3.30 Ga high-silica intraplate volcanic–plutonic system of the Gavião Block, São Francisco Craton, Brazil: Evidence of an intracontinental rift following the creation of insulating continental crust

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A B S T R A C T

High-silica rhyolites having U–Pb zircon ages of 3303 ± 11 Ma occur along the eastern border of the Gavião Block (Brazil) associated with the Contendas-Mirante and Mundo Novo supracrustal belts. Unlike many Archean greenstone sequences, they are not interlayered with mafic to intermediate units. Instead, they belong to an inter-related plutonic–volcanic system, together with granitic massifs having similar zircon crystallization ages of ca. 3292 ± 3 Ma and 3328 ± 3 Ma and plotting along the same geochemical trends as the rhyolites. The rhyolites show well-preserved primary volcanic features such as magma flow textures and euhedral phenocrysts. High emplacement temperatures are indicated by petrographic evidence (β-quartz phenocrysts), zircon saturation temperatures (915–820 °C) and geochemical data, especially high SiO2 (74–79 wt.%) together with elevated Fe2O3(T) (~3 wt.%), MgO (0.5–1.5 wt.%) and low Al2O3 (~11 wt.%). The rhyolites show homogeneous trace element ratios (La/YbN 4.8 ± 1.8; EuN/Eu* ~0.55; Sr/Y ~0.7) and negative εHf(3.3 Ga) from 0 to −7, indicating derivation from a single crustal source for both occurrences. Specifically, the rhyolites would have derived from extraction and eruption of highly silicic residual liquid formed by crystallization of granitic magma in a relatively shallow (~10 km) reservoir, now represented by the granite massifs. The granite magma was formed by melting or differentiation of material similar to the diorite gneiss that occurs regionally. The 3.30 Ga volcanic–plutonic systems formed after a period of crustal growth and stabilization of a thick continental lithosphere, represented by massive 3.40–3.33 Ga TTG and medium to high-K calc-alkaline magmatism in the Gavião Block. The 3.30 Ga-old rhyolites and granites would therefore have formed in an intraplate tectonic setting after the formation and stabilization of new continental crust, and accordingly would represent the first stages of rifting and continental break-up. Intraplate magmatism and intracrustal differentiation processes took place on Earth at 3.3 Ga and produced magmas that were distinct from Archean TTGs, questioning the reliability (or at least the uniqueness) of “intraplate models” to explain the origin of the latter.

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1. Introduction

The building of the Earth’s continental crust is a complex process that includes the generation of primary mafic melts in the mantle and water-assisted chemical differentiation to more felsic compositions (e.g., Arndt, 2013). Differentiation of mafic magmas from the mantle with recycling of the cumulate or residue to the mantle accounts for the felsic bulk composition of the continental crust (Cawood et al., 2013; Rudnick, 1995), and the preservation of the resulting felsic blocks accounts for long-term crustal growth (Condie et al., 2011; Hawkesworth et al., 2009).

However, the geodynamic processes that have accounted for continental crust formation, stabilization and preservation over time are still a matter of considerable debate (Arndt, 2013; Hawkesworth et al., 2010, 2013; Kröner et al., 2013), especially for the Archean period (4.0–2.5 Ga) for which any interpretation is complicated by a more fragmentary, limited and ambiguous geological record (Artemieva, 2006; Gerya, 2014). In particular, early Archean continents were formed in two distinct settings; as intraplate volcanic centers driven by narrow mantle upwellings or in volcanic arcs associated with subduction-induced mantle-melting zones (e.g., Van Kranendonk, 2010). Predicting the dominant type in hybrid types of crust is still unclear and has been a
subject of contention (e.g., Bédard, 2006; Bédard et al., 2013; de Wit et al., 1992; Martin et al., 2005, 2014; Sizova et al., 2015; Smithies et al., 2003, 2009; Van Kranendonk et al., 2014; van Thienen et al., 2004; Zegers and van Keken, 2001).

Many models about the formation and evolution of the Archean crust are based on the field characteristics, petrology and geochemistry of tonalite–trondhjemite–granodiorite suites (TTGs) (Bédard, 2006; Hoffmann et al., 2014; Martin et al., 2005, 2014; Moyen and Martin, 2012; Smithies, 2000). Indeed, TTGs represent the dominant lithology of Archean crust (Arndt, 2013; Moyen, 2011), especially for the Neoarchean and Paleoproterozoic periods (4.0–3.2 Ga) because significant diversification of Archean granitoids, i.e. formation of crust-derived biotite, two-mica granites and mantle-derived sanukitoids, essentially took place after 3.0 Ga ago (e.g., Laurent et al., 2014a). However, several recent studies have emphasized that Eo- to Paleoproterozoic crustal segments often contain a diversity of granitoids that are not all akin to TTGs. Some examples are the ca. 4.02 Ga Idiwhaa tonalites in the Slave craton, NW Canada (Reimink et al., 2014); the ca. 3.60 Ga ferroan augen gneiss suite in the North Atlantic Craton of SW Greenland (Friend and Nutman, 2005; Nutman et al., 1996); as well as some components of the 3.66–3.45 Ga Ancient Gneiss Complex (Kröner et al., 2014) and the 3.48–3.43 Ga Tsawela gneisses (Hoffmann et al., 2016) of the eastern Kaapvaal Craton in southern Africa. Although subordinate in volume, these felsic rocks may represent important geodynamic events in the process of craton building and stabilization and must be considered along with TTG to depict the processes of Archean crust evolution. Specifically, petrogenetic studies of these non-TTG components showed that they likely formed in intraplate settings in which the crust is heated up by mantle upwelling and mafic magma underplating (Hoffmann et al., 2014; Nutman et al., 1984; Reimink et al., 2014).

Most intraplate volcanic systems, such as silicic large igneous provinces, bimodal volcanic systems, oceanic islands and oceanic plateaux, compositionally differ from magmas generated at plate boundaries, with the former being hotter, drier and induced by anhydrous decompression melting, and the latter are typically colder, H2O-rich and formed by fluid-driven flux melting (Bailey, 1983; Grove et al., 2012; Lee and Bachmann, 2014). Intraplate magmatism generally affects continental or oceanic interiors undergoing extension and is actively driven by narrow mantle plumes originating from relatively shallow asthenosphere upwelling (Anderson and King, 2014), or passively driven by plate motion (Ziegler and Cloetingh, 2004). The addition of underplating mafic magma in intraplate environment provides a “non-plate tectonics” way for crustal growth and thickening (Thybo and Artemieva, 2013), and may contribute to remelting of parts of the original crust. The continuous underplating of mafic magmas causes the progressive differentiation of the continental lithosphere and stabilization by lower crustal delamination events (e.g., Bédard, 2006; Johnson et al., 2013).

Therefore, in order to better constrain the role of these processes in the formation and evolution of continental lithosphere in the early Earth, it is necessary to better characterize and study the origin of non-TTG components in Eo- to Paleoproterozoic terranes. Felsic 3.30 Ga-old volcanic rocks and associated granitoids occur at the eastern margin of Gavião Block, São Francisco Craton, in association with the Contendas-Mirante and Mundo Novo supracrustal belts (Marinho et al., 1994a, 1994b; Peucat et al., 2002). They are geologically and petrologically unrelated to classical TTGs, yet their petrogenesis is poorly understood. Here, we describe the petrology, whole-rock geochemistry, zircon U–Pb ages and Lu–Hf isotopic signatures of these felsic and plutonic rocks. Geochemical modeling is used to investigate the nature of the source and magmatic evolution of the volcanic–plutonic system. The combined results are used to (i) discuss the tectonic environment of the volcanic–plutonic system related to the evolution of the Gavião Block; and (ii) explore the general role that intraplate magmatism might play in craton formation and evolution.

