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Protracted thermal evolution of the Neoproterozoic Araçuaí hot orogen (SE Brazil): consequences for rheology, strain distribution, and deformation analysis

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5 Alain Vauchez[1]*, Maria Helena B.M. Hollanda[2], Patrick Monié[1], Mathieu Mondou[1, 2],
6 Marcos Egydio-Silva[2]

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8 [1] Géosciences Montpellier, Université de Montpellier & CNRS, Place E. Bataillon, 34095 Montpellier Cedex 05, France
10 [2] Instituto de Geociências, Universidade de São Paulo, Rua do Lago, 562, 05508-080 São Paulo, Brazil

11 *Abstract*:
12
13 The Araçuaí-Ribeira belt formed during the amalgamation of West Gondwana in the
14 late Neoproterozoic. Its evolution encompasses a main tectono-metamorphic peak at 600-580
15 Ma and a minor one, associated with the final collision with the Western Congo at 540-530
16 Ma. This belt holds characteristics of a hot-orogen: high thermal gradient (>30 °C/km),
17 pervasive partial melting of the middle crust, emplacement of huge volume of magmas
18 resulting from partial melting of the lower crust and underlying mantle, and slow cooling after
19 the peak of temperature. We report 21 new amphibole, biotite and muscovite $^{40}Ar/^{39}Ar$ ages,
20 which complement data already published by our group. Altogether, these data support slow
21 cooling (3-4 °C/Myr) during several tens of million years after the peak temperature
22 (~800 °C at ~600 Ma), followed by faster cooling (>10 °C/Myr) after the final amalgamation
23 of Western Gondwana. We estimate that ~30 Myr were needed to heat the middle crust to the
24 peak of temperature and that anatectic and plutonic bodies remained in the magmatic state
25 over at least 40 Myr. This protracted thermal evolution likely had major effects on the
26 rheology of the middle crust and on the tectonic evolution of this orogen. For instance,
27 correlation of U-Pb zircon crystallization ages and $^{40}Ar/^{39}Ar$ biotite cooling ages in the
28 anatectic core of the orogen denotes a diachronic thermal evolution likely related to a 3D
29 deformation characterized by successive upwelling of anatectic components, along a channel
30 crosscutting pre-existent fabric (channel-flow type). This study also highlights that classical
31 structural analysis techniques relying on changes in pressure or temperature conditions to
32 identify the succession of deformation phases, cannot be used to decipher the tectonic
33 evolution of hot, slowly cooling orogenic belts, where temperature varies weakly over tens of
34 millipn years, allowing diachronic episodes of deformation to occur under almost similar
35 pressure and temperature conditions.

* Corresponding author - email address: alain.vauchez@umontpellier.fr

1) Slow cooling of the Neoproterozoic Araçuaí hot orogen

2) Middle crust partially molten during 40-45 Ma

3) Rheology of the crust and strain distribution

4) Thermo-tectonic evolution

1.   Introduction:

Many orogens are characterized by an abnormally high geotherm (≥30 °C/km) down to the middle/lower crust (e.g., Peucat et al., 1999; Möller et al., 2000; Collins, 2002; Morisset et al., 2009; Högdahl et al., 2012; Turlin et al., 2018). A common characteristic of these "large-hot" orogens (Beaumont et al., 2010) is to have cooled very slowly; estimated cooling rates are usually << 10 °C/Myr. Thus, after temperature rose up to the metamorphic peak, it remained high over millions of years. The effects of this peculiar thermal evolution on the mechanical behavior of the lithosphere and on petrologic and geochemical systems are still not fully evaluated (e.g., Jamieson et al., 2011). Such an evolution contrasts with those of "small-cold", "alpine-type" orogens (Beaumont et al., 2010), in which temperature variations during and following orogeny are much faster. Cooling rates are in most cases high (several tens of °C/Myr) and, sometimes, even higher, in particular when fast exhumation follows orogen building (e.g., Zeck et al., 1992; Janots et al., 2009). Rapid temperature variations result in marked changes in the rheology of rocks over short periods of time and successive stages of the orogenic evolution are usually recorded as superimposed deformations occurring at different pressure and temperature conditions. In contrast, in a hot-orogen, temperature decreases very slowly; diachronic deformation episodes may occur under similar temperature conditions and, hence, result in similar rock microstructures and mineralogical assemblages. Consequently, it might be difficult to recognize contemporaneous deformations with contrasted kinematics from diachronous deformations separated by 10-15 Myr, for instance.

The processes that result in the formation of "hot orogens" are still poorly understood. Radiogenic heating of a highly thickened continental crust is regarded as the most efficient one (England and Thompson, 1986; Huerta et al., 1998; Vanderhaeghe et al., 2003; Clark et al., 2011; Clark et al., 2015). However, mechanical heating due to shearing or intrusion of multiple batches of mantle-derived magmas may also contribute to generate abnormally high temperatures in the middle to lower crust (Michaut and Jaupart, 2007; Nabelek et al., 2010).

69   At larger scales, it was suggested that slab-retreat might trigger upwelling of asthenospheric
70   mantle, which may contribute to heating of the overlying plate (e.g., Turlin et al., 2018 and
71   references therein).

72   Three processes may combine to lower the cooling rates in a collision zone:

73   - High temperature in the middle crust decreases its thermal conductivity and hence its
74   capacity to evacuate heat. A "heat reservoir" may thus form in the middle/lower crust,
75   allowing for maintenance of an abnormal thermal gradient over long periods of time (Michaut
76   and Jaupart, 2007; Whittington et al., 2009; Nabelek et al., 2010).

77   - High temperature in the middle to lower crust may trigger widespread partial melting
78   and emplacement of large volumes of magma in the middle crust. Upon cooling,
79   crystallization of these magmas releases latent heat of crystallization that compensates, at
80   least partially, conductive cooling (Michaut and Jaupart, 2011).

81   - The viscosity of a hot and partially molten middle crust is too low to sustain high
82   topography. When the weight due to orogenic topography exceeds the strength of the hot
83   middle crust, it triggers a gravity-driven flow in the low-viscosity middle to lower crust,
84   which reduces the topography and forms plateaus (e.g., Jamieson et al., 2011 and references
85   therein). This process partly compensates orogenic thickening of the crust through frontal
86   or/and lateral extrusion, limits erosion rates, and, consequently, hinders tectonic exhumation
87   and associated cooling.

88   In Southeast Brazil, the Araçuaí belt (Figure 1) displays evidence of high temperature
89   contractional deformation distributed almost homogeneously all over its internal domain on
90   which focus this study, which is ~200 km large and >500 km long. Pressure-temperature
91   estimates consistently point to synkinematic conditions between ~750 °C/500-600 MPa and
92   ≥800 °C/600-700 MPa (Petitgirard et al., 2009; Cavalcante et al., 2014; Moraes et al., 2015;
93   Richter et al., 2016), supporting a temperature gradient of 30-35 °C/km down to 25 km depth
94   at least. Cavalcante et al. (2018) have also substantiated that anatexis was active in the core of
95   the belt during ~30 Myr. However, cooling rates were poorly constrained. Only two
96   amphibole and three biotite ^{40}Ar/^{39}Ar ages from four samples, suggesting low cooling rate,
97   have been previously published (Petitgirard et al., 2009). Here, we report 21 new ^{40}Ar/^{39}Ar
98   ages in amphibole, muscovite and biotite from 16 samples representative of the different
99   tectonic units of the Araçuaí "hot orogen". These new data, combined with previously
100   published U-Pb zircon ages, thermo-barometric data, and structural data from field
101   measurements and Anisotropy of Magnetic Susceptibility (AMS) mapping (Vauchez et al.,
102   2007; Petitgirard et al., 2009; Mondou et al., 2012; Cavalcante et al., 2013, 2014, 2018)

103 constrain a protracted tectono-thermal evolution for this orogenic belt and confirm that
104 temperatures in the middle crust remained high over tens of Myr. Based on these results, we
105 discuss the implications of such an evolution for the strain repartition through space and time
106 in this orogenic segment and the difficulties in applying classical tectonic analysis to hot
107 orogens.

108 ## *2.   Structure of the Araçuaí belt and previous thermo-chronological data*

109   The Araçuaí belt (Figure 1) is a segment of a large orogenic system formed between the
110 São Francisco, Salvador, and Congo cratons during the convergence between the South
111 America and Africa proto-continents, which resulted in the formation of Western Gondwana
112 (Hasui et al., 1975; Almeida, 1977; Trompette et al., 1992; Pedrosa-Soares et al., 2007). The
113 bulk orogenic system comprises the Araçuaí-Ribeira and Cabo Frio-West Congo belts (Monié
114 et al., 2012; Egydio-Silva et al., 2018). The convergence lasted at least one hundred millions
115 years and diachronic orogenic climaxes have been substantiated for the Araçuaí-Ribeira
116 orogen (600-570 Ma; e.g., Nalini et al., 2000b; Mondou et al., 2012; Gonçalves et al., 2014;
117 Cavalcante et al., 2018) and for the West Congo belt (540-530 Ma; e.g., Monié et al. 2012).
118 The latter event involved thrusting of the "Cabo Frio Tectonic Domain" upon the eastern
119 Ribeira belt in the Rio de Janeiro province (Schmitt et al., 2004).

120   On the South American side, the Araçuaí-Ribeira belt molds the eastern to southeastern
121 boundary of the São Francisco craton (Vauchez et al., 1994; Egydio-Silva et al., 2018; Figure
122 1). The Ribeira belt forms the southern segment of the orogen. It is characterized by a
123 transpressional tectonic regime, which combined orogen-parallel transcurrent motion with
124 orogen-normal thrusting (Vauchez et al., 1994; Egydio-Silva et al., 2002). Seismic anisotropy
125 measurements (Heintz et al., 2003; Assumpção et al., 2006) suggest that orogen-parallel
126 shearing involved the whole lithosphere (Vauchez et al., 2012). Northward, the connection
127 with the Araçuaí belt is characterized by a curvature of the orogenic fabric from NE-SW to N-
128 S (Figure 1) and a progressive transition in strain regime to dominant westward thrusting of
129 allochtonous units onto the São-Francisco craton in the Araçuaí belt (Vauchez et al., 1994;
130 Oliveira et al., 2000; Egydio-Silva et al., 2005; Vauchez et al., 2007; Egydio-Silva et al.,
131 2018). These longitudinal variations in orogenic fabric and kinematics are likely related to the
132 termination of the São Francisco craton (Vauchez et al. 1994). Finally, the northern
133 termination of the Araçuaí belt is characterized by dominant northward thrusting of the
134 orogenic units onto the Salvador craton (e.g., Uhlein et al., 1998).

135   The central Araçuaí belt (Figure 2), where all samples used in this study were collected

136   (location of selected samples in Supplementary Material #1), is composed by four

137   allochthonous domains docked against the São Francisco craton (Oliveira et al., 2000;

138   Vauchez et al., 2007). These domains are from West to East:

139   - The *para-autochthonous metasedimentary cover* of the craton. Several U-Pb and

140    $^{207}Pb/^{206}Pb$ ages in zircons > 2 Ga obtained in this domain (e.g., Noce et al., 2000; Silva et al.,

141   2002) support that these rocks belong to the craton. Approaching the contact with the

142   allochthonous units, these rocks display a tectonic fabric parallel to the mylonitic fabric of the

143   westernmost allochtonous unit, suggesting that they have been both strongly deformed under

144   high temperature conditions during the Neoproterozoic orogeny. The foliation gently dips

145   eastward and the lineation trends dominantly EW (Figure 2), metapelites contain acicular

146   sillimanite, garnet, and K-feldspar and coarse-grained quartzites contain crystals of sillimanite

147   frequently included in overgrown quartz grains. Shear criteria, although infrequent,

148   consistently point to a top-to-west sense of shear.

149   - The *Western Mylonitic Unit (WMU,* Figure 2*)* is thrust westward onto the para-

150   autochthonous cover of the craton. This unit comprises mylonites derived from

151   metasediments, which display a foliation that gently dips eastward and bears an EW-trending

152   lineation formed under HT-LP conditions (Petitgirard et al., 2009). The similarity of

153   lithologies and deformation conditions with those of the para-autochthonous cratonic

154   formations hinders a precise definition of the basal contact of the WMU. The main

155   differences between the two units are the presence in the WMU of numerous top-to-west

156   shear criteria (Figure 3a-d), the absence of quartzites, the alternation of metapelites with

157   amphibolites, and the presence of large volumes (up to 40% locally) of anatectic

158   leucogranites in decimetric veins parallel to the mylonitic foliation (Figure 3b-d). There is no

159   evidence of strain localization at the base of the unit; strain is rather homogeneously

160   distributed across the WMU. Indeed, the microstructure of the HT mylonites is similar across

161   the whole unit, varying only as a function of mineralogy. Petitgirard et al. (2009) have

162   determined metamorphic conditions for the WMU mylonites based on garnet-biotite

163   exchange thermobarometers: core-core analyses (or biotite included in garnet) consistently

164   yield 760-780°C (±30°C) and rim-rim analyses, 730-760°C (±30°C) and pressure in the range

165   500-600 MPa. A few ages have been published for this unit. Using the Pb-evaporation

166   technique Noce et al. (2000) have dated two magmatic bodies (Brasilândia and Guarataia

167   plutons) from this unit (Oliveira et al., 2000), which yielded $^{207}Pb/^{206}Pb$ zircon ages of 595±3

168   and 574±2 Ma respectively, later on confirmed by Tadeschi et al. (2016). Petitgirard et al.

169   (2009) have dated two synkinematic leucocratic granites intercalated in the mylonites: one

170   yielded U-Pb ages of 577 ± 9 Ma for zircon and 572 ± 3 Ma for monazite and the other

171   yielded a zircon U-Pb age of 578 ± 3 Ma (Figure 4; Table 1). These authors have reported

172   $^{40}Ar/^{39}Ar$ ages of ca. 501 and ca. 495 Ma for amphiboles and ca. 474 and ca. 468 Ma for

173   biotites from the WMU and the associated Ibituruna syenite, respectively; these preliminary

174   data are integrated in our database (Table 1, Figure 7) and further commented (see

175   Supplementary Material #2 for amphibole AR-414 from the Ibituruna syenite).

176   - The *Central Plutonic Unit* (CPU; Figure 2) comprises huge volumes of synkinematic

177   garnet-bearing granodiorites and tonalites emplaced in metasediments (e.g., Mondou et al.,

178   2012; Narduzzi et al., 2017). These plutonic bodies (the "*São Vitor*" and "*Galileia*" batholiths)

179   display a consistent magmatic fabric (Figure 5) that evolves eastwards from a gently

180   eastward-dipping foliation bearing a dominantly EW-trending lineation to a steeply-dipping

181   to subvertical foliation bearing a lineation that switches from subvertical to subhorizontal

182   over intervals of a few tens of meters. In the easternmost part of the CPU, Mondou et al.

183   (2012) have substantiated the presence of a corridor in which the magmatic foliation is sub-

184   horizontal and bears a strong orogen-parallel, NS-trending magmatic lineation. In several

185   places, the granodiorites and tonalites contain elongated dioritic to gabbroic lenses (Vauchez

186   et al., 2007; Gonçalves et al., 2014; Narduzzi et al., 2017) implying in a contribution of

187   mantle-derived magmas to the formation of the batholiths (Figure 5c).

