting of immature metasediments (e.g. metagreywacke). This transition was made possible by the continental emergence and appearance of size-

able clastic sedimentary basins in the Neoarchean (the Rio das Velhas greenstone belt), subsequently buried to provide the more fertile metasedimentary protoliths required for the petrogenesis of the high-K magmas. It is important to note that the crust of the Belo Horizonte complex does not record this transition from medium- to high-K magmatism. Indeed, the voluminous Pequi and Florestal batholiths intruded in the west during the Mamona event have compositional characteristics very similar to those of TTGs, testifying of a somewhat different evolution for that complex.

3. Analytical techniques

Zircon grains from 30 samples were analysed for Lu-Hf isotopes. For 9 samples, Lu-Hf was analysed on zircons previously dated by Farina et al. (2015a). The remaining 21 samples were analysed for U-Pb during this study. Of these, 6 new samples were analysed, 8 samples previously studied by Romano et al. (2013) (labelled MR-) were re-dated using the same zircon mounts, and for the remaining 7 samples (previously dated by Farina et al. (2015a)), further U-Pb analyses were done on additional zircons picked from the same zircon separate and mounted separately for O isotope analyses. These additional zircons were mounted together with the zircon standards Temora (Valley, 2003) and 91500 (Wiedenbeck et al., 2004) and care was taken during polishing in order to minimize the effects of sample geometry and topography (Kita et al., 2009). For this mount, the O isotope analyses were done prior to U-Pb and Lu-Hf analyses to ensure that the O data were not compromised by the laser pits.

U-Pb analyses were first conducted at the Universidade Federal de Ouro Preto (UFOP) during two analytical sessions: the first one in June-July 2013 using an Ag-

ilent 7700 Quadrupole (Q)-ICP-MS coupled to a New Wave UP213 ($\lambda = 213$ nm) Nd:YAG laser, while the second session was done between May 2014 and July 2015 using a Thermo-Scientific Element 2 Sector Field (SF) ICP-MS coupled to a CETAC LSX-213 G2+ ($\lambda = 213$ nm) Nd:YAG laser. U-Pb data reduction was done using the GLITTER[®] software package (Van Achterbergh et al., 2001). Concordia diagrams were generated using the Isoplot/Ex 4 program (Ludwig, 2003). Following each session of U-Pb dating, Lu-Hf isotope analyses were carried out during two analytical sessions, following the methods by Gerdes and Zeh (2006, 2009). The first one was performed in August 2013 using a multi-collector (MC)-ICP-MS Thermo-Finnigan Neptune system coupled to a Resonetics RESolution M-50 193 nm Excimer laser at Goethe Universität Frankfurt (GUF) (Germany). The second session was conducted at UFOP between September 2014 and July 2015, using a multi-collector (MC)-ICP-MS Thermo-Scientific Neptune Plus system coupled to a Photon Machines 193 ($\lambda = 193$ nm) ArF Excimer laser ablation system. Oxygen isotopic compositions were determined by Secondary Ion Mass Spectrometry (SIMS) using a Cameca[®] IMS1270 multi-collector SIMS at the Edinburgh Materials and Micro-Analysis Centre (EMMAC, UK), following the methods described by Kemp et al. (2006, 2007). Laser spots for Lu-Hf analyses were drilled “on top” of the U-Pb and O spots or immediately beside, but always within the same zircon domain characterized by CL imaging.

The instrumental parameters used for U-Pb and Lu-Hf isotope analyses during both analytical sessions are shown in Table 1. A complete description of the analytical techniques is given in the Supplementary Material.

4. Results

A summary of all zircon U-Pb, Lu-Hf and O isotope data from this study is presented in Table 2 and 3 and Figures 2 and 3. Tables with the complete results are given in the Supplementary Material, along with concordia diagrams, zircon CL images and a full description of the geochronological data reduction applied for each sample. All uncertainties on the U-Pb analyses cited below are quoted at the 1σ level. All errors on the Lu-Hf analyses are quoted at the 2σ level.

Zircon textures together with U-Pb analyses indicate that ca. 50% of the studied samples are characterized by magmatic zircon crystals that do not show metamorphic overgrowth. Zircon grains from these samples contain fine-grained oscillatory zoning patterns and yield well-defined U-Pb concordia or weighted mean ages. Moreover, all zircons, including those affected by different degrees of Pb-loss within individual samples, have identical $^{176}\text{Hf}/^{177}\text{Hf}_i$ ratios, leading to nearly horizontal arrays when plotted versus time (Fig. 2). Zircons from ca. 30% of the samples show more complex structures in CL images, typically displaying core-rim relationships with bright zoned or homogeneous cores surrounded by darker rims either showing oscillatory zoning or, more rarely, structureless. The rims display similar to slightly higher $^{176}\text{Hf}/^{177}\text{Hf}_i$ ratios with respect to the related cores (Fig. 2), and were generally interpreted as related to a subsequent metamorphic event. The cores were either interpreted as reflecting the crystallization age of the sample (e.g. FQ2), or as inherited zircon xenocrysts either from the source (e.g. FQ14) or by assimilation during intrusion (e.g. FQ37). Finally, the remaining ca. 20% of the samples are characterized by homogeneously dark and/or structureless zircons in CL images. A subset of zircons display some locally recrystallized domains and/or disruption of concentric oscillatory zoning. Ages obtained from these grains were interpreted to date a later metamorphic event. Within individual samples, the Th/U ratios of the zircons can be

highly variable, and ca. 75% of the grains have ratios between 0.2 and 0.7. Moreover, except for 2 samples (FQ2 and FQ23) which display a systematic difference between high-Th/U cores and low-Th/U rims, cores and rims have similar Th/U ratios. Based on these considerations, this ratio was not used as criteria to distinguish different domains within a single grain nor as an argument to interpret an age. When possible, the α -decay damage accumulation of the zircons was calculated following Murakami et al. (1991), and used as an indicator of the degree of metamictization of the grains. Zircons showing α -decay doses $> 8 \times 10^{15}$ α/mg are considered highly damaged and those analyses were discarded. Overall, 9 out of 202 spot analyses were discarded. The α -decay doses of the remaining analyses range from 0.29 to 7.62×10^{15} α/mg , with an average of 2.84×10^{15} α/mg , testifying of the general good preservation of the zircon magmatic/metamorphic ages.

4.1 Geochronological results

The majority of the samples analysed during this study have already been dated by several authors, (see Table 3). For the seven samples previously dated by Farina et al. (2015a), and for which we have analysed additional zircons, we used the intrusion/metamorphic ages provided by these authors, because these were generally more concordant and more representative of the samples, because they were obtained from a larger number of grains. This is the case for samples FQ1 (2711 ± 3 Ma), FQ2 (2868 ± 10 Ma), FQ17 (2778 ± 2 Ma), FQ29 (2773 ± 2 Ma), FQ52 (metamorphic age; 2727 ± 11 Ma), FQ60 (2728 ± 16 Ma) and FQ74 (metamorphic age: 2638 ± 14 Ma). For five samples, the ages obtained during this study are identical, within error, to those from the literature. This is the case for samples FQ5 (2761 ± 11

Ma), MR31A (2716 ± 14 Ma), MR70G (2716 ± 6 Ma), MR259A (2721 ± 9 Ma) and MR51A (2708 ± 10 Ma) (Table 3). Five additional samples yielded slightly different ages than those obtained during previous studies ($< 1.5\%$ age difference): FQ6 (2779 ± 4 Ma), MR22A (2715 ± 2 Ma), MR87A (2646 ± 9 Ma), MR14A (2715 ± 3 Ma) and MR257A (2723 ± 8 Ma) (Table 3). For one sample (FQ20), we were not able to reproduce the published geochronological data. This sample corresponds to sample D07A from Lana et al. (2013), who obtained a SHRIMP magmatic age of 2918 ± 9 Ma, and a poorly defined metamorphic age of 2775 Ma. FQ20 yielded a majority of dark and structureless zircons, with a concordia age of 2723 ± 3 Ma which we interpreted as a metamorphic age of the rock.

In addition, we provide new LA-ICP-MS U-Pb ages for three, so far undated rocks. These consist of a banded gneiss crosscut by numerous leucogranitic sheets and dikes (FQ8), and a migmatitic gneiss (FQ14) from the Bação complex, as well as a foliated granitoid from the Caeté dome (FQ81). Zircons from FQ8 mostly display dark structureless centers, yielding a concordia age of 2612 ± 10 Ma, interpreted as a metamorphic age. A subset of zircon grains contain bright banded- or oscillatory-zoned inherited cores. Four discordant analyses on these cores gave older apparent ages that plot on a regression line yielding an age of ca. 2770 Ma for the protolith of the gneiss. CL images from zircons from FQ14 revealed a majority of dark and featureless grains. Ca. 40% of the zircons contain bright inherited cores, surrounded by dark homogeneous rims. The cores yielded concordant ages ranging from 2925 to 3472 Ma, clustered in four main populations at 2933, 3202, 3358 and 3465 Ma, while the rims gave a concordia age of 2692 ± 4 Ma. Sample FQ81 was collected from a foliated granitoid exposed in the Caeté dome (FQ81), where Machado and Carneiro (1992) had obtained a zircon U-Pb TIMS crystallization age of $2776 \pm 7/-6$ Ma from an

outcrop located several kilometres away. In CL images, zircons from FQ81 are mostly dark and structureless, and yielded a younger age of 2671 ± 10 Ma, interpreted as the age of a late metamorphic event.

Several samples analysed during this study yielded metamorphic ages significantly younger than the Mamona event in the SSFC (< 2680 Ma). Several authors have previously reported similarly young metamorphic ages in the Passa Tempo complex, located south of the Bonfim complex (2622 Ma, Campos et al., 2003), as well as in the Belo Horizonte complex (2670-2638 Ma, Farina et al., 2015a). It is important to note that some of these ages are contemporaneous to the magmatic ages obtained from small granitoid intrusions from the south of the SSFC (2612-2613 Ma, Noce et al., 1998; Romano et al., 2013). However, the significance of this spread of metamorphic ages remains elusive.

4.2 Lu-Hf isotopes

The **RVI event** is represented by four banded trondhjemitic gneisses, collected from the eastern Bação complex and the central Bonfim complex. The gneisses from the Bação complex yield $^{176}\text{Hf}/^{177}\text{Hf}_t$ of 0.28090-0.28110 (± 0.00002) corresponding to superchondritic ϵHf_t values between +0.7 and +5. By contrast, the rocks from the Bonfim complex have significantly lower $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios (Fig. 2a). Sample FQ40 contains several zircon cores with Hf isotope ratios ($^{176}\text{Hf}/^{177}\text{Hf}_t = 0.28087 \pm 0.00002$) identical to those of the gneiss FQ41 cropping nearby ($^{176}\text{Hf}/^{177}\text{Hf}_t = 0.28090 \pm 0.00002$), which corresponds to subchondritic ϵHf_t values between -1.1 and -3.4.

The **RVII event** is represented by six samples, five of them were collected within the Bação complex. They consist of two biotite-bearing banded gneisses, a

fine-grained plagioclase-rich gneiss and two weakly deformed Kfs-bearing granitoids collected from small domains within the gneisses. The last sample was collected in the eastern Bonfim complex within the Samambaia tonalite pluton, and consists of a medium-grained amphibole- and epidote-bearing granitoid. As for the RVI event, samples from the Bação complex yield systematically higher $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios ($0.28095\text{--}0.28107 \pm 0.00002$) than those from the Bonfim complex (0.28090 ± 0.00002) (Fig. 2b). It is worth noting that most of the rocks of the Bação complex exhibit superchondritic but slightly less radiogenic ϵHf_t values than the RVI gneisses. Three samples from the Bação complex display clear core/rim relationships, with rims dating a regional metamorphic event affecting the complex during the Mamona event, and yielding average $^{176}\text{Hf}/^{177}\text{Hf}_t$ of about 0.28102, corresponding to $\epsilon\text{Hf}_t \sim 1$. Four samples from the Bação complex contain inherited zircon cores dated between 2825 to 2920 Ma, with $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios overlapping those of the RVI gneisses, indicating the involvement of RVI rocks in the petrogenesis of RVII magmas.

The **Mamona event** is documented by twenty samples collected from the Bação, Bonfim and Belo Horizonte/Caeté complexes (Fig. 2). The rocks from the **Bação complex** consist of a small weakly-foliated biotite-poor granitic body intruded at the edge of the complex, two banded gneisses crosscut by numerous leucogranitic sheets and dikes and a migmatitic gneiss. Zircons from banded gneisses are mostly dark and structureless, interpreted as metamorphic. They yield $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios ranging from 0.28099 ± 0.00002 to 0.28102 ± 0.00003 . These values are similar to those obtained for the biotite-poor granite, and correspond to ϵHf_t values from -1 to -6 (Fig. 2c). Many zircons from the migmatitic gneiss (FQ14) contain bright inherited cores, surrounded by dark homogeneous rims. The cores yielded variable $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios ranging from 0.28101 to 0.28085, corresponding to highly scattered super-

chondritic to subchondritic ϵHf_t values (Fig. 2c).. The rims exhibit $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios of 0.28102 ± 0.00003 , indistinguishable from those obtained for the banded gneisses and the granite.

The Mamona event in the **Bonfim complex** is represented by eight samples: two small leucogranitic bodies, two weakly foliated aerially extensive granitoid phases from the Mamona and Souza Noschese batholiths, two small foliated granitoid bodies, a banded trondhjemitic gneiss cropping out in the centre of the dome and an amphibolitic dike. Zircons from all the granitoids show simple magmatic zoning patterns. All but one sample (FQ51, with $^{176}\text{Hf}/^{177}\text{Hf}_t = 0.28103 \pm 0.00002$) have identical $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios of 0.28095 ± 0.00003 (Fig. 2d). Zircons from the gneiss yielded comparatively higher Hf ratios of ~ 0.28101 , which corresponds to $\epsilon\text{Hf}_t \sim 3.8$, and were interpreted by Farina et al. (2015a) to date a metamorphic event. Zircons from the amphibolite sample display core/rim relationships. Hf ratios for the rims average 0.28096 which is identical to the values obtained for the majority of the granitoids from the complex, while those of the cores range from 0.28087 to 0.28096, which is similar to the values obtained for the RVI gneisses from the Bonfim complex.

The Mamona event in the **Belo Horizonte complex** is documented by seven samples: four plagioclase-rich granitoids collected from the Pequi and Florestal batholiths, one fine-grained banded gneiss exposed in the southern part of the complex, one Kfs-rich augen gneiss and a medium-grained leucogranite. One additional sample was collected from a foliated granitoid exposed in the Caeté dome (FQ81). These eight samples have $^{176}\text{Hf}/^{177}\text{Hf}_t$ ratios comprised between 0.28092 ± 0.00002 and 0.28098 ± 0.00002 , which corresponds to ϵHf_t values comprised between -6.2 and -1.7 (Fig. 2e).