2. Geological setting

The São Francisco Craton (Fig. 1) comprises Archean and Paleoproterozoic terranes of low-grade to granulitic metamorphic facies. The Gavião Block, which contains the oldest rocks of South America, forms the western segment of the São Francisco basement. It mainly consists of 3.40–3.33 Ga granitic gneiss complexes with TTGs and medium- to high-K calc-alkaline composition (Marchesin, 2015; Martin et al., 1997; Nutman and Cordani, 1993; Santos-Pinto et al., 1998, 2012; Silva et al., 2015; Zincone, 2016) and the investigated 3.30 Ga felsic volcanic rocks (Marinho et al., 1994b; Peucat et al., 2002).

We focus on two occurrences of ca. 3.30 Ga felsic volcanic rocks occurring along the eastern boundary of the Gavião Block, i.e., the north-trending, 600 km-long Contendas–Jacobina lineament according to Sabaté et al. (1990), and denominated herein as the Contendas–Jacobina shear zone (Fig 2A). This major strike-slip shear zone separates Paleoproterozoic low-to-medium-grade rocks of the Gavião Block from Neoarchean high-grade rocks of the Jequiti block and the Caralba/São José do Jacuípe complex. The Mundo Novo supracrustal belt (Fig 2B; Mascarenhas and Silva, 1994; Peucat et al., 2002) occurs in the northern part of the shear zone and represents an intracratonic basin, with felsic volcanism and a continental to marine sedimentary sequence (Loureiro and Santos, 1991). The 3.30 Ga Contendas Rhyolite occurs in the southern sector of the shear zone, partly over lain by the Paleoproterozoic Contendas-Mirante basin (Fig. 2C; Nutman et al., 1994; Zincone, 2016; Zincone and Oliveira, in review). The Jacobina basin, occurring west of the Mundo Novo belt, comprises siliciclastic units containing only 3580 to 3280 Ma detrital zircon grains, with the most abundant population at 3300 Ma (Barbuena et al., 2016; Magee et al., 2001; Mougeot, 1996; Teles et al., 2015), and was developed on the stable platform represented by >3.4 Ga gray gneiss of the Gavião Block (see Mascarenhas et al., 1998; Pearson et al., 2005 for a review). Similar 3350 to 3280 Ma zircon-bearing quartzite has been recently described in spatial association with the investigated rhyolites (Zincone and Oliveira, in review; Barbuena et al., 2016; Zincone, 2016).

Fig. 1. Simplified geological map of the São Francisco Craton (modified after Silva et al., 2015). The letters indicate the four main complexes within the Gavião Block: P: Porteirinha Complex; G: Gavião Complex; M: Mairi Complex; S: Sobradinho Complex. The red rectangle represents the area covered by Fig. 2.
3. Geological setting and petrography of the rhyolites and coeval plutonic rocks

The felsic volcanic rocks of Contendas (Entroncamento village) and Mundo Novo (Coqueiro farm) occur 200 km apart from each other (Fig. 2A). They were first recognized as part of the two different greenstone belt sequences, with the Contendas Rhyolite being part of the Contendas-Mirante supracrustal belt (Marinho et al., 1994a) and the Mundo Novo Rhyolite being part of the Mundo Novo supracrustal belt (Mascarenhas and Silva, 1994; Peucat et al., 2002). The contact between rhyolite and basement gneisses or metasedimentary rocks is not observed; the volcanic rocks occur as inliers within Paleoproterozoic cover (Fig. 2B–C). The rhyolites are not interlayered with any other meta-volcanic unit and the maximum metamorphic grade is lower greenschist facies.

The Contendas Rhyolite (CR; Fig. 2B) occurs in the vicinity of quartzite that exclusively holds ca. 3.3 Ga detrital zircon (sample TZD-6 of Zincone and Oliveira, in review; Zincone, 2016), but detailed field relationships cannot be established owing to the presence of younger cover. To the east, the rhyolite is flanked by a few meter wide phyllonite zone (quartz-sericite schist) and then by meter-size quartz veins that eventually separate the rhyolite from the Paleoproterozoic Contendas-
Mirante metasedimentary rocks. Quartz and feldspar phenocrysts are N-S oriented, possibly reflecting a primary igneous flow texture. To the east, the east dipping foliation (Sn: 090/58) associated with down dip lineation (Lx: 082/58) parallels the structural fabric of adjacent younger metasedimentary rocks, indicating that the rhyolite was affected by Paleoproterozoic deformation.
The Mundo Novo Rhyolite (MNR; Fig. 2C) has been described as dacite and rhyodacite with a microporphyritic and variolitic texture associated with pyroclastic, meta-volcaniclastic sediments (Loureiro and Santos, 1991; Peucat et al., 2002).

The rhyolites from both localities are mainly composed of 0.1 to 1-cm long, euhedral to subhedral white feldspars and purple-black β-quartz phenocrysts, together with chlorite, biotite, and opaques as accessory phases (Fig. 3a–d). Feldspar and β-quartz phenocrysts are embedded in a siliceous, fine-grained (0.05 mm), grayish blue, felsic groundmass of quartz, plagioclase, alkali feldspar, biotite (partially chloritized), tiny flakes of chlorite, and opaques (mostly magnetite) (Fig. 4A–C). Feldspar phenocrysts show compositional zoning with
K-rich inner regions and a sodic (albite-rich) outer region (Fig. 4D–E). Potassic feldspar shows inclusions of chlorite and apatite, and is partially altered to sericite, muscovite and calcite.

In most CR samples, primary igneous flow texture is retained (Fig. 4A–B, D–F), although there are some polycrystalline quartz aggregates with a granoblastic texture and feldspar phenocrysts displaying kink bands. In contrast, the MNR shows coarser groundmass, partially resorbed feldspar phenocrysts (Fig. 4G), and locally S/C fabrics defined by the preferred orientation of muscovite (Fig. 4H). Apart from such minor deformation and recrystallization features, the retention of crystals showing typical textures and composition of high-temperature igneous phases precludes any significant role played by subsolidus alteration and metasomatism (e.g., silicification) of the rocks, except for some MNR samples (see Section 5.1).

A distinct structure observed in CR samples is the presence of crystal-zone accumulation (Fig. 4I) and millimeter-size enclaves (Fig. 4A–C and J–L). The enclaves are coarser-grained (c. 0.1–1 mm) than the groundmass; they have zircon-bearing biotite selvage, and contain plagioclase phenocrysts (up to c. 1 mm), apatite, and quartz.

Plutonic rocks occur in the vicinity of both investigated rhyolites. The Boa Sorte meta-syenogranite occurs 30 km south of the MNR and is a pluton emplaced within the gray gneisses complex. It presents a weakly porphyritic texture with aplitic groundmass (Fig. 4I) and millimeter-size enclaves (Fig. 4G, 4J). The mineralogy of the Boa Sorte meta-syenogranite (sample TZD-268) consists of quartz (55%), microcline (25%), plagioclase (8%) and biotite (12%), as well as accessory zircon, apatite, epidote, allanite and calcite. The syenogranite is cut by a centimeter-scale, pink quartzofeldspathic dyke parallel to the main foliation (Fig. 3e).

