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## AUTHOR STATEMENT

**Cunha, I.R.V:** Conceptualization, methodology, petrography, and mineral chemistry, writing the manuscript and preparing the illustrations. **Dall’Agnol, R:** Conceptualization, supervision, reviewing, and editing. **Scaillet, B:** Supervision, reviewing, and editing. **Sousa, L.A.M:** Software - Map preparation, collaboration in petrography and iron oxides mineralogy.

Journal Pre-proof

# MAGMATIC EPIDOTE IN ARCHEAN GRANITOIDS OF THE CARAJÁS PROVINCE, AMAZONIAN CRATON, AND ITS STABILITY DURING MAGMA RISE AND EMPLACEMENT

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## ABSTRACT

In the Rio Maria and Sapucaia Domains of the Carajás Province, during the Mesoarchean, tonalite-trondhjemite-granodiorite (TTG) series (2.98-2.92 Ga; Colorado Trondhjemite) and sanukitoid rocks (~2.87 Ga; Rio Maria Suite of Ourilândia do Norte) were formed. They were followed, in the Neoproterozoic (2.75-2.73 Ga), by numerous stocks of granitoids similar to A-type granites (Vila Jussara Suite). Despite the compositional differences between these granitoids, epidote is a mineral phase common to all of them, with pistacite contents of 26-29 mol% in TTGs, 22-33 mol% in sanukitoids and 25-30 mol% in Neoproterozoic granitoids. The study of the dissolution kinetics of Archean epidotes of the Carajás Province reveals that the partial dissolution time of their crystals was ~4-10 years, with corresponding magma ascent rates of 4-8 km/year. Magma viscosities at liquidus temperature were estimated at  $10^{5.3}$  Pa s for TTG magma and  $10^{2.5}$  Pa s for sanukitoid magma, whereas monzogranitic magmas of the Vila Jussara Suite exhibited a viscosity of  $10^{4.4}$  Pa s. In contrast, the viscosity of tonalitic magma of the Vila Jussara Suite was  $10^{3.5}$  Pa s. The preservation of magmatic epidote in Archean granitoids requires that the plutons grew in an incremental way, similar to their Phanerozoic counterparts, with the stacking of small sills-dykes of about 100 m thickness each intruded in a rather cold crust, allowing fast cooling rates so as to prevent complete dissolution of epidote at the final emplacement level.

**Keywords:** magmatic epidote, Archean, Carajás Province, dissolution, viscosity, incremental growth

## 1. INTRODUCTION

In the Archean, the upper mantle temperature was estimated to be approximately 1600°C, near ~250°C above its current average temperature (Sizova et al., 2015). This was

37 due to the higher production of radiogenic heat and drastically affected the thickness and  
38 composition of the crust (Sizova et al., 2015 and references therein). The onset of the  
39 continental crust formation resulted from the cooling and differentiation of the Earth and was  
40 combined with continuous magma extraction from the mantle. This caused changes in the  
41 crustal growth rates and resulted in an increase of crustal volume. These processes began in  
42 the early Archean but were intensified at the end of the Mesoarchean and throughout the  
43 Neoproterozoic (Taylor and McLennan, 1985; Kramers and Tolstikhin, 1997; Belousova et al.,  
44 2010; Dhuime et al., 2011; Guitreau et al., 2012) when, in addition to tonalite-trondhjemite-  
45 granodiorite (TTG) associations, larger volumes of sanukitoids, hybrid granitoids and potassic  
46 granites were formed, resulting in rocks with more diversified geochemical signatures  
47 (Martin, 1987; Moyen et al., 2003; Jayananda et al., 2006; Barros et al., 2009; Feio et al.,  
48 2012; Moyen and Martin, 2012; Almeida et al., 2013; Laurent et al., 2014; Sizova et al., 2015;  
49 Bédard, 2017; Dall'Agnol et al., 2017; Silva et al., 2018). This geochemical variability is  
50 evidenced by marked variations in the continental crust. It occurred between 3.0 and 2.5 Ga,  
51 when more than 50% of the current continental crust volume was formed (Bédard, 2017).