4.3 O isotopes

Oxygen isotopic ratios were measured for 75 zircon grains from seven samples representative of the three main magmatic events, for which isotopic compositions (U-Pb and Lu-Hf) had already been obtained. Ca. 20% of the analyses fall within the range of $\delta^{18}\text{O}$ values in equilibrium with mantle-derived rocks ($\delta^{18}\text{O} = 5.3 \pm 0.3\text{‰}$) (Fig. 3). Ca. 30% analyses record slightly higher, more evolved compositions ($> 5.6\text{‰}$), with only 6 spots falling in the range of “supracrustal zircon” (6.5-7.5‰) as defined by Cavosie et al. (2005) (Fig. 3). This field indicates a range of magmatic $\delta^{18}\text{O}$ values that are elevated with respect to those from mantle zircons, thus requiring input from rocks whose compositions have been shifted towards higher $\delta^{18}\text{O}$ values by low-temperature processes at Earth’s surface. The upper limit of the field is based on the absence in the Hadean and Archean zircon record of values with $\delta^{18}\text{O} > 7.5\text{‰}$. One exception is a grain from FQ52 yielding an anomalously high $\delta^{18}\text{O}$ value of $10.88 \pm 0.35\text{‰}$. This grain is euhedral and shows oscillatory growth zoning under CL, and has an apparent $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2718 ± 17 Ma (96% concordant). This is much higher than the sample average, and higher than $\delta^{18}\text{O}$ reported in Archean zircons ($< 7.5\text{‰}$). Although there was no apparent crack or fracture at the surface of the grain, we suggest that the heavy O isotope value might originate from post crystallisation processes, as the gneiss has undergone metamorphism during the Mamona event. The remaining 50% of the analyses yielded $\delta^{18}\text{O}$ values below the range of mantle values ($< 5.0\text{‰}$). These low $\delta^{18}\text{O}$ values make up a significant proportion of the zircons analysed during the three magmatic events, with 15 spots falling in the compositional field of Archean “low $\delta^{18}\text{O}$ zircon” (2-4‰) from Hiess et al. (2011) (Fig. 3). Low $\delta^{18}\text{O}$ values in zircon indicate that the grains either crystallized from, or diffusive-

ly exchanged O with a low $\delta^{18}\text{O}$ melt or fluid (e.g. Bindeman et al., 2008; Gilliam and Valley, 1997; Hiess et al., 2011; Valley et al., 2005). Two gneisses from the Bação complex contain metamorphic rims surrounding oscillatory zoned grain interiors interpreted as crystallization features. For both samples, the $\delta^{18}\text{O}$ values obtained for the rims are identical to those obtained for the magmatic domains, indicating either that there was no change in O isotope between the magmatic and metamorphic magmas/fluids, or that the O isotope system was not disturbed during the metamorphic event. Additionally, the two following trends can be observed within this dataset: (1) the $\delta^{18}\text{O}$ values for individual samples increase with time from an average of $3.9 \pm 0.2\text{‰}$ at 2868 Ma (FQ2) to $5.4 \pm 0.4\text{‰}$ at 2638 Ma (FQ74), in concert with (2) a notable increase in the range of $\delta^{18}\text{O}$ composition per sample, from $\pm 1.6\text{‰}$ for FQ2 to $\pm 3.2\text{‰}$ for FQ74 (2SD) (Fig. 3).

5. Discussion

5.1 Origin of the Meso- and Neoarchean granitoids

5.1.1 Interpretation of the Hf vs. time arrays

The crustal evolution of the three basement complexes of the SSFC will be treated separately in the following discussion. The linear Hf array (Fig. 4), the presence in RVII rocks of RVI inherited zircon cores, as well as the presence in RVI and RVII rocks of metamorphic rims formed during the Mamona event indicate that the Bação complex underwent progressive crustal reworking throughout the Neoarchean time. These lines of evidence can be correlated with (1) the close relationships observed in the field between gneisses and small granitic intrusions (e.g. xenoliths of

gneisses within the granites), and (2) the geochemistry of the gneisses suggesting a derivation from mixing between a crustal component derived by melting of an older TTG and a more mafic juvenile end-member (Farina et al., 2015a). Indeed, these authors observed that these rocks have intermediate compositions between those of “true” TTGs and those of experimental melts derived from fluid-absent partial melting of TTGs. The RVI gneisses in the Bação complex were mostly generated through melting of a mafic crust, with only a limited proportion of reworked felsic crust. This is supported by: (1) overall geochemical compositions that are comparable to those of “true” TTGs, (2) the scarcity of inherited zircons, and (3) superchondritic Hf isotope compositions. Patchett and Arndt (1986) demonstrated that because of its higher Nd content compared to that of the mantle, the amount of continental crust needed to reduce the ϵ_{Nd} composition of a Proterozoic granite from +5 down to 0 is small (< 10%). Although this example was made using Nd isotopes, the same logic applies to the Hf system. Therefore we suggest that the volume of reworked component involved in the generation of the RVI gneisses from the Bação complex was probably < 10%.

It is interesting to note the large range in initial Hf isotopic compositions displayed by the RVI gneisses (ϵ_{Hf} between 0.0 and +6.5). This is rather uncommon in typical I-type granitoids, and could be explained in several ways, which include incomplete mixing between two isotopically different magmas (in this case, a crustal component and a more juvenile, mafic one), or as inherited from the source, as a result from incomplete homogenization of the magma during dissolution and crystallization of isotopically heterogeneous zircons (Villaro et al., 2012; Farina et al., 2014). The subsequent reworking of this crust during RVII, with some minor contribution from juvenile material produced granitoid magmas with lower (superchondritic-

ic to chondritic) $^{176}\text{Hf}/^{177}\text{Hf}$ isotope ratios and compositions that are “transitional” between those of a TTG-like end-member and a high-K one. Further reworking of this “transitional” crust during the Mamona event generated high-K granitoids with subchondritic Hf isotope compositions. One gneiss sample from the Bação complex (FQ14) displaying migmatitic structures with garnet-rich leucosomes provides evidence of inheritance of distinctly older crust (concordant zircons of up to ~ 3.47 Ga). The occurrence of inherited cores with ages clustering between 2933 and 3465 Ma suggest that the protolith for FQ14 was a (meta)sedimentary rock that underwent partial melting during the Mamona event.

In the Bonfim complex, the RVI event is represented by two gneiss samples (Alberto Flores gneiss), with low $\epsilon\text{Hf}_t = -1$ to -4 . Samples from the Alberto Flores gneiss show highly variable compositions which differ from those of “true TTGs” (e.g. higher Si, K, Rb and lower Al contents). This suggests the involvement of reworked continental crust in their petrogenesis (Farina et al., 2015a), a hypothesis which can be reconciled with their subchondritic Hf isotope data. During RVII, the Bonfim complex is intruded in the east by the Samambaia tonalite. This magma has an evolved Hf isotope composition ($\epsilon\text{Hf}_t = -2$ to -4) that falls within the ϵHf array defined for the Bonfim complex. However, a derivation from an older felsic crust formed during RVI for example can be ruled out, as it cannot account for the intermediate compositions of the Samambaia tonalite (i.e. SiO_2 ranging from 64 to 72 wt.%) (Carneiro, 1992; Farina et al., 2015a). Instead, its enriched Hf isotopic signature probably originates from assimilation of older felsic crust by a more mafic magma. The Mamona event is mostly represented by large high-K batholiths emplaced in the eastern and northern borders of the dome, namely the Mamona and Souza Neschese batholiths. These granites all have subchondritic ϵHf_t values ranging from -1 to -6 that plot on an array

defined by the gneisses of the Bonfim complex (Array 2, Fig. 4). This array suggests that high-K granitoids formed by reworking of the previously formed crust. However, the K and LILE content of these rocks is too high to be explained by a derivation from partial melting of the orthogneisses. Farina et al. (2015a) therefore argued that the large high-K batholiths emplaced during the Mamona event require a source that is more enriched in K and LILE, and more fertile than the average continental (medium-K) crust present in the SSFC at the time, suggesting their derivation by melting of metasedimentary protoliths such as metagreywackes.

In the Belo Horizonte complex, the Mamona event is marked by the coeval intrusion of granitoids with very different field and geochemical features: (1) in the east, two voluminous batholiths (Pequi and Florestal) are characterized, as opposed to those emplaced in the Bonfim complex, by homogeneous sodic compositions presenting some similarities with TTGs (e.g. typically high Al_2O_3 , CaO , Na_2O , Sr, LREE and low K_2O , Rb, Y, HREE), and (2) smaller granitoid domains/dikes and leucogranites, showing a wider range of more enriched compositions. The petrology and geochemistry of the former suggest a derivation from a hydrous mafic rock. Assuming this is the case, their low Hf isotope signatures can be explained in two ways: (1) partial melting of an old mafic crust. This however implies that this mafic precursor was extracted from the depleted mantle long before remelting to produce these magmas. In particular, considering an average $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ for a mafic crust (Nebel et al., 2007), this corresponds to > 600 m.y. of crustal residence time (Fig. 4), or (2) partial melting of a mafic crust but with assimilated portions of much older crust.

The existence of several Archean basement complexes with distinct ϵHf vs. time arrays (Fig. 4) suggests that these terranes underwent partly/largely different histories. In this scenario, it is possible to imagine that the basement presently ex-

posed in the SSFC consists of a collage of different micro-continental blocks amalgamated during a late collisional stage.

5.1.2 Secular trend: from sodic to potassic magmatism

We have combined the Hf isotope data presented here with geochemical results obtained by Farina et al. (2015a) for sixteen igneous samples. The results are presented in Fig. 5, where ϵHf_t for each sample is plotted against some key geochemical features ($\text{K}_2\text{O}/\text{Na}_2\text{O}$, Th and Sr/Y), representative of the trends observed here. Magmas produced during RVI share similar features with TTGs, such as high Na_2O , $(\text{La}/\text{Yb})_N$, low K_2O , Th, U contents, and high Sr/Y, usually interpreted to reflect the depth at which those TTG magmas are produced. However, it is important to note that small differences do occur between “true” TTG magmas and those of the SSFC (e.g. lower Sr/Y, Fig. 5). During RVII, the magmas display a wider range of more evolved compositions (e.g. $\text{K}_2\text{O}/\text{Na}_2\text{O}$ up to 1, Th up to 20 ppm), however still overlapping the compositional field of TTGs. The magmas produced during the Mamona event all have subchondritic Hf isotope signatures but heterogeneous trace element compositions. In particular, we observe systematic differences between magmas forming large batholiths in the Belo Horizonte complex which are similar to TTGs, and those emplaced as the large batholiths of the Bonfim complex, which present the highest K/Na and lowest Sr/Y compositions of this dataset. Between these two end-members, we observe a range of (much less voluminous) magmas with intermediate compositions and different petrological features (from sodic Plag-rich granites to Kfs-rich granitoids, Farina et al., 2015a).

Overall, the most striking feature is the relatively continuous evolution trend with decreasing age (and decreasing ϵHf_t), from compositions close to those of TTGs to more enriched granitoids. In particular, we observe a progressive enrichment in High Field Strength Elements (HFSE) with time (e.g. Th from 6 to 30 ppm). These trends cannot be explained as a result of different degrees of fractionation from a similar source, as these magmas almost all have very similar high SiO_2 contents (> 70 wt.%). HFSE are highly immobile elements, and are usually considered as good tracers of source enrichment. We infer that these trends reflect crustal maturation and differentiation via progressive reworking at shallower crustal levels. In addition, it is important to note that the wide range of geochemical compositions characteristic of Mamona granitoids mirrors the diversity of processes and sources involved in the generation of these magmas, which cannot be appreciated using Hf isotopes alone. The trends documented here are not exclusive to the SSFC, and similar ones are well described in the Yilgarn and Pilbara cratons of Western Australia for example (Griffin et al., 2004; Ivanic et al., 2012). There, nearly continuous crustal melting over a protracted period of time has recorded a transition from sodic (TTG) to potassic magmatism, which is interpreted to reflect a similar scenario of progressive reworking.

5.1.3 Insights from O isotopes

Do these $\delta^{18}\text{O}$ values represent primary magmatic features?

The interpretation of O isotope data depends critically on whether the measured O isotope compositions reflect that of the magmas from which the zircons crys-

tallized, or if they are product of secondary alteration processes and isotopic exchange. In CL images, zircons generally display euhedral habits and oscillatory growth zonation. Th/U are between 0.02 and 0.8, and only 7 out of 75 spots analysed have Th/U < 0.1. However, a significant amount of zircons have low levels of U-Pb concordance (ca. 30% of zircons have concordance < 80%), which can cast doubts on the interpretation of the isotope compositions as primary values. Valley et al. (1994) observed that Pb-loss (indicated by higher levels of discordance) is always associated with resetting of O isotope ratios, shifting $\delta^{18}\text{O}$ values up to 2‰ lower than their primary values. However, these authors observed that there is no apparent proportional relationship between the degree of concordance and the shift in O isotope ratio (Valley et al., 1994, their Fig. 4). In other words, elevated levels of discordance are not necessarily coupled with highly disturbed O isotope ratios. For most samples in this dataset, no correlation is observed between $\delta^{18}\text{O}$ and Th/U, $^{204}\text{Hg} + \text{Pb}$) or the degree of concordance (%), which suggests that $\delta^{18}\text{O}$ values and radiation doses U-Pb systematics cannot be directly associated (Fig. 6). Overall, we argue that most of the zircons from the São Francisco craton analysed in this study have preserved their primary isotopic signatures and that these reflect that of the magmas they crystallized from.

We note however that two samples (FQ52 and FQ60) show a slight positive correlation between $\delta^{18}\text{O}$ and the level of concordance of the zircons (Fig. 4c). Zircons from FQ52 sometimes display faint oscillatory zoning disrupted inwards from the grain boundary by recrystallized domains. The edges of the grains commonly contain inclusions and/or fractures suggesting that the zircons were affected by fluid-dominated recrystallization (as described by Hoskin and Black, 2000), probably during the metamorphic event dated by this sample at 2727 ± 11 Ma. By contrast, the

zircons analysed from FQ60 are mostly euhedral and display clear oscillatory zoning patterns, with no sign of pervasive alteration in CL or BSE images. In this case, it is not clear what process caused the Pb-loss and the disruption of the Hf system. Together, this suggests that FQ52 and maybe FQ60 underwent some degree of post-magmatic alteration causing a shift towards some of the low $\delta^{18}\text{O}$ values observed in the most discordant zircons.

Secular trend of increasing $\delta^{18}\text{O}$ values: evidence for supracrustal reworking

If they can fingerprint involvement of crustal material, Hf isotopes alone do not allow determining whether contamination occurred via (1) source mixing, whereby the recycling of subducted materials into the source reservoir influences the isotopic signature of the resulting magmas, or (2) crustal contamination, that is contamination of the magma during ascent and/or emplacement by interaction (assimilation) with the crust it intrudes. In modern arcs, stable (e.g. O) and radiogenic (e.g. Nd, Hf, Sr) isotopes have been used widely to discriminate between these contamination processes (Appleby et al., 2010; Kemp et al., 2006; Peck et al., 2000).