The 3.35 Ga Boa Vista dome (Nutman and Cordani, 1993) occurs in the southern part of the Contendas-Mirante basin (Zincone and Oliveira, in review; Zincone, 2016). The northeastern portion is largely composed of biotite meta-granite (represented by sample TZD-199A) cut by albite granitic dykes (Fig. 3g).

The Calderão diorite gneiss (Fig. 3h) is located between the Sete Voltas and Boa Vista domes, specifically 2.5 km west of the Boa Vista meta-granite (sample TZD-199A) and 75 km south of the CR. The Calderão diorite gneiss shows a compositional banding characterized by centimetric layers of plagioclase (75 to 80%), biotite (15–20%) and quartz (5–10%) alternating with thinner bands richer in quartz (15–20%), with less plagioclase (70–75%) and biotite (5–7%). Magnetite, titanite, epidote, zoisite, zircon and apatite are accessory phases.

4. Zircon U–Pb–Hf studies and whole-rock geochemistry

Analytical methods are summarized in Appendix A.

4.1. U–Pb geochronology

Four samples were investigated for geochronology: sample TZD-31B from the CR, sample PO-119 from the MNR, and samples TZD-268 and TZD 199A from the coeval Boa Sorte and Boa Vista granites. Zircon grains in all four samples show similar shapes (perfectly euhedral, translucent, colorless to light pink stubby prisms ranging from 150 to 400 μm and having aspect ratio of 3:1) and showing a concentric oscillatory zoning with no overgrowths (Fig. 5a). As pointed out by Marinho et al. (1994b), the (100) prism is better developed than (110), and the (101) pyramid is more developed than (211), which are characteristic features of primary magmatic crystals of high crystallization temperature (850 °C), typical of volcanic rocks (Pupin, 1980).

Zircon Th/U ratios are ~0.5 in the rhyolites and slightly higher (0.7–0.8) in the granites. From sample TZD-31B, 44 spots on 39 zircon grains were analyzed, yielding a Concordia age of 3304.3 ± 8.2 Ma (MSWD = 0.68; n = 35) (Fig. 6), interpreted as the time of CR magmatic crystallization. This age matches the 3304 ± 31 Ma age obtained by Marinho et al. (1994b).

From sample PO-119, 30 spots from 30 zircon grains were analyzed, yielding a Concordia age of 3303.7 ± 11 Ma (MSWD = 3.1; n = 30) (Fig. 6). This age is within uncertainties identical to the existing SHRIMP U–Pb age of 3305 ± 9 Ma (Peucat et al., 2002) and is interpreted as reflecting the magmatic crystallization of the MNR.

From sample TZD-268, 46 spots on 41 zircon grains were analyzed, yielding a Concordia age of 3292 ± 3 Ma (MSWD = 4.8; n = 32) (Fig. 6) which is interpreted as the igneous crystallization age of Boa Sorte biotite meta-granite. Seven zircon grains present indistinguishable ages on cores and rims, except from one single grain (Table S1 zr 3a and 3b; Fig. 5c). From sample TZD-199A, 40 zircon grains were analyzed, yielding an upper intercept age of 3324 ± 53 Ma (MSWD = 3.2; n = 30) and an identical but more precise Concordia age of 3327.7 ± 3 Ma (MSWD = 5; n = 25), the latter of which is interpreted as the crystallization age of the Boa Vista biotite meta-granite.

4.2. Zircon Hf isotopes

HF-in-zircon analyses were carried on the MNR sample PO-119 (39 spots) and the CR sample TZD-31B (40 spots). The 176Hf/177Hf(3.3 Ga) ratios overlap for both rhyolites, but are very heterogeneous (ranging from 0.280474 to 0.280659). Most of the corresponding εHf(3.3 Ga) values are negative with MNR analyses showing εHf(3.3 Ga) ranging between −0.35 and −6.26 (TDM from 3.57 and 3.87 Ga). Similarly, the CR zircons show εHf(3.3 Ga) between +0.35 and −6.57 (with one outlier at −9; TDM from 3.54 to 3.88 Ga and up to 4.02 Ga).

4.3. Whole-rock geochemistry

The CR and MNR samples show strikingly similar major and trace element signatures. All samples are silica oversaturated with more than 43% normative quartz and SiO2 ranging from 76.5 to 78.7 wt.% for CR and 77.8 to 79.3 wt.% (except two samples at 74–75 wt.%) for MNR. The two groups have low Al2O3 (<11 wt.%) contents and K2O/Na2O ratios between 0.5 and 1 (Fig. 8). The MNR samples display higher MgO (average 1.07 wt.% with Mg# = 0.44) and lower CaO (average 0.18 wt.%), the CR (average MgO 0.48 wt.% with Mg# = 0.28; average CaO = 0.73 wt.%). Intriguingly, the most silica-rich samples from the MNR are also the richer in MgO. The two rhyolite units are also depleted in ferromagnesian elements (Fe2O3(T) + MgO + MnO + TiO2 < 4.4 wt.%), with relatively high Cr (from 5 to 24 ppm, average 13 ppm) and low Ni (<1 ppm) contents. The CR and MNR compositions overlap extensively in terms of Al2O3 (10.7%), Fe2O3(T) (3.0%), Na2O (3.4%), K2O (2.7%), and P2O5 (0.02%).

On classification diagrams using major and trace elements, the samples fall in the sub-alkaline rhyolite field (Fig. 9A–B). In the CIPW-normative Ab–An–Or classification diagram, the samples show a narrow compositional range from rhyolite to quartz keratophyre (Fig. 9C). In the diagrams proposed by Frost et al. (2001), CR samples classify as metaluminous to slightly peraluminous (molecular [Al2O3 / (CaO + Na2O + K2O)] = 0.94–1.1), ferroan calcic rocks whereas the MNR samples are slightly to strongly peraluminous (A / CNK = 1.07–1.45), ferroan to magnesian calcic rocks (Fig. 10). All the samples match the medium-K calc-alkaline fields on the AFM diagram and SiO2 vs. K2O plot.

On a primitive mantle-normalized multi-element diagram both rhyolites display strong depletions in Sr and Ti, and moderate Nb–Ta negative anomalies. In addition, they are depleted in Rb relative to Th and enriched in Ba relative to Sr (Fig. 11A). The CR has higher Sr (c. 615 ppm) and Ba (ca. 50 ppm) contents than the MNR (Sr 550 ppm; Ba 33 ppm). The REE patterns (Fig. 12) of both rhyolite units show no difference and present slight enrichment of the light REE relative to heavy REE (average La/Yb(N) ratio = 4.75) and marked negative Eu anomalies (Eu/Eu* = 0.55). Light REE are moderately fractionated (La/Sm(N) = 2.7–4.4; average 2.8) and the heavy REE profile is nearly flat (Gd/Yb(N) = 1.1–1.3).
Fig. 5. Representative cathodoluminescence images of zircons from the studied volcanic and plutonic rocks of the Gavião Block. a) Contendas Rhyolite (sample TZD-31). b) Mundo Novo Rhyolite (sample PO-119). c) Boa Sorte Granite (sample TZD-268). d) Boa Vista Granite (sample TZD-199A). Spots are labeled with $^{207}\text{Pb}/^{206}\text{Pb}$ age (in Ma), Hf model age (in Ga), $\varepsilon_{\text{Hf}}(t)$ and spot number (between brackets) (except for (c and d), which has not been analyzed for Hf isotopes). U–Pb ages are all within 5% of concordance and uncertainties are 2$\sigma$. White scale bar is 50 $\mu$m.
The composition of coeval plutonic rocks regionally associated with the rhyolites (Boa Sorte, Boa Vista and Meiras granites and granodiorites) and the Calderão diorite gneiss are presented in Figs. 8 to 11. These rocks clearly plot along the same chemical trends as the rhyolites. The Boa Sorte and Meiras granite has the most striking compositional similarity to the rhyolites, both in major and trace elements, albeit having slightly lower K_2O (K_2O/Na_2O = 0.4). The older Boa Vista granite (3328 ± 3 Ma, sample TZD-199A in this paper) and the Meiras granite and granodiorite define sublinear trends in Harker and other major-element plots, with compositions intermediate between the Calderão diorite and the rhyolites (Figs. 8–10). They are calc-alkaline, intermediate to felsic rocks ranging from 65 to 72 wt.% SiO_2 and have high Fe_2O_3(T), CaO and Na_2O (CaO/Na_2O ratio of 0.5–0.6) contents. Like the rhyolites, they have slightly peraluminous and ferroan compositions (Fig. 10) and similar trace element and REE patterns (Fig. 11B), with slightly higher Rb, Th, Nb–Ta, Sr, La and Ti contents and lower Ba and HREE contents.

