

1 **Evidence for a complex accretionary history preceding the amalgamation of**
2 **Columbia: The Rhyacian Minas-Bahia Orogen, southern São Francisco**
3 **Paleocontinent, Brazil**

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19 **Highlights**

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21 . Diachronous Collision of the São Francisco Palecontinent

22 . Protracted orogenic cycle, from pre-subduction to late orogenic collapse is recognized
23 within the Minas-Bahia Orogen

24 . A large continental landmass amalgamated during the Rhyacian (ca. 2.1-2.05 Ga)

25 . Archean-Paleoproterozoic basement units reworked during the Brasiliano Orogeny

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33 **Keywords:**

34 São Francisco Craton, TTG-Sanukitoid, Orogenic collapse, Ribeira Belt, Columbia
35 Supercontinent

36

37 **Abstract**

38 The Minas-Bahia orogeny juxtaposed Archean crustal fragments and Paleoproterozoic
39 magmatic arcs to form the São Francisco-Congo Paleocontinent by the Rhyacian (ca. 2.05
40 Ga). Unravelling the Minas segment of the Minas-Bahia Orogenic Belt (MBO) is an
41 important key to understanding the role of the São Francisco-Congo Paleocontinent in the
42 construction of the Columbia Supercontinent. The Orosirian (ca. 1.9 Ga) final
43 amalgamation of Columbia was preceded by a complex history of accretion of Archean
44 nuclei and Proterozoic magmatic arcs. We present new whole-rock element geochemistry
45 and isotopic (Rb-Sr, Sm-Nd) data and U-Pb ages for granitoid rocks of the main basement
46 complexes located in the southern part of the São Francisco cratonic tip, which displays
47 varied degrees of Neoproterozoic reworking related to the Brasiliano orogeny. Published
48 data for the Campo Belo Complex, the Mineiro Belt and the Piedade Block are combined
49 with the new data set to propose an integrated model for the tectonic evolution of the
50 Minas segment of the MBO. This evolutionary model documents a complete
51 Paleoproterozoic orogenic cycle, from subduction with terrane accretion to collision,
52 followed by late-orogenic collapse. Subduction started diachronously between ca. 2.4 Ga
53 and 2.2 Ga involving various Archean nuclei and Paleoproterozoic magmatic arcs that
54 were later amalgamated during two collisional events at ca. 2.10 and 2.05 Ga. The oldest
55 tonalite-trondhjemite-granodiorite (TTG) to sanukitoid magmatic suites transition are of
56 Neoproterozoic age in the Piedade block, and of Paleoproterozoic age in the Mineiro belt and
57 Mantiqueira complex, apparently indicating a diachronous onset of plate tectonic
58 processes in different crustal segments. The petrogenesis, geochronology and isotopic
59 signatures of these granitoid rocks provide important evidence towards understanding the
60 periodicity of tectonic processes associated with the supercontinent cycle throughout
61 Earth history.

62 **1. Introduction**

63

64 Magmatic rocks typical of the Archean (> 2.5 Ga) and Paleoproterozoic (2.5 to 2 Ga),
65 such as tonalite-trondhjemite-granodiorite (TTG) association, sanukitoid and K-rich
66 granites, are considered key indicators of the tectonic mechanisms that took place during
67 early Earth history (e.g. Sun et. al., 2020; Laurent et. al., 2019; Moyen & Laurent, 2018;
68 Rajesh et. al., 2018; Laurent, et. al., 2014; Moyen & Martin, 2012; Næraa et. al., 2012;
69 Heilimo et. al., 2010; Halla et. al., 2009). The petrogenesis, geochronology and isotopic
70 signatures of these granitoid rocks provide important evidence towards understanding the
71 periodicity of tectonic processes associated with the supercontinent cycle throughout
72 Earth history.

73 Various geochemical, isotopic and geochronological proxies can be used to determine the
74 geodynamic settings of contrasting tectono-stratigraphic terranes and hence to develop
75 tectonic models for the episodic assembly and break-up of supercontinents (Nance, 2019;
76 Murphy & Nance, 2012; Condie, 2011; Hawkesworth et. al., 2010; Santosh & Zhao,
77 2009; Rogers & Santosh, 2002; Zhao, et. al., 2002).

78 The existence of the Columbia supercontinent during Paleo- and Mesoproterozoic times
79 has been proposed with basis on paleomagnetic data as well as geological links between
80 different continental blocks (e.g. Caxito et. al., 2020b; D'Agrella-Filho et. al., 2020;
81 Iaccheri & Bargas, 2020; Wu et. al., 2020; D'Agrella-Filho & Cordani, 2017; Meert &
82 Santosh, 2017; Xu et. al., 2014; Condie, 2013). Though the reconstruction of Archean
83 and Proterozoic supercontinents is challenging due to the incomplete geologic record and
84 reworking during later orogenic processes, their recognition can aid understanding of
85 tectonic processes and mantle dynamics in those ages (e.g. Cawood et. al., 2016; Heron

86 et. al., 2015; Cawood & Buchan, 2007; Collins & Pisarevsky, 2005; Lowman & Jarvis,
87 1995).

88 The Paleoproterozoic orogenic belts of Brazil have crucial role in reconstructive models
89 of the Columbia supercontinent, through the convergence and amalgamation of older,
90 Archean continental fragments and new crust accretion through subduction. These
91 orogenic belts are exposed both within the major cratons, such as the São Francisco, and
92 as reworked basement inliers within the surrounding Neoproterozoic orogens. One of the
93 most studied Paleoproterozoic orogenic belts, and focus of this work, is the Minas-Bahia
94 Orogen (MBO), with the northern (Bahia) and southern (Minas) segments (Figure 1b).

95 The tectonic evolution of the MBO in the southern São Francisco Craton is reasonably
96 constrained by a host of new data recently produced, as recently summarized by Teixeira
97 et. al. (2017a). For a better understanding of this important Paleoproterozoic orogen as a
98 whole, we present new geochemical, U-Pb geochronological, and isotopic Rb-Sr and Sm-
99 Nd data from the basement inliers within the adjacent Neoproterozoic orogenic systems
100 (Figure 1). We also propose new correlations with other cratonic blocks involved in the
101 assembly of the Columbia supercontinent, e.g. the North China block, indicating a
102 diachronous onset of plate tectonic processes in different crustal segments.

103 **2. Tectonic Framework of the São Francisco Paleocontinent**

104 The São Francisco-Congo Craton comprises various Paleoproterozoic orogenic belts and
105 Archean nuclei that were largely unaffected by Neoproterozoic reworking during the west
106 Gondwana supercontinent assembly (Figure 1a, b) (Alkmim & Teixeira, 2017; Caxito et.
107 al., 2017; Teixeira et. al., 2017a; Valeriano et. al., 2017). However, these Neoproterozoic
108 orogens contain slivers of reworked Paleoproterozoic and Archean basement associations
109 that are comparable in most respects to those within the São Francisco-Congo Craton.

110 (Alkmim & Teixeira, 2017; Heilbron et. al., 2017b; Alkmim et. al., 2017) (Figure 1b).
111 We therefore refer to the broader São Francisco Paleocontinent, which incorporates all
112 the Archean and Paleoproterozoic units (Figure 1b).

113 The northern portion of the São Francisco Paleocontinent comprises a number of tectono-
114 stratigraphic terranes that were diachronously amalgamated during the Minas-Bahia
115 Orogeny (MBO): the Archean Gavião and Jequié blocks; the Paleoproterozoic Itabuna-
116 Salvador-Curaçá Belt, characterized by both recycled and juvenile crustal additions; and
117 the mainly juvenile Paleoproterozoic Buerarema Complex (Bersan et al., 2018; Aguilar
118 et al., 2017; Barbosa & Barbosa, 2017; Cruz et al., 2016; Silva et al., 2016; Barbosa &
119 Sabaté, 2004; Silva et al., 2002; Brito Neves et al., 1999; Oliveira et al., 2010) (Figure
120 1b).

121 The southern (Minas) segment of the MBO (Figure 2) comprises Paleoproterozoic
122 magmatic arcs and the Archean Piedade block that were accreted to the northwest onto a
123 large Archean TTG-greenstone belt crustal domain (Bruno et. al., 2020; Cutts, et. al.,
124 2020; Degler et. al., 2018; Albert et. al., 2016; Barbosa et. al., 2015; Teixeira et. al., 2015;
125 Goulart et. al., 2013; Lana et. al., 2013; Heilbron et. al., 2010, Duarte et. al., 2004;
126 Teixeira et. al., 2000, Teixeira et. al., 1996) (Figure 1b).

127 **3. Geological Context of the Study Area**

128 The study area comprises basement inliers exposed in the external thrust-fold belts of the
129 adjacent Neoproterozoic orogenic systems, represented in the west by the southern part
130 of the ESE-verging Brasília Belt (630-605 Ma) and in the east by the younger NW-
131 verging (620 to 530 Ma) Ribeira Belt (eg. Heilbron et. al., 2017b, Trouw et. al., 2013;
132 Valeriano et. al., 2008).