52 TTGs are the oldest and largely dominant felsic plutonic association in Archean times.  
53 Accordingly, their origin and relationship with tectonic regimes have been widely discussed  
54 (Moyen and Martin, 2012; Sizova et al., 2015; Moyen and Laurent, 2018; Smithies et al.,  
55 2019). Conversely, high-Mg monzodiorites and granodiorites, known as sanukitoids, are less  
56 voluminous components of late Archean terranes, formed from interactions between the  
57 peridotitic mantle and a component rich in incompatible elements (Martin et al., 2005;  
58 Oliveira et al., 2010, 2011; Heilimo et al., 2010; Laurent et al., 2014; Sizova et al., 2015).  
59 Considerable volumes of leucogranodioritic and granitic rocks were also formed in the final  
60 stages of the Archean (Jayananda et al., 2006; Almeida et al., 2013, 2017; Silva et al., 2018).  
61 This process of granitoid magma migration enabled the transfer of heat and mass from the  
62 lower to upper crust, resulting in the chemical differentiation of the continents and thermal  
63 and mechanical interactions during magma rise and emplacement (Brown and Solar, 1988).

64 In this context, it is relevant to determine variations of the viscosity of silicate magmas,  
65 because this parameter plays a key role in the transport dynamics, type of eruption and rates  
66 of physicochemical processes of natural magmas (Giordano et al., 2008), being crucial for a  
67 quantitative understanding of igneous processes (Persikov, 1991). The viscosity strongly  
68 depends on the composition of the melt, particularly its SiO<sub>2</sub> content, which determines its

69 degree of polymerization, but also on temperature, content and nature of volatile species,  
70 cation types, among other factors (Cruden and Weinberg, 2018).

71 The Carajás Province (CP) presents a wide diversity of granitoid rocks formed from the  
72 Mesoarchean to the Neoproterozoic (~3.0 to 2.57 Ga). The present study focuses on three  
73 different types of granitoids that occur in the Sapucaia Domain of this province and at its  
74 border with the Rio Maria Domain. Such granitoid rocks include tonalites and trondhjemites  
75 of TTG associations (Santos et al., 2013; Teixeira et al., 2013; Silva et al., 2014; Leite-Santos  
76 and Oliveira, 2014) and sanukitoids (Gabriel and Oliveira, 2014; Santos and Oliveira, 2016;  
77 Silva et al., 2018), both of Mesoarchean age (2.93 a 2.85 Ga), as well as ferroan to magnesian  
78 Neoproterozoic granitoids (~2.75 to 2.73 Ga), geochemically akin to A-type granites (Dall'Agnol  
79 et al., 2017; Silva et al., 2020).

80 Despite the geochemical and petrological differences among TTGs, sanukitoids and  
81 Neoproterozoic granitoids, magmatic epidote is a common mineral phase in these rocks. It is  
82 known that epidote stability strongly depends on the physicochemical parameters prevailing  
83 during magma evolution (Brandon et al., 1996; Keane and Morrison, 1997; Sial et al., 2008;  
84 Brasilino et al., 2011; Nagar and Korakoppa, 2014; Long et al., 2019). Hence, this study aims  
85 at, on the basis of the textural and chemical-mineralogical characterization of the epidote  
86 present in Mesoarchean and Neoproterozoic granitoids of the CP, estimating epidote partial  
87 dissolution time scale, the ascent rates and initial viscosity of each granitoid magma, and the  
88 ability of the magma to preserve epidote during pluton emplacement and crystallization in  
89 upper crust. Our study allows us to more accurately assess the physicochemical conditions of  
90 the formation of these granitoid magmas and to better understand the evolution of magmatic  
91 epidote. In the following, we start by a brief overview of factors controlling epidote stability  
92 in granitoid magmas, and then present the data obtained in the study area and their  
93 implications.