The Calderão diorite is an intermediate (c. 58 wt.% SiO_2) metaluminous and alkali-calcic rock of ferroan afiniti (Fig. 10). It is poorer in K_2O and SiO_2 and richer in Fe_2O_3(T), MgO, CaO and Al_2O_3 than the granite and rhyolites. Although it has comparable profiles in multi-element diagrams and REE patterns, the Calderão diorite is also distinguished from the more felsic rocks by its lower contents in highly incompatible elements, less pronounced Nb, Ta, Sr and Ti troughs in primitive mantle normalized trace element diagrams, the absence of a negative Eu anomaly (Eu_N/Eu* = 0.92) and lower HREE contents (La/Yb_N = 18.8).

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**Fig. 6.** Concordia diagrams (206Pb/238U vs. 207Pb/235U) for LA-ICP-MS zircon analyses on samples from the Contendas (13°40’10.40"S; 40°54’22.46"W) and Mundo Novo (11°53’36.79"S; 40°29’25.08"W) rhyolites and from the Boa Sorte (12°14’1.82"S; 40°34’27.43"W) and Boa Vista (14°22’38.59"S; 40°44’47.77"W) granites, the plutonic equivalent of the rhyolites. Error ellipses are 2σ.

**Fig. 7.** Composite Hf isotope evolution diagram. (A) 40 analyses from Contendas Rhyolite; (B) 30 analyses from Mundo Novo Rhyolite. ^{176}Lu/^{177}Hf evolution lines are based on assumed ^{176}Hf/^{177}Hf crustal ratio of 0.0113 (Bouvier et al., 2008). The ^{176}Lu decay constant of Söderlund et al. (2004) and Scherer et al. (2001) was used in the calculation of εHf(3.3 Ga) values.
5. Discussion

5.1. Element mobility

Some MNR samples are strongly peraluminous compared with other rhyolites (Fig. 10). These samples are systematically characterized by the presence of shear planes filled with recrystallized muscovite (see Fig. 4H and Section 3) and growth of green biotite and chlorite. Their peculiar composition might thus reflect superimposed metasomatism due to focused fluid-flow regimes during later tectonothermal events. The fluid-dominated alteration modified some of their mineralogy and major element chemistry, but did not affect the trace element
patterns (see Fig. 11). Consequently, the composition of these MNR samples will not be considered in petrogenetic modeling below.

Aside from the most deformed samples, element mobility due to post-crystallization events seems to be minor or even absent, which is in line with preserved igneous textures and mineralogy. Specifically, the lack of any clear negative correlation between silica, other major element, and trace element contents argues against any role played by silicic contribution in the generating the high-SiO2 content of these rocks.

5.2. Zircon saturation

The investigated rhyolites lack zircon xenocrysts and core–rim relationships, which we interpret as reflecting that the magma was hotter than the zircon saturation temperature (Tzr). In this case, Tzr is a minimum estimate for the actual magma temperature, assuming that the bulk rock composition reflects that of the magma (Boehnke et al., 2013; Watson and Harrison, 1983). Under this premise, the Tzr is 867–915 °C for the CR and 833–868 °C for the MNR. These can therefore be considered as “hot” (> 850 °C) magmas. In addition, this implies that the well-preserved prismatic, oscillatory-zoned zircons with concordant U–Pb ages are definitely of igneous origin, related to zircon saturation in the late stages of magma cooling and crystallization.

5.3. Petrogenetic modeling

5.3.1. Comparison with experimental data: the granite–rhyolite connection

In order to constrain the differentiation processes and the source of the rhyolites, we compared their major element composition with a compilation of experimental melts (n = 357) relevant for crustal melts (i.e. P < 1.5 GPa; T < 1200 °C) and from a wide range of starting materials, including tholeiitic to calc-alkaline basalts and andesites (Beard and Lofgren, 1991; Martel et al., 1999; Patiño-Douce and Beard, 1995; Rapp and Watson, 1995; Sisson et al., 2005; Tatsumi and Suzuki, 2009; Wolf and Wyllie, 1994), granitoids (Bogaerts et al., 2006; Klimm et al., 2003; Qian and Hermann, 2013; Singh and Johannes, 1996; Watkins et al., 2007), and sedimentary rocks (Montel and Vielzeuf, 1997; Patiño-Douce and Beard, 1996; Patiño-Douce and Harris, 1998; Patiño-Douce and Johnston, 1991; Pickering and Johnston, 1998; Skjerlie and Johnston, 1996; Vielzeuf and Holloway, 1988). This database was filtered to take into account only melts with >70 wt.% SiO2. The results of this comparison for selected major elements are presented in Fig. 13. Clearly, the rhyolites have an unusual composition compared to experimental melts. Indeed, the rhyolites are clearly richer in SiO2 than most leucogranitic liquids and, at that level of silica content, show higher Fe2O3(T), MgO and lower Al2O3, and CaO contents (Fig. 13). Lack of evidence for late- or post-emplacement silicic contribution (see Sections 3 and 5.1), the very high SiO2 contents of the rhyolites, and their tight compositional range (Figs. 8 and 13) indicate that they represent near-liquid compositions. In fact, their SiO2 contents are similar to those of matrix glass found in modern silicic volcanic rocks (78–80 wt.%; Fig. 13) (e.g., Bégue et al., 2014), which supports this hypothesis. In contrast, the samples of the Boa Sorte and Boa Vista granites show compositions that are closer to the field of experimental leucogranites in all diagrams (Fig. 13), especially samples TZD-268 (SiO2 = 72.6 wt.%) and WMS-20B (SiO2 = 71.5 wt.%). These granites and the rhyolites share compositional affinities (elevated Fe2O3(T), and MgO and low Al2O3 contents at a given level of SiO2; Figs. 8–11) and
define rough linear trends in Harker diagrams. We therefore propose that they are cogenetic and specifically that the rhyolites represent differentiated liquids from a parental composition close to that of the granites. This suggests two-step mechanism with (1) derivation of the granites by partial melting of crustal rocks or crystallization of a mafic magma and (2) further differentiation of the granitic bulk composition to form the rhyolites. This two-step model is consistent with petrogenetic models of recent high-SiO$_2$ rhyolites, indicating that these represent extracted liquids from crystal mushes formed in magma chambers of intermediate-felsic composition (Bégue et al., 2014; Dufek and Bachmann, 2010; Graeter et al., 2015).