133 Broadly from northwest to southeast, the MBO comprises the following tectonic domains
134 (Figure 2b):

135 i) a foreland domain where the Archean crust is better preserved from the
136 Paleoproterozoic orogeny. It encompasses Meso- to Neoarchean gneissic-granitic
137 greenstone terranes comprising TTG rocks, migmatites, K-rich granitoids,
138 metaultramafic and supracrustal units (Figure 2b) covered by rift to passive margin basin,
139 represented by the (2.6 – 2.1 Ga) Minas Supergroup, and syn- to post-collisional
140 sedimentary successions (e.g. Teixeira et. al., 2017a; Farina et. al., 2015; Romano et. al.,
141 2013; Lana et. al., 2013, Machado et al., 1996; Teixeira & Figueiredo, 1991);

142 ii) the Siderian to the Rhyacian accreted terrane, referred to as Mineiro Belt,
143 consists of magmatic arc associations of juvenile to crustal contaminated high Ba-Sr,
144 TTGs, sanukitoids, hybrid granitoid rocks and associated supracrustals (Araujo et al.,
145 2019a; Cardoso et al., 2019; Moreira et al., 2018; Teixeira et al., 2015; Ávila et al., 2014;
146 Seixas et al., 2013; Noce et al., 2000.);

147 iii) the accreted Piedade Block is interpreted as an Archean microcontinent,
148 juxtaposed against the Mantiqueira Complex by the Ponte Nova Shear Zone, a
149 Paleoproterozoic suture zone (Bruno et al., 2020; Peres et al., 2004). The rocks of the
150 Piedade Block are Neoarchean TTGs and sanukitoids and associated mafic bodies, whose
151 isotopic signatures indicates reworking of a pre-existing continental crust, and minor
152 Paleoproterozoic granitoid rocks with post-collisional signatures (Bruno et al., 2020;
153 Silva et al., 2002);

154 iv) and the Mantiqueira and Juiz de Fora complexes, both accreted
155 Paleoproterozoic magmatic arcs (Bruno et al., 2020; Cutts et. al., 2020; Araújo et al.,
156 2019b; Kuribara et. al., 2019; Degler et. al., 2018; Heilbron et. al., 2017b; Heilbron et.

157 al., 2010; Noce et. al., 2007; Duarte et. al., 2004; Silva et. al., 2002; Brueckner et. al.,
158 2000). Geochemical and isotope data indicates juvenile to crustal contaminated granitoid
159 rocks including TTGs, sanukitoids, post-collisional alkaline, within-plate tholeiitic basic
160 rocks and peraluminous granitoid rocks with crystallization ages ranging from ca. 2.2 Ga
161 to ca. 2.0 Ga, for the Mantiqueira Complex, and ca. 2.4 Ga to ca. 2.0 Ga, for the Juiz de
162 Fora Complex (Figure 2b).

163 **4. Field and Petrographic Features of the Sample Units**

164 The studied basement units are reworked during the Neoproterozoic tectono-
165 metamorphic event, which obliterated most of the primary textures. The metamorphic
166 grade varies from green schist to high amphibolite facies with partial melting from the
167 autochthonous towards the allochthonous domains of the studied area.

168 **4.1 Piedade Block**

169 In the studied outcrops of the Piedade Block, the most common lithology is a biotite
170 orthogneiss with a well-developed metamorphic foliation, locally crosscut by layers of
171 metabasic rocks (Figure 3a, b). The felsic biotite orthogneiss layers are dominated by
172 inequigranular granoblastic texture, with quartz, plagioclase and K-feldspar and rounded
173 mass intergrowths of epidote and clinozoisite, which might represent late replacement of
174 plagioclase during epidote facies low-grade metamorphism (Figure 4a). The metabasic
175 rock usually comprises metric to decametric layers of massive to finely banded
176 metagabbro metamorphosed at epidote-amphibolite facies, with relict igneous
177 clinopyroxene surrounded by coronas of poikiloblastic intergrowth between hornblende
178 and quartz (Figure 4 b, c). Another lithotype is a muscovite -biotite granitoid. Biotite is
179 scarce and related to coarse-grained muscovite crystals, defining a (proto)-mylonitic
180 foliation (Figure 4 d, e). A well-foliated (hornblende)-biotite orthogneiss is also present,

181 with K-feldspar porphyroclasts (Figure 3c). Hornblende is rare and larger than the matrix
182 and presents fine-grained inclusions of opaque minerals. It is often green-brownish in the
183 core and green in the rim (Figure 4f), with late muscovite and chlorite. A banded
184 orthogneiss with migmatitic textures characterized by hornblende-bearing leucosomes
185 surrounded by hornblende-rich melanosomes, biotite and plagioclase is another lithotype
186 (Figure 3d).

187 **4.2 Campo Belo Complex**

188 The main lithotype found in the Campo Belo Complex, within the Parautochthonous
189 Domain is a medium-grained biotite orthogneiss (Figure 3e, f) It is characterized by
190 biotite orthogneiss with the foliation defined by the orientation of brownish biotite. The
191 texture is inequigranular, with microcline porphyroblasts, plagioclase is scarce, and
192 quartz can be found in flaser textures. Titanite is subhedral and parallel to the main
193 foliation (Figure 5a). Epidote occurs as coronas around plagioclase (Figure 5b). The
194 granodioritic biotite orthogneiss main mineralogy is constituted by granular feldspars and
195 quartz in the matrix, with the foliation defined by the orientation of biotite, late muscovite,
196 and accessory epidote with allanite inclusions (Figure 5c, d).

197 **4.3 Mineiro Belt**

198 The main lithology of the Mineiro Belt found in the study area is a banded granodioritic
199 hornblende-biotite orthogneiss with migmatitic textures, exposed in the
200 Parautochthonous Domain (Sample 51B) (Figure 3g) and a granodioritic hornblende-
201 biotite orthogneiss (Figure 3h) found in the Autochthonous Domain (Sample 42). In the
202 Autochthonous domain it is mainly a coarse-grained metagranodiorite metamorphosed at
203 greenschist facies. The main mafic mineral is a greenish biotite with no orientation. There
204 are intergrowths between chlorite and zoisite, with the latter usually in coronas with spike

205 acicular texture (Figure 6a). Rutile interpreted as an igneous relict and always has coronas
206 of titanite that are interpreted as metamorphic (Figure 6b). The other variety found in the
207 Parautochthonous Domain of the Mineiro Belt shows a well-defined folded layering of
208 felsic and mafic bands, metamorphosed at epidote-amphibolite facies, where felsic bands
209 are dominated by granoblastic plagioclase and interstitial quartz, with minor K-feldspar,
210 and mafic layers are dominated by nematoblastic hornblende (Figure 6c) and biotite
211 oriented in the axial plane of the folds, defining a new foliation (Figure 6c). There are
212 some mafic bands where subhedral epidote is developed in the hornblende rims (Figure
213 6c, d) and titanite is fine- to medium-grained with subhedral texture (Figure 6d). Rutile
214 is rare and can be found included in titanite (Figure 6d).

215 **5. Analytical Methods**

216 Lithochemical, Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-
217 ICPMS) U-Pb geochronology and thermal ionisation mass spectrometry (TIMS)
218 radiogenic isotope data were obtained from a range of samples from the Piedade Block,
219 the Campos Gerais Complex, and the Mineiro Belt. Major and trace elements composition
220 analyses, including rare-earth elements (REE) were carried out by Activation
221 Laboratories Ltd (Actlabs, Ancaster, Canada); the cathodoluminescence (CL) and
222 Backscattered electron (BES) imagery were performed in a Scanning Electron
223 Microscopy (SEM) at the School of the Environment, Geography and Geosciences
224 University of Portsmouth, UK; U-Pb data were obtained in the geochronology
225 laboratories of the University of Portsmouth (UK), and the Sm-Nd and Sr isotope data
226 were acquired by the Laboratory of Geochronology and Radiogenic Isotopes (LAGIR)
227 of the Rio de Janeiro State University (Brazil). The analytical procedures of the several
228 techniques and laboratories involved in this work can be found in Supplementary Material
229 A.

230 **6. Results**

231 **6.1 Geochemistry**

232 Fifteen samples were selected for major and trace element composition analyses
233 (Supplementary Material B). Five compositional groups of acid and intermediate rocks
234 (TTG, sanukitoid, hybrid, bitotite- two-mica and alkaline groups) and one compositional
235 group of basic rocks (tholeiitic metabasic) have been discriminated based on their
236 geochemical affinities.