## 94 **2.STABILITY OF EPIDOTE IN MAGMAS AND TEXTURAL CRITERIA FOR ITS** 95 **IDENTIFICATION**

96 Epidote is a key indicator of variations in crystallization parameters (Schmidt and  
97 Thompson 1996) because it is extremely sensitive to changes in P-T- $fO_2$  conditions and it  
98 may be destabilized and completely dissolved during emplacement (Long et al., 2019). Based  
99 on experiments conducted over the past few decades, the role of epidote during magmatic  
100 crystallization is relatively well understood, and experimental results have confirmed the

101 paragenesis of natural rocks produced under different P-T- $f_{O_2}$  conditions (Schmidt and Poli,  
102 2004).

103 Experimental studies demonstrate the stability of epidote above 300 MPa; however,  
104 depending on the composition and oxygen fugacity, the minimum pressure for epidote  
105 crystallisation may vary (Schmidt and Poli, 2004). Crawford and Hollister (1982), based on  
106 the intersection of the solidus curve with that of epidote (Liou, 1973), proposed that in H<sub>2</sub>O-  
107 saturated granites, magmatic epidote requires a minimum pressure of ~600 MPa. In  
108 granodioritic and tonalitic magmas, epidote is stable at pressures between 800 and 1000 MPa  
109 at temperatures near the solidus curve (Naney 1983; Liou 1993; Schmidt and Thompson,  
110 1996). However, epidote may crystallise at pressures as low as ~300 MPa at  $f_{O_2}$  near the  
111 Hematite-Magnetite buffer, while at Ni-NiO, it is only stable at pressures  $\geq 500$  MPa (Naney,  
112 1983; Schmidt and Thompson, 1996). In general, epidote stability increases under more  
113 oxidizing  $f_{O_2}$  conditions (Liou, 1973; Schmidt and Thompson, 1996; Poli and Schmidt, 2004;  
114 Schmidt and Poli, 2004; Oliveira et al., 2010), and the pistacite contents characteristic of  
115 magmatic epidotes observed in experiments are mainly found between hematite-magnetite  
116 (HM) and nickel-NNO buffers (Liou, 1973). Oscillatory zones in epidote crystals may reflect  
117 changes in composition or in oxygen fugacity (Franz and Liebscher, 2004; Pandit et al.,  
118 2014). At low-pressures, outside the stability field of epidote, magnetite is usually the main  
119 Fe<sup>3+</sup>-containing phase (Drinkwater et al., 1991) and its modal content increases in rocks  
120 without epidote crystallization (Hammarstrom and Zen 1992; Schmidt, 1993).

121 The following equilibrium appears to control epidote stability in granitoids (Schmidt  
122 and Poli, 2004):

123 hornblende + K-feldspar + plagioclase + magnetite + H<sub>2</sub>O = epidote + biotite + quartz

124 Archean granitoids described below exhibit textural facies consistent with the  
125 concomitant formation of epidote and biotite, both of which are magmatic. This relationship  
126 can be explained by the peritectic reaction below (e.g., Sial et al. 1999).

127 plagioclase + amphibole + melt = biotite and epidote

128 Experimental studies in tonalitic compositions show that epidote + biotite can be  
129 formed through peritectic hornblende resorption without completely consuming the  
130 hornblende in the system, decreasing only its abundance proportionally to the modal increase  
131 in biotite + epidote (Schmidt and Poli, 2004 and references therein).

132 Textural and compositional criteria used to distinguish between magmatic and  
 133 secondary epidotes, include (1) euhedral to subhedral crystals with prismatic or hexagonal  
 134 shapes; (2) mineralogical associations with primary hornblende and biotite; (3) epidote with  
 135 oscillatory zoning, allanite core and irregular contact with quartz and plagioclase; (4) pistacite  
 136 content ranging from 25 to 35%; and (5) TiO<sub>2</sub> content ≤ 0.137% (Liou, 1973; Tulloch, 1979;  
 137 Zen and Hammarstrom, 1984; Zen, 1985; Evans and Vance, 1987; Dawes and Evans, 1991;  
 138 Schmidt and Poli, 2004; Sial et al., 2008).