The first trend in the O isotope data is the steadily increasing maximum $\delta^{18}\text{O}$ throughout the Neoarchean, from $\sim 5.3\text{‰}$ at 2.87 Ga to $\sim 7.8\text{‰}$ at 2.64 Ga (Fig. 3). This follows the general trend defined by a global O isotope dataset (Roberts and Spencer, 2014; Valley et al., 2005). In the global dataset, this secular rise of magmatic O isotope ratios is explained by a combination of modifications in sediment composition, availability, weathering and burial, originating from a change in tectonic styles at the Archean-Proterozoic boundary (Valley et al., 2005). In this study, the appearance of high $\delta^{18}\text{O}(\text{Zrc})$ ($> 6.5\text{‰}$) in rocks at ~ 2.7 Ga indicates reworking of

supracrustal lithologies during the last Mamona event, this either being sedimentary rocks (10 to 40‰) or altered volcanics (20‰) (Eiler, 2001). Archean sediments, dominated by greenstone belt assemblages made of volcanoclastics, pyroclastics and less mature sediments record lower $\delta^{18}\text{O}(\text{WR})$ ratios compared to their modern counterparts (Longstaffe and Schwarcz, 1977; Veizer and Mackenzie, 2003). Using an average of 15‰ for Archean sediments (average of Archean sandstones and shales from Valley et al., 2005), we estimate that 10-20% of sedimentary contaminant would be required to increase the $\delta^{18}\text{O}$ of a normal igneous rock and zircons by 1-2‰, the amount required to reach the range of $\delta^{18}\text{O}$ measured here. The question remains to determine whether these high $\delta^{18}\text{O}$ result from upper-crustal contamination or via direct sediment reworking. While we cannot completely rule out the possibility that these high $\delta^{18}\text{O}(\text{Zrc})$ values originate from assimilation of upper-crustal lithologies, the involvement of metasediments is in agreement with the model proposed earlier for the petrogenesis of these rocks.

The origin of the low $\delta^{18}\text{O}$ magmas

About 50% of all O isotope analyses have values below the range of mantle zircon compositions ($< 5.3 \pm 0.3\text{‰}$, Fig. 3). Low $\delta^{18}\text{O}$ values indicate that the zircons either crystallized from, or exchanged O via diffusion with a low $\delta^{18}\text{O}$ melt or fluid. The only low $\delta^{18}\text{O}$ materials are meteoric and seawater and materials that have undergone alteration with these fluids at high temperatures. Eiler (2001) indicated that the lower gabbroic portion of oceanic crust preserves a low $\delta^{18}\text{O}(\text{WR})$ (0-5‰) acquired during high-temperature interaction with percolating water at the ridge. We propose two different possible interpretations to account for the systematic presence of low $\delta^{18}\text{O}$ zircons in all three magmatic events.

The first model stems from the fact that all these rocks derived from a diversity of rather well-characterized sources and processes (Farina et al., 2015a; this study), and that perhaps a single unique mechanism is unlikely to account for the low $\delta^{18}\text{O}$ values observed in all of them. We therefore consider them separately. The magmas emplaced during the RVI event (FQ2) have $\delta^{18}\text{O}$ values that range from ~ 3 to 5% . This rock is inferred to result from mixing between a juvenile material and a TTG-derived melt (Farina et al., 2015a). The superchondritic ϵHf_t values displayed by the zircons in FQ2 indicates only a minor component of evolved crust within this magma ($< 10\%$). These first-order observations suggest that $> 90\%$ of the source of this rock was a juvenile rock, and that it had to account for the O isotope compositions observed. The low $\delta^{18}\text{O}(\text{Zrc})$ values of FQ2 can be explained as a result of partial melting of a lower and relatively young $\delta^{18}\text{O}$ -depleted gabbroic oceanic crust. The resulting melts will then mix with a small proportion of a continental crustal component, possibly originating by local wall-rock melting of the surrounding TTG crust. The RVII magmatic zircon population, documented by sample FQ29 and a few inherited cores displaying mantle-like values, is centred around mantle-like $\delta^{18}\text{O}$ values. It is worth noting that zircons from FQ29 display a small spread towards low $\delta^{18}\text{O}$ values, with two grains plotting in the field of “low $\delta^{18}\text{O}$ zircons”, and relatively clustered evolved Hf compositions ($\epsilon\text{Hf}_t = -1.0$ to -4.2). Derivation of these rocks by reworking of RVI continental crust to explain its evolved Hf composition, as may be suggested by the Hf linear array, can be ruled out on the basis that this existing crust had SiO_2 contents greater than $70 \text{ wt.}\%$ and was therefore incapable of giving rise to some of the more mafic compositions ($\text{SiO}_2 = 65\text{-}70 \text{ wt.}\%$) recorded in the Samambaia samples (Carneiro, 1992; Farina et al., 2015a). Moreover, field and geochemical observations suggest the involvement of a mantle source component for the Samambaia tonalite

(Farina et al., in prep.), which is in agreement with the observed mantle-like $\delta^{18}\text{O}$ compositions. In this context, the unradiogenic Hf isotope composition of sample FQ29 could be explained either by assimilation or source contamination by an evolved and slightly $\delta^{18}\text{O}$ -depleted crustal component. It is interesting to speculate on the origin of the low $\delta^{18}\text{O}$ magmatic zircons observed in rocks from the Mamona event. Samples emplaced at that time all record $\delta^{18}\text{O}$ values above those of the mantle, attributed to the reworking of $\delta^{18}\text{O}$ -enriched metasediments. However, this process does not account for the $\delta^{18}\text{O}$ -depleted zircons. Firstly, we should address the question of the nature of the fluid responsible for the $\delta^{18}\text{O}$ -depleted compositions. If the O composition of seawater is assumed to be constant and equivalent to its present-day value ($\delta^{18}\text{O} \sim 0\text{‰}$, Gregory, 1991), then high-temperature exchange with seawater would require very large volumes of seawater to explain the low $\delta^{18}\text{O}$ observed. Instead, meteoric fluids have compositions between 0 and -55‰ (Bindeman, 2011), a range that is more plausible to produce the $\delta^{18}\text{O}$ -depleted compositions. In recent magmatic systems reporting low $\delta^{18}\text{O}$ magmas, the latter are interpreted to originate within shallow sub-volcanic magma chambers where assimilation of hydrothermally altered wall rock has occurred (e.g. Bindeman and Valley, 2001; Monani and Valley, 2001; Wotzlaw et al., 2012). Where these low $\delta^{18}\text{O}$ zircons occur in the Archean, similar settings (shallow-level geothermal systems) have been proposed (Hiess et al., 2011; Hollis et al., 2014). In such environments, the emplacement of new granitoid magmas effectively drives groundwater into the crust through fracture networks where these meteoric fluids heterogeneously interact with wall rocks, lowering their O isotope compositions (Taylor, 1977). The isotopic heterogeneity observed in some samples (e.g. FQ74) is consistent with such an environment. These systems generally require that magmas are emplaced at relatively shallow levels, to have ac-

cess to meteoric water. In general, low $\delta^{18}\text{O}$ values are scarce, particularly amongst Archean rocks (Valley et al., 2005), where they have only been reported in southwest Greenland (Hiess et al., 2011) and northern Australia (Hollis et al., 2014) (see inset in Fig. 3), testifying of either their rarity, or a lack of preservation of such materials. Therefore, this model may be hampered by the fact that it requires low $\delta^{18}\text{O}$ material to be generated at three consecutive times in the same area, which may be considered unlikely for this has not been documented elsewhere.

Alternatively, the systematic presence of these low $\delta^{18}\text{O}$ values could be interpreted as resulting from a single hydrothermal alteration event, producing a significant amount of $\delta^{18}\text{O}$ -depleted material whose O isotope signature is then carried through the subsequent crustal reworking events. In this scenario, we propose that the source of the RVI gneisses, essentially a juvenile mafic rock, most probably oceanic crust, was contaminated by a significant amount of meteoric water during formation at a submarine rift zone, generating dynamic hydrothermal systems (see Eiler, 2001). Moreover, this is in agreement with the consensus that TTGs commonly derive from a partially hydrated mafic crust (e.g. Rapp and Watson, 1995). The systematic presence of these values in rocks from all three magmatic events highlights both the lack of efficient magma mixing, as well as the fact that reworking is not swamped by addition of new crust through the subsequent important magmatic events. In this hypothesis, the low $\delta^{18}\text{O}$ signatures act as an effective tracer for crustal reworking in the SSFC, much in agreement with the Hf isotope data.

Overall, although both models propose slightly different interpretations of these low $\delta^{18}\text{O}$ values, the data definitely indicates the interaction between meteoric water and crustal rocks at least once in the Mesoarchean and possibly also during the Neoarchean Mamona event in the SSFC. Given the proposed importance of oce-

anic crust in the formation of TTG magmas, it is perhaps surprising that low $\delta^{18}\text{O}$ TTGs have not been more commonly reported.

5.2. The growth of the continental crust in the SSFC

5.2.1 Archean depleted mantle evolution

Model ages are commonly used to produce estimates of crustal growth events (e.g. Bennett and DePaolo, 1987). This is based on the premise that model ages reflect the age of extraction from a depleted mantle reservoir, with zircon crystallization occurring only later on. Current models suggest that the depletion of the upper mantle from a chondritic reservoir started very early in Earth's history and was extrapolated more or less continuously, resulting in a linear $^{176}\text{Hf}/^{177}\text{Hf}$ array to reach its current composition, constrained by present-day MORB. Although some rocks show local evidence for early (Hadean) mantle depletion (e.g. suprachondritic ϵHf_t values recorded in Eoarchean rocks from West Greenland, Pilbara and Barberton, Amelin et al., 2000), there is little data supporting derivation of early continental crust from a long-term Hf-depleted mantle reservoir. In fact, zircons from felsic rocks with ages > 3.5 Ga from a number of Archean terranes have maximum Hf isotope compositions similar to chondritic values (Amelin et al., 2000; Kemp et al., 2009b; Lancaster et al., 2014, 2015; Naeraa et al., 2012; Satkoski et al., 2013; Zeh et al., 2009). Similarly in the SSFC, there is a striking lack of zircon data with superchondritic Hf compositions, particularly observed within the detrital record (Fig. 4b). Koglin et al. (2014) and Moreira et al. (2016) argued that the subchondritic ϵHf_t compositions of zircons from the SSFC reflect intense reworking episodes of the crust, starting as early as ~3.4-

3.5 Ma. An alternative view would be that the zircon record does not necessarily reflect crustal reworking, but instead argues against the presence of a strongly depleted mantle beneath the continental crust. Using a compilation of ~13000 Lu-Hf analyses on zircons, Guitreau (2012) proposed a model of evolution for the depleted mantle that fits the maximum Hf compositions of zircons through time. This model is defined by a period of only minor increase of ϵHf_i for the first 1 Gy, followed by a period from ~3.5 to ~2.5 Ga of enhanced differentiation of the depleted mantle, followed after ~2.5 Ga by a constant increase, although slightly less steep, until a present-day value of $\epsilon\text{Hf}_{\text{today}} = +18$. This model corresponds broadly to the “two-stage” evolution model discussed by Zeh et al. (2009) (their Fig. 12). It implies that for the first half of the history of the Earth, the source of the continental crust is not as depleted as proposed in the more commonly used models (e.g. that of Blichert-Toft and Puchtel, 2010). Applied to the SSFC, the model of Guitreau (2012) fits exceptionally well with the detrital zircon data (Fig. 4b). In addition, this model reduces the crustal residence time of zircons > 2.9 Ga from an average of 400 down to 200 m.y. Different tectonic settings in the Archean – with higher radiogenic heat production resulting in a large number of small unstable microplates and more dynamic tectonics – are more easily reconciled with shorter crustal residence times. For the next section, we tentatively use this depleted mantle model from Guitreau (2012) to discuss the crustal evolution of the SSFC.

5.2.2 Crustal growth

In order to discuss continent formation, understanding the growth of the continental crust involves evaluating the ratio between new magmatic additions directly

extracted from the mantle (juvenile) and ones that originate from remelting of older crust (reworking). The main challenge when trying to determine the distribution of crustal growth using detrital zircon is to correct for the data that represents crustal reworking in order to calculate a crustal generation curve. Temporal Hf isotopic trends have been used in several regional studies (i.e. Boekhout et al., 2013; Kemp et al., 2009a) in order to estimate the varying degrees of mantle input in the generation of granitic magmas. In this study, the regional magmatic evolution of the SSFC was explored using U-Pb and Lu-Hf data from detrital zircon from recent publications focused on the Neoarchean supracrustal successions of the Maquiné Group (Moreira et al., 2016) and Moeda Formation (Koglin et al., 2014; Martínez Dopico, unpublished). Modelling of the degree of juvenile growth was largely based on the calculation techniques applied by Belousova et al. (2010). Five zircons with $\epsilon\text{Hf}_t > 2$ $\epsilon\text{Hf}_{\text{DM}}$ were rejected. In total, 1178 analyses were used for modelling. All the crustal T_{DM2} ages for these zircons were recalculated according to the depleted mantle evolution model of Guitreau (2012) for which we have graphically determined the equation for this age interval (4.0-2.5 Ga) (Fig. 7b). For all the other parameters (Lu decay constant, CHUR parameters and average $^{176}\text{Lu}/^{177}\text{Hf}$ of the crust), we have adopted the same values as indicated in the Analytical Techniques section of the Supplementary Material. The results of the modelling are summarised in Figure 7.

The distribution of U-Pb detrital zircon ages along with an integral curve are plotted in Fig. 7a. The three major peaks observed at ca. 2.88, 2.79 and 2.72 Ga correspond to the RVI, RVII and Mamona events, respectively. Their presence reflects the good preservation of Neoarchean rocks in the SSFC. The existence of continental crust older than 2.9 Ga, although it is absent in the field, is evidenced by the Hf model ages of the detrital zircons. The dark blue line in Fig. 7a represents the cumu-

lative curve of crustal model ages obtained for this dataset. It suggests that if the generation of new crust was predominant during the Paleoproterozoic, the importance of crustal reworking increased after 2.9 Ga. However, the Hf model ages on their own do not address the possibility that some of these zircons were produced from a mixture between a radiogenic component and an older crustal material with a lower Hf isotope composition. Payne et al. (2016) estimated that only a small proportion (14%) of Hf model ages actually provide a meaningful indicator of the timing of crustal growth, the rest of the model ages likely resulting from mixtures of melt derived from multiple mantle and crustal sources. Recent studies have used different methods in an attempt to remove this “mixed signal” and thereby accurately evaluate the proportion of juvenile material added to the crust at each step of its evolution (Belousova et al., 2010; Dhuime et al., 2012; Kemp et al., 2007). Zircons with juvenile Hf compositions are defined as falling in a range of $\pm 2\epsilon$ units (or $\pm 0.75\%$) around the depleted mantle composition (Griffin et al., 2014). Using this simple definition, we have calculated the proportion of juvenile material produced at each time step, and obtained a cumulative curve (the green line in Fig. 7a) that is comparable to the one obtained from model ages, only slightly shifted towards older ages. Although debatable, these two simplistic approaches to estimate the growth of the continental crust in the SSFC indicate that most of it (ca. 95%) was already generated by 2.9 Ga (Fig. 7a), and that crustal reworking was dominating over net juvenile additions after 2.9 Ga. This conclusion is consistent with the geochemical data from granitoids of the SSFC, as discussed in section 5.1.