237 **6.1.1 TTG**

238 Sample 50, from the Piedade Block, is characterized as silica-rich (71.8 wt.%), with high
239 Na₂O contents (4.4 wt.%) and Na₂O/K₂O ratio (1,7), low Mg# (37) and is poor in
240 ferromagnesian elements (Fe₂O₃ + MgO + MnO + TiO₂ = 3.7%). It plots on the calc-
241 alkaline series on the AFM diagram (Figure 7a) and on the granite field on the SiO₂ vs.
242 Na₂O+K₂O (TAS) diagram (Figure 7b). The sample is classified as slightly peraluminous
243 ($1.0 \leq A/CNK \leq 1.1$), medium-K calc-alkaline rocks and magnesian (Figure 8a,b,c). It
244 plots in the overlap field of different groups in the ternary diagram of Figure 8d. The TTG
245 affinity, of this one sample, is corroborated by high concentrations of incompatible
246 elements Ba (1365 ppm), Sr (459 ppm), low Y (6ppm), high Sr/Y ratio (76.5), and low
247 concentrations (under detection limit ≤ 20 ppm) of compatible trace elements (Ni, Cr). In
248 the REE chondrite-normalized diagram, it shows enrichment of light rare earth elements
249 (LREE) compared to the heavy rare earth elements (HREE), with high (La/Yb)_N (106)
250 and positive Eu anomalies (Eu/Eu* = 1.26; Figure 9a). In the primitive mantle diagram
251 (Figure 9b), the sample has peaks at Ba, Pb and Zr and troughs at P, Nb and Ti.

252 **6.1.2 Sanukitoids**

253 The second compositional group comprises samples 42, 51B, 52A and 52B from the
254 Mineiro Belt. This group has SiO₂ contents varying from 60.32– 68.15 wt.% (Figure 7a)
255 and plots in the intermediate and acid field of the TAS diagram and on the calc-alkaline
256 series on the AFM diagram (Figure 7b). The samples have been classified as
257 metaluminous (Figure 8a), medium-K calc-alkaline rocks in the SiO₂ vs. K₂O diagram
258 (Figure 8b), magnesian (Figure 8c) and sanukitoids in the ternary diagram 2*A/CNK vs.
259 Na₂O/K₂O vs. 2*FMSB (Figure 8d). They are rich in ferromagnesian oxides ($5 \leq \text{FeOt} +$
260 $\text{MgO} + \text{MnO} + \text{TiO}_2 \leq 25$ wt.%) and CaO. All have high concentrations of compatible
261 trace elements Ni (30-110 ppm), Cr (50-130 ppm) and V (60-152 ppm), in incompatible
262 elements (Ba = 430 – 1251 ppm and Sr = 236 - 1010 ppm), and high Mg# (47-58) for
263 their silica content, corroborating their sanukitoid classification. The chondrite-
264 normalized REE patterns show enrichment in LREE relative to HREE with moderate
265 (La/Yb)_N (8 - 23) and slightly negative to positive Eu anomalies (Eu/Eu* = 0.98 – 1.11)
266 (Figure 9c). In the primitive mantle diagram, all the samples show enrichment in large
267 ion lithophile elements, with peaks in Ba, K, Pb and Sr and troughs at P, Ti, and Nb
268 (Figure 9d).

269 **6.1.3 Biotite-/Two-Mica Granitoids**

270 Samples 65, 68A, 69A and 70B, from the Piedade Block, comprise the third
271 compositional group. They are characterized by high silica contents (SiO₂ > 71.95 wt.%),
272 low concentrations of ferromagnesian oxides ($\text{FeO}_t + \text{MgO} + \text{MnO} + \text{TiO}_2 \leq 4$ wt.%) and
273 CaO (< 3 wt.%), moderate Al₂O₃ ($13.55 \leq \text{Al}_2\text{O}_3 \leq 15.05$ wt.%) leading to high
274 Al₂O₃/(FeOt+MgO). They plot on the calc-alkaline series (Figure 7a) and on the granite
275 field on the TAS diagram (Figure 7b). They have peraluminous affinity (A/CNK ≥ 1),
276 medium to high-K calc-alkaline signatures and plot in the magnesian field of the SiO₂ vs.
277 FeO_t / (FeOt + MgO) diagram (Figure 8a, b, c). In the ternary diagram of Laurent et al.,

278 (2014), the samples plot on the Biotite- Two-mica granites (Figure 8d. The samples have
279 low high field strength elements (HFSE) and transition element contents (such as Zr and
280 V) and Ba/Rb and Sr/Y ratios. REE chondrite-normalized diagram show moderate
281 fractionation of REE patterns ($10.11 \leq \text{La}_N/\text{Yb}_N \leq 36.81$) but significant negative Eu
282 anomalies ($\text{Eu}_N/\text{Eu}^* = 0.25\text{--}0.71$) (Figure 9e). The primitive mantle diagram (Figure 9f)
283 shows high contents in highly incompatible elements such as Rb and Th.

284 **6.1.4 Hybrid Granitoids**

285 Samples 58B, from the Piedade Block, and 48A and 48B from the Campo Belo Complex,
286 plot on the calc-alkaline series (Figure 7a) and have high SiO_2 (wt. 69.15-71.02%),
287 therefore they plot on the granite field on the SiO_2 vs. $\text{Na}_2\text{O}+\text{K}_2\text{O}$ diagram (Figure 7b).
288 They have high Al_2O_3 (15.23-17.14%), moderate Na_2O (3.42-4.69%) and K_2O contents
289 (2.52-3.63%; and therefore $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios, ranging from 0.94-1.86), and low MgO
290 (0.09-0.79%), Cr and Ni (below detection limits) contents are also characteristic of the
291 compositional group. They are peraluminous, medium to high-K calc-alkaline rocks and
292 straddle the ferroan and magnesian fields of the SiO_2 vs. $\text{FeO}_t / (\text{FeO}_t + \text{MgO})$ diagram
293 (Figure 8a, b, c). In the petrogenetic indicator ternary diagram (Figure 8d), the samples
294 plot in the hybrid granites field, reflecting a mixed parentage, likely a mix of felsic
295 magmas and crustal melts (Laurent et al., 2014). The REE chondrite-normalized diagram
296 (Figure 9g) shows enrichment of LREE relative to HREE, with moderate to high (La/Yb)
297 N , ranging from 15.5 to 209. The samples also have slightly negative to positive Eu
298 anomalies ($\text{Eu}_N/\text{Eu}^* = 0.71\text{--}2.15$), high Sr (400 to 508 ppm) and low Y (<1 to 6 ppm)
299 contents, resulting in high Sr/Y ratios of 68–400 (obs.: the higher value of 400 is a
300 minimum estimation, since Y concentration in sample 58B is below the detection limit).
301 In mantle-normalized trace element diagrams (Figure 9h), they are characterized by

302 enrichment of LILE (large-ion lithophile elements), depletion of Nb and Ta, and positive
303 Pb and negative Ti anomalies.

304 **6.1.5 Alkaline Rock**

305 Sample 58A from the Piedade Block is significantly different, as it displays high contents
306 of K₂O (5.08 wt.%) and other incompatible elements (Ba 5729 ppm, Sr 1235 ppm)
307 together with a low Na₂O/K₂O ratio (0.73). In the AFM diagram it plots on the calc-
308 alkaline series, however on the TAS diagram, the sample straddles the subalkaline and
309 alkaline series (Figure 7 a, b). It has high contents of Zr, Nb, Ce and Y corroborating its
310 alkaline signature. Chondrite-normalized REE patterns show high LREE and low HREE,
311 leading to a highly fractionated REE pattern ($La_N/Yb_N \leq 232.97$) and a small negative Eu
312 anomaly ($Eu_N/Eu^* = 0.93$) (Figure 9i). In the primitive mantle diagram, there is clear
313 enrichment in LILEs (Ba, Rb and K) and a negative Ti anomaly (Figure 9j).

314 **6.1.6 Tholeiitic Metabasic Rocks**

315 The metabasic rocks of the study area, represented by samples 68B and 70A from the
316 Piedade Block, are considered as one compositional group. All samples are classified as
317 tholeiitic in an AFM diagram (Figure 7a) and, on the TAS diagram (Figure 7b), they plot
318 on the subalkaline/tholeiitic series. Chondrite-normalized REE shows a flat pattern (1.32
319 $\leq La_N/Yb_N \leq 1.39$) and small negative to absent Eu anomalies ($0.97 \leq Eu_N/Eu^* \leq 1$),
320 indicating an E-MORB (enriched mid-ocean ridge basalts) signature (Figure 10a). They
321 have low TiO₂ (wt.% <2), moderate Mg# (51-52), high contents of CaO, Al₂O₃, and
322 therefore high Al₂O₃/TiO₂ ratios. In the N-MORB-normalized diagram (Figure 10b),
323 despite the metamorphism, the group shows signs of enrichment in incompatible elements
324 such as K, B, La and Rb peaks at Ba, K, Pb and Nd. In the tectonic discriminant diagram

325 of Meschede (1986; Figure 10c) the samples plot in the N-type MORB field. In the Zr vs.
326 Ti diagram of Pearce (1982; Figure 10d), the samples also plot in the MORB field.

327 **6.2 U-Pb Geochronology**

328 Following geochemical analysis, zircon LA-ICPMS U-Pb geochronology data were
329 obtained from twelve samples which covered the range of different geochemical
330 signatures across a wide geographical area, including the Piedade Block, the Mineiro Belt
331 and the Campo Belo Complex (Figure 2). Tables with analytical results are provided in
332 Supplementary Material C.