We further compared the composition of the granites to that of the experimental liquids to identify their origin. We selected sample TZD-268 of the Boa Sorte granite for use in modeling, because zircon U–Pb dating shows unequivocally that it is coeval with the rhyolites (Fig. 6). We used a simple approach in which we calculated the sums of squared residuals ($\Sigma R^2$) between the composition of the Boa Sorte granite and those of all experimental melts of our compilation (all compositions being normalized to 100%). Most $\Sigma R^2$ are very high (10–50), except for melts in the low-pressure (~0.2 GPa) crystallization experiments of a calc-alkaline andesite from Martel et al. (1999), for which 10 experiments out of 15 yield $\Sigma R^2$ between 4 and 10 and always plot close and
parallel to the granite–rhyolite trend in Harker diagrams (Fig. 13). The experimental conditions and starting materials used by these authors thus best reproduce the natural conditions for the origin of the Boa Sorte granite. Importantly, the starting materials used by Martel et al. (1999) are very close in composition to that of the Calderão diorite.

Although the lack of geochronological data for the Calderão diorite, prevents the confirmation of a genetic link with the Boa Sorte granite and the rhyolites, its composition (represented by sample TZ-185) represents the best estimate for the parental material of these felsic rocks and will therefore be used below for modeling purposes.
5.3.2. Major element modeling

We used least-squares mass balance calculations on major elements to check if the inferred two-step petrogenetic model (Calderão diorite, sample TZ-185, → Boa Sorte granite, sample TZ-268, → MNR and CR rhyolites) was plausible. We considered only the following major elements: SiO₂, Al₂O₃, Fe₂O₃T, MgO, CaO, Na₂O, K₂O, and TiO₂. P₂O₅ was only used to determine the amount of fractionating apatite (Table 1).

From the experiments of Martel et al. (1999) relevant to our sample compositions (see Section 5.3.1), the possible fractionating phases are plagioclase, orthopyroxene, amphibole, ilmenite, magnetite, and apatite. Biotite was added as a possible fractionating phase, because the Calderão diorite gneiss is twice as rich in K₂O (1.8 wt.%) than the starting compositions of Martel et al. (1999) (0.9–1.0 wt.%). For mineral compositions (see Section 5.3.1), the possible fractionating phases are starting compositions of Martel et al. (1999) (0.9 wt.%), oxides and apatite; and literature data for biotite (Laurent et al., 2014b). The compositions used in the models yielding the best results are reported in Table 1.

For the first fractionation step (diorite to granite), the model that yields the best results (R² = 0.11) corresponds to a solid assemblage (Table 1). Although acceptable (R² = 0.77), the match is not as good by plagioclase (~60%) and biotite (~38%) with minor ilmenite and apatite (~<1%) (Table 1). The proportion of liquid (c. 30%) is relatively consistent, yet somewhat lower (~40–60%) than obtained in experiments from Martel et al. (1999).

For the second fractionation step, we used the Boa Vista granite as a starting composition and the average composition of the rhyolites as the final liquid to be modeled. Because of the small differences in the composition of both end-members, the models generally yield higher R². The best case is when the fractionating solid assemblage is dominated by plagioclase (~60%) and biotite (~38%) with minor ilmenite and apatite (Table 1). Although acceptable (R² = 0.77), the match is not as good as for the first step, especially for CaO, Na₂O and K₂O (Table 1). However, fractionation of plagioclase, biotite and apatite explains well the presence of these minerals in close association as small enclaves within the rhyolites (Fig. 4D–I). Such enclaves would thus represent cumulate fragments, as is commonly observed in modern high-SiO₂ rhyolite bodies (Ellis et al., 2014; Graeter et al., 2015).

It must be noted that the entire process modeled here is consistent with the formation of large volumes of rhyolites by segregation of residual melt within long-lived, crystallizing mush zones, as described in modern settings in which granitoids represent the leftovers of large rhyolitic eruptions (Bachmann and Bergantz, 2004, 2008; Gelman et al., 2014; Lipman and Bachmann, 2015). Moreover, the model predicts that the bulk proportion of fractionated crystals relative to the initial amount of melt is c. 50% (c. 30% for the first step and c. 20% for the second step; Table 1). This is consistent with the results of numerical models showing that the probability of residual melt extraction in crystal mushes (and thus ascent and eruption of high-SiO₂ rhyolites) dramatically increases as soon as the crystal fraction reaches c. 50%, irrespective of the initial melt composition and size of the magma chamber (Dufek and Bachmann, 2010).

5.3.3. Trace element modeling

The two-step model inferred from experimental constraints and major element modeling was tested using trace element modeling. We used a range of partition coefficients (Laurent et al., 2013) for the models, which are available in Table S4 of the supplementary material.

For the first step, we used the same approach as for major elements, considering the trace element composition of the Calderão diorite gneiss as representative of the parent material. We modeled the trace element signature of liquids formed by either (i) fractional crystallization or (ii) partial melting from this parent composition.

The Rayleigh equation was used to model fractional crystallization; fractionating phases and their proportions were derived from major element modeling of the first differentiation step (Table 1), with addition of trace amounts of allanite (0.05%) and zircon (0.1%). Results are presented in Fig. 14a. The calculated liquids provide a fairly good match to the composition with the trace element pattern of the Boa Sorte granite, which is taken to represent the evolved liquid. In particular, they reproduce well the troughs in Ba, Sr and Eu observed in the latter, as well as the REE profile, including the elevated HREE concentrations (Fig. 14a).

Table 1

<table>
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<tr>
<th>Step 1 (fractionation from diorite to granite)</th>
<th>Starting composition</th>
<th>Final composition</th>
<th>Plagioclase</th>
<th>Biotite</th>
<th>Amphibole</th>
<th>Oxide</th>
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<td>Martel et al. (1999)</td>
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<td>60.50</td>
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<tr>
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<tr>
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R² = 0.11

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<th>Amphibole</th>
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<tr>
<td>CaO</td>
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<td>1.50</td>
<td>1.26</td>
<td>6.25</td>
<td>0.13</td>
<td>10.46</td>
<td>0.00</td>
</tr>
<tr>
<td>Na₂O</td>
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<td>3.32</td>
<td>2.85</td>
<td>8.17</td>
<td>0.26</td>
<td>1.81</td>
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<tr>
<td>K₂O</td>
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<td>2.77</td>
<td>3.24</td>
<td>2.67</td>
<td>10.07</td>
<td>0.23</td>
<td>0.00</td>
</tr>
</tbody>
</table>

R² = 0.77

R² is the sum of squares residuals between the calculated liquid and the expected final composition.

* Determined by calculating the amount of fractionating apatite necessary to balance the difference in P₂O₅ contents between starting and final compositions.
To model partial melting, we considered that melting of the biotite-rich Calderão diorite gneiss would take place as non-modal, incongruent melting such as $\text{Bt + Plag + Qtz} \rightarrow \text{Amp + Liq.}$, as observed in experimental studies (Helz, 1976; Martel et al., 1999; Sisson et al., 2005). We used the incongruent melting equation of Hertogen and Gijbels (1976) and a starting modal composition calculated by mass balance for sample TZ-185 (Table 1). Experiments of Martel et al. (1999) were used to estimate the phase relationships during melting, especially the stoichiometry of the melting equation ($50\% \text{Bt} + 15\% \text{Plag} + 35\% \text{Qtz} \rightarrow 20\% \text{Amp} + 80\% \text{Liq.}$). The results are presented in Fig. 14b. Clearly, the match is not as good as for the fractional crystallization model (Fig. 14a), especially regarding Sr–Eu concentrations that are too high in the models, and HREE concentrations that are too low (Fig. 14b). This results from the significant contribution of plagioclase to the liquid composition and retention of HREE in peritectic amphibole.