Assuming that the generation of Archean granitoids involved interactions between juvenile and crustal sources, the average proportion of new crust through time can be estimated using two-components mixing calculations. This requires some as-

assumptions to be made about the geochemical and isotopic compositions of the mantle and crustal end-members. Here we estimated the isotope composition of the crustal end-member from an “integral crust” calculation as described by Belousova et al. (2010). This integral crust is calculated so that the average signature of new crust is added successively to that of an older and more evolved crust. This method is similar to the calculation of crustal model ages but with an interpolation projected forward instead of backwards. This provides a more realistic estimate for the average crustal composition than the use of a single value. Ages younger than 2.60 Ga and older than 3.45 Ga are largely under-represented in this dataset, therefore the calculations are statistically more subject to bias and the results obtained for these periods will not be discussed further. The juvenile fraction between a depleted mantle (that of Guitreau, 2012) and a crustal component (given by the integral curve) for bins of 20 Ma is calculated using mean Hf concentrations of 2.31 (Kelemen et al., 2003) and 4.0 ppm (average of TTG, Laurent et al., 2014) respectively. If we consider a truly depleted mantle with $[Hf]_{DM} = 0.157$ ppm (Workman and Hart, 2005), the calculated juvenile contribution shifts towards higher proportions. However as discussed in the previous section, the presence of a true depleted mantle beneath the SSFC in the Archean is questionable. Moreover, even if the quantitative proportions of juvenile material depend on this assumption, the temporal trends will remain the same. The results shown in Fig. 7b (red line) indicate that 20-70% of the melts generated at all times were juvenile. The following trends can be observed: (1) there is a general decrease in the juvenile contribution to the magmas over time (red box in Fig. 7b), dropping from an average of 55% during the Paleo- and Mesoarchean to ~40% at 2.9 Ga, and down to ~20% at around 2.7 Ga. We suggest that the reason for this general trend is a change in geodynamics at 2.9 Ga, with a transition from island arc to conti-

mental arc (see discussion in the next section). Additionally, this drop of mantle input coincides with the increase of sediment input to the generation of crustal granitoids as suggested by the O isotope data, (2) we observe some significant variations of the juvenile input during the Paleoproterozoic, with rapid drops of up to 30% immediately followed by rapid increases. Although similar variations have been interpreted to reflect tectonic switching from regimes dominated by compression punctuated by extensional events (e.g. Collins, 2002; Collins et al., 2011; Kemp et al., 2009a), it is likely that some of these apparent variations of mantle input can simply be an artefact related to the scarcity of the Paleoproterozoic record, largely underrepresented in this dataset.

Overall, the detrital zircon record indicates a change of geodynamic processes at ca. 2.9 Ga. During the Paleoproterozoic, processes of renewed juvenile crust generation were operating, and although we have little direct evidence, it was certainly characterized by relatively high mantle contribution to magmatic episodes. At ca. 2.9 Ga there is a net increase in the extent of crustal reworking (further supported by geochemical arguments, Farina et al., 2015a) in the transition to a regime with balanced growth and destruction.

5.3 A geodynamic model for the SSFC

The combined dataset of magmatic and detrital zircons indicates that the SSFC was continuously affected by magmatic activity from ca. 3.50 to 2.65 Ga. Any geodynamic model chosen to represent the evolution of this portion of the crust must account for all of the features within this combined U-Pb-Hf-O dataset, as well as the

field, petrographic and geochemical evidence collected by Farina et al. (2015a). Below we discuss the magmatic and geodynamic evolution of the SSFC.

5.3.1 Pre-2.9 Ga

The early evolution of the SSFC is represented solely by detrital zircons, displaying continuous and homogeneous subchondritic ϵHf_t values between ca. 3.5 Ga and 2.9 Ga. The lack of Paleoarchean rocks in the field precludes any precise interpretation on the nature of the crust that was formed at the time. However, similar isotopic values have been described in TTGs from other Archean terranes (Moyen and Martin, 2012). Additionally, the modelling discussed in section 5.2 indicates that: (1) the majority of the crust (~90%) was originally formed before 2.9 Ga, and (2) the juvenile proportion of newly formed crust averages ~55% over that period, before decreasing notably after 2.9 Ga. Evidence is missing however in the SSFC to further discuss Paleoarchean tectonics and to explain the long-lived production of juvenile continental crust in the SSFC.

5.3.2 Post-2.9 Ga

The proposed Neoarchean evolution model comprises two stages, the first one coincides with the RVI and RVII events, and the second one corresponds to the Mamona event.

During **stage I** (Fig. 8a), the model must account for: (1) reworking of Paleoarchean felsic crust evidenced by older model ages in post-2.9 Ga magmas, the erosion and deposition of this crust into Neoarchean and Paleoproterozoic sedimentary

basins, (2) the general geochemical trends indicating a progressive HFSE enrichment of the source of these magmas, (3) lesser juvenile contribution to the magmatism, and (4) the systematic differences (field, geochemical and isotopic) observed between the three complexes. These data can be reconciled with a continental collision model, during which the accretion of various proto-continent leads to a tectonically thickened crust which undergoes progressive reworking. This scenario also accounts for the metamorphic event between 2.78 and 2.73 Ga recorded in the SSFC (Farina et al., 2015a; Lana et al., 2013).

Following this episode of crustal thickening, we suggest that **stage II** (Fig. 8b) marks a modification of the tectonic regime into extensional or non-compressional settings, during which slab break-off or retreat, lithospheric delamination or late- to post-orogenic gravitational collapse can trigger further melting of lower crustal lithologies (e.g. Duretz and Gerya, 2013; van Hunen and Allen, 2011). In this scenario, the compression of local geotherms at the base of the crust promotes renewed partial melting and regional metamorphism of the lower crust (Sandiford and Powell, 1986) and upper mantle. This is further supported by the existence of mafic-intermediate amphibolitic dikes emplaced during this time. We suggest that the magmatism produced during this event generates three types of granitoids. First, the local remelting of older (RVI and RVII) gneisses (the medium-K magmas of Farina et al., 2015a) generates small volumes of magmas with a wide range of compositions reflecting the compositional and isotopic heterogeneity of the source(s). These rocks are currently exposed in all three complexes as decimetre- to meter-scale granitic veins and dikes, small plutons (General Carneiro, Santa Luzia, Ibirité, Brumadinho) and domains closely associated with the gneisses as well as leucocratic veins and dikes. Secondly, in the Bonfim complex, the reworking at depth of metasediments

produced large volumes of biotite and two-mica granitoids (Mamona and Souza Neschese batholiths). And finally in the Belo Horizonte complex, the generation of the Pequi and Florestal batholiths primarily reflects the remelting of a metabasaltic source, generating magmas that later assimilated some older crust. This final event is synchronous with the complete depletion of the lower crust in heat-producing elements and the subsequent stabilization of the SSFC (Lana et al., 2013; Romano et al., 2013; Sandiford and McLaren, 2002).

Such a scenario has been proposed for a number of analogous Archean terranes such as for example the Yilgarn and Greenland cratons, interpreted as collages of several crustal blocks formed during arc magmatism, terrane accretion and collisional orogeny (e.g. Czarnota et al., 2010; Windley and Garde, 2009). Additionally, Laurent et al. (2014) recently reviewed the temporal evolution of several well-characterized Archean terranes, identifying a two-stage sequence. First, a long-lived period of TTG magmatism, followed by the generation of a range of more enriched granitoids emplaced during a shorter event. These authors tentatively proposed that this pattern reflects a global geodynamic model of subduction and subsequent continental collision, taking place on a planetary scale between 3.0 and 2.5 Ga, as a result of the progressive cooling of the Earth. The zircon dataset presented in this study can easily be reconciled with such scenario, although it does not provide direct evidence for it. 2.9 Ga marks a clear transition in the SSFC into a period where crustal reworking has dominated over net juvenile magmatic additions. By analogy, we infer that this change relates to the onset of accretionary and collisional events in the SSFC, leading to oceanic closure and progressive amalgamation of Paleoarchean proto-continental blocks.

6. Conclusion

The conclusions that arise from this U-Pb-Hf-O isotope study on Neoarchean granitoids and gneisses from the SSFC are as follows:

- Samples from the three complexes (Bação, Bonfim and Belo Horizonte) plot on distinct crustal evolution arrays, suggesting the involvement of Paleoarchean crust in their generation, which occurs via different processes (remelting, mixing, assimilation and/or source contamination). Combined with field data, this suggests that these complexes represent terranes with different identities (different proto-continent and/or levels of the crust) accreted together during a late collisional stage.
- The present Hf dataset combined with whole-rock geochemistry indicates: (1) a continuous differentiation trend from sodic (medium-K) to potassic (high-K) magmatism with age, reflecting progressive enrichment of the source in HFSE and probable shallowing of the depth of melting, and (2) that Hf isotopes alone do not represent the diversity of rocks and processes evidenced here.
- O isotope results indicate: (1) a secular trend towards high $\delta^{18}\text{O}$ values confirming the involvement of metasediments in the petrogenesis of Neoarchean high-K granitoids, (2) the presence of upper-crustal level hydrothermal systems during the Meso- and Neoarchean magmatism in the SSFC.
- Isotopic modelling, based on available Hf from detrital zircons record complements this dataset, indicating a major change in the evolution of the SSFC at 2.9 Ga. This change marks the transition between a Paleoarchean regime possibly dominated by TTG production and net crustal growth, and a Neoar-

clean regime that is dominated by reworking processes, producing a wide variety of granitoid magmas with highly scattered radiogenic isotope compositions. We attribute this transition to the onset of continental collision in the SSFC.

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TABLES

Table 1. Operating conditions and instrumental settings for the U-Th-Pb and Lu-Hf analyses.

Table 2. Detailed summary of zircon U-Pb, Lu-Hf and O isotope results.

Table 3. Summary of ages and isotopic compositions for samples analysed in this study. Data are from Farina et al. (2015a), Lana et al. (2013), Romano et al. (2013).

FIGURES

Figure 1. Geological map of the southern São Francisco craton with sample locations. Abbreviations: F – Florestal, M – Mamona, P – Pequi, Sa – Samambaia, SN – Souza Neschese. Inset: simplified map of the São Francisco craton, showing the lo-

cation of the exposed Archean provinces and the bordering Neoproterozoic orogenic belts. The box indicates the Quadrilátero Ferrífero. Modified from Alkmim and Marshak (1998).

Figure 2. $^{176}\text{Hf}/^{177}\text{Hf}_t$ vs. apparent $^{207}\text{Pb}/^{206}\text{Pb}$ age diagrams for each magmatic event described in Section 2. Filled symbols represent analyses with >95% concordance, open symbols are analyses with <95% concordance. Grey horizontal arrows indicate Pb-loss trends within individual samples. The Depleted Mantle (DM) (full line) composition is from Guitreau (2012), while the dashed line represents the most radiogenic DM composition from Blichert-Toft and Puchtel (2010) (see text for explanation).

Figure 3. $\delta^{18}\text{O}$ vs. intrusion age diagram. Filled symbols represent O analyses on zircons with >90% concordance, open symbols are zircons with <90% concordance. Fields for Hadean and Archean “supracrustal zircon” and Archean “low $\delta^{18}\text{O}$ zircon” are from Cavosie et al. (2005) and Hiess et al. (2011) respectively. The “mantle zircon” field is $5.3 \pm 0.3\text{‰}$ as defined in Valley et al. (1998). Inset: $\delta^{18}\text{O}$ values of dated zircons from this study, plotted against a compilation from Valley et al. (2005) and Zeh et al. (2014). Zircons from the Greenland (Hiess et al., 2011) and Northern Australian (Hollis et al., 2014) cratons where low $\delta^{18}\text{O}$ values were reported are plotted for comparison.

Figure 4. εHf_t vs. age diagrams for: (a) igneous and metamorphic zircons (this study). The diagonal grey arrows indicate crustal evolution trends derived from the Hf data of this study, using an average $^{176}\text{Lu}/^{177}\text{Hf}$ of 0,0113 for the average continental crust. The arrays 1 and 2 are fitted through the average compositions of the granit-

oids and gneisses of the (1) Bação and (2) Bonfim + Belo Horizonte complexes respectively, indicating that the complexes evolved as different terranes or portions of the crust, and (b) detrital zircons (data from Koglin et al., 2014; Moreira et al., 2016; Martínez Dopico, unpublished) of the SSFC. The Depleted Mantle (DM) evolution is as in Fig. 2.

Figure 5. Plots of ϵHf_t vs. geochemical parameters, showing general trends of enrichment with decreasing age and ϵHf_t of the magmas (grey arrows). The grey fields represent TTG compositional fields from Moyen (2011).

Figure 6. $\delta^{18}\text{O}$ vs. U-Th-Pb systematics diagrams. The lack of apparent correlations indicates that the $\delta^{18}\text{O}$ values are not products of secondary alteration and can instead be considered as primary features (see text).

Figure 7. (a) U-Pb age distribution for detrital zircons from the SSFC (grey bars; right scale) (data from Koglin et al., 2014; Moreira et al., 2016; Martínez Dopico, unpublished). Cumulative/integral curves of U-Pb ages (light blue) and of crustal T_{DM} ages (dark blue). The green line represents the integrated curve obtained from zircons defined as “juvenile” (with $\epsilon\text{Hf}_t > 0.75 * \epsilon\text{Hf}_{\text{DM}}$). (b) ϵHf_t vs. age diagram. The integral crust (grey line) is calculated following the method of Belousova et al. (2010). The proportion of juvenile component (red curve; right scale) is calculated in reference to the depleted mantle evolution from Guitreau (2012) and the integral crust (see text for explanation).

Figure 8. Sketch illustrating the evolution of the Archean continental crust exposed in the SSFC. The diagrams are explained in more details in Section 5.3. (a) From 2.9 Ga, the accretion of several proto-continents progressively leads to a tectonically thickened continental nucleus that undergoes differentiation, producing the medium-K magmas characteristic for the RVI and RVII events. The felsic volcanism associated with RVII is recorded in the greenstone belt sequence; (b) During the post-collisional Mamona event (< 2.75 Ga), the induced inflow of asthenospheric mantle beneath the orogenic prism triggers further melting of lower crustal and upper mantle lithologies.