333 **6.2.1 Piedade Block**

334 The results of sample 50, a hornblende biotite gneiss from the TTG compositional group
335 from the Piedade Block (Figure 2), are complex with thirty-six analyses of zircon cores
336 and rims arranged into two discordia lines. (Figure 11a). The most common zircon
337 morphology is prismatic (subhedral), with rounded grains, that display fine igneous
338 oscillatory zoning in the cores surrounded by homogeneous bright and dark rims. The
339 first discordia line defined by cores with fine igneous oscillatory zoning and dark
340 metamorphic rims yielded an upper intercept of 2776 ± 28 Ma and a lower intercept of
341 1930 ± 37 Ma (Figure 11a). Six analysis of cores yielded a concordia age of 2691 ± 18
342 Ma, interpreted as the crystallization age (Figure 11b); the analysis of the dark rims in the
343 spots#20 and #34, with 98% and 99% of concordance, yielded $^{207}\text{Pb}/^{206}\text{Pb}$ age of $2021 \pm$
344 30 Ma and 2043 ± 29 Ma, respectively, and are interpreted as metamorphic ages. The
345 second discordia line, defined by cores with fine igneous oscillatory zoning and bright
346 metamorphic rims, yielded an upper intercept of 2705 ± 13 Ma and a lower intercept of
347 592 ± 44 Ma (Figure 11a). Three analyses of the bright metamorphic rims yielded a
348 concordia age of 578 ± 3 Ma, interpreted as the Neoproterozoic metamorphic overprint
349 (Figure 11c).

350 Zircon grains from Sample 65, a felsic orthogneiss of the biotite- two-mica compositional
351 group collected close to the Ponte Nova Shear Zone (Figure 2), are prismatic to subhedral
352 displaying inherited cores, fine igneous oscillatory zoning, and bright metamorphic rims.
353 Eighteen analyses define a discordia with an upper intercept of 2011 ± 30 Ma and a lower
354 intercept of 540 ± 27 Ma, interpreted as igneous crystallization and metamorphic
355 overgrowth ages, respectively (Figure 11d). The analysis of nine inherited cores are
356 shown as blue ellipses in the concordia diagram and have $^{207}\text{Pb}/^{206}\text{Pb}$ ages from 2767 ± 9
357 Ma (spot #10) with Pb loss towards the Neoproterozoic metamorphic overprint. (Figure
358 11 d).

359 Sample 68A is also a biotite gneiss of the biotite- two-mica compositional group of the
360 Piedade block (Figure 2). The most common morphology of the zircon grains is prismatic
361 subhedral with fine igneous oscillatory zoning, sometimes showing inherited cores
362 (Figure 11e). The grains are mostly fractured in the BSE images and are translucent to
363 opaque with light to deep yellow colors. The analysis of an inherited core yielded a
364 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2497 ± 36 Ma (spot #35). Thirteen analyses yield a discordia line with
365 upper intercept at 2051 ± 30 Ma and lower intercept at 449 ± 44 Ma, the latter interpreted
366 as Pb loss during Neoproterozoic metamorphism. Seven analyses yielded a concordia age
367 of 2048 ± 28 Ma, interpreted as the igneous crystallization age (Figure 11e).

368 Sample 70A, a tholeiitic metabasic rock from the Piedade Block (Figure 2), displays
369 translucent prismatic euhedral to subhedral, with a few rounded, zircon grains, varying
370 from light to dark brown colours. Six discordant analyses of inherited cores show
371 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2718 ± 14 Ma with 93% of concordance (spot #23) to $^{207}\text{Pb}/^{206}\text{Pb}$ age
372 of 2458 ± 13 Ma with 90% of concordance (spot#6) (Figure 11f). Sixteen analysis of fine
373 igneous oscillatory zoning of the prismatic grains yielded a concordia age of 2053 ± 18
374 Ma, interpreted as the crystallization age (Figure 11f).

375 Sample 70B, a biotite orthogneiss of the biotite- two-mica compositional group, collected
376 in the same outcrop of Sample 70A (Figure 2), has the most common morphology of the
377 zircon grains, namely prismatic subhedral showing cores and rims with igneous
378 oscillatory zoning (Figure 11g). The grains are commonly fractured in the BSE images
379 and mostly translucent with light to deep brown colors. Twenty core analyses scatter
380 along the Concordia, varying from a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2866 ± 35 Ma (spot #16; 98% of
381 concordance) to $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2110 ± 44 Ma (spot #14; 91% of concordance) are
382 interpreted as inheritance ages. Four analyses (spots #11, #15, #20, #26) in zircon grains
383 with igneous oscillatory zoning yield a concordia age of 2059 ± 35 Ma, interpreted as the
384 crystallization age (Figure 11g).

385 Zircon grains from sample 58A, a hornblende biotite gneiss from the alkaline
386 compositional group from the Piedade Block (Figure 2), are rounded to subhedral with
387 fine igneous oscillatory zoning. Thirty-three analyses of concordant to sub concordant
388 grains vary from $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2130 ± 31 Ma (spot #17) with 96% of concordance to
389 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2021 ± 32 Ma (spot #25) with 91% of concordance. The average
390 $^{207}\text{Pb}/^{206}\text{Pb}$ mean age of 2058 ± 6 Ma is within error of the upper intercept age of $2078 \pm$
391 6 Ma (all zircons) and interpreted as the crystallization age (Figure 12a).

392 Zircon grains from sample 58B, a biotite-muscovite orthogneiss from the hybrid granitoid
393 compositional group collected in the same outcrop of Sample 58A, are subhedral to
394 rounded, translucent to opaque brownish colors with oscillatory zoning in the cores and
395 bright metamorphic rims. Thirty-three analyses yield a discordia with an upper intercept
396 of 2054 ± 24 Ma, interpreted as the crystallization age, and a lower intercept of 538 ± 47
397 Ma, interpreted as indicative of the Neoproterozoic metamorphic overprint (Figure 12b).

398 **6.2.2 Mineiro Belt**

399 Sample 42, a metagranodiorite of the sanukitoid compositional group collected in the
400 Autochthonous Domain (Figure 2), yielded an upper intercept of 2139 ± 40 Ma with the
401 most common morphology of the zircon grains being prismatic subhedral with fine
402 igneous oscillatory zoning (Figure 13a). The grains are mostly translucent with light to
403 deep yellow colors. The average mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the 32 analyses, with at least
404 95% of concordance, give an age of 2114 ± 10 Ma, interpreted as the crystallization age
405 (Figure 13b).

406 Sample 51B was an amphibolitic gneiss of sanukitoid composition collected in the
407 Parautochthonous Domain (Figure 2). Thirty analyses of subhedral to euhedral prismatic
408 zircon grains, displaying fine igneous oscillatory zoned cores and bright rims yield an
409 upper intercept of 2117 ± 24 Ma and a lower intercept of 551 ± 36 Ma, interpreted as
410 crystallization and metamorphism ages, respectively (Figure 13c). Five discordant
411 analyses of bright cores surrounded by igneous oscillatory zoning textures are interpreted
412 as inheritance (Figure 13c).

413 Zircon grains from sample 52B, a hornblende biotite orthogneiss from the sanukitoid
414 compositional group also collected in the basement thrust sheets of the Brasilia Belt
415 (Figure 2), are mostly translucent, varying from light to dark yellow colours and euhedral
416 to subhedral showing fine igneous oscillatory zoning in the cores and bright rims (Figure
417 13d). Sixty analyses yield a discordia line with an upper intercept of 2152 ± 11 Ma,
418 interpreted as the crystallization age, and a lower intercept of 577 ± 32 Ma, interpreted as
419 metamorphism related to Neoproterozoic orogeny (Figure 13d).

420 **6.2.3 Campo Belo Complex**

421 Zircon grains from sample 48A, a biotite gneiss from the hybrid granitoid compositional
422 group of the Parautochthonous Domain (Figure 2), are prismatic, subhedral to euhedral,

423 and show dark cores surrounded by fine igneous oscillatory zoning textures and small
424 bright rims (Figure 14b). The analyses of the cores are interpreted as inheritance, with
425 $^{207}\text{Pb}/^{206}\text{Pb}$ ages from 3007 ± 90 Ma (spot #16; with 98% of concordance) to an age of
426 2621 ± 93 Ma (spot #4; with 88% of concordance). The analyses of the igneous oscillatory
427 zoned areas and the bright rims yield a discordia line with an upper intercept of $2165 \pm$
428 39 Ma and a lower intercept of 585 ± 67 Ma, interpreted as corresponding to the ages of
429 crystallization Neoproterozoic metamorphism, respectively (Figure 14b). Five analyses
430 of concordant to sub concordant grains yielded a concordia age of 2090 ± 24 Ma,
431 interpreted as a better constraint on the crystallization age of the rock (Figure 14b).

432 The results of sample 48B, a biotite gneiss of the hybrid compositional group collected
433 in the same outcrop of Sample 48A (Figure 2), are complex with twenty-seven analyses
434 of cores and rims in prismatic brownish zircon grains, showing dark cores and rims with
435 igneous textures, being interpreted to correspond to inheritance and crystallization,
436 respectively. The analysed cores show $^{207}\text{Pb}/^{206}\text{Pb}$ ages from of 2794 ± 36 Ma (spot #19;
437 with 91% of concordance) to 2009 ± 53 Ma (spot #16; with 89% of concordance), being
438 interpreted as inheritance, showing a Pb loss trend towards the Neoproterozoic (Figure
439 14c). Eighteen analyses of the rims yield a discordia line with upper intercept at $2096 \pm$
440 59 Ma, and a lower intercept of 931 ± 390 Ma, interpreted as the ages of crystallization
441 and Neoproterozoic metamorphism, respectively.