188   Metamorphic conditions in the metasedimentary country rocks of the CPU are still

189   poorly constrained due to poor outcropping conditions, but they tend to decrease eastwards as

190   supported by the occurrence of staurolite and, locally, of muscovite in garnet-sillimanite

191   bearing metapelites of the easternmost part of the CPU. These minerals, indicative of

192   temperatures ≤ 650 °C (Nalini et al., 2015), are only present in the easternmost CPU.

193   Several ages have been obtained for CPU granitoids close to the contact with the WMU

194   (Figure 4). Zircons from a tonalite sill yielded a U-Pb age of 580 ± 8 Ma (Petitgirard et al.,

195   2009). Two leucocratic veins interlayered with mylonitic metasediments yielded zircon U-Pb

196   ages of 587 ± 5 and 579 ± 8 Ma (Mondou et al., 2012). Zircons from a sample of tonalite

197   collected ~80 km south of the previous locations and ascribed to the CPU western boundary

198   yielded an age of 597 ± 4 Ma (Gonçalves et al., 2014).

199   The São Vitor tonalite was first dated at 576 ± 5 and 569 ± 9 Ma (Noce et al., 2000;

200   ^{207}Pb/^{206}Pb ages on zircons) then, Mondou et al. (2012) obtained one U-Pb and two ^{207}Pb/^{206}Pb

201   ages on zircons of 583 ± 4, 583 ± 4, 582 ± 6, and 585 ± 4 Ma, respectively.

202   The Galileia tonalitic/granodioritic batholith was dated at 594 ± 6 Ma (Nalini et al.,

203   2000b; ^{206}Pb/^{238}U on zircons). Subsequently, Mondou et al. (2012) have dated samples from

204   domains of this batholith presenting contrasted fabrics. U-Pb zircon ages from these domains

205   with different kinematics are fairly consistent between 579 and 583 Ma.

206   The muscovite-bearing Palmital peraluminous granite was dated at 582±2 Ma (Nalini et

207   al., 2000a) and the cordierite-garnet-biotite-bearing Wolf leucogranite at 582±5 Ma (Noce et

208   al., 2000) and 585 ± 4 Ma (U-Pb on zircons; Mondou et al., 2012). These two granites

209   (Figures 2, 4) likely result from melting of aluminous metapelites contemporaneous to the

210   melting of the lower crust and underlying mantle that formed the tonalitic/granodioritic

211   magmas.

212   - *Intermediate, discontinuous kinzigitic unit* (Figures 2 and 6). The steeply dipping,

213   medium temperature, staurolite-bearing metasediments that crop out in the easternmost part

214   of the CPU are separated from the gently dipping Eastern Anatectic Unit (EAU; see below)

215   by a discontinuous unit of partially molten kinzigites with a steeply-dipping foliation

216   (Oliveira et al., 2000; Mondou et al., 2012). In contrast with metasediments of the

217   easternmost CPU, these kinzigites contain biotite, garnet, cordierite, sillimanite, but no

218   staurolite or muscovite. The tectonic contact between these kinzigites and the CPU is

219   therefore characterized by a metamorphic gap, despite the similar subvertical fabric recorded

220   by the two units. This suggests that the intermediate kinzigitic unit, initially equilibrated at

221   deeper levels, has been thrust upon the CPU and that, subsequently, the tectonic contact and

222   associated fabric have been verticalized.

223   - The *Eastern Anatectic Unit* (EAU, also known as "*Carlos Chagas leucogranite*";

224   Figures 2 and 6) is more than 300 km long and 50 to 100 km wide. It comprises peraluminous

225   metatexites, diatexites, and leucogranites containing quartz, K-feldspar, garnet, biotite,

226   sillimanite, ± rutile, ± cordierite (e.g., Cavalcante et al., 2013). Some garnet-bearing

227   leucogranites are almost free of biotite (Figure 6c). Rather than a batholith, the EAU

228   represents the anatectic core of the Araçuaí belt, which was intruded by several late orogenic

229   charnockitic plutons (~520 Ma). This unit is characterized by a widespread, penetrative

230   magmatic fabric marked by a magmatic foliation that dominantly dips gently to moderately in

231   various directions due to open folds that affected an initial low-angle foliation (Cavalcante et
232   al., 2013). Detailed mapping of magnetic foliations and lineations across the EAU revealed a
233   complex flow field, characterized by a western domain where gently dipping foliations bear
234   lineations that progressively rotate from SW in the south to NW in the north (Figure 2) and by
235   an eastern domain where subvertical NS-trending foliations bear sub-horizontal, orogen-
236   parallel lineations (Cavalcante et al., 2013). The eastern boundary of the anatectic unit is
237   defined by a progressive transition to migmatitic kinzigites (Figure 6d) in which are
238   intercalated layers of orthopyroxene-bearing granulites.

239   Cavalcante et al. (2014, 2018) using the Ti-in-quartz (TitaniQ), the Zr-in-rutile
240   geothermometers in anatexites and Fe-Mg exchange geothermobarometers in neighbouring
241   kinzigites obtained temperatures of 780-800 °C and pressures in the range of 650-700 MPa;
242   these P-T conditions are interpreted as those under which crystallization of anatexites began
243   during cooling. These thermobarometric estimates are in good agreement with those obtained
244   for the migmatitic kinzigites by Munhá et al. (2005) using Fe-Mg exchange
245   geothermobarometers and, more recently, by Richter et al. (2016) from phase equilibrium
246   modeling. Such P-T conditions are consistent with the presence of cordierite (Harley et al.,
247   2002; Pattison et al., 2003; Harley, 2004) in the studied rocks and also of rutile that,
248   according to Le Breton and Thompson (1988), crystallizes at T ≥760 °C. Evidence of
249   destabilization of restitic biotites in several samples further supports temperatures >800 °C
250   (Le Breton and Thompson, 1988; Vielzeuf and Holloway, 1988; Pattison et al., 2003). As
251   highlighted by Cavalcante et al. (2014), such high temperatures probably allowed for melt
252   fractions >30%, which may have drastically reduced the viscosity of the anatectic middle
253   crust (Rosenberg and Handy, 2005; Vanderhaeghe, 2009). Moreover, the contrast in peak
254   equilibrium temperatures supports a major tectonic contact between the central (CPU) and
255   eastern (EAU) units.

256   Vauchez et al. (2007) reported a SHRIMP U-Pb zircon age of 575 ± 3 Ma for a diatexite
257   from the EAU. Additional U-Pb SHRIMP zircon ages from anatectic granite, diatexites, and
258   metatexites substantiate a broad variability in U-Pb zircon ages across the EAU (Cavalcante
259   et al., 2018). Several diatexites and anatectic granites yielded ages in the range 597-593 Ma,
260   whereas the remaining anatectic samples were dated between 589 and 572 Ma. Ages are on
261   average slightly older in the northwest and younger in the southeast, but the trend is not clear-
262   cut (Figure 5). Richter et al. (2016) reported a similar age range (595-570 Ma) for samples
263   from the easternmost "*Nova Venecia*" kinzigitic unit of the Araçuaí belt. These data indicate

264   that anatexis was already underway in the middle crust (~25 km) of the Araçuaí belt by

265   ~600 Ma and that it remained active until at least 570 Ma (Cavalcante et al., 2018).

3. $^{40}Ar/^{39}Ar$ ages

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267   Ten samples were collected across the CPU: two synkinematic tonalites from the

268   contact between this domain and the WMU, five synkinematic granodiorites/tonalites from

269   the São Vitor and Galiléia complexes showing a well developed magmatic fabric, one

270   synkinematic leucogranite (Wolf granite), one muscovite-bearing peraluminous granite from

271   the Palmital plutonic body, and one orthogneiss from the country rock (Figure 7). The U-Pb

272   zircon ages of six of the selected plutonic rocks are close to 580 Ma (Mondou et al., 2012).

273   Biotite was separated and analyzed for the 10 samples; amphibole was analyzed in four

274   tonalites, and muscovite in the peraluminous Palmital granite (see Table 1).

275   Six samples from the EAU were selected among the 12 recently dated by Cavalcante et

276   al. (2018): four anatectic granites, one diatexite and one metatexite (Figure 4, Table 1). Their

277   U-Pb zircon ages range between 597 and 572 Ma. All selected samples display a well-

278   developed magmatic foliation and are free of solid-state deformation (Cavalcante et al., 2013).

279   Only biotite was dated since amphibole was not found in anatexites.

280   The location of the samples for which $^{40}Ar/^{39}Ar$ ages were obtained in this study is

281   shown on Figure 7, which also integrates $^{40}Ar/^{39}Ar$ ages obtained previously (Table 1, 2

282   amphibole and 3 biotite ages from Petitgirard et al., 2009). Samples from the CPU were

283   analyzed at Geosciences Montpellier (University of Montpellier, France) and those from the

284   EAU at the Geochronological Research Center of the University of São Paulo (Brazil).

285    $^{40}Ar/^{39}Ar$ diagrams are shown on Figures 9 and 10. Full analytical data are available in the

286   Supplementary Material #3 and #4.

287   In the CPU, $^{40}Ar/^{39}Ar$ cooling ages on amphibole, muscovite, and biotite show a larger

288   dispersion than the U-Pb ages on zircon; they range from about 550 to 470 Ma (Figure 7).

289   This dispersion reflects both the difference of closure temperatures between amphibole and

290   micas and the variability from sample to sample. Amphiboles from tonalite/granodiorite

291   samples provide nearly undisturbed age spectra with plateau ages ranging from 549.2 ± 5.6

292   Ma to 500±5 Ma (Figure 8). Biotites from the São Vitor and Galileia tonalites/granodiorites

293   yielded $^{40}Ar/^{39}Ar$ ages between 489.5 ± 5.4 Ma and 474.5 ± 4.4 Ma; those of the Wolf

294   leucogranite 491.9 ± 5.8 Ma and those from the Palmital granite and one orthogneiss sample

295   overlap at 484.1 ± 4.7 Ma. Muscovite from the Palmital granite yield plateau age of 487.6 ±

296   4.5 Ma for more than 90% of the released ^{39}Ar (Figure 8). Two samples of tonalite collected

close to the contact between the WMU and the CPU yield older cooling ages for amphibole (549.2 ± 5.6 and 541.6 ± 7.7 Ma) and biotite (532.2 ± 5.4 and 507.5 ±5.3 Ma). Although these ages are reliable from the analytical point of view, their geological significance remains unclear since for neighboring samples collected on both sides of this contact amphibole and biotite yield younger cooling ages in the range 510-470 Ma.

In the EAU, biotite ages span from 485.8 ± 4.4 Ma to 458.5 ± 4.2 Ma with the oldest age obtained for the northernmost sample and the youngest one for the southernmost sample (Figure 9; Supplementary Material #4). This distribution is correlated with the variation in U-Pb zircon ages across this domain.

4. *Discussion*

4.1. Origin and duration of the heating episode

Considering the current crustal thickness of ~40 km (Assumpção et al., 2013) and the paleopressures of 0.6-0.7 GPa recorded by rocks currently exposed in the central Araçuaí orogen (Petitgirard et al., 2009; Cavalcante et al., 2014; Richter et al., 2016), the orogenic crustal thickness was likely >60 km. Temperatures of 740 to ≥800 °C were recorded at 20-25 km depth (Petitgirard et al., 2009; Cavalcante et al., 2014; Richter et al., 2016) suggesting an average thermal gradient >30 °C/km within the upper and middle crust (likely lower in the lower crust). Throughout the orogen, the magmatic bodies are either embedded in metapelitic rocks or formed by local anatexis of metapelitic rocks. This supports that the orogenic middle crust of the Araçuaí belt, down to 25 km depth at least, was mostly composed by metapelites with some amphibolite and quartzite intercalated layers. Although the content in heat-producing elements of these metapelites is not known, such a middle crust likely had a rather high radiogenic heat production, which probably contributed significantly to heating (Clark et al., 2011; Clark et al., 2015). In response to heating, it underwent prograde metamorphic reactions that produced free water favoring extensive partial melting (Weinberg and Hasalová, 2015).

The presence of diorite/gabbros mingled within the granodioritic/tonalitic bodies of the CPU (Figure 5c) indicates that partial melting also affected the underlying mantle (Mondou et al., 2012; Gonçalves et al., 2014). Heating in the lithospheric mantle may be due to enrichment in heat-producing elements resulting from metasomatic reactions (Neves et al., 2008) during pre-orogenic rifting of the lithosphere along the eastern border of the São Francisco craton (Cunningham et al., 1996; Uhlein et al., 2007).

329   Available U-Pb ages on zircons are remarkably consistent across the various geological
330 units that shape the central part of the Araçuaí orogen. Similar ages have been reported in the
331 northern (Siga Jr et al., 1987) and southern portions of the belt, including its connection with
332 the Ribeira transpressional belt (Silva et al., 2005; Bento dos Santos et al., 2010). Ages
333 obtained in the EAU (Cavalcante et al., 2018) and in the kinzigites that bound it eastwards
334 (Richter et al., 2016; Melo et al., 2017) support that by ~600 Ma the middle/lower crust (0.6-
335 0.7 GPa) of the Araçuaí orogen had reached temperatures ~800 °C and was already partially
336 molten, producing garnet-rich aluminous magmas. Following this metamorphic peak, huge
337 volumes of synkinematic granodioritic/tonalitic magmas were emplaced in the CPU. They
338 systematically display a magmatic fabric that, although variable through the whole CPU, is
339 always consistent with the fabric of the country rock, and their ages are in the range 585-575
340 Ma independently of the orientation of their magmatic fabric (Mondou et al., 2012). In this
341 plutonic complex, only a few ages older than 600 Ma have been obtained (Silva et al., 2005).

342   The data support that contractional tectonics and crustal thickening started significantly
343 earlier than 600 Ma. A rough estimate may be obtained based on data from the EAU
344 considering heating rates in the range of 5-20 °C/Myr (Harris et al., 2000 and references
345 therein) or 8-15 °C/Myr, as recently deduced from allanite and monazite dating in the central
346 Alps by Janots et al. (2009). Assuming an initial temperature of ~400 °C at 20-25 km depth
347 (McKenzie et al., 2005), heating up to 800 °C would have required between 27 and 50 Myr.
348 This would imply that deformation producing thickening and prograde metamorphism in the
349 Araçuaí belt started before 630 Myr ago. Such a delay between the initiation of the orogeny
350 and the occurrence of widespread partial melting is in agreement with estimates from models
351 for the southern India developed by Clark et al. (2015). It is also consistent with the time
352 intervals (20-60 Myr) between the initiation of collision and melting of the middle/lower
353 crust under peak metamorphism conditions, estimated for the southern Tibet and the High
354 Himalayas or the Canadian Cordillera (e.g., Vanderhaeghe and Teyssier, 2001b and
355 references therein; Zhang et al., 2004; Guo and Wilson, 2012).

356   The granodioritic/tonalitic bodies of the CPU yield consistent ages in the range 590-570
357 Ma. These plutonic units display a magmatic fabric parallel to the solid-state fabric of their
358 country rocks, of the mylonites in the WMU, and of the intermediate kinzigitic unit between
359 the CPU and EAU. This supports synkinematic emplacement of huge volumes of
360 granodioritic and tonalitic magmas in the middle crust (Vauchez et al., 2007; Mondou et al.,
361 2012) during the temperature climax, the magmas likely resulting from partial melting of the

362 lower crust, with minor contribution from mantle melting, as evidenced from mixing with
363 dioritic and gabbroic magmas (Gonçalves et al., 2014).