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Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Table 1

	U-Th-Pb Session 1	U-Th-Pb Session 2	Lu-Hf Session 1	Lu-Hf Session 2
Instrument	Agilent 7700	Element 2	Thermo-Finnigan Neptune	Thermo-Scientific Neptune+
Scan mode	E-Scan	E-Scan	Static	Static
Scanned masses	204, 206, 207, 208, 232, 238	202, 204, 206, 207, 208, 232, 238	172, 173, 175, 176, 177, 178, 179, 180	172, 173, 175, 176, 177, 178, 179, 180
Mass resolution	300	300	300	300
Dwell time	10 (204), 15 (206), 40 (207), 10 (208), 10 (232), 15 (238) ms	4 (202), 4 (204), 14 (206), 20 (207), 10 (208), 10 (232), 14 (238) ms	0.52 s	0.52 s
Integration time	0.9 s	0.9 s	1 s	1 s
Background	20 s	19 s	-	-
Ablation time	40 s	30 s	36 s	36 s
Carrier gas	1.0 L/min He (+ 1.0 L/min Ar)	0.5 L/min He (+ 1.0 L/min Ar)	0.6 L/min He (+ 0.9 L/min Ar, 0.007 L/min N)	1.3 L/min He (+ 1.0 L/min Ar, 0.15 L/min N)
Laser	213 nm New Wave	CETAC LSX-213 G2+	Resonetics RESolution M-50	Photon Machines 193
Spot size	30 μm	20 μm	40 μm	40-50 μm
Laser settings	10 Hz, 6-8 J/cm ²	10 Hz, 3.5 J/cm ²	6 Hz, 4 J/cm ²	5 Hz, 3 J/cm ²
Cell volume	Low (teardrop)	Low (teardrop)	Low (Two volume cell)	Low (Two volume cell)

Table 2

Sample spot			Grain descrip- tion ^b	α -decay events/ mg (x 10 ¹⁵ α /mg)	Th/ U	²⁰⁴ (H g + Pb)	²³⁸ U/ ²⁰⁶ Pb	1 σ	²⁰⁷ Pb/ ²⁰⁶ Pb	1 σ	Pb/P b age (Ma)	1 σ	Con c. (%)	¹⁷⁶ Lu/ ¹⁷⁷ Hf	2 σ	¹⁷⁶ Hf/ ¹⁷⁷ Hf _t	2 σ	ϵ_{Hf}	2 σ	$\delta^{18}\text{O}_{\text{VSM}}$ ow (‰)	2 σ
Lu-Hf	U-Pb _a	O																			
Bação complex FQ1																					
14_1	14		p m os	4.81	0.4 6	21	2.031	0.01 5	0.1874	0.001 7	2719	1 5	97.1	0.00159	1 1	0.28098 8	2 8	-1.7	1. 0		
16_1	16		p m os	3.94	0.7 4	27	1.909	0.01 5	0.1869	0.001 8	2715	1 6	100. 0	0.00158	1 0	0.28108 7	3 9	1.8	1. 4		
22_1	22		p m os	3.80	0.8 2	46	2.075	0.01 9	0.1854	0.002 1	2702	1 9	96.5	0.00167	1 1	0.28099 5	2 2	-1.8	0. 8		
39_1	39		p m os	5.08	0.3 6	31	1.952	0.01 5	0.1865	0.002 1	2711	1 8	99.1	0.00173	1 1	0.28102 8	2 7	-0.4	1. 0		
58b_1	58		p m os	2.96	0.3 4	9	1.917	0.01 7	0.1860	0.002 4	2707	2 1	100. 0	0.00088	6	0.28098 1	3 6	-2.2	1. 3		
75_1	75		p m os	3.15	1.1 3	5	1.903	0.01 6	0.1872	0.002 4	2718	2 1	100. 1	0.00156	1 0	0.28102 4	2 6	-0.4	0. 9		
12a_2	15	8	p m os		0.2 8	0	2.073	0.01 5	0.1853	0.002 1	2701	1 9	96.4	0.00111	8	0.28100 4	3 8	-1.5	1. 4	7.00	0.1 9
15a_2	18	7	p m os		0.2 4	474	2.331	0.01 6	0.1786	0.002 4	2640	2 2	92.0	0.00187	1 1	0.28100 7	1 4	-2.9	0. 5	6.42	0.1 8
15b_2	18	7	p m os		0.2 4	474	2.331	0.01 6	0.1786	0.002 4	2640	2 2	92.0	0.00187	1 1	0.28101 3	2 0	-2.6	0. 7	6.42	0.1 8
17a_2	19	6	p m os		0.8 0	568	3.431	0.02 5	0.1665	0.001 9	2522	1 9	74.4	0.00144	1 6	0.28096 0	2 4	-7.3	0. 8	4.11	0.3 0
16a_2	21		p m os		0.6 6	239	2.329	0.01 7	0.1816	0.002 0	2667	1 8	91.4	0.00124	9	0.28095 1	1 6	-4.2	0. 6		
16b_2	22		p m os		0.6 0	0	3.258	0.02 4	0.1661	0.001 9	2519	1 9	77.5	0.00109	7	0.28096 6	3 3	-7.1	1. 2		
18a_2	29		p m os		0.5 9	529	2.136	0.01 8	0.1863	0.002 8	2710	2 5	94.7	0.00175	1 1	0.28095 0	1 6	-3.3	0. 6		
18b_2	29		p m os		0.5 9	529	2.136	0.01 8	0.1863	0.002 8	2710	2 5	94.7	0.00121	8	0.28092 7	1 9	-4.1	0. 7		
20a_2	30	5	p m os		0.4 4	150	2.816	0.02 0	0.1671	0.001 8	2528	1 8	85.1	0.00094	6	0.28099 8	1 7	-5.8	0. 6	6.24	0.2 1
21a_2	32	3	p m os		0.5	292	7.198	0.05	0.1311	0.001	2112	2	48.0	0.00151	1	0.28102	2	-	0.	4.76	0.2

				6			8			7			3			1			2			5			14.6			9			1		
23b_2	34	4	p m os	0.04	60	2.916	0.024	0.1687	0.0023	2544	22	82.8	0.00182	13	0.280963	25	-6.79	0.9	5.28	0.22													
22a_2	37	1	p m os	0.17	494	4.503	0.031	0.1248	0.0015	2026	21	76.4	0.00272	19	0.281014	17	-16.99	0.6	7.34	0.19													
26b_2	43	11	p m os	0.18	581	6.552	0.050	0.1389	0.0019	2213	23	48.6	0.00205	13	0.281023	18	-12.22	0.6	5.75	0.26													
24b_2	47	12	p m os	0.71	0	2.778	0.026	0.1696	0.0040	2554	38	85.2	0.00135	8	0.281014	22	-4.68	0.8	4.65	0.27													
25b_2	47	12	p e os	0.71	0	2.778	0.026	0.1696	0.0040	2554	38	85.2	0.00128	8	0.281016	17	-4.56	0.6	4.65	0.27													
13a_2	49		p e os	0.58	287	4.967	0.036	0.1451	0.0017	2289	20	61.4	0.00122	7	0.280998	20	-11.44	0.7															
FQ2																																	
24_1	24		p m os	2.95	0.51	13	1.7792	0.014	0.2059	0.0022	2874	17	100.0	0.00418	35	0.281020	28	3.10	1.0														
27a_1	27		p m os	0.98	0.51	38	1.7739	0.017	0.2042	0.0026	2860	21	100.4	0.00167	11	0.281016	31	2.61	1.1														
27b_1	27		p m os	0.98	0.51	38	1.7739	0.017	0.2042	0.0026	2860	21	100.4	0.00079	5	0.280991	19	1.77	0.7														
43_1	43		p m os	0.80	0.34	28	1.7636	0.016	0.2090	0.0025	2898	19	100.0	0.00157	10	0.281006	19	3.27	0.7														
45_1	45		p m os	2.56	0.45	0	1.7983	0.014	0.2030	0.0020	2850	16	100.0	0.00241	15	0.281060	29	4.00	1.0														
53_1	53		p e rx	2.09	0.32	24	1.9291	0.020	0.1848	0.0032	2696	29	99.9	0.00134	10	0.281040	31	-0.41	1.1														
56b_1	56		p m os	2.37	0.52	3	1.7658	0.014	0.2083	0.0024	2892	18	100.0	0.00302	18	0.281050	23	4.68	0.8														
57_1	57		p m os	2.24	0.47	21	1.7874	0.015	0.2049	0.0022	2866	17	100.0	0.00245	15	0.280995	25	2.09	0.9														
61_1	61		p m os	2.39	0.44	0	1.7941	0.014	0.2033	0.0022	2853	18	100.1	0.00225	14	0.281037	18	3.26	0.6														
62_1	62		p m os	2.85	0.50	33	1.7864	0.015	0.2048	0.0022	2865	17	100.0	0.00196	12	0.281048	22	3.98	0.8														
67_1	67		p m os	2.34	0.59	41	1.7683	0.016	0.2078	0.0027	2888	21	100.0	0.00523	34	0.281106	25	6.59	0.9														
69_1	69		p m os	2.67	0.4	14	1.7902	0.01	0.2044	0.002	2861	1	100.	0.00315	3	0.28097	3	1.3	1.														

70_1	70		ane os	2.87	0.03	12	1.9301	0.018	0.1843	0.0031	2692	62	0100.	0.00246	44	80.28101	53	-1.3	31.2					
71_1	71		pe os	4.79	0.02	0	2.8981	0.027	0.1718	0.0023	2575	22	84.9	0.00269	21	0.281035	25	-3.4	0.9					
73_1	73		ov m h	2.50	0.47	9	1.7879	0.015	0.2045	0.0027	2862	21	100.0	0.00125	88	0.281018	17	2.7	0.6					
74_1	74		p m os	2.35	0.49	8	1.7894	0.014	0.2049	0.0023	2866	18	99.9	0.00294	18	0.281002	24	2.3	0.8					
75_1	75		per x	3.88	0.07	0	1.9289	0.016	0.1854	0.0021	2702	18	99.8	0.00262	24	0.281025	25	-0.8	0.9					
76_1	76		pe h	3.48	0.21	0	2.6047	0.022	0.1867	0.0027	2713	24	86.4	0.00242	20	0.281012	23	-1.0	0.8					
32_2	15	10	p m os		0.21	163	3.0019	0.024	0.1747	0.0021	2603	20	79.5	0.00191	12	0.280985	17	-4.5	0.6	3.93		0.22		
40_2	19		ov m os		0.28	53	2.1265	0.016	0.1902	0.0022	2744	19	94.2	0.00115	7	0.281038	16	0.7	0.6					
41_2	26	11	per x		0.04	50	3.2286	0.023	0.1598	0.0018	2454	19	79.9	0.00109	7	0.281030	23	-9.4	0.8	4.80		0.20		
42_2	28	12	p m os		0.36	36	2.3211	0.017	0.1880	0.0021	2724	18	90.2	0.00257	23	0.281018	23	-0.5	0.8	4.64		0.14		
39_2	32	1	per x		0.03	186	7.0636	0.057	0.0910	0.0013	1446	28	78.7	0.00175	12	0.281060	22	-28.6	0.8	3.11		0.19		
37_2	39	2	p m os		0.42	0	1.9508	0.018	0.2033	0.0033	2853	26	96.0	0.00233	16	0.281034	22	3.1	0.8	3.06		0.14		
36_2	40		p m os		0.24	81	2.3268	0.018	0.1807	0.0023	2659	21	91.6	0.00267	17	0.280989	33	-3.1	1.2					
35_2	44	6	per x		0.02	0	3.0307	0.022	0.1621	0.0019	2478	20	82.6	0.00112	7	0.281076	20	-4.2	0.7	3.77		0.21		
27_2	46		p m os		0.43	0	1.9391	0.017	0.2036	0.0032	2855	25	96.3	0.00189	14	0.281019	24	2.6	0.9					
29_2	48	8	pe os		0.04	95	2.7267	0.020	0.1649	0.0020	2507	21	87.4	0.00170	10	0.281019	41	-5.6	1.4	4.62		0.14		
FQ13																								
23	23		p c h	5.80	0.32	10	1.7194	0.013	0.2093	0.0017	2900	13	101.1	0.00181	11	0.280883	26	-1.2	0.9					
26	26		p m h	1.33	0.49	0	1.8431	0.018	0.1965	0.0026	2797	22	100.0	0.00127	8	0.281030	27	1.6	1.0					
28	28		p m os	1.98	0.67	22	1.8597	0.016	0.1957	0.0018	2790	15	99.7	0.00133	10	0.281029	33	1.4	1.2					

33	33	p m os	1.30	0.5 6	13	1.9248	0.01 9	0.1854	0.002 0	2702	1 8	99.9	0.00127	8	0.28103 6	3 0	-0.4	1. 1
34	34	p m h	0.90	0.5 7	22	1.8415	0.01 6	0.1953	0.002 1	2788	1 7	100. 2	0.00088	6	0.28101 0	2 2	0.7	0. 8
38	38	p m os	1.49	0.6 8	7	1.8400	0.01 5	0.1964	0.001 8	2796	1 5	100. 0	0.00143	9	0.28105 9	2 9	2.7	1. 0
39	39	p m os	0.87	0.5 2	9	2.0158	0.01 9	0.1832	0.002 2	2682	2 0	98.2	0.00100	8	0.28104 6	2 5	-0.5	0. 9
49	49	p m os	2.24	0.3 4	5	1.8515	0.01 5	0.1947	0.001 8	2783	1 5	100. 0	0.00082	8	0.28102 2	2 3	1.0	0. 8
50	50	p m os	5.44	0.4 1	27	1.8452	0.01 6	0.1947	0.001 6	2783	1 4	100. 2	0.00079	1 0	0.28102 7	2 1	1.2	0. 8
57	57	p m h	3.00	0.1 1	3	1.8375	0.01 5	0.1969	0.001 6	2800	1 4	100. 0	0.00081	7	0.28094 5	1 5	-1.3	0. 6
58	58	p m h	3.83	0.0 9	22	1.8238	0.01 5	0.1968	0.001 6	2800	1 4	100. 4	0.00084	7	0.28090 9	2 1	-2.6	0. 7
59	59	p m os	1.11	0.5 8	26	1.8363	0.01 6	0.1938	0.002 0	2775	1 7	100. 6	0.00118	8	0.28104 4	3 0	1.6	1. 1
61	61	p m os	1.49	0.6 9	19	1.8431	0.01 5	0.1958	0.002 0	2791	1 6	100. 1	0.00189	1 3	0.28107 5	3 1	3.1	1. 1
68	68	p m h	1.67	0.6 7	9	1.8433	0.01 5	0.1959	0.001 8	2792	1 5	100. 0	0.00129	9	0.28104 9	2 7	2.2	1. 0
72	72	p m os	0.65	0.3 9	0	1.8428	0.02 0	0.1958	0.002 7	2791	2 3	100. 1	0.00140	1 4	0.28102 5	3 4	1.3	1. 2
83	83	p m os	1.25	0.5 6	7	1.7996	0.01 5	0.1964	0.002 0	2796	1 7	101. 1	0.00132	8	0.28105 2	1 7	2.4	0. 6
84	84	p m os	1.33	0.6 4	14	1.8645	0.01 6	0.1930	0.001 9	2768	1 6	100. 0	0.00100	8	0.28103 2	2 3	1.0	0. 8
FQ17																		
10aI_1	10_a	ov m os	5.40	0.0 6	23	1.9114	0.01 7	0.1883	0.001 8	2727	1 6	99.7	0.00125	9	0.28106 2	3 0	1.1	1. 1
10bI_1	10_a	ov m os	5.40	0.0 6	23	1.9114	0.01 7	0.1883	0.001 8	2727	1 6	99.7	0.00102	9	0.28105 6	2 8	0.9	1. 0
11I_1	11_a	eq m os	5.76	0.0 5	5	1.9041	0.01 5	0.1893	0.001 9	2736	1 7	99.7	0.00082	5	0.28106 0	2 7	1.3	1. 0
14I_1	14_a	p m os	5.13	0.0 5	9	1.9076	0.01 5	0.1874	0.001 8	2719	1 6	99.9	0.00067	4	0.28101 5	2 2	-0.7	0. 8
15I_1	15_a	p m os	0.46	0.3 3	0	1.9055	0.03 0	0.1884	0.004 3	2728	3 7	99.8	0.00044	3	0.28099 9	2 3	-1.1	0. 8
19II_1	19_b	p e os	4.00	0.0 7	12	1.9006	0.01 5	0.1881	0.001 8	2726	1 6	100. 0	0.00036	2	0.28103 5	1 8	0.2	0. 7