442 **6.3 Isotope data**

443 Eleven samples were selected based on their geochemical affinities and different
444 tectonostratigraphic terranes. Samples from the Piedade Block, Mineiro Belt and Campo
445 Belo Complex show a wide variety of $\epsilon\text{Nd}(t)$ values, varying between juvenile (+1.3) to
446 strongly evolved (-16), at their respective crystallization ages (Supplementary Material
447 D). For sample 69, the crystallization age used for $\epsilon\text{Nd}(t)$ calculations was that obtained

448 for other samples from the same lithogeochemical group and tectonostratigraphic terrane.
449 In the $\epsilon\text{Nd}(t)$ vs crystallization age diagram (Figure 15), the different isotopic fields of
450 the Archean and Paleoproterozoic terranes previously published for the studied area are
451 shown in order to compare with the results obtained in this study.

452 The diversity of Initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, vary widely, from 0.700 to 0.788
453 and 0.5091 to 0.5107, and relate to the respective lithogeochemical group of the studied
454 samples, despite their crystallization age. Lower initial $^{87}\text{Sr}/^{86}\text{Sr}$ correlated with the
455 chondritic $\epsilon\text{Nd}(t)$, reproduces the mantle isotopic reservoir that originated the tholeiitic
456 metabasic and sanukitoid groups, while the higher values of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are also
457 correlated with negative $\epsilon\text{Nd}(t)$ values and likely represent reworking of pre-existing
458 continental crust, represented by the biotite- two-mica and hybrid compositional group.

459 **7. Discussions**

460 **7.1 Petrogenetic Implications**

461 Combining isotopic information and whole-rock geochemistry of igneous rocks is a
462 powerful tool for understanding igneous petrogenesis and related tectonic processes
463 (Moyen, 2019; Laurent et. al., 2011), providing different petrogenetic scenarios for the
464 studied lithogeochemical groups. The initial ($t = \text{U-Pb}$ crystallization age) ϵNd vs.
465 $^{87}\text{Sr}/^{86}\text{Sr}$ diagram of Figure 16 highlights the role of continental crust contamination in
466 the generation of magmatic suites. In this context, different rock assemblages show varied
467 degrees of mantle participation in their origins, from the least contaminated, mantle-
468 derived TTG (Sample 50 of the Piedade Block) and sanukitoid (Samples 42, 51B and 52B
469 of the Mineiro Belt) suites, to the more contaminated magmatic suites of hybrid, alkaline
470 and biotite-bearing or two-mica granitoid rocks. Younger tholeiitic basic magmatism,
471 represented by sample 70A, shows a subcontinental lithospheric mantle source (Figure
472 16).

473 The combination of initial ($t =$ crystallization ages) ϵNd vs whole-rock element
474 compositions confirm the varied degrees of mantle interaction of the aforementioned
475 lithochemical groups (Figure 17 a, b). This is exemplified in the binary plot of #mg
476 vs. initial ϵNd and of #mg vs. initial $^{87}\text{Sr}/^{86}\text{Sr}$. The highest ϵNd is correlated with the
477 highest #mg, showing the mantellic signatures of the aforementioned rocks and the
478 highest $^{87}\text{Sr}/^{86}\text{Sr}$ with the lowest #mg, indicating a more continental signatures for the
479 samples.

480 The sanukitoid magmas were likely sourced from a fluid metasomatized mantle wedge,
481 reinforcing recent studies about the repetition in time and space of this type of magmatism
482 in the geological record, implying subduction of sediment followed by its interaction with
483 the mantle wedge (Bruno et al., 2020; Laurent et. al., 2019; Moreira et. al., 2018, Laurent
484 et. al., 2014). Remarkably, the samples show mantle affinity and isotopic signatures close
485 to chondritic compositions (Figure 15) and in the Sr + Ba (ppm) vs. (FeO+MgO wt.%)
486 diagram (Figure 17c) the source of the sanukitoid group is characterized as the melting
487 of enriched mantle. Among the Palaeoproterozoic sanukitoids, only one sample (51B) has
488 a negative ϵNd value below -5, the other two (Sample 42, 52B) have a less evolved Nd
489 signature. The latter probably reflects the involvement of coeval sediments (~2.1 Ga)
490 acting as metasomatic agents in the mantle wedge. Furthermore, the first could be
491 accounted for by an Archaean sedimentary contribution due to overlap with crustal arrays
492 in the ϵNd space (Figure 16) and the presence of inherited zircon grains (Figure 13c). The
493 Archaean TTG (sample 50) shares the melting of enriched mantle (Figure 17c) of the
494 sanukitoid suite and show elevated MgO, sharing the characteristics of the Low Al and
495 High-HREE, what favors a geodynamic setting with shallow low-pressure source
496 possibly with a mantle involvement (Halla et. al., 2009).

497 Samples of this study classified as two-mica granitoids have overall the lowest ϵNd values
498 and the highest $^{87}\text{Sr}/^{86}\text{Sr}$ (initial). Because of the peraluminous geochemical affinity of
499 these samples and low $\text{FeO}+\text{MgO}$ (wt.%), it is assumed that the samples generated by
500 partial melting of crustal lithologies, therefore implying the presence of a significant
501 sedimentary contribution to the protolith (Figure 17 c, d) (e.g. Bucholz et al., 2018). This
502 interpretation is supported by the presence of inherited grains with a range of
503 Paleoproterozoic and Archaean ages (Figure 11 d, e, g), likely sourced during partial melt
504 of the mid-crust.

505 A distinct lithochemical group found in the study area is represented by the hybrid
506 granitoids (Samples 48A and 48B of the Campo Belo Complex and Sample 58B of the
507 Piedade Block) classified after the ternary diagram of Laurent et. al. (2014). They are
508 classified as late- to – post-collisional and despite the relatively high K_2O content, the
509 chemical signatures of these rocks resembles those of adakitic rocks (high Na_2O , Al_2O_3 ,
510 Sr, Eu contents, low Yb and Y contents, enrichment in LREEs and LILEs and depletion
511 in HREE, high Sr/Y and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ and La/Yb(N) ratios (Drummond and Defant, 1990,
512 Martin et al., 2005). It is highlighted that their crystallization ages are different, ca. 2.10
513 Ga for samples 48A and 48B of the Campo Belo Complex and ca. 2.05 Ga of the Piedade
514 Block, representing the same geological process at different times, since they are related
515 to different collisional events.

516 Several models have been proposed for the petrogenesis of adakitic granitoids, including:
517 (1) assimilation–fractional crystallization from a parental basaltic magma (Rooney et al.,
518 2011; Castillo et al., 1999), (2) partial melting of a young subducted oceanic slab
519 (Castillo, 2012; Martin et al., 2005), (3) partial melting of delaminated lower continental
520 crust with mantle interaction (Wang et al., 2006; Xu et al., 2006; Gao et al., 2004; Kay
521 and Kay, 2002; Xu et al., 2002), and (4) partial melting of thickened mafic lower crust

522 (Huang et al., 2009; Gao et al., 2004; Chung et al., 2003; Xiong et al., 2003; Kay and
523 Kay, 2002; Atherton and Petford, 1993). The following observations with respect to the
524 studied rocks seem pertinent. Although fractionation of garnet, amphibole and plagioclase
525 from a basaltic magma could produce residual melts with adakitic compositions, our
526 samples have narrow ranges of MgO and SiO₂. Thus, some form of direct generation by
527 crustal partial melting is more likely the dominant process involved in the petrogenesis
528 of these rocks. Low MgO, Mg#, Cr and Ni contents contrast with those expected for
529 adakites derived from slab-derived magmas or delaminated lower continental crust with
530 mantle interaction (Castillo, 2012; Martin et al., 2005). On the other hand, chemical and
531 petrographic characteristics are indicative of high-pressure melting, with garnet and no
532 plagioclase in the residue (as positive Sr, Eu anomalies, high Sr/Y ratio, low Y and Yb
533 contents. Thus, the rocks presented here have affinities with adakites derived from
534 thickened lower crust (Liu et. al., 2020).

535 The crystallization of the post-collisional granitoids can indicate a delamination of the
536 lower crust and lithospheric mantle providing an asthenosphere inflow in a favorable
537 configuration for melting, responsible for the bi-modal magmatism found in the
538 Paleoproterozoic samples of Piedade Block (eg. Laurent et. al., 2019).