364   Younger ages obtained on several late- and post-collisional granitic bodies (Silva et al.,
365 2002, 2005; Mondou et al., 2012) highlight that magmatic pulses occurred over a very long
366 period of time until, at least, 500 Ma. Among the late magmatic bodies, the "Ibituruna"
367 Syenite in the CPU (U-Pb zircon age of 530 Ma; Petitgirard et al., 2009) and the "Padre
368 Paraíso" charnockitic suite in the EAU (U-Pb zircon ages of ~520 Ma; Noce et al., 2000;
369 Mondou et al., 2012) still have a poorly understood origin. Their compositions imply
370 nevertheless that rather high temperature conditions subsisted in the lower crust and
371 underlying mantle up to these late stages of the evolution of the Araçuaí belt. It is worth
372 noticing that the Ibituruna syenite (~530 Ma) has a magmatic fabric (foliation and lineation)
373 parallel to the solid-state fabric in the surrounding HT mylonites (Petitgirard et al., 2009),
374 whereas the charnockites crosscut the tectonic fabric of the Araçuaí belt (e.g., Oliveira et al.,
375 2000; Mondou et al., 2012). Interestingly, these late magmatic events in the Araçuaí belt are
376 coeval with the main tectono-metamorphic event in the West Congo belt in Angola (Monié et
377 al., 2012) and in the "*Cabo Frio Tectonic Domain*" (Schmitt et al., 2004). They might,
378 therefore, record limited reactivation of the Araçuaí belt in response to the formation of the
379 "Congo-Cabo Frio" belt during the final collision between the African and South American
380 proto-continents at 540-530 Ma.

381   **4.2. Cooling rates**

382   In this section, we will only consider U-Pb and ^{40}Ar/^{39}Ar ages obtained by our group
383 (Table 1; Vauchez et al., 2007: Petitgirard et al, 2009; Mondou et al., 2012; Cavalcante et al.,
384 2018; this work), for which the location of samples, the definition of the rock type and, in
385 case of magmatic rocks, the synkinematic or post-kinematic emplacement and absence of
386 solid-state reworking are well known. This choice does not alter the results, since there is a
387 good agreement between the ages obtained in different studies for rocks from the Araçuaí belt.

388   Since our goal was to estimate the cooling temperature at the regional scale, to define
389 the minimum starting temperature of the cooling path we have used the equilibrium
390 temperatures estimated for the metamorphic country rocks (~750 °C; Petitgirard et al., 2009)
391 and the U-Pb zircon ages of synkinematic magmas in the WMU (577-578 Ma; Petitgirard et
392 al., 2009). For the anatexites of the EAU, we have considered U-Pb zircon ages obtained
393 across the domain (597-570 Ma; Cavalcante et al., 2018) and the maximum temperatures
394 estimated for the anatectic rocks using the Ti-in-Quartz (~800 °C, Cavalcante et al., 2014) and

Zr-in-Rutile thermometers (790 °C, Cavalcante et al., 2018). These temperatures are not necessarily representative of the metamorphic peak conditions, since the magmas were already crystallizing and thus cooling, but they rather provide a robust value for the starting temperatures of the cooling path. Finally, for the youngest anatexites of the EAU (570-578 Ma), we have considered that their Ti-in-Zircon temperatures (720-740 °C; Cavalcante et al., 2018) represent the minimum value of the equilibrium temperature at this time.

Figure 10 summarizes all ages obtained by our group for the Araçuaí belt. Zircon U-Pb ages from synkinematic leucogranites in mylonites from the WMU, granodiorites/tonalites from the CPU, and anatexites from the EAU spread from ~600 Ma to ~570 Ma with a clear concentration around 580 Ma (19/29 ages between 575-585 Ma, 8/29 between 585 and 600 Ma), supporting that the metamorphic peak, partial melting, and tonalitic/granodioritic magmatism occurred during this period of time. To estimate a mean cooling rate, we use the median value of 584 Ma ($\sigma = 7$) calculated based on 27 zircon ages (excluding the late Ibituruna syenite AR-414 and the orthogneiss AR-722). Temperature and pressure estimates support that, by ~600-580 Ma, metamorphic temperature at 20-25 km depth was in the range of 750 °C (WPU, Petitgirard et al., 2009) to ≥800 °C (EAU; Cavalcante et al., 2014). These values crudely define the minimum starting temperature of the cooling path (Figure 10).

Almost all dated minerals are of magmatic origin, except the recrystallized biotite extracted from orthogneiss AR-722, which has an U-Pb zircon age of 2103 ± 11 Ma (Mondou et al., 2012), but yielded similar $^{40}Ar/^{39}Ar$ age as the magmatic biotites. $^{40}Ar/^{39}Ar$ ages on amphibole and micas record, therefore, cooling through the respective closure temperatures. A first estimate of the cooling rate was obtained considering closure temperatures of 550 ± 30 °C for amphibole (Dahl, 1996), 400 ± 20 °C for muscovite (Harrison et al., 2009), and 325 ± 20 °C for biotite (Harrison et al., 1985). However, for very low cooling rates, the isotopic re-equilibration may continue at temperatures lower than the theoretical closure temperature (Dodson, 1973). We have thus recalculated the closure temperatures for amphibole, muscovite and biotite taking into account the estimated cooling rate (5 °C/Myr for amphibole and 15 °C/Myr for biotite and muscovite) and the grain size (Brandon et al., 1998; McDougall and Harrison, 1999; Braun et al., 2006; Reiners and Brandon, 2006). Depending on grain size, estimated closure temperatures range from 485 to 526 °C for amphibole, 288 to 338 °C for biotite and 416±4 °C for the single analyzed muscovite (Table 1).

Cooling rates inferred for various scenarios are shown on Figure 10:
- The first one (curve 1) is calculated using: 1) average ages computed considering all samples (except the syenite AR-414 and the orthogneiss AR-722 as stated

429   above), i.e., 584 Ma for U-Pb zircon age, 510 Ma and 487 Ma for ^{40}Ar/^{39}Ar

430   amphibole and biotite ages respectively, 2) the average closure temperatures of

431   amphibole (508 °C) and biotite (316 °C). This results in cooling rates of

432   3.9 °C/Myr with an initial temperature (T_i) of 800 °C (curve 1) between 584 and

433   510 Ma (3.3 °C/Myr with T_i = 750 °C) and of 8.4 °C/Myr between 510 and 487 Ma.

434   - For the second scenario (curve 2), we excluded the two samples that yielded

435   significantly older amphibole and biotite ^{40}Ar/^{39}Ar ages, and used average closure

436   temperatures of 504 °C for amphibole and of 319 °C for biotite. The obtained

437   cooling rates are of 3.5 °C/Myr between 584 and 499 Ma with T_i = 800 °C

438   (2.9 °C/Myr with T_i = 750 °C) and of 9.7 °C/Myr between 499 and 480 Ma.

439   - Even considering only the oldest ages obtained for amphibole (average: 546 Ma)

440   and biotite (average: 520 Ma), cooling rates still remain low (curve 3): 7.6 °C/Myr

441   between 583 and 546 Ma with T_i = 800 °C (6.3 °C/Myr with T_i = 750 °C). Cooling

442   rates below the closure temperature of amphibole range between 16.1 °C/Myr (with

443   AR-590 biotite age of 532 Ma) and 5.8 °C/Myr (with AR-1057 biotite age of 508

444   Ma) and 8.7 °C/Myr for the average age of biotites (520 Ma).

445   Local cooling rates for the EAU can be estimated using biotite ^{40}Ar/^{39}Ar ages of the six

446 samples for which Cavalcante et al. (2018) obtained U-Pb zircon ages between 597 and 572

447 Ma (Table 1; Figures 5, 9). Biotite from these samples yield ^{40}Ar/^{39}Ar ages from 486 to 459

448 Ma and closure temperatures from 307 to 338 °C. Considering the oldest anatectic

449 leucogranites of the EAU (AR-1133; curve 4 on Figure 10), which has U-Pb zircon age of

450 597 ± 3 Ma and ^{40}Ar/^{39}Ar biotite age of 486±4, the TitaniQ temperature for this unit of

451 ~800 °C (Cavalcante et al., 2014) as initial temperature, and 307 °C as the closure

452 temperature of biotite, we obtain an average cooling rate of ~4.4 °C/Myr during the ≥110 Myr

453 that followed the crystallization of zircons. Sample AR-1349, for which an equilibrium

454 temperature of 780-790 °C was estimated using Zr-in-rutile data (Cavalcante et al., 2018),

455 yields U-Pb zircon age of 583 Ma and ^{40}Ar/^{39}Ar biotite age of 476 Ma (closure T: 323 °C),

456 pointing to an average cooling rate of ~4.3 °C/Myr. For the youngest anatectic sample (AR-

457 1370), the ^{40}Ar/^{39}Ar biotite age is 469 Ma with a closure temperature of ~ 316 °C, the U-Pb

458 zircon age is 572 Ma, and a minimum ambient temperature of ~720 °C may be assumed from

459 Ti-in-zircon data (Cavalcante et al., 2018). This points to cooling by ~404 °C in 103 Myr and,

460 therefore, to an average cooling rate of 3.9 °C/Myr. These cooling rates estimated for the

461 EAU rely on ^{40}Ar/^{39}Ar biotite ages and closure temperatures only. They lack intermediate

462 constraints and therefore consider a linear cooling between the crystallization of zircons and

463 the closure of the K/Ar system in biotite. Taking into account the difference in cooling rates

464 above and below the closure temperature of amphibole obtained for rocks from the WMU and

465 the CPU, a cooling rate significantly lower than 4 °C/Myr may also be expected for the EAU

466 during the first tens of million years after the peak of metamorphism.

467 In summary, U-Pb, $^{40}Ar/^{39}Ar$ ages and the metamorphic temperature estimates from the

468 three allochthonous units consistently suggest low cooling rates (~3 to 8 °C/Myr; likely 3-

469 5 °C/Myr) during at least the first 40 Myr after initiation of cooling (based on the oldest

470 $^{40}Ar/^{39}Ar$ ages of amphiboles, Am3 on Figure 10), more probably 70-80 Myr (based on the

471 average amphibole $^{40}Ar/^{39}Ar$ age, Am1 on Figure 10), or even >100 Myr in the EAU. These

472 data also imply that cooling was faster (at least 11-15 °C/Myr) after the closure of the

473 amphibole K/Ar system, and possibly even faster (>20 °C/Myr) approaching the closure

474 temperature of muscovite, although this last inference needs further investigation since it

475 relies on the single muscovite age of our database. Moreover, as discussed above, the

476 estimated cooling rates consider linear trends between the different time markers

477 (crystallization of zircon and the closure of $^{40}Ar/^{39}Ar$ system in amphibole and biotite). Yet,

478 cooling was probably lower right after the metamorphic peak, especially during progressive

479 solidification of magmatic rocks, due to the release of the latent heat of crystallization

480 (Michaut and Jaupart, 2011), and has progressively increased through time. However,

481 although no evidence has been found yet, episodes of reheating cannot be ruled out.

482 The present estimates of cooling rates in the Araçuaí belt are in rather good agreement

483 with those proposed by Bento dos Santos et al. (2010; 2014) for anatexites, granulites, and

484 charnockites from the northern Ribeira belt. These authors propose a two-step thermal

485 evolution with initial cooling rates of 1-5 °C/Myr during 50-90 Myr, followed by faster

486 cooling by 8-30 °C/Myr and possibly up to 100 °C/Myr, based on diffusion of Fe^{2+} and Mg

487 between garnet and biotite inclusions. Although the approaches are different, estimates

488 obtained in both areas consistently point to a very slow cooling during several tens of million

489 years after the major high temperature deformation that shaped the Araçuaí and the northern

490 Ribeira belts.

491 $^{40}Ar/^{39}Ar$ ages of 498-493 Ma for amphibole and of 487 to 466 Ma for biotite obtained

492 in the West Congo belt in Angola (Monié et al., 2012) are similar to those obtained in the

493 Araçuaí belt, suggesting that, although the age of the orogenic/metamorphic peak is different

494 in these two belts (Araçuaí : 600-580 Ma; West Congo ~540-530 Ma), they shared a common

495 cooling history after the final amalgamation of Western Gondwana.

4.3. Effect of protracted high temperature conditions in the middle crust on its rheology and strain repartition

Under "hot orogen" conditions (high temperatures and low cooling rates in the middle and lower crust), we may expect specific rheological behaviors. When the temperature gradient reaches equilibrium, it is almost homogeneous over large distances both laterally and vertically (e.g., Whittington et al., 2009). Rheological contrasts between different crustal rock types are minimized, and strain localization efficiency, which is inversely proportional to stress and thus to temperature, becomes much lower. Strain tends to be distributed over large volume of rocks (Vanderhaeghe and Teyssier, 2001b; Vauchez et al., 2007). Consequently, transitions between orogenic domains should be much smoother than under lower temperature conditions (Vanderhaeghe and Teyssier, 2001a). This is particularly true when the middle to lower crust undergoes widespread partial melting, because this process may lead to a decrease in viscosity by several orders of magnitude (e.g., Rosenberg and Handy, 2005; Rosenberg et al., 2007; Cavalcante et al., 2014). The presence of even a small proportion of melt in a rock significantly modifies its viscosity. For instance, Rosenberg and Handy (2005) pointed out through reinterpretation of experimental data on partially molten systems that the strength of a rock undergoes a drastic drop during the first ~7% of partial melting. They also argued that, symmetrically, the strength of magmas deforming during crystallization remains low until they reach a proportion of solid-phase of ~90% and then increases rapidly up to the solidus. When cooling is slow, as it is the case in the Araçuaí orogen, high melt fractions subsist for long times and, by consequence, partially crystallized magmatic bodies or partially molten anatexites may accommodate most of the strain, either gravity-induced or due to far-field forces, during most of the duration of the continental collision.

The three main allochthonous units of the Araçuaí orogen are characterized by the presence of large volumes of synkinematic magmas intermingled with metasedimentary rocks: leucogranites interlayered with mylonites in the western unit, granodiorites/tonalites in the central unit, and anatexites in the eastern unit. Such large volumes of magmas in the middle crust certainly had a major impact on the rheology of the orogenic lithosphere. Using the cooling rates discussed in the previous section, we may roughly estimate the time lapse between emplacement and solidification of the magmas. For instance, the oldest anatectic granite of the EAU, emplaced at ~597 Ma in the middle crust at a minimum temperature of 800 °C yields a mean cooling rate of ~4 °C/Myr, suggesting that the solidus temperature (~630 °C) was attained at least 40-45 Myr after its emplacement, i.e., at about 550-555 Ma. Time lapses in the range of 30-40 Myr are obtained considering the synkinematic

530   leucogranites of the WMU and the tonalite and granodiorites of the CPU emplaced at ~580

531   Ma in mylonites or metasediments equilibrated at ~750 °C, which yield cooling rates of 3-

532   4 °C. The consistent estimates for the different units of the Araçuaí belt support that the large

533   volumes of magmas emplaced or formed in the three allochthonous units allowed for

534   magmatic flow and hence maintained low strengths in the middle crust for ~40 Myr after

535   initiation of cooling. Considering the suggestion by Ackerson et al. (2018) that granitic and

536   granodioritic magmas may contain a significant volume of interstitial magma down to

537   temperatures as low as ~500 °C, the above value may be still an underestimate. Indeed, in the

538   Grenville Province in North America, cooling rates in the range 2.3-4.4 °C/Myr and

539   maintenance of supra-solidus conditions over ≥70 Myr have been substantiated (Turlin et al.,

540   2018). The presence today, as deduced from seismic profiles (Nelson et al., 1996), of a

541   partially molten middle crust beneath Tibet, which results from partial melting initiated 20-25

542   Myr ago (Zhang et al., 2004; Guo and Wilson, 2012), illustrates that a long-lived partially

543   molten crust may have been present in orogens of a variety of ages.