23I_1	23_a		p m os	6.40	0.4 3	16	1.8891	0.01 5	0.1896	0.001 7	2739	1 5	100. 0	0.00061	4	0.28098 7	1 9	-1.2	0. 7			
30I_1	30_a		p m os	5.40	0.3 5	9	1.8539	0.01 4	0.1941	0.001 8	2778	1 5	100. 1	0.00119	9	0.28101 4	2 2	0.6	0. 8			
32I_1	32_a		p m os	1.48	0.0 6	6	1.8577	0.01 6	0.1944	0.002 1	2780	1 7	99.9	0.00091	1 0	0.28096 3	1 7	-1.1	0. 6			
33I_1	33_a		p m os	4.21	1.0 9	9	1.8560	0.01 5	0.1942	0.001 9	2778	1 6	100. 0	0.00110	1 1	0.28098 9	2 5	-0.3	0. 9			
33II_1	33_b		p m os	4.55	0.2 0	20	1.9454	0.01 8	0.1850	0.002 5	2698	2 2	99.5	0.00085	6	0.28101 8	2 9	-1.1	1. 0			
35II_1	35_b		p c h	7.34	0.4 1	48	1.7840	0.01 5	0.2055	0.002 2	2871	1 7	100. 0	0.00166	1 4	0.28091 3	2 4	-0.8	0. 9			
42I_1	42_a		p m h	3.64	0.2 5	59	1.8551	0.01 4	0.1946	0.001 8	2782	1 5	99.9	0.00055	4	0.28097 9	2 0	-0.5	0. 7			
43all_1	43_b		p m h	3.10	0.1 6	25	1.9035	0.01 7	0.1874	0.002 4	2720	2 1	100. 0	0.00059	4	0.28099 9	2 1	-1.3	0. 7			
43bII_1	43_b		p m h	3.10	0.1 6	25	1.9035	0.01 7	0.1874	0.002 4	2720	2 1	100. 0	0.00081	6	0.28100 6	1 5	-1.0	0. 6			
50II_1	50_b		p c os	4.54	0.4 6	58	1.8170	0.01 5	0.1998	0.002 4	2825	1 9	100. 0	0.00094	6	0.28093 4	2 1	-1.1	0. 7			
51II_1	51_b		p e h	7.38	0.5 4	24	1.8810	0.01 6	0.1906	0.002 4	2747	2 0	100. 0	0.00057	4	0.28100 0	1 8	-0.6	0. 6			
52II_1	52_b		p m os	3.66	0.2 9	43	1.8892	0.01 6	0.1900	0.002 1	2742	1 8	99.9	0.00076	5	0.28101 2	2 1	-0.3	0. 7			
56II_1	56_b		p m os	4.33	0.3 6	3	1.8983	0.01 7	0.1910	0.002 3	2751	1 9	99.5	0.00148	1 0	0.28104 5	2 2	1.1	0. 8			
61II_1	61_b		p c h	2.28	0.3 3	46	1.7882	0.01 9	0.2039	0.003 9	2858	3 1	100. 1	0.00121	8	0.28103 2	1 9	3.2	0. 7			
48_2	12	8	p m os		0.3 2	38	1.9276	0.01 5	0.1847	0.002 0	2695	1 8	100. 0	0.00071	6	0.28104 2	2 9	-0.3	1. 0	5.76	0.1 5	
47_2	14	7	p m h		0.5 1	310	3.3735	0.02 4	0.1841	0.002 0	2690	1 8	70.0	0.00175	2 1	0.28098 8	1 8	-2.4	0. 6	4.00	0.2 2	
45_2	15	5	p e os		0.2 4	0	2.0449	0.01 6	0.1859	0.003 0	2706	2 6	97.0	0.00113	7	0.28101 5	1 8	-1.0	0. 6	6.29	0.1 9	
46_2	16	6	p m os		0.3 9	160	2.4549	0.01 8	0.1755	0.001 9	2611	1 8	90.1	0.00072	6	0.28099 1	1 6	-4.1	0. 6	5.46	0.2 1	
44_2	17		p m os		0.6 3	83	1.9270	0.01 4	0.1865	0.002 1	2711	1 8	99.7	0.00143	1 1	0.28098 5	1 8	-2.0	0. 7			
43_2	18		p e os		0.0 4	157	2.0333	0.01 4	0.1868	0.002 1	2714	1 8	97.0	0.00137	1 0	0.28099 8	1 8	-1.4	0. 6			
52_2	19		p m os		0.3	153	1.8932	0.01	0.1870	0.002	2716	1	100.	0.00123	1	0.28100	2	-1.2	0.			

Bonfim Complex

FQ29																	
20_2a	15	p m os	0.3 1	24	2.0304	0.01 6	0.1926	0.002 1	2764	1 8	96.1	0.00089	7	0.28088 3	1 5	-4.3	0. 6
19_2a	16	p m h	0.2 3	46	1.8707	0.01 5	0.1935	0.002 1	2772	1 7	99.8	0.00072	5	0.28088 1	1 5	-4.2	0. 5
21_2a	17	p n os	0.2 1	0	1.8834	0.01 4	0.1949	0.002 5	2784	2 0	99.2	0.00121	8	0.28091 7	2 1	-2.7	0. 8
23_2a	21	p m os	0.2 5	71	1.7063	0.01 3	0.1967	0.002 5	2799	2 0	103. 6	0.00134	9	0.28088 9	1 2	-3.3	0. 4
24_2a	21	p m os	0.2 5	71	1.7063	0.01 3	0.1967	0.002 5	2799	2 0	103. 6	0.00084	5	0.28089 5	1 6	-3.1	0. 6
49_2a	31	p m os	0.4 2	10	2.1532	0.01 6	0.1943	0.002 5	2779	2 1	93.2	0.00153	1 1	0.28090 2	1 8	-3.3	0. 6
25_2a	32	p m os	0.3 0	80	1.8770	0.01 5	0.1917	0.002 2	2757	1 8	99.9	0.00058	4	0.28090 7	1 3	-3.7	0. 5
26_2a	33	p m os	0.2 1	75	2.1377	0.01 6	0.1895	0.002 4	2738	2 0	94.4	0.00137	9	0.28089 4	1 7	-4.6	0. 6
27_2a	33	p m os	0.2 1	75	2.1377	0.01 6	0.1895	0.002 4	2738	2 0	94.4	0.00169	1 5	0.28090 2	1 9	-4.3	0. 7
29_2a	34	p m os	0.2 7	7	1.8559	0.01 5	0.1945	0.002 1	2781	1 8	100. 0	0.00129	8	0.28090 0	1 6	-3.4	0. 6
28_2a	35	p m os	0.2 2	0	1.8534	0.01 4	0.1947	0.002 2	2782	1 8	100. 0	0.00173	1 5	0.28090 1	1 7	-3.3	0. 6
30_2a	36	p m os	0.3 5	49	1.8528	0.01 5	0.1922	0.002 1	2761	1 7	100. 4	0.00138	8	0.28088 8	1 7	-4.3	0. 6
31_2a	48	p m os	0.5 6	36	1.8794	0.01 4	0.1929	0.002 4	2767	2 0	99.7	0.00165	1 2	0.28090 2	2 2	-3.6	0. 8
36_2a	50	p m os	0.4 6	112	1.8526	0.01 6	0.1959	0.002 7	2792	2 3	99.8	0.00111	8	0.28091 9	1 7	-2.4	0. 6
34_2a	52	p e os	0.2 6	0	2.1081	0.01 8	0.1938	0.002 8	2774	2 3	94.2	0.00078	5	0.28090 2	1 4	-3.5	0. 5
38_2a	69	p m os	0.4 2	32	2.5144	0.02 3	0.1939	0.003 9	2776	3 3	86.6	0.00124	1 0	0.28090 1	1 5	-3.5	0. 5
39_2a	70	p m os	0.2 8	0	2.0041	0.01 7	0.1926	0.002 9	2765	2 4	96.7	0.00136	9	0.28089 7	1 6	-3.8	0. 6
72_2b	11	p e os	0.1 9	181	2.3455	0.01 4	0.1907	0.002 0	2748	1 7	89.0	0.00130	8	0.28092 2	2 4	-3.4	0. 8
73_2b	16	p e os	0.1 7	481	2.3599	0.01 4	0.1940	0.002 1	2776	1 8	88.0	0.00082	6	0.28092 2	1 8	-2.7	0. 6
74_2b	17	p m os	0.1 9	132	1.8802	0.01 3	0.1967	0.002 5	2799	2 1	99.0	0.00142	9	0.28092 1	2 2	-2.2	0. 8

78_2b	19		p m os	0.1 9	220	2.0579	0.01 3	0.1953	0.002 0	2787	1 7	94.8	0.00133	8	0.28091 7	2 7	-2.6	0. 9		
77_2b	20		p m os	0.2 0	0	1.9227	0.01 2	0.1954	0.002 0	2788	1 6	98.1	0.00169	1 0	0.28094 3	2 6	-1.7	0. 9		
76_2b	21		p m os	0.2 1	106	2.4086	0.01 5	0.1903	0.001 9	2745	1 6	87.7	0.00164	1 1	0.28091 5	2 2	-3.7	0. 8		
75_2b	26		p m os	0.1 7	357	1.8789	0.01 4	0.1958	0.002 3	2792	1 9	99.1	0.00149	9	0.28092 8	2 0	-2.1	0. 7		
12_2b	32		p m os	0.2 1	0	2.0933	0.01 3	0.1977	0.002 1	2808	1 7	93.5	0.00224	1 7	0.28094 9	2 2	-1.0	0. 8		
11_2b	34		p e os	0.1 8	365	2.6075	0.01 6	0.1910	0.002 0	2750	1 7	83.2	0.00159	1 0	0.28090 8	1 7	-3.8	0. 6		
80_2b	35	8	p m os	0.2 6	289	2.5051	0.01 6	0.1915	0.002 0	2755	1 7	85.3	0.00112	7	0.28091 0	1 6	-3.6	0. 6	4.95	0.2 3
16_2b	47	7	p e os	0.1 5	0	2.0930	0.01 3	0.1948	0.002 0	2783	1 7	94.1	0.00145	9	0.28089 3	1 9	-3.6	0. 7	5.29	0.2 5
15_2b	49		p m os	0.2 6	133	2.6635	0.01 7	0.1904	0.002 1	2746	1 8	82.2	0.00147	9	0.28090 6	1 6	-4.0	0. 6		
13_2b	51	9	p e os	0.1 9	312	2.1952	0.01 4	0.1981	0.002 2	2811	1 8	91.0	0.00143	9	0.28090 5	1 6	-2.5	0. 6	3.46	0.1 5
14_2b	53	10	p m os	0.2 3	226	2.0130	0.01 4	0.1947	0.002 1	2782	1 8	95.9	0.00111	8	0.28088 6	1 7	-3.8	0. 6	4.90	0.1 4
21_2b	59	11	p m os	0.1 5	324	1.9404	0.01 3	0.1966	0.002 1	2798	1 7	97.5	0.00118	9	0.28091 5	1 8	-2.4	0. 6	6.03	0.2 1
20_2b	62	14	p m os	0.1 2	177	1.9064	0.01 3	0.1966	0.002 1	2799	1 7	98.3	0.00060	8	0.28097 2	1 6	-0.4	0. 6	4.73	0.1 8
19_2b	64		p m os	0.1 9	0	2.1561	0.01 4	0.1987	0.002 4	2815	1 9	91.8	0.00152	1 1	0.28092 4	2 3	-1.7	0. 8		
18_2b	66	3	p m os	0.1 5	0	2.1031	0.01 4	0.1959	0.002 1	2792	1 8	93.6	0.00124	1 1	0.28092 3	1 7	-2.3	0. 6	5.23	0.2 6
27_2b	69	5	p m os	0.2 4	103	1.8554	0.01 3	0.1949	0.002 2	2784	1 8	100. 0	0.00142	9	0.28090 0	2 0	-3.3	0. 7	4.86	0.1 5
28_2b	69	5	p m os	0.2 4	103	1.8554	0.01 3	0.1949	0.002 2	2784	1 8	100. 0	0.00160	1 2	0.28094 6	1 6	-1.7	0. 6	4.86	0.1 5
25_2b	71	6	p m os	0.2 0	354	2.8988	0.02 0	0.1810	0.002 0	2662	1 8	79.8	0.00125	8	0.28090 9	2 0	-5.8	0. 7	5.80	0.1 4
26_2b	71	6	p m os	0.2 0	354	2.8988	0.02 0	0.1810	0.002 0	2662	1 8	79.8	0.00100	6	0.28089 4	2 0	-6.4	0. 7	5.80	0.1 4
24_2b	74	13	p m os	0.2 0	0	2.1463	0.01 4	0.1943	0.002 1	2779	1 7	92.9	0.00118	8	0.28092 2	1 7	-2.6	0. 6	5.04	0.2 2
22_2b	76		p m os	0.1	158	2.1085	0.01	0.1967	0.002	2799	1	93.3	0.00134	8	0.28091	1	-2.4	0.		