139 In granitic systems, allanite crystallizes at temperatures significantly above solidus, it  
 140 can be later shielded by epidote rims and finally isolated epidote crystals can also occur  
 141 (Gieré and Sorensen, 2004). Dawes and Evans (1991) proposed that allanite cores surrounded  
 142 by epidote form due to magma impoverishment in rare earth elements during crystallization,  
 143 which inhibits allanite crystallization and enables epidote epitaxial growth. However, the  
 144 oxidation state of Fe when inserted in the crystalline structure of allanite is controlled by the  
 145 reaction:



147 Based on this reaction, Gieré and Sorensen (2004) suggested that the stability of allanite  
 148 in magmas may be more dependent on T and  $f\text{H}_2$  than on the content of rare earth elements.

### 149 **3.REGIONAL GEOLOGY**

150 Located in the southeast section of the Amazonian craton, the Carajás Province is one of  
 151 the main Archean terranes of this craton. The CP is inserted in the Archean domain of the  
 152 Central Amazonian Province (Tassinari and Macambira 2004), or alternatively, according to  
 153 the model proposed by Santos (2003), it represents an independent Archean province (Fig.  
 154 1a). In recent decades, various tectonic segmentation models have been proposed for the CP  
 155 (Souza et al., 1996; Althoff et al., 2000; Dall'Agnol et al., 2006, 2013). Vasquez et al. (2008),  
 156 in line with Santos (2003), proposed naming the southern section of the province as the Rio  
 157 Maria Domain (RMD) and its central and northern sections as the Carajás Domain (CD). In a  
 158 more recent model, Dall'Agnol et al. (2013) maintained the RMD but subdivided the CD into  
 159 the Sapucaia Domain (SD), Canaã dos Carajás Domain (CCD) and Carajás Basin (CB). These  
 160 three domains would have evolved differently from the RMD (Fig. 1b). This interpretation  
 161 was not followed by Costa et al. (2020), who suggested that dome-and-keel structures  
 162 controlled the evolution of both the RMD and the southern section of the old CD, which  
 163 would be an extension of that domain.



164 The Rio Maria Domain is a Mesoarchean terrane (3.0 to 2.86 Ga), predominantly  
165 consisting of greenstone belt sequences (3.0 to 2.90 Ga; Macambira and Lancelot, 1996;  
166 Souza et al., 2001) and TTG granitoids (2.98 to 2.93 Ga; Almeida et al. 2011, 2013), in  
167 addition to sanukitoids, high Ba-Sr granitic-to-leucogranodioritic suites and potassic  
168 leucogranites (2.87-2.86 Ga; Althoff et al., 2000; Dall'Agnol et al. 2006; Almeida et al., 2010,  
169 2013; Oliveira et al., 2011; Silva et al. 2018).

170 The Sapucaia Domain (2.95 to 2.73 Ga) comprises TTG rocks (Santos et al., 2013;  
171 Leite-Santos and Oliveira, 2014; Silva et al., 2014), sanukitoids (Gabriel and Oliveira, 2014;  
172 Gabriel et al., 2014), high Ba-Sr leucogranodiorites and potassic leucogranites (Teixeira et al.,  
173 2013; Leite-Santos and Oliveira, 2016), showing strong lithological similarities to the RMD  
174 (Fig. 1c). However, SD rocks were intensely deformed during the Neoproterozoic (2.75-2.73 Ga)  
175 and intruded by granitoids of the Vila Jussara Suite (Dall'Agnol et al., 2017; Silva et al.,  
176 2020) and potassic leucogranites (Leite-Santos and Oliveira, 2016), both of Neoproterozoic age.