544   Viscosity increase upon cooling may generate a strength inversion between the

545   magmatic bodies and their country rocks. Nabelek and Liu (2004) suggested, for instance,

546   inversion of shear strength between a mica schist and a dry granite at a temperature of ca.

547   530 °C depending weakly on strain rate. Based on the cooling rates estimated from our data,

548   strength inversion between metapelitic country rocks and magmatic bodies might have

549   occurred as late as 60 Myr after emplacement of the magmatic bodies. The magmatic bodies

550   were therefore the most deformable part of the middle crust during several tens of million

551   years and, considering their large volume, likely accommodated most of the strain imposed by

552   both plate tectonics and gravity. Upon cooling and crystallization of the magmas, the strength

553   of the magmatic rocks became higher than the strength of the metasedimentary country rocks

554   and the late pulses of deformation must have been accommodated mostly in metasediments.

555   The viscosity increase during cooling in a hot orogen may also lead to a progressive

556   decrease of the contribution of gravity-driven forces to regional deformation so that the strain

557   regime becomes dependent on far-field forces only. Strengthening of the middle crust may

558   also have allowed more effective transmission of tectonic forces toward the external domains

559   of the orogen, which are, in the present case, the remobilized eastern border of the São

560   Francisco craton.

4.4. Tectonic evolution of the Araçuaí belt

The slow cooling rates estimated in this study and their possible rheological consequences require reconsidering the tectonic evolution of the Araçuaí belt. As stated above, a characteristic of this orogenic segment is that strain was almost evenly distributed across the tectonic units described previously and occurred under high temperature conditions in the presence of syntectonic magmas. However, detailed mapping revealed that the tectonic fabric is variable across the belt. In the CPU and EAU in particular, the foliation and lineation form a complex pattern supporting a 3D deformation (Mondou et al., 2012; Cavalcante et al., 2013). Even if domains with different tectonic fabrics display consistent HT conditions of deformation and similar U-Pb ages of synkinematic magmas, the question of their contemporaneity remains open, since the present study substantiates that HT conditions were maintained during several tens of Myr. In partially molten rocks a new deformation event tends to erase evidence of any previous one. Fast diffusion, allowing for easy grain boundary migration, may have the same effect in HT metamorphic rocks. Under such conditions, it is difficult to establish a reliable chronology since classical techniques of structural analysis are therefore less efficient than in faster cooling orogenic belts.

Interestingly, despite the homogeneity of rock types, equilibrium temperatures, and systematic deformation in the magmatic state throughout the whole EAU, geochronological data denote a diachronous tectonothermal evolution within this unit. There is a correlation between the southeastward decrease in zircon ages evidenced by Cavalcante et al. (2018) and the variation in $^{40}Ar/^{39}Ar$ biotite ages across the EAU. Samples with the oldest U-Pb ages were the first ones to reach the closure temperature of biotite and samples with the youngest U-Pb ages also yield the youngest $^{40}Ar/^{39}Ar$ biotite ages (Figure 11). Only the $^{40}Ar/^{39}Ar$ age of biotite from the anatectic granite AR-535 does not follow this tendency, displaying a younger than expected biotite $^{40}Ar/^{39}Ar$ age. This might be due to the vicinity of a large massif of charnockite emplaced at ~520 Ma. The presence of this hot magmatic body may have reduced the cooling rate in the surrounding rocks during its solidification. There is no evidence of a correlation between U-Pb and Ar/Ar ages in the WMU and CPU, suggesting that the EAU underwent a specific thermotectonic evolution.

The variations in U-Pb and Ar/Ar ages support that, although not far from each other, some parts of the EAU started to crystallize and cooled down well before other parts. This might be due to diachronic migration of granites generated in the anatectic middle crust. However, considering the complex strain pattern mapped across the EAU (Cavalcante et al., 2013), the variation in thermal evolution in the EAU is in good agreement with a "channel-

flow like" deformation during which, due to gravity forces, some parts of the anatectic core of the belt were thrust upwards before or more efficiently than other parts (Cavalcante et al. 2018). The western EAU anatexites, in an initially shallower position, crystallized zircon and reached the closure temperature of biotite before the initially deeper eastern EAU. The regional variation in ages may also mean that the transcurrent fabric mapped in the eastern EAU (Cavalcante et al., 2013; Figure 2) has been formed after the gently-dipping foliations of the western part of the EAU, even if strain was, in both cases, accommodated by magmatic flow in the anatexites.

The "channel-flow like" interpretation for the deformation of the EAU is in agreement with the large-scale structure of the belt and with the crosscutting relationships of the contacts between the different units (Figure 12), which suggest diachronic deformations under supra-solidus temperature across the belt. The anatexites of the EAU cover discordantly the other allochthonous units. In the northernmost part of the Araçuaí belt, they are directly atop the para-autochthonous HT metasediments of the São Francisco craton (Figure 12a), while in the central part of the Araçuaí belt, the EAU is thrust over the subvertical contact between the western CPU and the higher temperature intermediate kinzigitic unit (Figure 12b). This supports earlier thrusting of the intermediate kinzigitic unit over the CPU, then verticalization of the contact and, finally, thrusting of the anatexites of the EAU onto the steeply-dipping structures of the central and intermediate units. No evidence of significant solid-state deformation associated with thrusting of the EAU over the other units has ever been observed in the various EAU anatexites, which are solely characterized by a strong magmatic fabric. This suggests that these events, although diachronic, occurred before full solidification of the magmatic bodies in the different units and that, at all times, deformation was concentrated in the partially molten (or crystallized) rocks.

5.   *Conclusions*

21 new $^{40}Ar/^{39}Ar$ ages of amphibole, muscovite and biotite from 16 samples collected in the central (CPU) and eastern (EAU) units of the Araçuaí belt, together with $^{40}Ar/^{39}Ar$ and U-Pb ages and P-T data previously published by our group (Vauchez et al., 2007; Petitgirard et al., 2009; Mondou et al., 2012, Cavalcante et al., 2013, 2014, 2018) provides constrains on the thermal history of the middle crust in the central part of the Araçuaí hot orogen. U-Pb zircon ages support that by ~600 Ma the middle to lower crust, heated to temperatures of ~800 °C, was already partially molten and that production of anatectic magmas persisted until ~570 Ma at least (Cavalcante et al., 2018). Most of the granodioritic/tonalitic magmatism of

628   the CPU, as well as the synkinematic leucogranites within the mylonitic metasediments of the

629   WMU, has been emplaced during this period (Nalini et al., 2000a,b; Noce et al., 2000;

630   Petitgirard et al, 2009; Mondou et al., 2012; Gonçalves et al., 2014).

631   Altogether these data point to a protracted orogenic evolution marked by:

632   - Initiation of crustal thickening and heating of the middle crust at about 630 Ma,

633   - Heating of the middle crust to ≥800 °C and initiation of pervasive partial melting at

634   ~600 Ma,

635   - Partial melting remained active until at least ~570 Ma; it affected the middle and

636   lower crust and also the mantle,

637   - Slow cooling (3-5 °C/Ma) of the middle crust, which reached temperatures of

638   ~500 °C at 510-500 Ma and of ~300 °C at ~480 Ma.

639   - Magmatic rocks maintained high melt fractions for several tens of Myr (>40 Myr)

640   during which they accommodated most of the imposed deformation by magmatic

641   flow. A rough estimate suggests that they reached the solidus temperature

642   (~630 °C) at ~545-555 Ma.

643   - The anatexites of the EAU underwent a diachronic evolution: those from the

644   northwestern part crystallized and cooled down earlier than those from the

645   southeastern part. However, estimated cooling rates across the EAU are similar.

646   This diachronic evolution is consistent with a "channel flow like" tectonic model

647   compatible with the 3D deformation of anatexites substantiated by Cavalcante et al.

648   (2013, 2018).

649   Such a protracted thermal evolution had major effects on the rheology of the orogenic

650   crust, on the strain repartition across the belt, and thus on the tectonic evolution of the

651   Araçuaí belt. During progressive heating of the orogenic crust (prograde metamorphism),

652   deformation was already active and crustal thickening was accommodated in the

653   metasedimentary sequence, but the microstructures recording these early stages of

654   deformation are impossible to differentiate from those formed at the peak of metamorphism in

655   presence of melt and during subsequent deformation under almost stable temperature

656   conditions. Initiation of partial melting at ~600 Ma and syntectonic injection of large volumes

657   of granodioritic/tonalitic magmas between 590 and 570 Ma strongly decreased the viscosity

658   of the middle crust, allowing gravity-induced deformation to combine with far-field

659   compressive forces to produce a 3D deformation regime (Cavalcante et al., 2013). This may

660   explain that northwestwards, the tectonic flow recorded in the middle crust rotate from

661 dominantly eastward onto the São Francisco craton to northward onto the Salvador craton
662 (e.g., Uhlein et al., 1998).

663   Similar $^{40}Ar/^{39}Ar$ cooling ages are obtained for biotite in both the Araçuaí belt in Eastern
664 Brazil and the West-Congo belt in Angola (Monié et al., 2012) despite the large differences in
665 the age of the orogenic climax and in tectonic evolution. This suggests that the late thermal
666 evolution in the Araçuaí-Ribeira orogeny is shared by the West-Congo - Cabo Frio belt
667 during the final stages of continental amalgamation. This might explain the late emplacement
668 of some atypical magmatic bodies (syenites, charnockites) in the Araçuaí belt.

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Sample number and type	U/Pb *Myr* (*=^{207}Pb/^{206}Pb)	^{40}Ar/^{39}Ar *Myr*			Closing T *°C*			Cooling rate *°C/My*
	Zr	*Am*	*Bt*	*Ms*	*Am*	*Bt*	*Ms*	
western mylonitic unit	*All data from (1)*							
AR-684 leucocratic melt	577±9							
AR-935 leucocratic melt	578±3*							
AR-562 Mylonite		501±5	474±5		488±7	310±11		~3/8
AR-86 leucocratic melt			468±4			318±12		
AR-414 Ibituruna syenite	530±1	505±5			485±6			
Contact zone Derribadinha	*All data from (2)*	This work						
AR-264 Tonalite	581±4							
AR-590 Tonalite		549±6	532±5		513±7	288±9		7/13
AR-940 leucocratic melt	587±5							
AR-87 leucocratic melt	579±8							
AR-1057 Tonalite	583±4*	542±8	508±5		521±8	293±8		5.6/6.6
Central plutonic unit	*All data from (2)*	*This work*						
AR-1009 SV To/Gd		501±9	490±5		526±6	321±9		
AR-800 SV To/Gd	582±6*		485±5			320±10		4.4
AR-968 SV To/Gd	585±7							
AR-957 Wolf leucogranite	585±4		492±6			326±5		4.6
AR-787 Galiléia To/Gd	582±6							
AR-705 Galiléia To/Gd	579±4		475±4			311±9		4.4
AR-815 Galiléia To/Gd	583±4							
AR-717 Galiléia To/Gd	581±4*	500±5	481±5		517±4	311±9		2.9/12
AR-747 Galiléia To/Gd			482±4			325±6		
AR-722 orthogneiss	2103±11		484±5			307±5		
AR-648 Palmital granite			484±5	488±5		306±6	416±4	M-B:27
Eastern anatectic unit	*All data from (3) except AR548*	*This work*						
AR-408 Diatexite	597±8							
AR-455 Anatectic granite	594±4							
AR-1133 Anatectic granite	597±3		486±4			307±5		4.4
AR-1145 Diatexite	593±3		477±4			328±3		4.1
AR-1315 Anatectic granite	597±4							
AR-664 Anatectic granite	589±6		474±4			307±5		4.1
AR- 1349 metatexite	583±5		476±4			323±3		4.3
AR-1358 Anatectic granite	585±4							
AR-535 Anatectic granite	575±3		459±4			338±7		3.3
AR-1119 Diatexite	578±4							
AR-1332 Anatectic granite	572±9							
AR-548 Anatectic granite	575±4 *(4)*		472±4 *(1)*			315±5		3.9
AR-1370 Anatectic granite	572±4		469±4			316±4		3.9

Table 1: Summary of U-Pb and ^{40}Ar/^{39}Ar ages obtained from rocks collected in the Central Araçuaí belt. Samples are organized from West to East. (1) is for ages published in Petitgirard et al. (2009), (2) for those published in Mondou et al., (2012), (3) for those