539 **7.2 Southern São Francisco Paleocontinent Tectonic Model**

540 Within the units of the São Francisco Paleocontinent, the ca. 2.4 to 2.07 Paleoproterozoic
541 granitoid rocks with high Ba-Sr, sanukitoid and TTG signatures suggest a complex and
542 diachronous subduction history (Bruno et. al., 2020; Cutts, et. al., 2020; Araújo et. al.,
543 2019b; Moreira, et. al., 2018; Degler et. al., 2018; Heilbron et. al., 2017b; Teixeira et.
544 al., 2015, Seixas et. al., 2013). Late- to post-collisional granitoid rocks between ca. 2.1
545 and 2.06 Ga and an important metamorphic record at ca. 2.05 to 2.04 Ga, with the
546 development of associated foreland basins marked the final transition to extensional

547 collapse (Cutts, et. al., 2019; Alkmim & Teixeira, 2017; Aguilar Gil et. al., 2015; Vlach
548 et. al., 2003; Brueckner et. al., 2000; Machado et. al., 1992) (Figure 18).

549 A complete orogenic cycle from subduction to collisional stages, followed by late
550 orogenic lithospheric collapse with associated magmatism is recognized within the
551 different tectono-stratigraphic terranes of the Minas segment of the MBO (Figure 18).
552 Its Siderian-Rhyacian evolution, which led to the amalgamation of the São Francisco-
553 Congo Paleocontinent, is characterised by the diachronous accretion of Archean blocks
554 with cordilleran (e.g. Mantiqueira Complex) and intra-oceanic juvenile accretionary arcs
555 (e.g. Mineiro Belt and Juiz de Fora Complex). Based on the geochronological, isotopic
556 and lithogeochemical constraints we propose a tectonic model for the Minas-Bahia
557 orogen. Available data for the Archean complexes of the southern São Francisco
558 Paleocontinent show that Meso- and Neoproterozoic polycyclic terranes are mainly
559 composed of TTG, migmatites and potassic granitoid suites (e.g. Alkmim & Teixeira,
560 2017). Based on description by Bruno et. al., (2020) and on the results of this study
561 (Sample 50), the Neoproterozoic Piedade Block is composed of TTGs and sanukitoids, with
562 subordinated within-plate alkaline basalts. The existence of an Archean microcontinent,
563 represented by the Mantiqueira Complex, is indicated by the presence of Archean
564 inherited zircon grains in Paleoproterozoic arc-related granitoid rocks (eg. Heilbron et.
565 al., 2010) and by the Nd and Sr isotope signatures indicative of a mixed crust-mantle
566 origin (Bruno et al., 2020).

567 From 2.4 to 2.1 Ga (Figure 19a), the crystallization of juvenile magmatic arc rocks of in
568 the Mineiro Belt and of the granitoid rocks of the Juiz de Fora Complex, are interpreted
569 to represent the early, accretionary stages of the southern portion of the MBO (Moreira
570 et. al., 2018, Teixeira et al., 2017, Barbosa et. al., 2015; Ávila et. al., 2014; Ávila et. al.,
571 2010, Heilbron et. al., 2010, Teixeira & Figueiredo, 1991).

572 As the magmatic arc system was constructed, an east-directed subduction system of ca.
573 2.2 – 2.1 Ga is interpreted as responsible for the generation of rocks with mixed mantle-
574 crust isotopic signature in the Juiz de Fora Complex, and of coeval TTGs and sanukitoid
575 suites in the Mineiro Belt (Samples 42, 51B and 52B) and Mantiqueira Complex (e.g.
576 Bruno et. al., 2020; Degler et. al., 2018; Heilbron et. al., 2017b; Ávila et. al., 2014;
577 Duarte et. al., 2004) (Figure 19a). For the Mineiro Belt, our interpretation follows the
578 model recently proposed by Araujo et al. (2019a) in which a double-sided subduction
579 was responsible for the origin of magmatic arc plutonic rocks and supracrustal units. For
580 the Juiz de Fora magmatic arc, eastward subduction is necessary to explain the
581 geochemical zoning described by Heilbron et al (2010). Double-sided subduction for the
582 Mineiro Belt also seems necessary to explain the ca. 2.1-2.05 Ga late-to post-collisional
583 granitoids observed in the Piedade block and to account for the dome-and-keel structure
584 that affected the foreland of the São Francisco paleocontinent (eg. Cutts et. al., 2018).

585 This stage in geologic record (ca. 2.1 – 1.9 Ga) is marked by the generation of high Ba-
586 Sr granitoid rocks in the Mineiro Belt, Mantiqueira and the Juiz de Fora Complexes
587 (Figure 19b) (Bruno et. al., 2020, Araújo et. al., 2019b; Moreira et. al., 2018; Heilbron
588 et al., 2010, Duarte et. al., 2004). The ca. 2.1 Ga accretion of these terranes against the
589 Archean Campo Belo complex took place along a suture zone referred as the Jeceaba-
590 Bom Sucesso lineament (Moreira et. al., 2018, Alkmim & Teixeira, 2017, Ávila et. al.,
591 2010).

592 The ensuing late- to post-collisional magmatism generated ca. 2.05 Ga high-K granitoid
593 rocks and adakites in the Campo Belo complex (samples 48A and 48B). An important
594 tectono-metamorphic episode has been recognized at ca. 2.05 – 2.04 Ga by Heilbron et
595 al. (2010) and a granulite facies metamorphic event at ca. 2.05 Ga by Bruno et. al. (2020)
596 at the Ponte Nova suture zone. The latter is thought to correspond to the collision

597 between the Mantiqueira Complex and the Piedade Block, which took place coeval to
598 the development of the Abre Campo suture zone between the Mantiqueira and Juiz de
599 Fora Complexes (Degler et. al., 2018; Heilbron et. al., 2010). This time interval is also
600 marked by the late- to post-collisional magmatism in the Piedade Block with the
601 crystallization of tholeiitic metabasic (Sample 70A) and alkaline rocks (Sample 58A)
602 (Figure 19d), which are diagnostic of extensional intraplate settings, with associated
603 adakites (sample 58B) and biotite-bearing or two-mica granites (Samples 65, 68A, 70B)
604 related to coeval melting of continental crust. Coeval 2.05 Ga post-collisional processes
605 in the foreland zone of this orogenic system are represented by ‘dome-and-keel’
606 tectonics in and by the sedimentation of quartzitic sediments of the Itacolomi Group
607 overlying the inverted Minas Supergroup passive margin successions (Cutts et. al., 2018;
608 Machado, et. al., 1996).

609 The ca. 1.9 Ga orogenic collapse stage in the southern São Francisco Paleocontinent is
610 marked by the presence of within-plate alkaline basalts with OIB-like signatures and of
611 alkaline rocks (Figure 19b) (Bruno et. al., 2020, Cutts, et. al., 2019, Heilbron et. al.,
612 2010).

613 **7.3 Regional Correlations and implications for the assembly of the Columbia** 614 **supercontinent**

615 The determination of precise ages and isotopic signatures of Archean and
616 Paleoproterozoic tectono-stratigraphic terranes in different cratons is an important tool in
617 understanding the processes that led to the formation of the first supercontinents
618 (Terentiev & Santosh, 2020; D’Agrella-Filho et. al., 2020; Iaccheri & Bargas, 2020; Wu
619 et. al., 2020; D’Agrella-Filho & Cordani, 2017; Meert & Santosh, 2017; Xu et. al., 2014;
620 Condie, 2013; Mitchell et. al., 2012/ Rogers & Santosh 2009). Additional to the
621 contribution made by paleomagnetism, correlations between the periods of major juvenile

622 crustal growth and recycling play an important role in such reconstructions (e.g. Cawood,
623 et. al., 2016; Condie, 2013; Santosh, 2009). Several authors (Evans & Mitchell, 2011;
624 Yakubchuk, 2010; Zhao et. al., 2004) postulate the assembly of the Columbia
625 supercontinent during the ca. 1.9 Ga global peak in accretionary orogenesis but others
626 suggest a later assembly at ca. 1.6 Ga (e.g. Terentiev & Santosh, 2020; Caxito et. al.,
627 2020b; Meert & Santosh, 2017; Pisarevsky et. al., 2014; Rogers & Santosh, 2009).

628 Within the units of the São Francisco Paleocontinent, the ca. 2.4 to 2.07 Paleoproterozoic
629 granitoid rocks with high Ba-Sr, sanukitoid and TTG signatures suggest a complex and
630 diachronous subduction history (Bruno et. al., 2020; Cutts, et. al., 2020; Araújo et. al.,
631 2019b; Moreira, et. al., 2018; Degler et. al., 2018; Heilbron et. al., 2017b; Teixeira et. al.,
632 2015, Seixas et. al., 2013). Late- to post-collisional granitoid rocks between ca. 2.1 and
633 2.06 Ga and an important metamorphic record at ca. 2.05 to 2.04 Ga, with the
634 development of associated foreland basins marked the final transition to extensional
635 collapse (Cutts, et. al., 2019; Alkmim & Teixeira, 2017; Aguilar Gil et. al., 2015; Vlach
636 et. al., 2003; Brueckner et. al., 2000; Machado et. al., 1992).