23_2b	77	12	p m os	4 0.1 8	144	1.7882	4 0.01 2	0.1990	2 0.002 2	8 2818	1 101. 0	0.00114	9	5 0.28090 7	8 1 9	-2.2	6 0. 7	5.11	0.1 8
FQ41																			
14	14		p m os	0.2 5	134	2.0479	0.01 7	0.1988	0.002 5	8 2816	1 9	0.00128	8	5 0.28088 3	8 1 8	-3.1	6 0. 6		
13	15		p m os	0.3 2	51	1.8742	0.01 3	0.2029	0.002 3	8 2850	1 8	0.00099	6	5 0.28089 0	8 2 1	-2.1	7 0. 7		
15	17		p e os	0.2 9	133	2.4565	0.02 1	0.2009	0.002 4	8 2834	1 9	0.00138	1 0	5 0.28090 7	8 2 0	-1.9	7 0. 7		
16	17		p e os	0.2 9	133	2.4565	0.02 1	0.2009	0.002 4	8 2834	1 9	0.00116	7	5 0.28091 2	8 6	-1.7	6 0. 6		
12	18		p m os	0.3 9	63	2.1878	0.01 7	0.2034	0.002 6	8 2853	2 0	0.00134	1 0	5 0.28090 8	8 1 9	-1.4	7 0. 7		
19	21		p m os	0.3 1	58	1.8235	0.01 4	0.2029	0.002 3	8 2850	1 8	0.00111	7	5 0.28090 3	8 1 7	-1.6	6 0. 6		
17	23		p m os	0.3 4	138	2.0642	0.01 7	0.1957	0.002 4	8 2791	1 9	0.00124	8	5 0.28088 5	8 2 5	-3.7	9 0. 9		
18	24		p m os	0.3 8	53	2.2154	0.01 7	0.1981	0.002 1	8 2810	1 7	0.00137	8	5 0.28091 9	8 1 9	-2.0	7 0. 7		
20	25		p m os	0.2 7	241	2.4523	0.02 1	0.2039	0.002 5	8 2857	2 0	0.00191	1 4	5 0.28089 9	8 1 4	-1.6	5 0. 5		
23	32		p m os	0.3 8	17	2.4668	0.01 9	0.1955	0.002 1	8 2789	1 8	0.00124	1 2	5 0.28090 8	8 2 7	-2.9	9 0. 9		
22	39		p m os	0.2 9	43	1.9341	0.01 5	0.2029	0.002 3	8 2850	1 8	0.00102	6	5 0.28090 9	8 2 5	-1.4	9 0. 9		
25	58		p m os	0.3 9	55	1.8561	0.01 4	0.2024	0.002 4	8 2846	1 8	0.00151	1 1	5 0.28089 4	8 2 1	-2.1	7 0. 7		
26	62		p m os	0.2 5	88	1.8318	0.01 4	0.1970	0.002 8	8 2802	2 2	0.00150	1 1	5 0.28088 6	8 2 2	-3.4	8 0. 8		
27	62		p m os	0.2 5	88	1.8318	0.01 4	0.1970	0.002 8	8 2802	2 2	0.00106	1 1	5 0.28090 6	8 1 8	-2.7	6 0. 6		
28	63		eq m os	0.4 0	51	2.0871	0.01 6	0.2020	0.002 4	8 2842	1 9	0.00120	8	5 0.28090 7	8 2 1	-1.7	7 0. 7		
29	66		p e os	0.3 1	46	2.0283	0.01 5	0.2024	0.002 9	8 2846	2 3	0.00105	7	5 0.28089 0	8 1 9	-2.2	7 0. 7		
30	67		eq e os	0.2 4	10	1.8288	0.01 6	0.2055	0.002 6	8 2870	2 1	0.00124	8	5 0.28086 7	8 7	-2.4	6 0. 6		
FQ52																			
11_2a	13_7		p m h	0.2	0	1.8873	0.01	0.1903	0.003	2745	2	99.9	0.00097	7	5 0.28099	8 1	-0.8	0.	

12_2a	13_7		p m h	6	0	1.8873	0.01	0.1903	0.003	2745	2	99.9	0.00102	1	0.28098	1	-1.3	0.		
				6			7		0		6			0	1	6		6		
13_2a	16_2		p m os	0.3	47	2.0083	0.01	0.1895	0.002	2738	1	97.2	0.00120	7	0.28100	1	-0.8	0.		
	3			5			5		3		9				2	9		7		
15_2a	17_7		p m h	0.2	118	2.0043	0.02	0.1908	0.003	2749	2	97.0	0.00060	4	0.28098	2	-1.1	0.		
				2			1		2		7				6	0		7		
16_2a	19_2		p m os	0.2	5	1.8867	0.01	0.1905	0.002	2747	1	99.9	0.00070	7	0.28098	1	-1.3	0.		
	3			1			4		1		8				3	8		6		
17_2a	21_2		p m os	0.1	71	1.8939	0.01	0.1856	0.002	2704	1	100.	0.00079	5	0.28096	1	-2.7	0.		
	3			6			5		0		8	6			8	8		6		
19_2a	24_2		p m os	0.2	0	1.9875	0.01	0.1887	0.002	2731	1	97.8	0.00069	5	0.28096	2	-2.2	0.		
	3			6			5		1		8				6	0		7		
26_2a	44_2		p e os	0.1	28	1.9433	0.01	0.1904	0.002	2746	2	98.5	0.00083	5	0.28097	2	-1.4	0.		
	3			7			7		4		0				7	3		8		
20_2a	58_7		p m os	0.2	41	1.9175	0.01	0.1876	0.003	2721	2	99.7	0.00061	4	0.28099	2	-1.4	0.		
				6			6		3		8				6	0		7		
21_2a	61_7		p m os	0.3	0	1.9482	0.01	0.1892	0.003	2735	3	98.6	0.00073	5	0.28100	2	-0.8	0.		
				5			8		6		1				1	4		8		
22_2a	64_7		p m h	0.1	7	1.8905	0.02	0.1900	0.004	2742	3	99.9	0.00032	2	0.28097	1	-1.8	0.		
				7			3		0		4				0	9		7		
27_2a	83_7		p m os	0.2	27	1.9622	0.02	0.1891	0.004	2735	3	98.3	0.00086	1	0.28100	1	-0.7	0.		
				4			0		1		5			0	5	9		7		
52_2b	18	6	p m os	0.6	193	2.0731	0.01	0.1873	0.002	2718	1	96.0	0.00186	1	0.28101	2	-0.7	1.	10.88	0.3
				7			7		0		7			7	5	9		0		5
53_2b	18	6	p m os	0.6	193	2.0731	0.01	0.1873	0.002	2718	1	96.0	0.00172	1	0.28103	2	-0.2	0.	10.88	0.3
				7			7		0		7			1	1	1		7		5
54_2b	20		p m os	0.7	133	2.1038	0.01	0.1872	0.002	2718	2	95.4	0.00118	1	0.28101	3	-0.6	1.		
				4			9		6		3			0	8	1		1		
56_2b	27	3	p m os	0.4	62	2.3487	0.01	0.1725	0.001	2582	1	93.0	0.00176	1	0.28101	2	-3.8	0.	4.48	0.2
				8			8		9		8			1	8	1		7		9
55_2b	29	4	p m h	0.4	235	6.8733	0.05	0.1062	0.001	1735	2	67.7	0.00150	2	0.28105	4	-	1.	2.88	0.2
				8			4		2		0			4	0	0	22.	4		9
																	3			
58_2b	35	9	p m os	0.4	190	4.2757	0.03	0.1574	0.001	2428	1	64.8	0.00142	1	0.28102	2	-7.2	0.	4.17	0.3
				0			5		7		8			2	5	0		7		1
59_2b	48	8	p m os	0.4	52	2.1764	0.01	0.1819	0.002	2670	2	94.7	0.00066	4	0.28099	1	-2.5	0.	6.73	0.3
				4			9		3		1				8	4		5		4
60_2b	48	8	p m os	0.4	52	2.1764	0.01	0.1819	0.002	2670	2	94.7	0.00121	1	0.28100	2	-2.3	0.	6.73	0.3
				4			9		3		1			0	2	0		7		4
61_2b	50		p m os	0.6	232	3.3519	0.02	0.1556	0.001	2408	1	79.3	0.00165	1	0.28102	2	-7.7	0.		

Belo Horizonte complex
FQ60

15_2a	145		p m h	0.5 5	118	1.9164	0.01 5	0.1866	0.002 2	2713	1 9	99.9	0.00070	5	0.28096 7	1 8	-2.6	0. 7		
19_2a	153		p m h	0.3 1	312	2.4612	0.02 0	0.1832	0.002 6	2682	2 3	89.4	0.00071	4	0.28096 8	2 0	-3.2	0. 7		
20_2a	159		p m h	0.2 9	20	2.2232	0.01 7	0.1857	0.002 3	2705	2 0	93.3	0.00084	5	0.28097 3	1 4	-2.5	0. 5		
21_2a	159		p m h	0.2 9	20	2.2232	0.01 7	0.1857	0.002 3	2705	2 0	93.3	0.00082	6	0.28094 8	1 8	-3.4	0. 7		
22_2a	160		p m h	0.4 3	0	3.3584	0.02 9	0.1676	0.002 1	2534	2 0	80.2	0.00056	4	0.28096 8	2 8	-6.7	1. 0		
24_2a	165		p m os	0.3 3	154	1.9724	0.01 6	0.1735	0.002 6	2592	2 3	101. 1	0.00083	5	0.28100 0	2 9	-4.2	1. 0		
25_2a	166		p m os	0.1 0	198	2.8405	0.02 3	0.1668	0.002 3	2526	2 3	86.7	0.00084	7	0.28095 2	2 2	-7.5	0. 8		
16_2a	172		p m h	0.5 4	0	1.9085	0.01 6	0.1883	0.002 6	2728	2 3	99.8	0.00051	4	0.28096 4	1 8	-2.3	0. 6		
71_2b	13	13	p m os	0.1 8	0	6.2996	0.04 9	0.1380	0.001 6	2202	1 9	51.3	0.00140	1 4	0.28101 3	1 7	-12. 9	0. 6	2.86	0.3 3
69_2b	15		p e h	0.5 6	135	2.1648	0.01 7	0.1877	0.002 2	2722	1 9	93.8	0.00073	5	0.28109 6	3 3	2.2	1. 2		
72_2b	20		p m os	0.3 4	29	1.9434	0.01 6	0.1858	0.002 3	2705	2 0	99.4	0.00100	6	0.28093 2	2 0	-4.0	0. 7		
73_2b	30	1	f m os	0.2 2	337	2.8275	0.02 5	0.1884	0.003 0	2728	2 6	79.3	0.00155	1 6	0.28101 1	2 2	-0.7	0. 8	3.12	0.2 7
74_2b	34	5	p m os	0.3 0	577	4.3016	0.03 4	0.1864	0.002 1	2710	1 8	53.7	0.00084	6	0.28095 7	2 5	-3.0	0. 9	3.09	0.3 3
77_2b	43	14	f m os	0.3 0	144	1.9481	0.01 6	0.1859	0.002 1	2706	1 9	99.3	0.00042	3	0.28092 1	2 0	-4.4	0. 7	7.04	0.3 3
76_2b	44	15	f m os	0.2 9	22	1.9203	0.01 6	0.1855	0.002 2	2703	1 9	100. 0	0.00054	3	0.28092 4	2 2	-4.3	0. 8	6.60	0.2 9
78_2b	45	7	ov c h	0.1 2	91	1.8973	0.01 6	0.1961	0.002 4	2794	2 0	98.6	0.00025	3	0.28099 7	1 9	0.4	0. 7	5.41	0.4 0
79_2b	46	8	ov c h	0.2 0	299	1.8430	0.01 6	0.1940	0.002 6	2776	2 2	100. 4	0.00028	2	0.28098 1	1 7	-0.6	0. 6	5.34	0.3 7
11_2b	48		p m os	0.2 5	731	2.6782	0.02 2	0.1891	0.002 4	2734	2 1	82.2	0.00118	7	0.28092 0	1 8	-3.7	0. 6		
80_2b	50	4	p m os	0.2 5	310	3.8220	0.03 1	0.1723	0.002 0	2580	1 9	66.1	0.00125	8	0.28100 0	2 5	-4.5	0. 9	5.01	0.2 9
14_2b	54	9	p m os	0.2 2	266	5.0446	0.04 0	0.1547	0.001 9	2399	2 0	56.0	0.00196	1 5	0.28098 5	1 7	-9.3	0. 6	3.84	0.3 7

FQ74																	
18_2a	16	p m h	0.5 3	0	2.2873	0.01 8	0.1784	0.002 2	2638	2 1	93.4	0.00066	5	0.28099 4	1 9	-3.4	0. 7
23_2a	17	b m h	0.6 9	197	2.4164	0.01 8	0.1620	0.002 0	2477	1 8	94.5	0.00132	1 6	0.28094 2	1 5	-9.0	0. 5
21_2a	18	b m h	0.8 6	112	1.9859	0.01 6	0.1806	0.002 1	2658	1 9	99.4	0.00077	1 1	0.28096 1	1 6	-4.1	0. 6
22_2a	18	b m h	0.8 6	112	1.9859	0.01 6	0.1806	0.002 1	2658	1 9	99.4	0.00071	8	0.28095 7	1 8	-4.2	0. 6
25_2a	20	p m h	0.6 7	35	2.0394	0.01 7	0.1797	0.002 2	2650	2 0	98.3	0.00100	7	0.28097 4	1 8	-3.8	0. 6
26_2a	20	p e h	0.6 7	35	2.0394	0.01 7	0.1797	0.002 2	2650	2 0	98.3	0.00070	5	0.28099 7	1 6	-3.0	0. 6
19_2a	34	p m h	0.7 2	7	2.1548	0.01 8	0.1777	0.002 2	2632	2 1	96.2	0.00122	1 3	0.28097 5	2 2	-4.2	0. 8
20_2a	34	p m h	0.7 2	7	2.1548	0.01 8	0.1777	0.002 2	2632	2 1	96.2	0.00086	8	0.28096 3	2 0	-4.6	0. 7
16_2a	35	b m o s	0.7 5	7	1.9880	0.01 5	0.1781	0.002 2	2636	2 0	99.8	0.00120	9	0.28100 8	2 6	-2.9	0. 9
24_2a	39	p e o s	0.7 3	4	2.1173	0.01 7	0.1791	0.002 2	2644	2 0	96.7	0.00099	1 7	0.28096 1	2 3	-4.4	0. 8
27_2a	43	b e h	0.6 3	138	2.0353	0.01 6	0.1752	0.002 0	2608	1 8	99.3	0.00075	6	0.28096 9	2 0	-5.0	0. 7
28_2a	43	b e h	0.6 3	138	2.0353	0.01 6	0.1752	0.002 0	2608	1 8	99.3	0.00155	1 6	0.28096 6	2 2	-5.1	0. 8
29_2a	55	p m h	0.8 8	0	2.1397	0.02 4	0.1772	0.003 5	2627	3 3	96.6	0.00141	2 0	0.28099 4	2 4	-3.6	0. 9
31_2a	55	p m h	0.8 8	0	2.1397	0.02 4	0.1772	0.003 5	2627	3 3	96.6	0.00111	1 3	0.28096 0	1 8	-4.8	0. 6
32_2a	74	p m h	0.6 1	68	2.1565	0.02 0	0.1790	0.003 1	2644	2 9	95.9	0.00119	9	0.28095 6	2 2	-4.6	0. 8
33_2a	74	p m h	0.6 1	68	2.1565	0.02 0	0.1790	0.003 1	2644	2 9	95.9	0.00141	1 3	0.28095 4	2 0	-4.7	0. 7
38_2a	96	b m h	0.5 9	23	1.9935	0.01 6	0.1794	0.002 6	2647	2 3	99.4	0.00115	1 4	0.28096 4	2 1	-4.2	0. 7
39_2a	96	b m h	0.5 9	23	1.9935	0.01 6	0.1794	0.002 6	2647	2 3	99.4	0.00089	9	0.28095 9	2 0	-4.4	0. 7
33_2b	12	b m h	0.5 6	18	2.0267	0.02 2	0.1764	0.003 0	2619	2 8	99.3	0.00098	7	0.28097 6	2 4	-4.5	0. 9
32_2b	14	b e o s	0.3 2	66	3.0582	0.02 4	0.1597	0.001 7	2452	1 8	82.9	0.00316	2 2	0.28100 4	2 2	-7.4	0. 8