published in Cavalcante et al., 2018 and (4) for those published in Vauchez et al., 2007. The last column shows individual cooling rates calculated for samples for which U-Pb and $^{40}Ar/^{39}Ar$ ages of amphibole and/or biotite have been obtained, and for which an initial temperature of the country-rock may be inferred (for sample AR-562 and AR-590 an age of 578 Ma and 582 Ma respectively has been assumed taking into account their vicinity and similarity with dated samples). Cooling rates are indicated as: above closing T of amphibole/ between amphibole and biotite closing T. For sample AR-648, MB is for cooling rate between muscovite and biotite closing T.
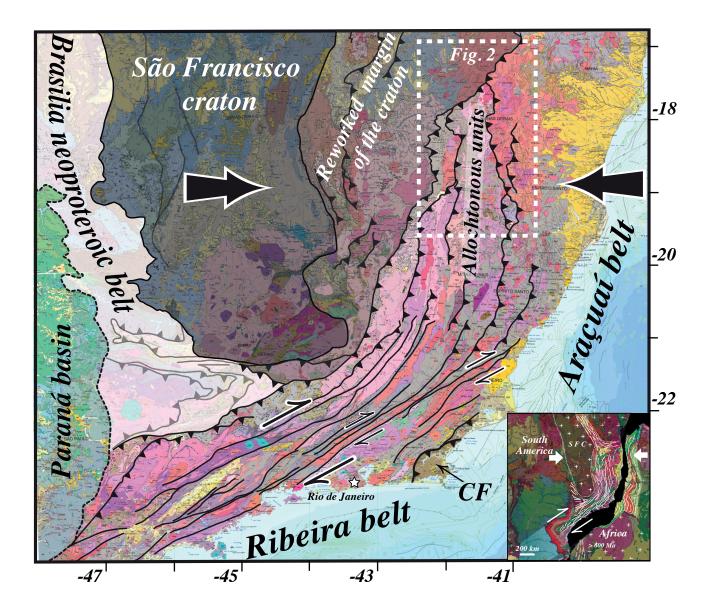
Figure 1: Schematic map of the neoproterozoic tectonic belts in eastern Brazil. The Brasilia belt is docked against the western and southwestern boundaries of the São Francisco (SF) craton. The Araçuaí and Ribeira belts formed along the eastern and southeastern boundary of the craton. The transition between these two orogenic segments, marked by a clear change in tectonic regime from contractional to tranpressional, is correlated with the southern termination of the craton. The eastern part of the São Francisco craton was strongly remobilized during the formation of the Araçuaí belt. Deformation in the Cabo Frio tectonic domain (CF) is younger than in the Araçuaí-Ribeira belt and likely represents the westernmost part of the early Cambrian Angola-Congo orogen. The underlying tectonic map is from (Cordani et al., 2016). The area delimited by doted white lines shows the location of Figure 2. Insert at the lower right corner is a reconstitution of the Ribeira-Araçuaí and Congo orogens before the South Atlantic opening (Vauchez et al., 2007).
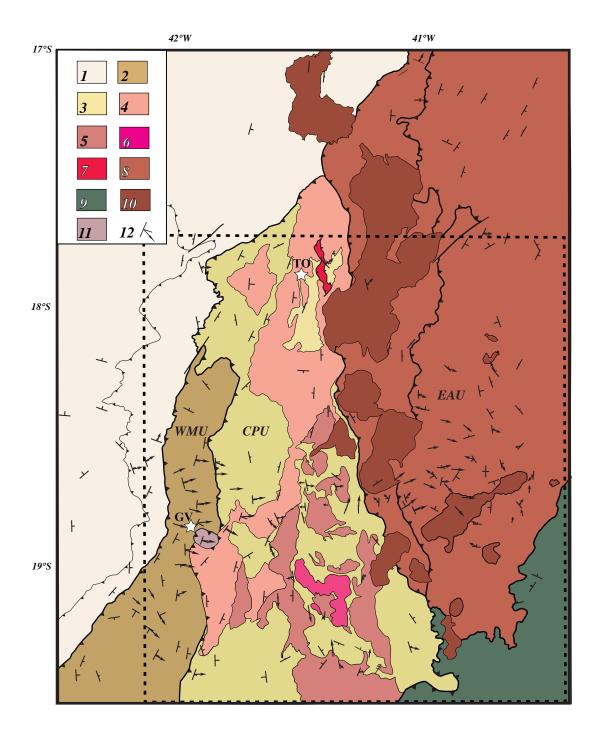
Figure 2: Simplified map of the central Araçuaí belt modified from Vauchez et al., 2007 showing (1) the para-autochthonous cover of the São Francisco cratons and the three allochthonous units: the western mylonitic unit (WMU, 2), the Central Plutonic Unit (CPU, 3-7) and the Eastern Anatectic Unit (EAU, 8-9). In the CPU, (3) is for metasediments, (4) for the "São Vitor" tonalite/granodiorite, (5) for the "Galileia" tonalite/granodiorite, (6) for the "Palmital" aluminous granite, (7) for the "Wolff" leucogranite. In the EAU: (8) anatexites and (9) kinzigites. (10) is the late "Padre Paraíso" charnockite. (11) foliation and lineation either measured in the field or inferred from AMS (Petitgirard et al., 2009; Mondou et al., 2012; Cavalcante et al., 2013).
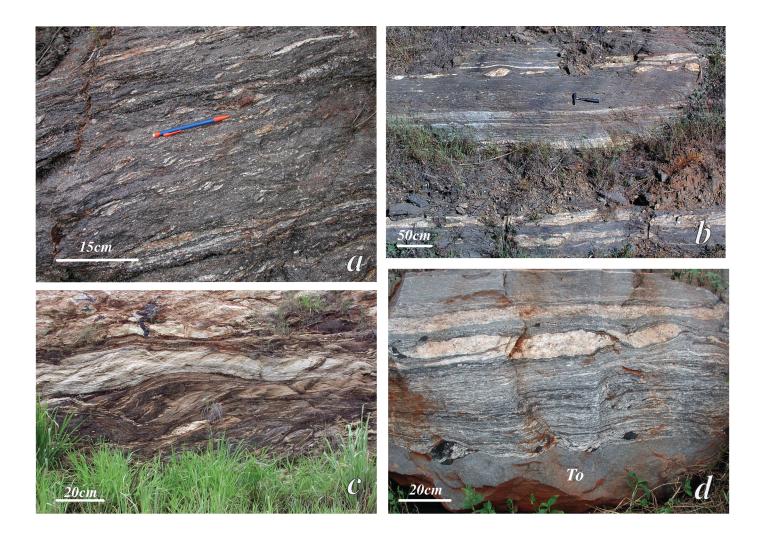
Figure 3: Typical outcrops of the Western Mylonitic Unit (a-c) and of its boundary with the Central Plutonic Unit (d). For the four pictures West is on the right. a-c show metasediments injected by leucocratic granite and shear criteria that indicate top-to-west sense of shear. In (a) lenses are leucogranite. (d) shows intercalations of leucogranite, migmatitic metasediments, tonalite (To) free of solid-state deformation and several asymmetric black lenses resulting from the disruption of an amphibolite layer.
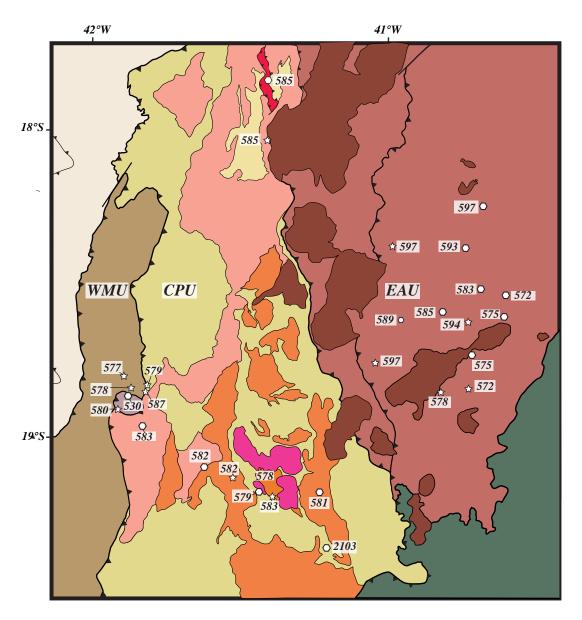
Figure 4: Location and U-Pb ages (Ma) of zircons used in this study. Results (and references) are summarized in table 1.

Figure 5: Pictures of outcrops typical of the Central Plutonic Unit. Tonalites/Granodiorites showing a well-developed magmatic foliation, gently-dipping in the western CPU (a-b) and that progressively becomes subvertical eastward (c-d). In (c) the subvertical magmatic foliation is underlined by elongated dioritic lenses and in (d) by lenses and discontinuous layers of biotite rich granite.
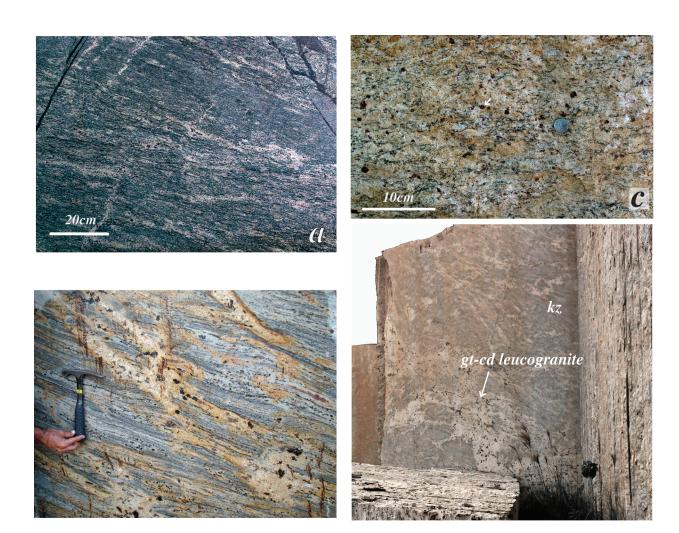
Figure 6: Pictures of outcrops typical of the Eastern Anatectic Unit: metatexites (a), diatexites (b) and garnet-rich leucocratic granite (c), all displaying a well-developed foliation, which is magmatic in the leucosome and solid-state in the restitic parts (when present). (d) shows the anatectic kinzigites (kz) at the eastern boundary of the EAU, in which leucogranitic melt focus and forms lenses and veins. Dark lenses in the leucogranite contains garnet + biotite ± cordierite.
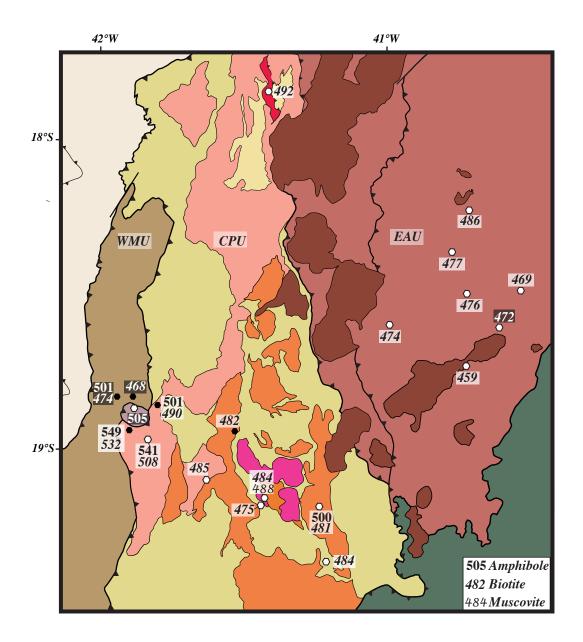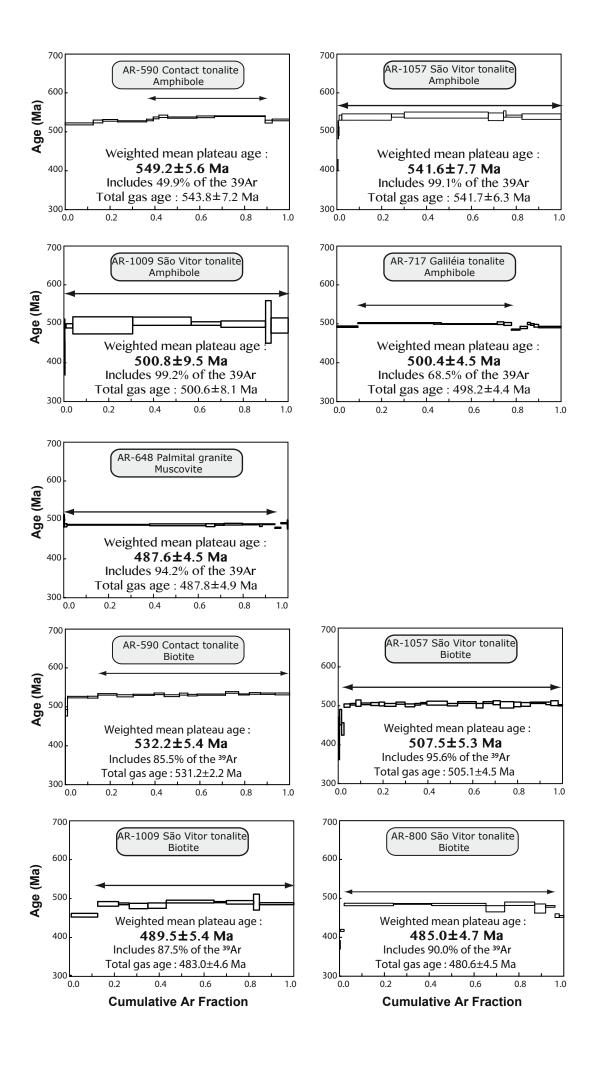
Figure 7 : Location and ^{40}Ar/^{39}Ar ages of amphibole, biotite and muscovite used in this study. Numbers in black are ages obtained in this study and those in white are previously published ^{40}Ar/^{39}Ar ages (4 samples). Diagrams for newly dated samples are shown on Figures 8 and 9. Analytical data are available in supplementary material #3 and #4.
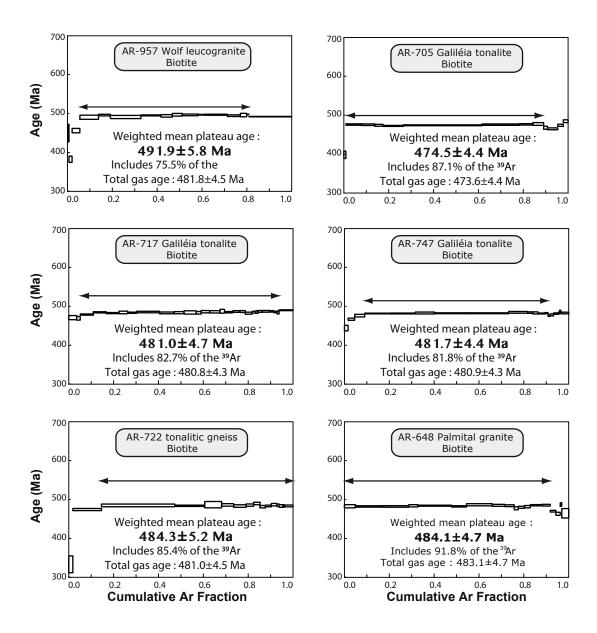
AR-590 Contact tonalite
Amphibole

Weighted mean plateau age :
549.2±5.6 Ma
Includes 49.9% of the 39Ar
Total gas age : 543.8±7.2 Ma

AR-1057 São Vitor tonalite
Amphibole

Weighted mean plateau age :
541.6±7.7 Ma
Includes 99.1% of the 39Ar
Total gas age : 541.7±6.3 Ma

AR-1009 São Vitor tonalite
Amphibole

Weighted mean plateau age :
500.8±9.5 Ma
Includes 99.2% of the 39Ar
Total gas age : 500.6±8.1 Ma

AR-717 Galiléia tonalite
Amphibole

Weighted mean plateau age :
500.4±4.5 Ma
Includes 68.5% of the 39Ar
Total gas age : 498.2±4.4 Ma

AR-648 Palmital granite
Muscovite

Weighted mean plateau age :
487.6±4.5 Ma
Includes 94.2% of the 39Ar
Total gas age : 487.8±4.9 Ma

AR-590 Contact tonalite
Biotite

Weighted mean plateau age :
532.2±5.4 Ma
Includes 85.5% of the 39Ar
Total gas age : 531.2±2.2 Ma

AR-1057 São Vitor tonalite
Biotite

Weighted mean plateau age :
507.5±5.3 Ma
Includes 95.6% of the 39Ar
Total gas age : 505.1±4.5 Ma

AR-1009 São Vitor tonalite
Biotite

Weighted mean plateau age :
489.5±5.4 Ma
Includes 87.5% of the 39Ar
Total gas age : 483.0±4.6 Ma

AR-800 São Vitor tonalite
Biotite

Weighted mean plateau age :
485.0±4.7 Ma
Includes 90.0% of the 39Ar
Total gas age : 480.6±4.5 Ma

Age (Ma)