For the second differentiation step (granite to rhyolite), we only considered fractional crystallization as a possible genetic link because the rhyolites and Boa Sorte granite are coeval and rhyolites contain enclaves of minerals corresponding to the fractionating phases (as constrained by major element modeling). The observed decrease of Th/U ratios in zircons between the Boa Sorte granite ($\sim 0.7$; sample TZD-268) and the rhyolites ($\sim 0.5$) also supports a crystal fractionation model, where zircon co-crystallizes with a phase competing for Th such as allanite. The Rayleigh fractionation equation was used, and the amount and proportions of fractionating phases were deduced from the least-square calculations based on major elements (Table 1). As with the first step, we added trace amounts (0.1%) of zircon and allanite to the fractionated solid, to account for the observations mentioned above. The calculated liquid compositions yield a very good fit to the average compositions of the rhyolites (Fig. 15), with the exception of Rb and Pb concentrations that are too high and too low, respectively, in the models. Given the high mobility of these two elements, we consider that these discrepancies are minor and do not detract from the efficacy of the model as a whole.

### 5.4. Interpretation of Hf isotope data

The large variations observed in zircon $\varepsilon_{\text{Hf}(3.3 \text{ Ga})}$ (more than 5 $\varepsilon_{\text{Hf}(3.3 \text{ Ga})}$ units for each sample, Fig. 7) most likely reflect heterogeneous
Hf isotopic composition of the magma from which zircon crystallized. The results of geochemical modeling suggest that differentiation of a parent dioritic liquid formed a granitic magma, crystallized at relatively shallow (<10 km) reservoirs and from which highly silicic residual liquid was extracted to erupt as rhyolites. In the scope of this model, the heterogeneous Hf isotopic composition of the rhyolites could result from two (non-exclusive) scenarios: (i) the source (i.e. diorite parent) itself had a heterogeneous Hf isotopic composition (e.g., Appleby et al., 2010; Belousova et al., 2006; Kemp et al., 2008; Kröner et al., 2013, 2014; Villaros et al., 2012); or (ii) the granites and rhyolites were formed by disequilibrium partial melting of the zircon-bearing diorite source, with zircon specifically remaining in the residual phase, which is able to produce large variations in Hf isotopic compositions of the resulting melts (Gerdes and Zeh, 2009; Laurent and Zeh, 2015; Tang et al., 2014).

Despite the variation, almost all zircon $\varepsilon_{Hf}(3.3\text{ Ga})$ values in the rhyolites are negative. In this respect, the two possible differentiation mechanisms of dioritic material for the origin of granites and rhyolites (i.e., partial melting or fractional crystallization) have different implications. The crystal fractionation model is the simplest: it implies that the latter may have had a heterogeneous composition (e.g., Appleby et al., 2010; Belousova et al., 2006; Kemp et al., 2008; Kröner et al., 2013, 2014; Villaros et al., 2012). If partial melting is considered, then it means that the diorite may have had a juvenile isotope composition at the time of emplacement and, in that case, be much older (>3.5 Ga).

In either case, the negative $\varepsilon_{Hf}(3.3\text{ Ga})$ of zircons in the studied rhyolites indicates that the magmatism investigated is the result of melting of older crustal components, potentially as old as 3.5–4.0 Ga according to the Hf model ages. No rocks with this age range have so far been found in the São Francisco Craton, and even detrital and xenocrystic zircon grains of this age are scarce (see Section 5.6 for discussion).

5.5. Comparison with other Archean granitoids and tectonic settings

The studied granites and rhyolites have high SiO$_2$ (~77.7 wt.%) and low Na$_2$O (~3.4 wt.%; K$_2$O/Na$_2$O ~0.8), Al$_2$O$_3$ (~10.7 wt.%), Sr/Y (~0.7), and La/Yb$_{N3}$ (~4.7), and strong Sr and Eu depletion (Sr ~47 ppm; Eu/Eu* ~0.55). These characteristics are completely different compared with major and trace element features of the most common Archean granitoid types, such as “sanukitoid” suites, crust-derived granites (see Laurent et al., 2014a and references therein), and Archean TTGs (e.g., Condie, 2014; Martin et al., 2014; Moyen, 2011), precluding any affinity with these rocks (Fig. 16). This observation, and the fact that the rhyolites are younger than the 3.40–3.33 Ga TTG and medium- to high-K calc-alkaline magmatism of the Gavião Block (Marchesin, 2015; Martin et al., 1997; Nutman and Cordani, 1993; Santos-Pinto et al., 1998, 2012; Silva et al., 2015; Zincone, 2016), contrasts strikingly with the relationship observed in most Paleoarchean terranes, where the felsic volcanics erupted in the greenstone belts are similar in composition to, and coeval with the associated TTG gneisses (Agangi et al., 2015; Benn and Moyen, 2008; Kohler and Anhaeusser, 2002; Moyen et al., 2007; Sanchez-Garrido et al., 2011; Smithies et al., 2007).

The granite and rhyolites have a distinctive trace element signature that makes the assessment of their tectonic setting even more ambiguous than it normally is for the Archean period (e.g., Bédard, 2006; Moyen and Martin, 2012). This exotic geochemistry, such as the higher Y and Yb$_{N3}$ values and lower Sr/Y and La/Yb than classical arc magmas (Basalt–Andesite–Dacite–Rhyolite or “BADR” suites), post-Archean...
granitoids, Archean TTGs, and the average crust itself is illustrated in Fig. 16A–B. On primitive mantle-normalized multi-element diagrams, the granite and rhyolites show negative anomalies of Nb–Ta, Sr and Ti, accompanied by relatively fractionated LREE patterns, which are characteristics that resemble the typical geochemical signature of continental arcs (e.g., Pearce, 1982; Figs. 11 and 16C). However, those negative anomalies, together with the very low Sr/Y ratio relative to both continental crust and modern calc-alkaline series, instead reflect an extensive differentiation process (melting or fractional crystallization) where plagioclase is ubiquitous as the main fractionating phase retaining Sr. These features therefore cannot be used to infer the tectonic scenario where melting took place (Moyen, 2011; Moyen and Martin, 2012). Similarly, Nb–Ta and Ti anomalies can be explained by retention of Fe–Ti oxides in the residue, rather than necessarily advocating a subduction environment (Willbold et al., 2009). Overall, the trace element signature of the rhyolites only indicates extensive plagioclase fractionation, thus depicting shallow melt segregation (<10 kbar; Moyen, 2011) and/or low H2O activity in the system (Beard and Lofgren, 1991; Bogaerts et al., 2006; Tatsumi and Suzuki, 2009).

On the other hand, the rhyolites present a ferroan affinities similar to “A-type” magmas and, just like them, elevated HFSE and HREE contents together with low Al2O3 and CaO (Frost and Frost, 2011; Frost et al., 2001, Whalen et al., 1987). Although care must be taken when applying classical tectonic discrimination diagrams to Archean rocks because the compositional fields in these plots were determined using Phanerozoic granites (Bédard, 2006; Condie, 2015; Moyen and Martin, 2012), our samples consistently correspond to magmas formed away from convergent plate margins in all discrimination diagrams. They lie in the fields of within-plate granites in the diagrams of Pearce et al. (1984) and Harris et al. (1986) (Fig. 17A–B), classify as the “A2” subtype of A-type granites in the ternary Y–Nb–Ce diagram of Eby (1992) (Fig. 17C) and in the combined field of continental rift and ocean-island magmas in the linear discriminant analysis (LDA) based diagrams of Verma et al. (2012, 2013) (Fig. 17D–E). Altogether, these consistent observations suggest that the granites and rhyolites were formed in an intraplate environment.