637 The southern Archean complexes of the São Francisco Paleocontinent (e.g. Bonfim, Belo
638 Horizonte, Santa Bárbara), and the reworked Campo Belo Complex may represent the
639 southern continuation of the Archean Gavião Block (Figure 20a). Minor Archean
640 continental blocks such as the Piedade, Guanhões, Guararema, Jequié and Serrinha were
641 welded together as a result of Paleoproterozoic accretionary tectonics and were intruded
642 by large volumes of plutonic rocks generated by crustal recycling and addition of juvenile
643 arcs. Parts of the Minas-Bahia Orogen display more accretionary characteristics, such as
644 the Mineiro Belt, the Juiz de Fora, and the Buerarema Complex. Others, such as the
645 Itabuna-Salvador-Curaçá Belt and the Mantiqueira Complex, show both juvenile
646 additions within Cordilleran tectonic settings (Figure 20b).

647 Our proposed tectonic scenario involves the coeval development of passive margins
648 around the Archean blocks and outboard juvenile arcs at ca. 2.6 Ga, followed by a period
649 of arc-related magmatism (with both Cordilleran and intra-oceanic signatures), building
650 the Paleoproterozoic orogenic belt at ca. 2.4 to 2.06 Ga (Figure 20). All the various crustal
651 fragments were diachronously amalgamated between 2.1 and 2.04 Ga, with the
652 development of foreland basins and metamorphic core complexes. The orogenic cycle
653 was terminated by bimodal magmatism and extensional deformation between 2.02 to 1.9
654 Ga, altogether interpreted as related to slab detachment and orogenic collapse (Bruno et.
655 al., 2020; Cutts, et. al., 2020; Araújo et. al., 2019b; Cutts et. al., 2019; Moreira, et. al.,
656 2018; Degler et. al., 2018; Alkmim & Teixeira, et. al., 2017; Heilbron et. al., 2017b;
657 Teixeira et. al., 2017a; Teixeira et. al., 2015, Aguilar Gil et. al., 2015; Seixas et. al., 2013;
658 Vlach et. al., 2003; Brueckner et. al., 2000; Machado et. al., 1992).

659 The assembly of the São Francisco Paleocontinent by the amalgamation of Archean
660 blocks and Rhyacian orogens before ca. 2.0 Ga is similar in its timing and mechanism to
661 the formation of the West Africa Block/Central African Block/Transamazonian
662 (Terentiev & Santosh, 2020; Caxito et. al., 2020a; D'Agrella-Filho et. al., 2017; Cordani
663 et al., 2013), both prior to the main orogenic period that consolidated the Columbia
664 supercontinent (Figure 21a). Similar accretionary processes had terminated by ca. 1.95
665 Ga in the Khondalite Belt of North China (Li et. al., 2020; Liu et al., 2012; Santosh, 2010).
666 However, based on paleomagnetic poles in mafic dykes (ca. 1.7 Ga), that crop out in most
667 of the building blocks of Columbia, and the geometry of Paleoproterozoic belts,
668 D'Agrella-Filho et. al., (2020) proposed that the Rio de la Plata, São Francisco-Congo,
669 North China, and India blocks were all connected as part of Columbia by this time (Figure
670 21b). Zhao et. al. (2004) suggests the proximity of the São-Francisco Congo and the North

671 China cratons due to Statherian dyke swarms and related volcanic rocks, supported by
672 paleomagnetic poles (Xu et. al., 2020; Salminen et. al., 2016).

673 **8. Concluding Remarks**

674 The reworked basement units of the external segments of the Neoproterozoic Brasília
675 and Ribeira Orogenic Belts, SE Brazil, are considered to be part of the Minas-Bahia
676 Orogenic Belt, despite the Brasiliano metamorphic overprint (ca. 600-560 Ma).
677 Integrating new data with previous contributions a protracted orogenic cycle, from
678 subduction to collisional stages and a subsequent orogenic collapse with associated
679 magmatism. Cordilleran magmatic arcs that developed along the margins of Archean
680 microcontinents (e.g. Mantiqueira Complex) and intra-oceanic juvenile accretionary arcs
681 (e.g. Mineiro Belt and Juiz de Fora Complex) were diachronously amalgamated with
682 Archean nuclei to form the São Francisco-Congo Paleocontinent, between 2.1 to 2.05
683 Ga. The collisional episodes were followed by tectonic collapse, associated with bimodal
684 magmatism between 2.05 and 1.98 Ma. The presence of Archean and Paleoproterozoic
685 TTG, sanukitoids and adakitic rocks point to the importance of subduction processes as
686 driving forces building the São Francisco-Congo Paleocontinent.

687 Finally, our contribution to unravelling the Minas-Bahia Orogenic Belt, in the context of
688 supercontinent cycles, suggests that a large continental landmass was amalgamated
689 during the Rhyacian (ca. 2.1-2.05 Ga). Together with Baltica, Amazonia and West
690 Africa, the São-Francisco Congo is surrounded by a broad Paleoproterozoic (pre-
691 Columbia) oceanic realm. After the orogeny, Orosirian bimodal magmatic successions
692 point to an extensional tectonic setting, hampering any direct connections with Columbia
693 before ca. 1.9 Ga.

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1201 **Figures Caption**

1202 **Figure 1:** Tectonic framework of the São Francisco Craton basement. a) South American
1203 Platform and Gondwana configuration adapted from D'Agrella-Filho & Cordani (2017).
1204 b) Archean blocks and Paleoproterozoic magmatic arcs of the São Francisco
1205 Palecontinent. Modified from Bruno et al., 2020; Bersan et al., 2020; Degler et al., 2018;
1206 Alkmin & Teixeira, 2017; Barbosa & Barbosa, 2017.

1207 **Figure 2:** a) Neoproterozoic compartmentation of the basement units associated to the
1208 São Francisco Craton; b) Geological map and compiled U-Pb data of the southern São
1209 Francisco Palecontinent reworked during the Brasiliano/Pan-African orogeny (Modified
1210 from Bruno et al., 2020; Alkmin & Teixeira 2017; Heilbron et al., 2017; Peres et al.,
1211 2004). Compiled U-Pb data from: Bruno et al., 2020; Cardoso et al., 2019; Pinheiro et al.,
1212 2019; Moreira et al., 2018; Teixeira et al., 2015; Seixas et al., 2013; Seixas et al., 2012;
1213 Corrêa Neto et al., 2012; Heilbron et al., 2010; Ávila et al., 2014; Noce et al., 2007;;
1214 Duarte et al., 2004; Quéméneur et al., 2003; Silva et al., 2002).

1215 **Figure 3:** Field photographs of representative lithologies and outcrops of the Piedade
1216 Block, Campos Gerais Complex and Mineiro Belt. a) Metric layer of amphibolite (68B)
1217 showing sharp contact with the country rock (68A - biotite gneiss) from the Piedade
1218 Block; b) Amphibolite (70A) from the Piedade Block; c) Megacrystic Biotite -Muscovite
1219 Gneiss (58A) from the Piedade Block; d) Hornblende-Biotite Gneiss (50) from the
1220 Piedade Block e) Biotite orthogneiss (48A) from the Campo Belo Complex; f) Biotite
1221 orthogneiss (48B) from the Campo Belo Complex; g) Granodioritic Hornblende-Biotite
1222 orthogneiss (51B) from the Mineiro Belt; i) Dioritic Hornblende-Biotite orthogneiss (42)
1223 from the Mineiro Belt.

1224 **Figure 4:** Photomicrographs from thin sections of the studied orthogneisses and
1225 metabasic rocks of the Piedade Block. a) Sample 70B; b) Sample 70A; c) Sample 70A
1226 under crossed nicols; d) Sample 58B; e) Sample 58A; f) Sample 58A.

1227 **Figure 5:** Photomicrographs from thin sections of the studied orthogneisses and
1228 metabasic rocks of the Campo Belo Complex. a) Sample 48A; b) Sample 48A; c) Sample
1229 48B; d) Sample 48B.

1230 **Figure 6:** Photomicrographs from thin sections of the studied orthogneisses of the
1231 Mineiro Belt. a) Sample 42; b) Sample 42; c) Sample 51B; d) Sample 51B.

1232 **Figure 7:** a) AFM diagram of Irvine and Baragar (1971) showing the intermediate and
1233 acid samples plotting in the calc-alkaline series and the basic rocks in the tholeiite series;
1234 b) TAS classificatory diagram (SiO_2 vs. $\text{Na}_2\text{O}+\text{K}_2\text{O}$) of Cox (1979).

1235 **Figure 8:** a) A/NK vs. A/CNK diagram (after Shand, 1943); b) SiO_2 vs. K_2O diagram
1236 (Peccerillo & Taylor, 1976); c) $\text{FeOt} / (\text{FeOt} + \text{MgO})$ vs. SiO_2 (Frost et al., 2001); d)
1237 Ternary classification diagram from Laurent et al. (2014). Vertices are: $2 \times \text{A/CNK}$
1238 (molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{K}_2\text{O} + \text{Na}_2\text{O})$ ratio); $\text{Na}_2\text{O}/\text{K}_2\text{O}$ and $2 \times (\text{FeO}_t + \text{MgO}) \times (\text{Sr} + \text{Ba})$
1239 wt.% (=FMSB).