Cumulative Ar Fraction

Figure 8: $^{40}Ar/^{39}Ar$ diagrams for amphibole, muscovite and biotite from samples collected in the Central Plutonic Unit. Diagrams are organized by mineral and for samples from West to East. Analytical data and methodology are available in supplementary material #3.
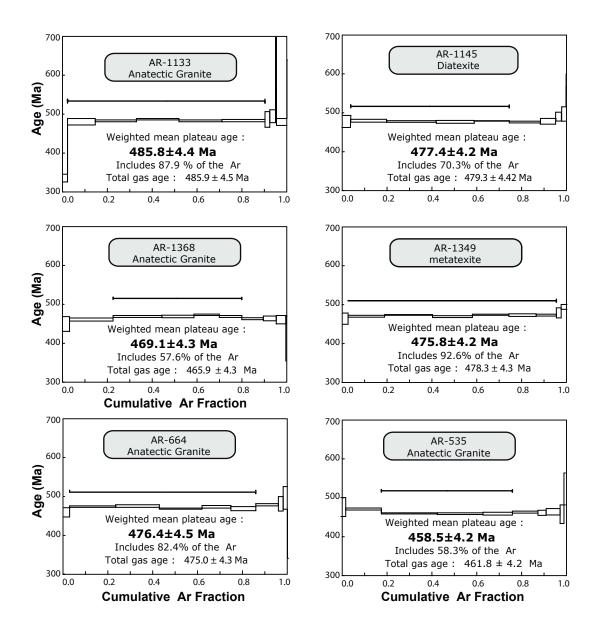
Figure 9: ^{40}Ar/^{39}Ar diagrams for biotite from samples collected in the Eastern Anatectic Unit. Diagrams are displayed for samples from North to South. Analytical data and methodology are available in supplementary data #4 .
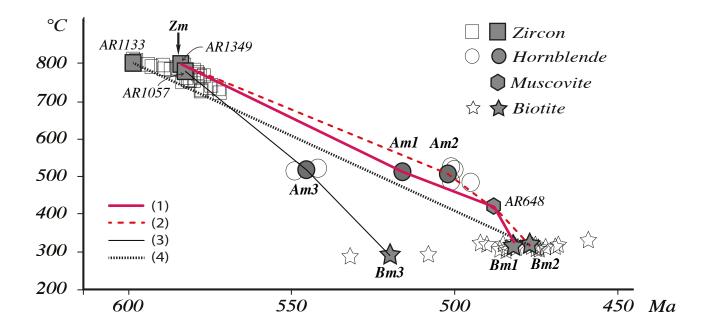
Figure 10: Figure 10: Diagram summarizing data and inferred thermochronological evolutions for the studied domain of the Araçuaí belt. Open symbols represents the whole set of data taken into account in this study. Symbols filled in grey represent either data for specific samples (zircons, muscovite) or average values computed for amphibole and biotite. Temperatures for zircons data have been roughly estimated from (a) the peak temperature (800 °C) and the oldest age (597 Ma) of the Eastern Anatectic Unit, (b) The Zn-in rutile temperature (780-790 °C) obtained for sample AR-1349 (583 Ma), (c) equilibrium temperatures (~750 °C) and ages (577-578 Ma) obtained for samples from the Western Mylonitic Unit, and (d) Ti-in-zircon temperature obtained for several samples from the EAU. Zm is average U-Pb age of zircons from synkinematic magmatic/anatectic rocks, Am are average $^{40}Ar/^{39}Ar$ ages of amphiboles with Am1 for all dated samples and Am2 and Am3 without and for the two oldest samples, respectively. Similarly, Bm are average $^{40}Ar/^{39}Ar$ ages of biotites with Bm1 for all dated samples and Bm2 and Bm3 without and for the two oldest samples, respectively. Line (1) represents the thermochronological evolution based on average values for all dated samples, (2) excludes the two oldest amphiboles and biotites, (3) is the cooling evolution based on the mean ages of the two oldest amphiboles and biotites, and (4) links the oldest zircons ages of the Eastern Anatectic unit to average age of biotites.
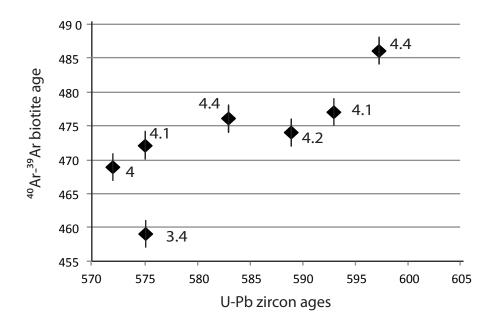
Figure 11: U-Pb zircon ages versus ^{40}Ar/^{39}Ar biotite ages in the Eastern Anatectic Unit. Numbers are for the cooling rate calculated for each sample (Table 1). This figure shows a tendency of the age of closure of the ^{40}Ar/^{39}Ar system in biotite to be correlated with zircon crystallization ages. Samples showing earlier crystallization of zircon also show earlier closure of the Ar/Ar system in biotite. This suggests a diachronous thermal evolution across the EAU.
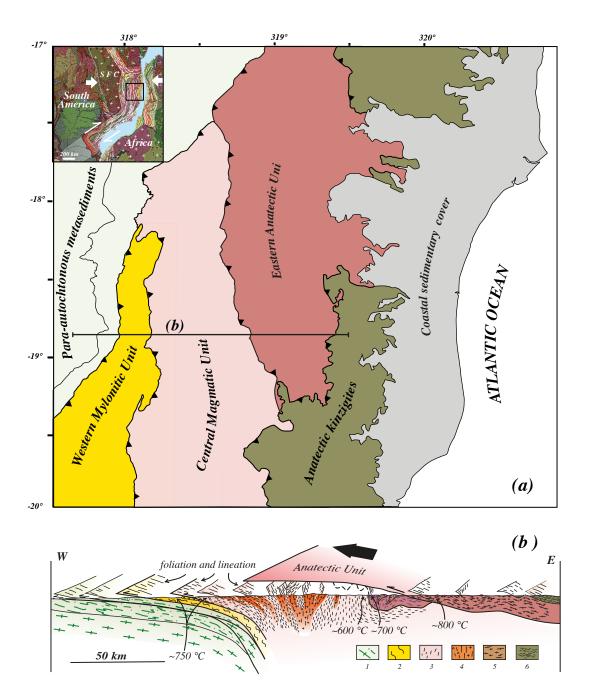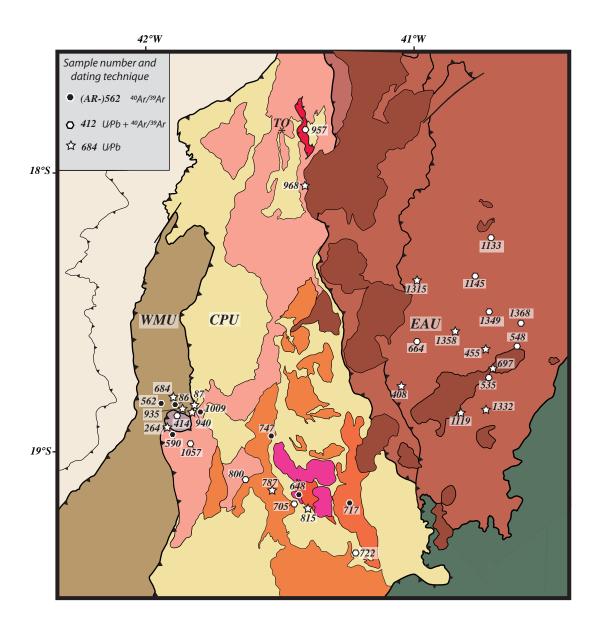
Figure 12: Schematic map and cross-section showing the Eastern Anatectic Unit thrust upon the Central and Western Units and even, in the northernmost part of the map, directly onto the para-autochthonous metasedimentary cover of the São Francisco craton. The cross-section also shows that the gently dipping basal contact of the Eastern unit crosscuts steeply dipping tectonic contacts and foliation in the eastern part of the CPU, suggesting that the tectonic emplacement of the EAU is later than the formation of this subvertical fabric, although it occurred before the anatexites reached solid-state.
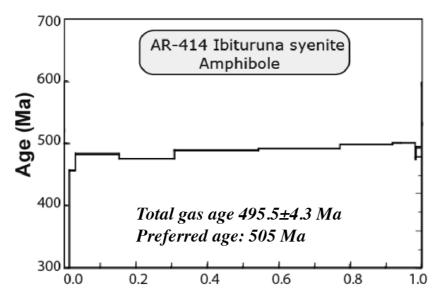
Supplementary material #1: Location of samples from which zircons and/or amphibole, biotite and muscovite have been dated. Results (and references for data already published) are summarized in Table 1, Figure 4 for zircons and Figure 7 for ^{40}Ar/^{39}Ar ages. In the main text, sample numbers are preceded by AR- (for instance 562 is AR-562). Legends of the map as in Figure 2.

Supplementary material #2:

$^{40}Ar/^{39}Ar$ data and diagram for sample AR-414 (Petitgirard et al., 2009).

Amphibole from this sample (Ibituruna syenite) did not yield a plateau age but a total gas age of 495.4 ± 4.3 Ma that should be considered as a minimum age for its closure for argon. Apparent ages slightly increase with increasing degassing temperatures up to apparent ages in the range 500-510 Ma corresponding to almost constant $^{37}Ar/^{39}Ar$ (Ca/K) and $^{38}Ar/^{39}Ar$ (Cl/K) compositions. We thus preferred taking into account the integrated age of 505 ± 5 Ma for the amphibole from this sample.

AR414 Ibituruna Syenite Amphibole					Data from Petitgirard et al.. 2009				
Step	$^{40}Ar/^{39}Ar$	$^{38}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$ (E-3)	$F^{39}Ar$ released	$\%^{40}Ar^*$	$^{40*/39}Ar_K$	Age (Ma)	±1 SD (Ma)
1	51.133	0.085	0.2841	133.399	0.99	22.90	11.71	186.1	2.2
2	44.631	0.074	0.1807	93.225	1.49	38.26	17.08	265.4	2.2
3	41.129	0.048	0.0574	30.448	3.08	78.08	32.12	470.5	0.6
4	34.864	0.040	0.0072	3.000	15.41	97.40	33.96	494.1	0.2
5	33.716	0.053	0.1632	0.863	30.92	99.21	33.45	487.6	0.3
6	34.722	0.115	0.9239	1.309	54.47	98.99	34.39	499.6	0.4
7	34.715	0.189	1.8392	0.875	77.15	99.52	34.59	502.1	0.2
8	35.299	0.218	2.2291	1.184	91.93	99.33	35.11	508.7	0.2
9	35.967	0.224	2.5280	3.007	94.47	97.90	35.27	510.7	0.4
10	35.832	0.247	2.7335	2.467	98.33	98.37	35.31	511.2	0.3
11	37.297	0.250	2.6180	13.641	98.63	89.56	33.46	487.7	1.7
12	41.118	0.249	2.6689	22.207	100.00	84.38	34.76	504.3	1.1

Petitgirard, S., Vauchez, A., Egydio-Silva, M., Bruguier, O., Camps, P., Monié, P., Babinski, M. and Mondou, M., 2009. Conflicting structural and geochronological data from the Ibituruna quartz-syenite (SE Brazil): Effect of protracted "hot" orogeny and slow cooling rate? Tectonophysics 477, 174-196.

Supplementary material #3:

Ar- Ar results for samples from the Central Plutonic Unit

Method:

These samples were collected for laser probe single grain ^{40}Ar-^{39}Ar studies at Geosciences Montpellier (Université de Montpellier, France). They were hand crushed and cleaned using ethanol and distilled water. Single grains of amphibole and mica about 500 μm in diameter were handpicked under binocular microscope and ultrasonically cleaned in acetone and distilled water. The grains were then packaged in aluminium foils and irradiated along with several flux monitors of MMhb-1 hornblende and TCR2-sanidine at McMaster Reactor (Canada) with an integrated power of 60 MWH. K_2SO_4 and CaF_2 salts were also irradiated to determine the correction factors for interference reactions from K and Ca. ^{40}Ar-^{39}Ar laser degassing was carried out at Geosciences Montpellier (France) using a 50W CO_2 laser. The gas cleaning was achieved in a low volume UHV gas purification system including a cold trap and two Ti/Zr getters. Argon isotope measurements were performed on a MAP215-50 noble gas mass spectrometer directly connected to the UHV purification system. Blank analyses were performed every three runs to eliminate the background contribution on the gas released from the sample itself. Individual grains were stepwise heated by increasing the laser beam intensity in successive increments until the final fusion. A plateau age is defined by three or more contiguous heating steps comprising 50% or more of the ^{39}Ar$_K$ released and overlapping at the two-sigma confidence level. Because of the high radiogenic content of the dated minerals and the clustering of data points near the ^{39}Ar-^{40}Ar abscissa, ^{36}Ar-^{40}Ar versus ^{39}Ar-^{40}Ar correlation plots do not provide meaningful information about the age and initial ^{36}Ar-^{40}Ar values.

Results:

Step	^{40}Ar/^{39}Ar	^{38}Ar/^{39}Ar	^{37}Ar/^{39}Ar	^{36}Ar/^{39}Ar (E-3)	F^{39}Ar released	%^{40}Ar*	$^{40*/39}$Ar$_K$	Age (Ma)	±1 SD (Ma)
AR-590 Derribadinha tonalite Amphibole									
1	1190.597	0.088	0.00163	115.344	0.01	97.14	1156.5	4440.84	959.0
2	525.881	0.414	37.06841	395.543	0.05	78.30	421.95	2871.623	139.9
3	40.087	0.055	0.07051	11.151	12.52	91.75	36.78	529.416	2.8
4	38.401	0.053	0.00000	3.618	16.75	97.17	37.32	536.056	3.8
5	38.749	0.053	0.00000	3.893	23.39	96.99	37.58	539.361	2.4
6	37.889	0.054	0.00000	1.733	36.33	98.61	37.36	536.607	1.5
7	38.485	0.05	0.00000	2.784	39.29	97.82	37.65	540.157	2.3
8	39.338	0.052	0.88118	4.345	41.82	96.86	38.13	546.099	2.4
9	40.045	0.053	0.22236	5.34	45.71	96.06	38.47	550.377	3.4
10	38.602	0.197	9.80187	4.637	58.67	98.32	38.2	546.953	1.8
11	38.792	0.22	8.05135	4.118	66.52	98.38	38.36	549.036	2.3
12	38.736	0.234	8.42765	3.584	89.15	98.86	38.5	550.762	1.2
13	37.921	0.149	4.41234	3.567	92.30	98.05	37.29	535.727	3.7

14	38.258	0.119	3.12708	2.99	100.00	98.26	37.67	540.452	2.3

AR-590 Derribadinha tonalite Biotite

1	260.267	0.279	0.85808	821.286	0.12	6.77	17.64	273.22	249.9
2	47.308	0.078	0.30582	47.638	0.99	70.26	33.24	484.733	8.4
3	37.783	0.055	0.00454	4.487	9.95	96.45	36.44	525.151	1.9
4	38.096	0.056	0.02386	5.614	14.51	95.61	36.42	524.927	2.4
5	37.547	0.054	0.00000	1.838	23.02	98.51	36.99	531.976	1.9
6	36.978	0.051	0.00894	0.905	30.26	99.24	36.70	528.325	2.8
7	37.316	0.052	0.00000	1.352	39.65	98.89	36.90	530.883	1.5
8	37.577	0.052	0.02104	1.507	44.11	98.78	37.12	533.591	2.0
9	37.213	0.052	0.00000	1.592	50.89	98.69	36.73	528.715	2.5
10	37.553	0.053	0.01094	1.671	54.61	98.65	37.04	532.675	2.6
11	37.395	0.054	0.00957	1.835	58.77	98.51	36.84	530.104	2.1
12	37.339	0.054	0.00419	1.298	71.42	98.93	36.94	531.374	1.7
13	38.200	0.054	0.02762	2.665	77.27	97.90	37.40	537.086	2.2
14	37.479	0.053	0.00051	1.614	82.08	98.69	36.99	531.947	1.5
15	37.740	0.053	0.00000	1.817	87.03	98.54	37.19	534.458	2.2
16	37.682	0.054	0.01606	1.812	93.55	98.54	37.13	533.775	1.7
17	37.827	0.052	0.00000	2.458	100.00	98.04	37.09	533.182	2.3