5.6. Implications for the evolution of the Gavião Block

The plutonic–volcanic system studied here was emplaced forthwith a period of important continental lithosphere formation, represented by scarce 4.1–3.5 Ga remnants and major 3.42–3.33 Ga magmatism. The 4.1–3.5 Ga event is recorded as a minor population of detrital zircon from the supracrustal sequence of the Gavião Block (Zincone and

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**Fig. 17.** Tectonic setting discrimination diagrams. A) Y + Nb vs. Rb (Pearce et al., 1984). The tectonic fields were represented by syn-collision (syn-COL), volcanic arc (VA), within-plate (WP), oceanic ridge granite (ORG), TTG (dashed line) and potassic granites (gray) fields are from Moyen (2011). B) Nb vs. Ga/Al (Whalen et al., 1987). C) Nb–Y–Ce (Eby, 1992). D and E) multi-dimensional discrimination diagrams obtained from linear discriminant analysis of natural logarithm-transformed ratios of immobile major and trace elements (Verma et al., 2012, 2013). The tectonic fields were represented by island arc (IA), continental arc (CA), collision (Col), continental rift (CR) and ocean-island (OI) combined together as within-plate.
The widespread magmatic activity is mainly constrained to the time span from 3.40 to 3.33 Ga (Barbosa et al., 2013; Guitreau et al., 2012; Martin et al., 1997; Nutman and Cordani, 1993; Silva et al., 2015; Zincone, 2016) and is represented by different 3.40–3.36 Ga TTG massifs (Martin et al., 1997; Mougeot, 1996; Nutman and Cordani, 1993; Santos-Pinto et al., 1998, 2012; Silva et al., 2015; Zincone, 2016) and 3.35–3.33 Ga medium- to high-K calc-alkaline magmatism (Marchesin, 2015; Zincone, 2016). The oldest Sete Volta TTGs are so far the best studied (Guitreau et al., 2012; Martin et al., 1997; Nutman and Cordani, 1993; Zincone, 2016) and were interpreted as partial melts of Archean tholeiites leaving a garnet amphibolite residue. The partial melting of the older TTGs to produce younger porphyritic granodiorites was interpreted as reflecting crustal thickening to between 30 and 45 km by mechanisms similar to modern continental collision (Martin et al., 1997). In terms of Hf isotopic evolution, only zircon grains older than 3.42 Ga have positive εHf(t) values, while 3.43–3.37 Ga grains from the Sete Volta massif yield εHf(t) ranging from +2 to −5.4 (Guitreau et al., 2012; Martin et al., 1997; Nutman and Cordani, 1993).

The Jacobina basin and the related quartzites surrounding the investigated rhyolites contain abundant 3.55–3.28 Ga-old detrital zircons (Barbuen et al., 2016; Magee et al., 2001; Teles et al., 2015) (Fig. 18). Among these, many grains crystallized in the 3.43–3.30 Ga time span and show negative εHf(t) values between −0.1 and −7.4 (Teles et al., 2015) that overlap with both the 3.30 Ga zircons from the rhyolite and the 3.43–3.33 Ga zircons from the TTGs. All lines of evidence suggest that the Jacobina basin and related quartzites represent an Archean supracrustal sequence deposited at a maximum age of ca. 3.28 Ga, and that the plutonic–volcanic system and TTGs of the Gavião Block were the main sources of detrital zircons and sediments (Fig. 18). The abundance of ca. 3.30 Ga-old zircons in those siliciclastic units suggests that the volcanic–plutonic association described here represents the remnants of an eroded system that could have originally been larger.

The scenario that best reconciles all observations is that the ca. 3.30 Ga granite–rhyolite system represents the initial stage of an intra-continental rifting following a major 3.43–3.33 Ga period of continental lithosphere formation and stabilization; with continental break-up and deposition of siliciclastic sequences with maximum deposition age of ca. 3.28–3.26 Ga. Remarkably, the plutonic–volcanic system and the siliciclastic metasediments both occur along the paleo-margin of the Gavião Block (i.e., along the Contendas–Jacobina shear zone), which supports that their emplacement was followed by continental break-up. Moreover, close relationships between continental rift settings and silicic volcanism have frequently been observed in early Paleozoic intraplate magmatism (e.g., Carlson Rhyolite volcanic field, Wichita igneous province (Hanson et al., 2013), modern extensional environments (e.g., Continental Rift: Yellowstone, EUA; Nash et al., 2006; Simakim and Bindeman, 2012; Drew et al., 2013, Huang et al., 2015, Rift Arc: Taupo Volcanic Zone, New Zealand; Allan et al., 2012; Béguel et al., 2014, Oceanic Rift: Iceland; Martin and Sigmarsson, 2007; Willbold et al., 2009), and specifically, represent the precursor stage for rift segment propagation events (e.g., Afar triple point; Lalit et al., 2003).

### 5.7. Global perspective

To summarize the interpretations above, the 3.30 Ga-old volcanic–plutonic system of the Gavião Block represents within-plate intracrustal magmatism following a burst of felsic crust magmatism during a period of transition between crustal stabilization and an early stage of continental break-up. Clearly, this model, together with the “crustal” zircon Hf isotopic signatures, suggests that reworking of a felsic protocrust and intracrustal differentiation processes did take place as early as 3.30 Ga in Earth’s history, earlier than the 3.2–3.0 Ga or later age range proposed previously by other works (e.g., Laurent et al., 2014a and references therein).

Remarkably, the plutonic–volcanic system described here is similar in composition to some older (late Hadean to Eoarchean) high-SiO2 ferroan felsic rocks showing high HREE contents and negative Eu anomalies, such as the 4.02 Ga Idiwha tonalitic gneiss associated with the Acasta Gneiss Complex in Canada (Reimink et al., 2014) and the ca. 3.62 Ga iron-enriched augen gneisses and diorite intrusions at the Iltsaq Gneiss Complex, southern West Greenland (Nutman et al., 1984, 1996) (Fig. 16). These rocks were respectively interpreted as reflecting shallow-level magmatic processes coupled with crustal assimilation of rocks previously altered by surface waters (Reimink et al., 2014) and differentiate from crustally-contaminated ferrodiorites and ferрогabbros ponded at the base of the crust (Nutman et al., 1984, 1996). In both cases, those rocks would have been formed in magmatically and/or tectonically thickened crust that underwent internal differentiation associated with intrusion of mafic mantle-derived magma in an intraplate extensional tectonic environment, similar to that inferred for the Gavião volcanic–plutonic system.

Such “intraplate” crustal differentiation processes appear to have been relatively common in the early Earth, even in the Eo- and Paleoarchean. It must be noted, however, that such intraplate magmatism did not produce TTGs but rather magmas of exotic, ferroan “A-type-like” compositions like the granites and rhyolites from the Gavião Block. This questions to some extent the validity (or at least the uniqueness) of “intraplate” models proposed to explain the genesis of TTGs (e.g., Bédard, 2006; Johnson et al., 2013; Zegers and van Keken, 2001).