177 The Canaã dos Carajás Domain consists of granites *sensu stricto* and charnockite  
178 associations (Fig. 1c). According to Feio et al. (2013), the CCD presents four main magmatic  
179 events, three of Mesoarchean age and one of Neoproterozoic age. The protolith of the Pium  
180 Complex (Pidgeon et al., 2000), the Bacaba Tonalite (~3.0 Ga; Moreto et al., 2011) and other  
181 rocks with similar ages were formed in 3.05-3.0 Ga. The Canaã dos Carajás Granite and the  
182 oldest rocks of the Rio Verde Trondhjemite were formed between 2.96 and 2.93 Ga. The  
183 Campina Verde Tonalite Complex, the Rio Verde Trondhjemite and the Cruzadão, Bom  
184 Jesus, Boa Sorte and Serra Dourada Granites crystallized from 2.87 to 2.83 Ga. The  
185 subalkaline granites of the Planalto Suite, the sodic granites of the Pedra Branca Suite and the  
186 charnockitic rocks of the Pium Complex were formed in the Neoproterozoic between ~2.75 and  
187 2.73 Ga (Feio et al., 2012; Santos et al., 2013; Rodrigues et al., 2014).

#### 188 **4. GEOLOGY, PETROGRAPHY AND GEOCHEMISTRY OF THE SELECTED** 189 **MESOARCHEAN AND NEOARCHEAN GRANITOIDS**

##### 190 **4.1 TONALITE-TRONDHJEMITE ASSOCIATION OF SAPUCAIA**

191 The Mesoarchean magmatism (2.93 to 2.85 Ga) exposed in the SD is marked by  
192 abundant TTG granitoids, predominantly trondhjemitic rocks with subordinate tonalites and  
193 granodiorites, grouped stratigraphically as undifferentiated TTGs and Colorado Trondhjemite  
194 (2.87 Ga; Silva et al., 2014). These TTG granitoids are distributed from Água Azul do Norte  
195 to Sapucaia, at the border between the Amazonian craton and the Araguaia Belt (Fig. 1c;  
196 Santos et al., 2013b; Silva et al., 2014; Leite-Santos and Oliveira, 2014). Structurally, these

197 rocks show E-W penetrative foliation and shear zones in the same direction. These rocks also  
198 exhibit compositional banding, S-C structures, drag folds and typical features of plastic flow,  
199 mostly indicating sinistral kinematics. In general, such rocks show porphyroclastic textures  
200 resulting from an intense recrystallization and mineral comminution process, marked by  
201 mylonitic-to-ultramylonitic tectonic fabrics (Santos et al., 2013b).

202 In mineralogical terms, the Sapucaia TT essentially consists of plagioclase and quartz,  
203 with highly subordinate microcline and with biotite as the main mafic phase, accompanied by  
204 epidote, opaques, allanite, titanite, apatite and zircon as accessory minerals. Geochemically,  
205 the granitoids are calc-alkaline with generally peraluminous compositions, no significant Sr  
206 and Eu anomalies, silica content ranging from 59 to 74 wt. %, relatively low ferromagnesian  
207 content ( $\text{Fe}_2\text{O}_3 + \text{MgO} + \text{MnO} + \text{TiO}_2 < 5$  wt. %), moderate Mg# (0.27-0.47) and high alumina  
208 values ( $\text{Al}_2\text{O}_3 > 15$  wt. %) (Santos et al., 2013b). The samples selected for analysis derive  
209 from the eastern section of the SD (Fig. 1c) and belong to the Colorado Trondhjemite. Santos  
210 et al. (2013b) reported the presence of magmatic epidote in these rocks and conducted a  
211 preliminary mineralogical study based on scanning electron microscopy and EDS analyses.