31_2b	16		p m os	0.4 0	55	2.4930	0.02 0	0.1754	0.002 0	2610	1 9	89.3	0.00134	9	0.28094 7	2 0	-5.7	0. 7		
40_2b	19	14	p m os	0.1 6	407	4.1825	0.03 6	0.1869	0.002 4	2715	2 1	55.4	0.00203	1 3	0.28098 0	1 5	-2.1	0. 6	5.46	0.3 6
39_2b	26		p e os	0.5 1	361	2.6748	0.02 1	0.1709	0.001 9	2566	1 8	86.8	0.00179	2 1	0.28096 8	1 9	-6.0	0. 7		
38_2b	27		p m rx	0.4 1	0	2.0689	0.01 6	0.1752	0.001 9	2608	1 8	98.6	0.00091	6	0.28102 2	1 1	-3.1	0. 4		
37_2b	30		p m rx	1.0 8	0	2.6683	0.02 2	0.1757	0.002 0	2613	1 8	85.7	0.00220	1 3	0.28099 1	1 3	-4.1	0. 5		
36_2b	33	1	p m os	0.6 5	245	4.0358	0.03 4	0.1625	0.001 9	2482	2 0	66.3	0.00136	9	0.28097 6	1 6	-7.7	0. 6	4.45	0.3 2
45_2b	35		p m os	0.7 8	0	2.8783	0.02 5	0.1749	0.002 3	2605	2 2	81.8	0.00223	1 4	0.28095 8	1 7	-5.4	0. 6		
44_2b	41	3	b m h	0.3 8	89	2.0282	0.01 7	0.1760	0.002 3	2615	2 2	99.4	0.00293	2 0	0.28100 6	1 8	-3.5	0. 6	6.46	0.3 7
43_2b	45		p m os	0.5 2	41	2.0034	0.01 7	0.1781	0.002 1	2635	2 0	99.5	0.00098	8	0.28095 0	1 8	-5.0	0. 7		
42_2b	47	12	p m os	0.5 4	262	2.0430	0.01 7	0.1742	0.002 1	2598	2 0	99.4	0.00142	9	0.28099 1	1 5	-4.4	0. 5	7.79	0.4 1
47_2b	49	11	p m os	0.6 1	0	2.5936	0.02 1	0.1775	0.002 0	2630	1 9	86.7	0.00106	6	0.28098 8	1 7	-3.8	0. 6	7.29	0.3 7
46_2b	60	9	p m h	0.6 4	0	6.5574	0.05 5	0.1248	0.001 6	2026	2 2	57.1	0.00307	1 9	0.28107 0	2 1	-14. 9	0. 8	2.50	0.3 4
50_2b	64		p m os	0.7 0	372	2.8111	0.02 6	0.1771	0.003 3	2625	3 0	82.6	0.00095	6	0.28098 2	1 9	-4.1	0. 7		
48_2b	69	10	p e os	0.4 4	0	2.4799	0.02 0	0.1707	0.002 0	2565	2 0	90.7	0.00150	1 0	0.28100 8	1 7	-4.6	0. 6	6.02	0.3 6
51_2b	70		p e os	0.6 3	31	2.9303	0.02 4	0.1639	0.002 0	2497	2 0	83.9	0.00069	4	0.28096 7	2 3	-7.6	0. 8		
52_2b	78	6	b m os	0.6 5	40	2.0243	0.01 7	0.1782	0.002 3	2636	2 1	99.0	0.00084	5	0.28097 4	1 9	-4.1	0. 7	4.50	0.3 3
53_2b	78	8	b m os	0.6 5	40	2.0243	0.01 7	0.1782	0.002 3	2636	2 1	99.0	0.00095	6	0.28100 2	1 6	-3.1	0. 6	3.87	0.2 7
54_2b	80	5	p e os	0.2 7	785	5.5334	0.04 8	0.1843	0.002 9	2692	2 6	36.9	0.00109	7	0.28096 8	1 6	-3.0	0. 6	5.87	0.4 2
55_2b	81		p e os	0.8 3	28	2.0064	0.01 7	0.1798	0.002 3	2651	2 1	99.1	0.00151	1 0	0.28095 4	1 7	-4.5	0. 6		

22	22	eq m os	2.33	0.4 6	37	1.9011	0.01 5	0.1860	0.001 9	2707	1 6	100. 4	0.00071	4	0.28096 7	3 3	-2.7	1. 2
34	34	p m os	3.40	0.6 7	40	1.9156	0.01 4	0.1874	0.001 8	2719	1 6	99.8	0.00087	6	0.28099 0	3 3	-1.6	1. 2
37	37	p m os	3.41	0.3 7	95	1.9304	0.01 6	0.1895	0.001 7	2738	1 5	99.0	0.00082	5	0.28098 9	2 9	-1.2	1. 0
42	42	p m os	1.39	0.4 0	30	1.9217	0.01 7	0.1882	0.002 1	2727	1 8	99.5	0.00057	5	0.28093 4	2 2	-3.4	0. 8
48	48	eq m os	2.19	0.3 9	52	1.9057	0.01 7	0.1876	0.002 2	2721	1 9	100. 0	0.00060	4	0.28094 5	2 7	-3.2	1. 0
51	51	p m os	1.54	0.3 5	39	1.9273	0.01 5	0.1878	0.001 7	2723	1 5	99.4	0.00058	4	0.28095 3	2 6	-2.8	0. 9
54	54	p m os	1.64	0.3 8	25	1.9204	0.01 5	0.1870	0.001 7	2716	1 5	99.7	0.00064	4	0.28092 9	2 8	-3.9	1. 0
55	55	eq m os	3.08	0.3 2	71	1.9088	0.01 7	0.1879	0.001 9	2724	1 7	99.8	0.00054	3	0.28093 5	3 3	-3.5	1. 2
60	60	f m os	0.78	0.5 7	30	1.9359	0.01 6	0.1877	0.001 9	2722	1 7	99.2	0.00087	5	0.28094 8	3 7	-3.1	1. 3
61	61	eq m os	2.32	0.3 2	40	1.9107	0.01 5	0.1875	0.001 8	2720	1 5	99.9	0.00093	1	0.28096 8	3 2	-2.4	1. 1
62	62	p m os	3.00	0.4 2	78	1.9206	0.01 6	0.1870	0.001 8	2716	1 6	99.7	0.00153	1	0.28096 3	2 4	-2.7	0. 9
64	64	p m os	4.32	0.4 5	54	1.9046	0.01 8	0.1875	0.002 3	2720	2 0	100. 0	0.00080	7	0.28095 7	2 8	-2.8	1. 0
66	66	p m os	2.24	0.2 7	109	1.9410	0.01 7	0.1882	0.002 4	2726	2 1	99.0	0.00037	3	0.28097 6	2 7	-1.9	0. 9

^a Numbers in *italic* refer to geochronological data published by Farina et al. (2015a)

^b Grain description: Habit: p - prismatic, f - fragment, eq - equant, b - blocky, ov - oval, an - anhedral; Analysis site: c - core, m - middle, e - edge; Zonation: os - oscillatory, h - homogeneous, rx - recrystallized

Table 3

Sample	UTM Lat	UTM Long	Classification ^a	Crystallisation age $\pm 2\sigma$ (Ma) ^b	Type ^c	N ₁ ^d	Inherited (Ma)	Metamorphic (Ma)	Average $^{176}\text{Hf}/^{177}\text{Hf}_{\text{int}} \pm 2\sigma^e$	Average $\epsilon\text{Hf}_{\text{int}} \pm 2\text{SD}^e$	N ₂ ^f	$\delta^{18}\text{O} \pm 2\text{SD}^g$	Published age	Reference ^h
Baço complex														
FQ1	643228	7746959	Kfs-rich	2711 $\pm$ 3	conc	6			0.280989 $\pm$ 23	-1.8 $\pm$ 2.3	22 of 22	5.5 $\pm$ 2.1		[1] - This study
FQ2	644685	7749318	Banded	2868 $\pm$ 10	wm	13		2705 $\pm$ 18	0.281009 $\pm$ 24	2.6 $\pm$ 2.1	19 of 19	3.9 $\pm$ 1.6		[1] - This study
FQ5 (= MR11)	625597	7762152	Plag-rich	2761 $\pm$ 11	wm	9	1 core at 3292		0.281002 $\pm$ 22	-0.2 $\pm$ 1.8	18 of 18		2744 $\pm$ 10	[2] - This study
FQ6 (= D12)	625629	7762157	Kfs-bearing	2779 $\pm$ 4	conc	3	2891 $\pm$ 3	2705 $\pm$ 7	0.280970 $\pm$ 23	-0.9 $\pm$ 2.2	3 of 3		2764 $\pm$ 10	[3] - This study
FQ8	628585	7760332	Kfs-bearing	-	wm		2770 $\pm$ 42	2612 $\pm$ 10						This study
FQ11	628157	7751740	Plag-rich	2790 $\pm$ 3	conc	6	2905 $\pm$ 17	2719 $\pm$ 14	0.281019 $\pm$ 18	1.1 $\pm$ 2.0	5 of 5			[1]
FQ13	628731	7748442	Banded	2790 $\pm$ 3	conc	17	ca. 2900		0.281030 $\pm$ 26	1.5 $\pm$ 1.2	12 of 14			[1]
FQ14	637895	7750487	Migmatite	2692 $\pm$ 4	conc	6	from ca. 2930 to ca. 3470		0.281020 $\pm$ 29	-1.2 $\pm$ 2.3	4 of 4			This study
FQ17	625643	7755180	Banded	2778 $\pm$ 2	conc	8	2862 $\pm$ 2	2732 $\pm$ 10	0.280982 $\pm$ 21	-0.2 $\pm$ 1.3	8 of 9	5.4 $\pm$ 1.6		[1] - This study
FQ20 (= D07A)	620091	7749557	Banded	-	conc		2 cores at 2829 and 2772	2723 $\pm$ 3					2918 $\pm$ 10	[3] - This study
FQ23	644680	7749530	Banded	2898 $\pm$ 12	wm	8		2783 $\pm$ 18	0.280977 $\pm$ 21	2.1 $\pm$ 3.3	8 of 8			[1]
Bonfim Complex														
FQ29	603409	7765331	Plag-rich	2773 $\pm$ 2	conc	10			0.280910 $\pm$ 18	-3.2 $\pm$ 1.3	42 of 42	4.9 $\pm$ 1.8		[1] - This study
FQ37	603293	7759954	Amphibolite	2719 $\pm$ 14	wm	9	2879 $\pm$ 14		0.280956 $\pm$ 24	-2.8 $\pm$ 2.9	9 of 9			[1]
FQ40	593263	7767241	Kfs-bearing	2854 $\pm$ 18	ui	4		2670 $\pm$ 15	0.280871 $\pm$ 20	-2.7 $\pm$ 1.1	4 of 4			[1]
FQ41	595299	7765480	Banded	2852 $\pm$ 16	wm	5			0.280898 $\pm$ 20	-1.8 $\pm$ 0.9	17 of 17			[1]
FQ51	603378	7741084	Plag-rich	2678 $\pm$ 10	wm	14			0.281026 $\pm$ 22	-1.3 $\pm$ 1.3	28 of 28			[1]
FQ52	582109	7761295	Banded	-	wm			2727 $\pm$ 11				4.6 $\pm$ 2.1		[1] - This study
MR31A	594650	7775356		2716 $\pm$ 14	wm	13			0.280958 $\pm$ 28	-2.8 $\pm$ 0.7	12 of 12		2700 $\pm$ 8	[2] - This study
MR22A (= FQ28)	602440	7769392	Kfs-rich	2715 $\pm$ 2	conc	22	3 cores at 3515, 3141 and 2835		0.280930 $\pm$ 26	-3.8 $\pm$ 1.8	7 of 7		2730 $\pm$ 8	[2] - This study
MR70G	602011	7727899		2716 $\pm$ 6	wm	33			0.280948 $\pm$ 28	-3.2 $\pm$ 1.5	20 of 20		2723 $\pm$ 7	[2] - This study
MR87A (= FQ50)	608581	7738759	Kfs-rich	2646 $\pm$ 9	wm	22	2734 $\pm$ 8		0.280960 $\pm$ 26	-4.4 $\pm$ 1.6	21 of 21		2613 $\pm$ 6	[2] - This study
MR14A (=FQ32)	608654	7756333	Kfs-rich	2715 $\pm$ 3	conc	9			0.280955 $\pm$ 28	-3.0 $\pm$ 1.6	7 of 7		2730 $\pm$ 7	[2] - This study

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FQ60	548149	7801529	Plag-rich	2728 ± 16	wm	6	2786 ± 29	0.280962 ± 21	-2.4 ± 2.0	17 of 18	4.6 ± 2.9	[1] - This study
FQ65 (= MR259A)	537885	7835546	Plag-rich	-	wm			2645 ± 8				[1]
FQ70	579027	7790240	Banded	-	conc			2713 ± 3				[1]
FQ74	555202	7842055	Kfs-rich	-	wm			2638 ± 14			5.4 ± 3.2	[1] - This study
FQ81	642637	7796689		-	conc			2671 ± 10				This study
MR257A	537893	7835513		2723 ± 8	wm	18		0.280958 ± 30	-2.7 ± 1.4	13 of 13	2706 ± 7	[2] - This study
MR259A	539550	7834072	Plag-rich	2721 ± 9	wm	15		0.280949 ± 29	-3.0 ± 2.2	9 of 9	2722 ± 7	[2] - This study
MR51A	621958	7803440		2708 ± 10	wm	10		0.280950 ± 27	-3.3 ± 1.3	18 of 19	2700 ± 8	[2] - This study

^a According to the classification of Farina et al. (2015a);

^b Intrusion age. Ages in bold indicate ages obtained during this study, regular font indicates ages already published by Farina et al. (2015a);

^c Type: conc - concordia, wm - weighted mean, ui - upper intercept;

^d Number of individual U-Pb spots used in age calculations;

^e $^{176}\text{Hf}/^{177}\text{Hf}_{\text{int}}$ and $\varepsilon\text{Hf}_{\text{int}}$ calculated using the intrusion age of each sample (excluding inherited cores or metamorphic domains);

^f Number of individual Lu-Hf spots used in the calculation of the average $^{176}\text{Hf}/^{177}\text{Hf}_{\text{int}}$ and $\varepsilon\text{Hf}_{\text{int}}$;

^g $\delta^{18}\text{O}$ calculated using the intrusion age of each sample, except for FQ52 and FQ74 where we used the metamorphic age

^h References: [1] - Farina et al. (2015a), [2] - Romano et al. (2013), [3] - Lana et al. (2013).

Highlights

- The Neoarchean basement represents different terranes with different identities.
- Continuous differentiation from Na- to K-rich magmatism reflects source enrichment.
- High $\delta^{18}\text{O}$ values reflect metasediments involvement in K-rich magma petrogenesis.
- Ubiquitous low $\delta^{18}\text{O}$ values indicate the presence of upper-level hydrothermalism.
- Model indicates a change at 2.9 Ga, reflecting the onset of continental collision.