### 6. Conclusions

The main conclusions of the present study can be summarized as follows:

- The 3.30 Ga rhyolites and granites of the Gavião Block represent the remnants of a vast, eroded volcanic–plutonic system.
- The rhyolites were derived from extraction and eruption of highly silicic residual liquid formed by crystallization of granitic magma in a relatively shallow (<10 km) reservoir.
- The granite may have formed by differentiation of mafic-intermediate material similar to the diorite gneisses that occur regionally.
- The zircon Hf isotopic signatures support that pre-existing felsic crust played a significant role in the origin of these magmas, either through contamination of the parental melts or involvement of an ancient (>3.5 Ga) crustal source component.

![Fig. 18. U–Pb LA-ICP-MS age distribution frequency of less than 5% discordant detrital zircon grains from metasedimentary rocks associated with the Gavião Block using Kernel density estimation plot and data from Teles et al. (2015), Zincone and Oliveira (in review) and Zincone (2016).](image-url)
The 3.30 Ga volcanic-plutonic systems postdate the 3.40–3.33 Ga TTG and medium- to high-K calc-alkaline magmatism related respectively to the thickening and relative stabilization of the Gavião lithosphere. As such, the 3.30 Ga magmatism likely occurred in an intracontinental tectonic setting, immediately following stabilization of newly formed continental crust and marking the early stage of continental break-up.

We argue that such intraplate magmatism and intracrustal differentiation processes took place as early as 3.3 Ga on Earth; and produced magmas that were distinct from the more widespread Archean TTGs, questioning the reliability of “intraplate models” to explain the origin of the latter.

Supplementary data to this article can be found online at doi:10.1016/j.lithos.2016.10.011.

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Appendix A. Analytical techniques

a. Zircon LA-ICP-MS U–Pb age dating

Small chips of riolites were collected on non-weathered outcrops and then crushed and ground by using a jaw crushe and disk mill. Panning and immersion into methylene iodide concentrated the heavy minerals. The diamagnetic zircon grains were separated using Frantz magnetic separator adjusted to a slope of 1°, and current values up to 1.5 A. More than 70 zircon grains per sample were handpicked under a binocular microscope. The grains were mounted in epoxy resin and polished to expose their interiors. The internal structures were revealed by cathodolumescence (CL) imaging using a LEO 430i (Zeiss Company) SEM equipped with an Oxford energy dispersive spectroscopy system and a Gatan Chroma CL detector. The images were taken under 15 kV accelerating voltage, working distance of 16 mm and probe current between 6 and 10 nA.

U–Pb ages were measured by laser ablation mass spectrometry using a Thermo Finnigan Neptune multicollector mass spectrometer coupled to a New Wave UP213 Nd:YAG laser (λ = 213 nm) at the University of Brasilia and using a Thermo Fisher Element XR sector field ICP-MS and a Photon Machines Excite 193 nm ultra-short pulse excimer laser ablation system (Analyte Excite WH) with a HelEx 2 volume cell at the Institute of Geoscience of Campinas University. At the University of Brasilia the typical laser settings were a spot size of 25 μm, a frequency of 11 Hz, and a fluence of approximately 0.8 J/cm2. Argon (approx. 0.90 L/min) and helium (approx. 0.40 L/min) were used to carry the ablated material from the ablation cell to the mass spectrometer. The international zircon GJ-1 (614 Ma;Jackson et al., 2004) was used in a standard-sample bracketing approach, accounting for mass bias, inter-elemental fractionation and drift correction, and the standard 91500 (1068 Ma; Wiedenbeck et al., 1995) was used as the secondary standard to check the accuracy of the corrections. Ratios were calculated according to the method described by Bührn et al. (2009). The masses 204Pb, 206Pb and 207Pb were measured with ion counters, while 238U was analyzed on a Faraday cup. The signal of 202Hg was monitored on an ion counter for the correction of the isobaric interference between 204Hg and 204Pb. At the University of Campinas the laser was regulated with a spot size of 25 μm, a frequency of 10 Hz, and a laser fluence of approximately 4.7 J/cm2. The acquisition protocol adopted was: 30 s of gas blank acquisition followed by the ablation of the sample for 45 s in ultrapure He. All U–Pb data were reduced off-line using loliite software (version 2.5) following the method described by Paton et al. (2010), which involves subtraction of gas blank followed by downhole fractionation correction comparing with the behavior of the 91500 reference zircon (Wiedenbeck et al., 1995). The Peixe zircon standard (564 ± 4 Ma, Dickinson and Gehrels, 2003) was used to check the correction. Common Pb correction was accomplished using VizualAge version 2014.10 (Petrus and Kamber, 2012). The resulting average 207Pb/206Pb age of 1108 ± 27 Ma for 91500 provides confidence on the data. Only ages from a single growth zone and avoiding irregular features such as cracks and inclusions were used. The complete data set of the U–Pb analyses is presented in the online data repository Table S1.

b. Zircon Lu–Hf isotope analyses

Zircon Hf isotope analysis was performed on a Nu Plasma HR MC-ICP-MS (Nu Instruments Ltd., UK) at the State Key Laboratory of Continental Dynamics at Northwest University, Xi’an, China. The Geolas 200M laser ablation system used consists of Com-Pex102 (193 nm ArF excimer laser, Lambda Physik) and optical system (MicroLas). For in situ Lu–Hf isotopic analyses, the interference of 176Hf on 178Hf was corrected by measuring the intensity of the interference-free 175Lu isotope and the recommended 176Lu/175Lu ratio of 0.02669 was applied (De Biévre and Taylor, 1993). Similarly, the interference of 176Yb on 176Hf was corrected by measuring 172Yb and using 176Lu/172Yb ratio of 0.5886 (Chu et al., 2002). Standard zircons 91500 and GJ-1 were used as the reference standards for calibration and monitoring the condition of analytical instrumentation. The detailed analytical technique was described by Yuan et al. (2008). 176Lu/177Hf evolution lines are based on assumed 176Hf/177Hf crustal ratio of 0.0113 (Bouvier et al., 2008). The 176Lu decay constant of Söderlund et al. (2004) and Scherer et al. (2001) was used in the calculation of εHf(t) values. The complete data set is presented in the online data repository Table S2.

c. Major and trace element analyses

Major, minor and trace element analyses on 20 samples were performed at the Institute of Geosciences of University of Campinas – UNICAMP. Major elements were determined on a PanAnalytical PW 2404 X-ray fluorescence spectrometer according to the method described by Vendemiatto and Enzweiler (2001). Glass disks were prepared by fusing of 1 g of pre-ignited sample with 6 g of lithium Li2O·Li2B4O7·Claisse 50:50 flux, for 15 min in Pt–Au crucibles. Loss on ignition (LOI) was measured by heating 4 g of sample at 1000 °C in pre-ignited porcelain crucibles for 1.5 h. The accuracy of results was checked by simultaneous analysis of the international standards BPR-1, and relative errors are 0.4–1.5% for major elements. Minor and trace elements were determined on a Thermo (Xseries2) quadrupole inductively coupled plasma mass spectrometer (ICP-MS), after total digestion with HF/HNO3 (Paar bomb, 5 days, 200 °C), similar to the procedure described by Navarro et al. (2008), and instrument conditions similar to Cotta and Enzweiler (2009). Quality control was performed by simultaneous analysis of the granite GS-N (Centre de Recherches Pétrographiques et Geochimiques), and relative errors are less than 10% deviation from the recommended values. The complete geochemical data set is presented in the online data repository Table S3.

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