AR-1057 São Vitor tonalite Amphibole   J= 0.009621

1	1318.154	5.167	0.01591	4055.33	0.01	9.09	119.79	1383.1	7131.5
2	137.46	1.089	0.00341	248.554	0.06	46.56	64	865.6	1790.6
3	87.653	1.146	0.00438	857.07	0.10	-188.96	-165.63	0.0	0.0
4	73.132	0.824	0.00216	49.913	0.19	79.81	58.37	804.0	1148.4
5	26.992	0.121	0.00062	0.002	0.47	99.94	26.98	416.3	14.2
6	33.716	0.136	0.00121	0.003	0.62	99.95	33.7	506.6	23.1
7	33.403	0	0.00064	0.002	0.90	99.95	33.39	502.5	11.7
8	36.065	0.038	0.00015	0	2.13	99.96	36.05	537.2	6.9
9	36.223	0.103	0.00001	0	24.26	99.96	36.21	539.2	8.1
10	36.476	0.083	0.00003	0	29.92	99.96	36.46	542.5	4.1
11	36.61	0.11	0.00000	0	67.39	99.96	36.59	544.2	7.7
12	36.231	0.073	0.00003	0	74.18	99.96	36.22	539.3	10.1
13	36.788	0.053	0.00014	0	75.42	99.96	36.77	546.5	8.0
14	36.393	0.09	0.00003	0	82.57	99.96	36.38	541.4	2.8
15	36.267	0.055	0.00001	0	100.00	99.96	36.25	539.8	6.7

AR-1057 São Vitor tonalite Biotite   J= 0.009621

1	110.008	0.856	51.77421	633.621	0.02	-66.68	-75.92	0.0	0.0
2	55.347	0.118	42.76951	156.805	0.05	22.05	12.55	205.7	635.9
3	19.962	0	27.29096	4.961	0.12	99.92	20.9	330.5	253.3
4	30.371	0.041	1.20390	11.343	0.44	89.21	27.12	418.2	54.5
5	32.03	0.03	0.00003	2.986	1.30	97.2	31.13	472.7	20.0
6	34.103	0.024	5.91843	19.592	2.40	84.28	28.85	442.0	16.0
7	33.375	0.021	0.00001	0	5.04	99.95	33.36	502.2	4.0
8	33.621	0.021	0.00001	0	7.65	99.95	33.61	505.4	2.4
9	33.872	0.022	0.00001	0.173	10.01	99.8	33.81	508.0	8.0
10	33.672	0.019	0.00001	0	14.42	99.95	33.66	506.1	2.1
11	33.639	0.017	0.00001	0	18.01	99.95	33.62	505.6	4.1
12	33.621	0.013	0.00001	0	20.48	99.95	33.61	505.4	4.8
13	33.401	0.025	0.68492	0	25.61	99.95	33.45	503.4	3.8
14	33.81	0.019	0.00001	0	29.71	99.95	33.79	507.9	2.6

15	33.405	0.016	0.15295	0	33.97	99.95	33.4	502.8	3.5
16	33.596	0.018	0.00995	0	36.51	99.95	33.58	505.1	1.9
17	33.77	0.019	0.47576	0	39.37	99.95	33.8	508.0	1.9
18	33.827	0.028	0.09567	0	48.56	99.95	33.82	508.2	5.0
19	33.52	0.018	0.00001	0	53.17	99.95	33.5	504.1	3.1
20	33.961	0.019	0.00000	0	61.48	99.95	33.95	509.9	2.6
21	33.56	0.021	0.00001	0.725	64.49	99.32	33.33	501.8	5.9
22	33.94	0.021	0.00001	0	68.98	99.95	33.92	509.6	2.6
23	33.895	0.019	0.00001	0	72.19	99.95	33.88	509.0	4.8
24	33.485	0.021	0.39021	0	78.31	99.95	33.51	504.1	8.5
25	33.807	0.018	0.03891	0.776	81.48	99.28	33.57	504.9	6.2
26	33.809	0.02	0.11646	0.905	85.65	99.19	33.54	504.5	4.8
27	33.911	0.02	0.00000	0	92.58	99.95	33.9	509.2	1.4
28	33.821	0.019	0.00001	0	94.72	99.95	33.81	508.0	3.3
29	33.943	0.02	0.00001	0.354	98.05	99.65	33.82	508.2	5.8
30	33.419	0.023	0.00001	0	100.00	99.95	33.4	502.8	1.4

AR-1009 São Vitor tonalite Amphibole				J= 0.009621					
1	240.903	2.011	0.00999	0.029	0.02	99.99	240.88	2163.5	187.3
2	110.055	0	0.00470	0.014	0.07	99.98	110.04	1302.6	104.3
3	31.356	0.142	0.00148	0.004	0.21	99.95	31.34	475.4	36.4
4	26.766	0	0.00195	0.006	0.31	99.94	26.75	413.1	43.5
5	13.629	0.094	0.00048	0.001	0.75	99.88	13.61	222.0	12.5
6	32.837	0.098	0.00007	0	3.84	99.95	32.82	495.1	5.7
7	32.942	0.092	0.00001	0	30.41	99.95	32.93	496.5	21.6
8	33.752	0.116	0.00001	0	56.55	99.95	33.74	507.1	10.3
9	33.243	0.104	0.00002	0	69.92	99.95	33.23	500.4	4.6
10	33.177	0.094	0.00001	0	89.86	99.95	33.16	499.6	7.7
11	36.04	0.126	0.00008	8.32	92.38	93.13	33.57	504.9	54.6
12	35.724	0.111	0.00749	9.601	100.00	92.02	32.87	495.7	19.3

AR-1009 São Vitor tonalite Biotite				J= 0.009621					
1	17.141	0.063	0.00031	48.478	0.21	16.34	2.8	48.0	246.5
2	19.604	0.054	2.54802	27.182	0.64	59.92	11.77	193.5	112.0
3	30.648	0.027	0.23931	2.129	12.46	97.96	30.03	457.8	4.8
4	33.241	0.033	0.56846	3.592	21.69	96.89	32.22	487.1	5.9
5	32.232	0.025	0.00002	0	26.50	99.95	32.22	487.1	2.4
6	32.84	0.028	0.48662	3.662	34.88	96.77	31.79	481.4	6.6
7	32.752	0.028	0.70427	3.193	43.07	97.23	31.86	482.4	6.6
8	32.715	0.028	0.20325	0.356	64.36	99.68	32.61	492.3	3.6
9	32.546	0.023	0.00001	0	69.86	99.95	32.53	491.2	1.3
10	32.668	0.024	0.08151	0.609	82.04	99.42	32.48	490.6	4.4
11	32.741	0.018	1.23529	1.025	84.66	99.31	32.54	491.4	20.1
12	32.509	0.025	0.40070	1.045	100.00	99.09	32.22	487.2	2.5

AR-800 São Vitor tonalite Biotite				J= 0.009621					
1	24.463	0.038	0.00042	0.001	0.24	99.93	24.45	381.1	11.0
2	27.11	0	0.00006	0	1.91	99.94	27.09	417.9	2.6
3	31.993	0.013	0.00000	0	24.01	99.95	31.98	483.9	3.2
4	32.104	0.01	0.03448	0	40.98	99.95	32.09	485.4	1.0
5	31.991	0.01	0.32973	0.101	65.29	99.94	31.98	483.9	2.8
6	31.174	0.01	1.67359	0.53	73.66	99.85	31.16	473.1	7.9
7	32.499	0.014	0.30455	1.315	87.06	98.83	32.12	485.9	4.3
8	31.573	0.015	0.92221	1.489	91.94	98.78	31.21	473.6	11.6
9	31.366	0.014	2.65175	0	96.19	99.95	31.6	479.0	2.1
10	29.342	0.012	5.49134	0	97.99	99.95	29.85	455.4	4.7

11	29.15	0.004	5.97273	0	100.00	99.95	29.7	453.4	2.3

AR-705 Galiléia tonalite Biotite J= 0.009143

1	47.49	0.035	0.09484	69.249	0.96	56.89	27.02	398.3	9.4
2	34.542	0.019	0.04145	5.079	7.90	95.62	33.03	476.1	1.7
3	34.124	0.019	0.03185	3.779	13.88	96.69	32.99	475.6	2.0
4	34.076	0.018	0.02898	4.019	17.51	96.48	32.88	474.1	2.4
5	35.064	0.017	0.00153	7.771	26.51	93.41	32.75	472.6	1.3
6	33.221	0.017	0.00252	1.093	61.59	98.98	32.88	474.2	1.1
7	33.451	0.016	0.00933	1.446	78.28	98.68	33.01	475.8	1.7
8	33.74	0.015	0.00433	2.203	84.24	98.03	33.07	476.6	1.8
9	33.62	0.015	0.04585	1.566	89.05	98.59	33.15	477.6	3.5
10	33.468	0.019	0.00000	3.643	90.78	96.74	32.38	467.8	3.9
11	33.088	0.016	0.00000	3.046	95.39	97.23	32.17	465.2	2.4
12	33.857	0.021	0.01916	3.274	97.83	97.1	32.88	474.1	2.7
13	34.062	0.022	0.30465	1.224	100.00	98.96	33.71	484.7	3.7

AR-957 Galiléia tonalite Biotite J= 0.009621

1	90.921	0.123	0.00007	220.372	0.47	28.36	25.79	399.8	45.0
2	28.051	0.133	0.00002	38.963	1.93	58.9	16.52	266.1	16.0
3	26.711	0.083	0.00001	0	5.44	99.94	26.7	412.4	11.3
4	31.753	0.099	0.00000	0	13.66	99.95	31.74	480.7	10.7
5	32.575	0.081	0.00001	0	19.06	99.95	32.56	491.6	4.1
6	31.829	0.095	0.00000	0	32.63	99.95	31.81	481.7	8.0
7	32.456	0.095	0.00000	0	41.55	99.95	32.44	490.1	2.4
8	32.39	0.087	0.74200	0.619	46.68	99.56	32.26	487.7	6.4
9	32.439	0.088	2.72211	0	51.05	99.95	32.69	493.3	6.8
10	32.654	0.063	0.00000	0	58.09	99.95	32.64	492.7	4.2
11	32.703	0.104	0.19190	0	61.17	99.95	32.71	493.6	1.7
12	32.701	0.069	0.00000	0	71.31	99.95	32.69	493.3	2.6
13	32.35	0.1	0.00001	0	77.13	99.95	32.33	488.7	4.4
14	32.634	0.087	1.00316	0.444	79.83	99.78	32.58	491.9	8.7
15	32.839	0.062	0.00003	0	80.90	99.95	32.82	495.1	1.0
16	31.988	0.12	0.18118	0	100.00	99.95	31.99	484.1	1.0

AR-747 Galiléia tonalite Biotite J= 0.009143

1	57.422	0.03	0.01364	91.03	1.61	53.13	30.51	443.9	7.1
2	39.955	0.022	0.02162	25.911	4.52	80.8	32.28	466.6	2.8
3	35.88	0.016	0.00738	9.687	9.02	91.98	33	475.7	3.3
4	34.273	0.015	0.00168	2.766	20.32	97.57	33.44	481.3	0.9
5	34.296	0.015	0.00014	2.832	31.62	97.51	33.44	481.3	1.1
6	33.764	0.015	0.01223	0.697	40.13	99.35	33.54	482.6	2.4
7	33.859	0.015	0.00433	1.196	56.64	98.91	33.49	481.9	1.1
8	33.859	0.015	0.00036	1.196	73.15	98.91	33.49	481.9	1.1
9	34.115	0.015	0.00573	1.423	81.64	98.72	33.68	484.3	2.2
10	34.308	0.015	0.00025	2.258	84.00	98.01	33.63	483.6	1.9
11	33.948	0.016	0.00000	1.262	86.30	98.86	33.56	482.8	2.7
12	34.072	0.016	0.00000	2.071	89.38	98.16	33.44	481.3	1.2
13	34.353	0.016	0.00000	2.806	90.79	97.54	33.51	482.1	1.5
14	33.582	0.014	0.01767	1.878	91.79	98.31	33.01	475.9	2.1
15	33.964	0.016	0.09576	2.674	93.24	97.65	33.17	477.8	2.6
16	34.171	0.015	0.04058	2.497	95.25	97.8	33.42	481.0	1.7
17	34.185	0.013	0.03524	2.226	95.88	98.04	33.51	482.2	2.4
18	35.065	0.012	0.04808	3.948	96.45	96.64	33.89	486.9	2.7
19	34.571	0.016	0.37912	3.552	100.00	97	33.54	482.6	2.1

AR-722 Orthogneiss Biotite — J= 0.009143

1	159.363	0.099	0.00000	464.193	1.87	13.92	22.18	333.1	21.7
2	34.741	0.014	0.00000	6.242	14.58	94.65	32.88	474.2	2.6
3	34.216	0.014	0.00000	1.643	47.17	98.54	33.72	484.7	3.5
4	33.938	0.014	0.01015	0.977	60.65	99.11	33.63	483.7	1.7
5	33.969	0.014	0.01127	0.317	68.16	99.68	33.86	486.6	8.2
6	34.18	0.013	0.00113	1.167	73.52	98.95	33.82	486.1	3.3
7	33.726	0.016	0.00313	0.389	79.06	99.61	33.6	483.3	3.3
8	33.956	0.014	0.00000	0.523	82.42	99.5	33.79	485.6	3.1
9	34.568	0.015	0.04584	1.497	85.25	98.68	34.11	489.8	3.2
10	34.858	0.01	0.04797	4.659	87.71	96.02	33.47	481.7	3.2
11	34.272	0.016	0.00315	2.179	90.25	98.08	33.61	483.5	2.2
12	34.487	0.01	0.00000	2.001	92.79	98.24	33.88	486.8	3.1
13	34.305	0.008	0.04174	1.852	93.90	98.37	33.75	485.1	3.2
14	33.628	0.013	0.00000	0	95.50	99.95	33.61	483.5	3.4
15	35.277	0.016	0.00000	5.575	96.17	95.29	33.61	483.5	3.1
16	33.888	0.015	0.00000	0.891	100.00	.18	33.61	483.4	2.4

AR-717 Galiléia tonalite Amphibole — J= 0.009143

1	34.57	0.087	2.81088	1.655	9.34	99.15	34.34	492.6	1.7
2	35.003	0.1	3.03178	0.741	43.23	99.96	35.07	501.7	1.0
3	36.44	0.098	2.87131	5.795	46.51	95.85	34.99	500.8	1.9
4	34.743	0.099	3.01849	0.501	71.53	99.96	34.87	499.3	0.8
5	35.207	0.097	3.19382	1.54	74.55	99.34	35.05	501.5	3.4
6	34.857	0.094	3.17437	0.909	77.87	99.87	34.88	499.4	3.6
7	33.844	0.093	3.06684	1.285	81.76	99.51	33.75	485.1	1.4
8	34.243	0.096	3.08908	1.2	84.98	99.6	34.17	490.5	2.8
9	34.695	0.089	3.04323	0	86.34	99.96	34.98	500.6	2.8
10	34.46	0.087	3.50846	0	87.96	99.95	34.79	498.2	2.2
11	34.269	0.101	4.06140	0.423	89.99	99.95	34.52	494.9	3.4
12	34.294	0.086	4.88181	1.663	100.00	99.59	34.26	491.6	1.8