1240 **Figure 9:** Left column (a) TTG Sample; c) Sanukitoid Group; e) Biotite- Two-Mica
1241 Group; g) Hybrid Granitoid Group, i) Alkaline Sample): Average chondrite-normalized
1242 REE patterns normalized after values from Boynton (1984); Right column (b) TTG
1243 Sample; d) Sanukitoid Group; f) Biotite- Two-Mica Group; h) Hybrid Granitoid Group,
1244 j) Alkaline Sample): Mantle-normalized multielement plots (McDonough and Sun,
1245 1995).

1246 **Figure 10:** a) Average chondrite-normalized REE patterns normalized after values from
1247 Boynton (1984); b) NMORB-normalized multielement plot (Sun & McDonough, 1989);
1248 c) Ternary Diagram Zr/4 vs. 2Nb vs. Y of Meschede (1986); d) Tectonic discriminant
1249 diagram Zr vs. Ti (Pearce, 1982).

1250 **Figure 11:** Concordia diagrams presenting zircon U-Pb results and Cathodoluminescence
1251 (CL) and Backscattered electrons (BSE) images of representative zircon grains for the
1252 rocks of the Piedade Block. Circles on the zircon grains indicate spot locations for U-Pb
1253 results. Paleoproterozoic and Archean data are shown in $^{207}\text{Pb}/^{206}\text{Pb}$ and Neoproterozoic
1254 data in $^{206}\text{Pb}/^{238}\text{U}$. Zircon codes refer to analytical ID in U-Pb data table in
1255 Supplementary Material C. Blue circles are indicative of inherited zircon grains. a) Two
1256 Discordia lines for Sample 50. Inset of Paleoproterozoic sub concordant rims; b)
1257 Concordia age for the Neoproterozoic metamorphic overprint of Sample 50; c) Concordia
1258 age for the Archean crystallization age of Sample 50; d) Sample 65; e) Sample 68A, inset
1259 of Concordia age for crystallization; f) Sample 70A, inset of Concordia age for
1260 crystallization; g) Sample 70B, inset of Concordia age for crystallization.

1261 **Figure 12:** Concordia diagrams presenting zircon U-Pb results and Cathodoluminescence
1262 (CL) and Backscattered electrons (BSE) images of representative zircon grains for the
1263 rocks of the Piedade Block. Circles on the zircon grains indicate spot locations for U-Pb
1264 results. Paleoproterozoic and Archean data are shown in $^{207}\text{Pb}/^{206}\text{Pb}$ and Neoproterozoic
1265 data in $^{206}\text{Pb}/^{238}\text{U}$. Zircon codes refer to analytical ID in U-Pb data table in
1266 Supplementary Material C. a) Sample 58A, inset of Paleoproterozoic Average Mean of
1267 the analytical results; b) Sample 58B.

1268 **Figure 13:** Concordia diagrams presenting zircon U-Pb results and Cathodoluminescence
1269 (CL) and Backscattered electrons (BSE) images of representative zircon grains for the
1270 rocks of the Mineiro Belt. Circles on the zircon grains indicate spot locations for U-Pb
1271 results. Paleoproterozoic and Archean data are shown in $^{207}\text{Pb}/^{206}\text{Pb}$ and Neoproterozoic
1272 data in $^{206}\text{Pb}/^{238}\text{U}$. Zircon codes refer to analytical ID in U-Pb data table in
1273 Supplementary Material C. Blue circles are indicative of inherited zircon grains. a)
1274 Sample 42; b) Average Mean of Sample 42 results; c) Sample 51B; d) Sample 52B.

1275 **Figure 14:** Concordia diagrams presenting zircon U-Pb results and Cathodoluminescence
1276 (CL) and Backscattered electrons (BSE) images of representative zircon grains for the
1277 rocks of the Campo Belo Complex. Circles on the zircon grains indicate spot locations
1278 for U-Pb results. Paleoproterozoic and Archean data are shown in $^{207}\text{Pb}/^{206}\text{Pb}$ and
1279 Neoproterozoic data in $^{206}\text{Pb}/^{238}\text{U}$. Zircon codes refer to analytical ID in U-Pb data table
1280 in Supplementary Material C. Blue circles are indicative of inherited zircon grains. a)
1281 Sample 48A, inset for concordia age; b) Sample 48B.

1282 **Figure 15:** Nd evolution vs time (crystallization ages) diagram. DM model from De
1283 Paolo, 1981. Different symbols for tectono-stratigraphic terranes and different colors for
1284 the lithochemical groups.

1285 **Figure 16:** Isotopic diagram for the Campo Belo Complex, Mineiro Belt and Piedade
1286 Block samples. ϵNd vs initial $^{87}\text{Sr}/^{86}\text{Sr}$ diagram (Sr and Nd isotopic systematics of the
1287 crust and mantle, horizontal grey band is the estimated ϵNd of the bulk silicate of Caro
1288 and Bourdon (2010); vertical grey band between dashed lines in their estimated bulk
1289 silicate Earth $^{87}\text{Sr}/^{86}\text{Sr}$)

1290 **Figure 17:** Binary plots of isotopic signatures. #mg vs. a) ϵNd (crystallization ages)/ b)
1291 $^{87}\text{Sr}/^{86}\text{Sr}$ (initial); c) Plot of incompatible element (Sr+Ba ppm) vs. FeO + MgO (wt.)

1292 (Laurent et. al., 2014); d) $\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{FeO} + \text{MgO} + \text{TiO}_2 - (\text{Na}_2\text{O} + \text{K}_2\text{O}) / (\text{FeO} +$
1293 $\text{MgO} + \text{TiO}_2)$ diagram (from PatiñoDouce, 1999).

1294 **Figure 18:** Table of synthesized results of this study (*) within the Minas segment of the
1295 São Francisco Palecontinent (Bruno et. al., 2020; Cutts, et. al., 2020; Araújo et. al.,
1296 2019b; Cutts et. al., 2019; Moreira, et. al., 2018; Degler et. al., 2018; Alkmim & Teixeira,
1297 et. al., 2017; eg. Heilbron et. al., 2017b, eg. Teixeira et. al., 2017a, eg. Albert et. al., 2016;
1298 Teixeira et. al., 2015, Aguilar Gil et. al., 2015; Farina et. al., 2015; Seixas et. al., 2013;
1299 Noce et. al., 2007; Vlach et. al., 2003; Silva et. al., 2002; Brueckner et. al., 2000; Machado
1300 et. al., 1992).

1301 **Figure 19:** Integrated tectonic evolution model for the Minas segment of the
1302 Paleoproterozoic Orogeny, southeast Brazil, as envisaged for the period between ca. 2.4
1303 and 1.9 Ga (Modified after Bruno et. al., 2020; Cutts et. al., 2020; Ávila et. al., 2014;
1304 Heilbron et. al., 2010; Noce et. al., 2007)

1305 **Figure 20: a)** Paleotectonic framework of the Sao Francisco Palecontinent at ca. 2.2 Ga,
1306 distinguished by individual Archean blocks and Paleoproterozoic accretionary orogenic
1307 belt and envisaged tectonic model for the Bahia segment (A-A') and the Minas segment
1308 (B-B') of the Minas Bahia Orogenic Belt (modified from Zincone et. al., 2020; Teixeira
1309 et. al., 2017a; Alkmim & Noce, 2006). UAC – Undivided Archean Complexes (e.g.
1310 Campo Belo, Belo Horizonte, Bonfim, Santa Bárbara); b) ϵNd vs. ϵSr diagram with
1311 distinct fields characterized by data from the Itabuna-Salvador-Curaçá Belt, Gavião,
1312 Serrinha and Jequié block (modified after Barbosa and Barbosa, 2017), Mantiqueira
1313 Complex (Bruno et. al., 2020); Piedade Block (Bruno et. al., 2020 and from this study);
1314 Mineiro Belt (Cardoso et. al., 2019; Teixeira et. al., 2015 and this study) and Juiz de Fora
1315 Complex (Araújo et. al., 2019b).

1316 **Figure 21: a)** Paleogeographic reconstruction at ca. 2.0 Ga (D'Agrella-Filho & Cordani,
1317 2017). Su – Superior Craton; S – Slave Craton; R – Rae cratonic fragment; H – Hearne
1318 cratonic fragment. KAR – Karelia Craton; K – Kola Craton; Pam – Proto-Amazonia; WA
1319 - West Africa; V-U – Volgo Uralia; AS – Sarmatia; CSF – Congo/São Francisco; KAL –
1320 Kalahari; RP – Rio de La Plata; BTS – Borborema Trans-Sahara. Dashed lines indicate
1321 later borders of Laurentia, Baltica and Amazonian craton; b) Paleogeographic
1322 reconstruction of Columbia Supercontinent at ca. 1.78 Ga (D'Agrella-Filho et. al., 2020).
1323 AC- Amazonia Craton; WA – West African Craton; BA – Baltica Craton; LA – Laurentia
1324 Craton; SIB – Siberia Craton; IN – India Craton; NC – North China Craton; LP – Rio de
1325 La Plata Craton; SF – São Francisco Craton; C – Congo Craton; NAU – North Australia
1326 Craton; EA – East Antarctica; SAU – South Australia Craton; WAU – Western Australia
1327 Craton.

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