AR-717 Galiléia tonalite Biotite — J= 0.009143

1	69.019	0.051	0.01228	123.733	3.52	47	32.44	468.6	4.8
2	41.894	0.034	0.00000	32.376	5.10	77.13	32.31	467.0	4.2
3	39.081	0.029	0.00000	20.501	10.99	84.46	33.01	475.8	1.4
4	37.388	0.03	0.00000	13.371	16.72	89.39	33.42	481.0	1.4
5	36.138	0.028	0.00294	9.623	22.38	92.09	33.28	479.2	1.3
6	36.239	0.027	0.00000	9.562	26.30	92.16	33.4	480.7	3.0
7	35.741	0.027	0.01162	8.044	30.38	93.31	33.35	480.1	1.3
8	34.784	0.027	0.00000	4.364	40.46	96.25	33.48	481.8	1.8
9	34.724	0.026	0.00451	4.722	46.91	95.94	33.31	479.7	2.1
10	34.831	0.024	0.00000	4.944	51.38	95.76	33.35	480.2	2.5
11	34.462	0.026	0.00000	3.546	56.80	96.91	33.4	480.8	3.9
12	34.525	0.025	0.01044	3.976	59.00	96.55	33.34	480.0	3.0
13	34.477	0.025	0.00000	3.118	62.83	97.28	33.54	482.5	2.7
14	34.058	0.027	0.00000	1.788	67.39	98.4	33.51	482.2	2.7
15	34.103	0.028	0.00000	1.581	69.44	98.58	33.62	483.6	4.6
16	33.701	0.026	0.00000	0.809	73.23	99.24	33.45	481.4	1.6
17	33.807	0.026	0.00000	0.726	76.31	99.32	33.58	483.0	2.0
18	34.128	0.026	0.00000	1.843	78.69	98.36	33.57	482.9	2.2
19	33.742	0.023	0.00778	1.58	81.84	98.57	33.26	479.0	2.4
20	33.921	0.026	0.05392	1.562	83.39	98.6	33.45	481.4	2.1
21	34.031	0.026	0.13242	1.878	84.41	98.35	33.47	481.7	3.2
22	34.255	0.027	0.07780	2.209	86.16	98.07	33.59	483.2	2.2
23	34.088	0.025	0.00000	1.579	89.48	98.59	33.61	483.4	2.0

24	34.182	0.023	0.00000	1.897	91.45	98.31	33.61	483.4	1.7
25	34.433	0.026	0.00000	3.116	92.87	97.28	33.5	482.0	1.8
26	34.898	0.026	0.06806	5.072	93.74	95.68	33.39	480.7	3.5
27	34.888	0.026	0.07908	3.617	100.00	96.91	33.81	486.0	1.0

AR-648 Palmital granite Muscovite		J= 0.009143							
1	44.588	0.045	0.00000	33.342	0.32	77.87	34.72	497.4	15.8
2	41.464	0.014	0.08912	25.844	1.44	81.56	33.82	486.1	4.0
3	34.49	0.012	0.00096	1.894	38.29	98.33	33.91	487.3	0.9
4	34.152	0.011	0.00000	0.74	63.38	99.31	33.92	487.3	2.3
5	34.333	0.013	0.03045	1.628	67.35	98.56	33.84	486.3	3.7
6	34.304	0.015	0.00066	1.067	71.51	99.04	33.97	488.0	1.9
7	34.057	0.011	0.00000	0	79.71	99.95	34.04	488.9	2.2
8	34.116	0.012	0.01782	0.308	87.49	99.69	34.01	488.5	1.2
9	33.864	0.011	0.04417	0.046	88.52	99.92	33.84	486.3	3.3
10	34.07	0.012	0.00000	0.058	94.25	99.9	34.04	488.8	0.9
11	33.329	0.009	0.01615	0	96.75	99.95	33.31	479.7	1.4
12	34.258	0.011	0.00364	0	99.74	99.95	34.24	491.4	1.5
13	34.001	0.014	0.00000	0	100.00	99.95	33.99	488.2	11.5

AR-648 Palmital granite Biotite			J= 0.009143						
1	52.46	0.026	0.00710	63.792	4.90	64.04	33.59	483.2	3.9
2	40.419	0.018	0.00657	23.181	17.64	83.02	33.55	482.7	1.7
3	37.645	0.017	0.00113	13.793	28.62	89.13	33.55	482.7	2.0
4	38.948	0.016	0.00414	17.756	32.78	86.49	33.69	484.4	2.8
5	36.093	0.017	0.00274	8.039	47.76	93.38	33.7	484.6	1.3
6	35.885	0.014	0.02312	7.814	54.41	93.53	33.56	482.8	1.6
7	35.43	0.015	0.00000	5.172	65.86	95.64	33.89	486.9	2.7
8	35.595	0.015	0.00000	6.099	71.54	94.89	33.78	485.5	2.5
9	35.949	0.014	0.02017	7.55	75.61	93.75	33.7	484.6	2.4
10	35.756	0.013	0.07064	8.661	77.47	92.81	33.19	478.1	3.8
11	35.479	0.016	0.02246	7.019	80.25	94.11	33.39	480.7	2.5
12	35.548	0.015	0.00000	6.866	82.88	94.25	33.5	482.1	2.9
13	35.359	0.016	0.00000	5.228	88.15	95.59	33.8	485.8	1.6
14	35.204	0.015	0.01110	4.678	91.77	96.03	33.81	485.9	2.2
15	32.629	0.012	0.01785	0	94.27	99.95	32.62	470.8	2.1
16	32.112	0.014	0.03756	0	96.17	99.95	32.1	464.3	3.8
17	36.991	0.015	0.00000	10.295	96.92	91.73	33.93	487.5	4.5
18	126.985	0.072	0.00000	320.696	100.00	25.36	32.2	465.6	11.8

Supplementary Material #4;

Ar- Ar results for samples from the Eastern Anatectic Unit

Method:

Samples from the Eastern Anatectic Unit were processed and analysed at the Geochronological Research Center of the University of São Paulo (Brazil). Separation and concentration consisted of hand crushing of centimetre-scale sample chips using steel pestle and mortar to reduce size to mm-scale, sieving for size fractions of #250-500, and final magnetic purification through a isodynamic Frantz® separator. Biotite was loaded into a 21-pit aluminum disk along with the Fish Canyon sanidine standard (FCs) arranged in a cross-like geometry. Neutron-flux irradiation was performed in the cadmium lined in-core (CLICIT) facility at the Oregon State University TRIGA reactor for 14 hours. After radioactive cooling (approximately 4 weeks), each sample was loaded on a Cu-disk as multigrain aliquots containing two or three irradiated biotite flakes and progressively step-heated by a solid state $Nd:YVO_4$ (532 nm) Verdi 6W Coherent laser. Data reduction including correction for nuclear interferences, background (blank) and mass discrimination, as well J-value determination and graphical representation were made using the Ar-ArCALC v2.5.2 software (Koppers, 2012). The ages were calculated relative to the age of 28.01 ± 0.04 Ma for the FCs standard (Phillips and Matchan, 2013a; Phillips and Matchan, 2013b), assuming the ^{40}K decay constant of Steiger and Jäger (Steiger and Jäger, 1977). The J-factor value for this irradiation is 0.00406 ± 0.00002.

Results:

Step	^{39}Ar released %	$^{40}Ar/^{39}Ar$	1σ	$^{37}Ar/^{39}Ar$	1σ	$^{36}Ar/^{39}Ar$	1σ	$^{39}Ar_K(\%)$	$^{40}Ar^*(\%)$	$^{40}Ar^*/^{39}Ar_K$	2σ	Age (Ma)	2σ
AR-535 Anatectic granite biotite													
1	2.1	75.104	0.426	0.0949	0.0321	0.0037	0.0071	2.09	98.53	74.00	4.30	475.06	24.30
2	18.0	72.939	0.018	0.0183	0.0059	-0.0006	0.0007	15.81	100.23	73.11	0.45	470.01	2.54
3	42.8	71.218	0.091	0.0171	0.0022	0.0003	0.0003	24.87	99.88	71.13	0.26	458.80	1.49
4	63.1	71.345	0.044	0.0118	0.0033	0.0012	0.0006	20.45	99.50	70.99	0.39	457.97	2.23
5	76.3	71.988	0.118	0.0203	0.0057	0.0033	0.0010	13.12	98.65	71.02	0.64	458.14	3.67
6	87.6	71.965	0.087	0.0785	0.0042	0.0008	0.0007	11.28	99.70	71.75	0.48	462.30	2.72
7	91.2	73.069	0.295	-0.1005	0.0072	0.0055	0.0016	3.70	97.75	71.42	1.11	460.44	6.30
8	97.5	72.664	0.144	-0.0329	0.0081	0.0027	0.0023	6.25	98.90	71.86	1.40	462.94	7.98
9	99.1	75.197	0.942	-0.0306	0.0213	0.0148	0.0063	1.67	94.14	70.79	4.17	456.83	23.78
10	100.0	82.607	0.533	-0.1886	0.0588	0.0009	0.0122	0.77	99.67	82.32	7.39	521.44	40.64
AR-664 Anatectic granite biotite													
1	2.8	74.623	0.162	0.0167	0.0195	0.0100	0.0034	2.74	95.98	71.63	2.04	461.60	11.59
2	23.2	73.705	0.100	0.0305	0.0007	-0.0020	0.0005	20.57	100.81	74.30	0.35	476.74	1.97
3	42.6	74.270	0.041	-0.0067	0.0049	-0.0007	0.0009	19.21	100.27	74.47	0.54	477.71	3.03
4	61.5	73.577	0.084	0.0019	0.0048	0.0008	0.0003	19.04	99.68	73.34	0.23	471.35	1.30
5	74.3	73.410	0.062	0.0058	0.0020	-0.0025	0.0010	12.68	101.02	74.16	0.64	475.97	3.60
6	85.2	73.691	0.098	-0.0301	0.0032	0.0012	0.0016	11.24	99.51	73.33	0.96	471.26	5.41
7	95.4	74.157	0.092	0.0581	0.0035	-0.0031	0.0009	9.94	101.24	75.08	0.58	481.15	3.24
8	97.3	73.385	0.774	-0.0296	0.0281	-0.0071	0.0049	2.05	102.88	75.50	3.35	483.48	18.83
9	99.6	78.937	0.837	-0.1303	0.0543	0.0023	0.0083	2.10	99.12	78.23	5.23	498.79	29.13
10	100.0	68.647	0.102	-0.1451	0.0947	0.0889	0.0156	0.44	61.33	42.09	9.31	285.31	58.38
AR1133 Anatectic granite biotite													
1	2.0	68.546	0.747	-0.0401	0.0538	0.0601	0.0021	1.85	73.83	50.60	1.62	337.89	9.85
2	14.1	76.158	0.025	-0.0042	0.0058	0.0021	0.0024	12.37	99.16	75.52	1.42	483.60	7.99
3	32.7	75.626	0.023	-0.0028	0.0024	-0.0009	0.0007	18.45	100.34	75.88	0.43	485.65	2.41
4	51.4	75.885	0.092	-0.0048	0.0024	-0.0025	0.0003	18.80	101.00	76.64	0.28	489.91	1.56
5	71.8	75.560	0.031	-0.0080	0.0008	-0.0011	0.0007	19.26	100.41	75.87	0.43	485.59	2.41
6	89.9	76.085	0.096	0.0030	0.0023	0.0001	0.0007	19.18	99.95	76.05	0.48	486.58	2.67

7	92.0	76.922	0.539	-0.0123	0.0164	0.0013	0.0057	2.09	99.49	76.53	3.57	489.26	20.02
8	94.5	77.429	0.277	0.0212	0.0221	-0.0021	0.0049	2.56	100.79	78.04	3.01	497.74	16.76
9	95.1	96.414	1.689	-0.0442	0.1504	-0.0416	0.0288	0.47	112.86	108.81	17.61	661.62	89.66
10	100.0	76.783	0.074	-0.0160	0.0135	0.0044	0.0026	4.72	98.28	75.46	1.56	483.27	8.78
AR-1145 Diatexite biotite													
1	3.8	76.081	0.461	0.0267	0.0152	0.0062	0.0045	3.76	97.58	74.24	2.84	476.40	15.99
2	19.7	74.885	0.098	0.0021	0.0036	0.0010	0.0008	16.04	99.61	74.59	0.49	478.38	2.78
3	41.7	74.068	0.056	0.0071	0.0018	-0.0005	0.0008	21.89	100.20	74.22	0.47	476.27	2.66
4	58.6	74.312	0.095	0.0352	0.0001	0.0008	0.0009	16.87	99.67	74.07	0.58	475.44	3.25
5	74.1	74.438	0.092	-0.0017	0.0032	-0.0002	0.0002	15.51	100.06	74.49	0.22	477.79	1.23
6	88.0	74.188	0.079	-0.0054	0.0063	0.0006	0.0003	13.97	99.75	74.00	0.26	475.07	1.45
7	94.9	74.099	0.070	-0.0330	0.0026	-0.0007	0.0024	6.67	100.29	74.31	1.46	476.81	8.26
8	97.2	78.537	0.256	0.0831	0.0015	0.0071	0.0031	2.62	97.31	76.43	1.91	488.71	10.68
9	99.1	75.991	0.233	-0.0308	0.0168	-0.0055	0.0056	1.92	102.16	77.63	3.37	495.42	18.82
10	100.0	103.780	3.899	-0.0479	0.2651	-0.0895	0.0691	0.22	125.76	130.51	42.35	768.85	203.22
AR-1349 Metatexite biotite													
1	2.6	75.148	0.048	0.0038	0.0193	0.0088	0.0044	2.41	96.52	72.53	2.61	466.73	14.82
2	18.7	73.699	0.043	-0.0035	0.0039	0.0003	0.0007	16.17	99.86	73.60	0.45	472.77	2.57
3	40.0	74.334	0.030	-0.0001	0.0033	0.0006	0.0003	21.43	99.77	74.16	0.18	475.97	1.04
4	57.9	73.908	0.031	0.0006	0.0036	0.0015	0.0008	17.85	99.41	73.47	0.48	472.07	2.73
5	74.3	73.914	0.012	-0.0028	0.0015	-0.0012	0.0005	16.36	100.50	74.28	0.29	476.65	1.64
6	86.4	74.232	0.069	-0.0049	0.0037	0.0003	0.0009	12.17	99.87	74.13	0.58	475.81	3.27
7	95.2	73.989	0.040	0.0236	0.0022	-0.0006	0.0006	9.04	100.24	74.17	0.39	475.98	2.18
8	97.6	73.159	0.180	-0.0378	0.0181	-0.0065	0.0038	2.19	102.65	75.09	2.30	481.21	12.97
9	100.0	78.291	0.116	-0.0526	0.0121	0.0012	0.0018	2.37	99.52	77.91	1.12	497.00	6.22
AR-1368 Anatectic granite biotite													
1	3.1	68.182	0.009	-0.0421	0.0207	-0.0045	0.0055	3.11	101.96	69.52	3.30	449.57	18.87
2	22.3	71.413	0.029	-0.0048	0.0052	-0.0005	0.0011	19.35	100.20	71.56	0.69	461.21	3.92
3	44.1	72.785	0.110	-0.0071	0.0048	-0.0005	0.0006	21.88	100.21	72.94	0.41	469.04	2.35
4	58.8	72.960	0.067	-0.0014	0.0039	0.0001	0.0009	14.26	99.98	72.94	0.56	469.08	3.18
5	69.9	72.692	0.061	-0.0072	0.0089	-0.0033	0.0005	11.30	101.34	73.67	0.30	473.16	1.71
6	79.9	73.675	0.053	-0.0049	0.0073	0.0025	0.0007	9.95	99.00	72.94	0.40	469.06	2.29
7	88.5	72.897	0.099	0.0083	0.0088	0.0034	0.0007	9.67	98.63	71.90	0.45	463.14	2.53
8	95.2	73.578	0.007	-0.0601	0.0050	0.0050	0.0019	5.59	97.95	72.07	1.11	464.11	6.28
9	99.2	71.936	0.090	-0.0225	0.0173	0.0015	0.0032	4.26	99.39	71.49	1.92	460.85	10.90
10	100.0	72.199	0.996	-0.1665	0.0926	0.0304	0.0163	0.63	87.41	63.10	9.86	412.45	57.65

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