#### 212 4.2 OURILÂNDIA DO NORTE SANUKITOID

213 The Ourilândia do Norte area is located near the borders of the Rio Maria, Canaã dos  
214 Carajás and Sapucaia domains in the CP. This area is marked by a diverse Mesoarchean  
215 magmatism, mainly characterized by leucogranites, sanukitoids and associated rocks (Santos  
216 and Oliveira, 2016; Silva et al. 2018). The Mesoarchean sanukitoids of the Ourilândia do  
217 Norte region are identified as a composite body, predominantly consisting of granodiorites  
218 and equigranular tonalites, with subordinate quartz diorite and quartz monzodiorite, which  
219 form elongated and deformed lenses along the E-W direction.

220 In petrographic terms, the sanukitoid rocks exhibit well-preserved igneous textures, but  
221 are also marked by protomylonitic to mylonitic tectonic fabrics. The mineral assemblage  
222 essentially consists of quartz, plagioclase, various proportions of K-feldspar and mafic  
223 clusters formed by amphibole, biotite and epidote. The main accessory minerals consist of  
224 epidote (zoisite and clinozoisite), allanite, titanite, zircon, apatite, magnetite and ilmenite.  
225 Geochemically, sanukitoids and associated rocks exhibit a highly variable  $\text{SiO}_2$  content,  
226 moderate  $\text{K}_2\text{O}$  values and high Mg#, Cr and Ni values. They have a predominantly  
227 metaluminous character and a calc-alkaline to high-K calc-alkaline signature (Santos and  
228 Oliveira, 2016; Silva et al., 2018). Two textural types of magmatic epidote were  
229 distinguished: crystals with zoned or unzoned allanite cores and euhedral to subhedral crystals

230 associated with biotite and hornblende, partly surrounded by them (Santos and Oliveira,  
231 2016).

#### 232 4.3 VILA JUSSARA SUITE

233 The Vila Jussara Suite (VJS) granitoids occur in the central and northern sections of the  
234 SD, stretching northward to the vicinity of the CCD and eastward to the border between the  
235 Amazonian craton and the Araguaia Belt (Figs. 1b, c). The Neoproterozoic granitic bodies  
236 (~2.75-2.73 Ga) of the VJS intrude the Mesoproterozoic units (Colorado Trondhjemite and  
237 Pantanal Leucogranodiorite; Teixeira et al., 2013; Silva et al., 2014; Dall'Agnol et al., 2017).  
238 These plutons are associated with shear zones and elongated in the E-W direction. Field,  
239 petrographic and microstructural evidence reveal that the VJS rocks were affected by  
240 heterogeneous deformational processes of varying intensity, exhibiting structures ranging  
241 from protomylonitic to mylonitic (Silva et al., 2020).

242 Four main petrographic varieties were identified in the VJS: biotite-hornblende  
243 monzogranite (BHMzG), biotite-hornblende tonalite (BHTnl), biotite monzogranite (BMzG)  
244 and hornblende-biotite granodiorite (HBGd). The different rock varieties of the VJS are  
245 characterized by a similar mineralogy, from essential tectosilicates (quartz, plagioclase and  
246 potassic feldspar) to mafic minerals (hornblende and/or biotite), and by the main primary  
247 accessory minerals, ilmenite, magnetite, apatite and zircon, which can also include titanite,  
248 allanite and epidote. Geochemically, the VJS consists of ferroan granitoids (reduced and  
249 oxidized A-type granites) and mildly magnesian granites (cf. Dall'Agnol et al., 2017; Sousa,  
250 2019). The four petrographic varieties of the VJS were formed under different conditions of  
251 oxygen fugacity. The reduced to moderately reduced BHMzG was formed under conditions  
252 varying from  $< \text{FMQ}$  to  $\text{FMQ} + 0.5$ . The oxidized BHMzG was generated under moderately  
253 oxidizing conditions (between  $\text{NNO}$  and  $\text{NNO}-0.5$ ), whereas the magnesian granites were  
254 crystallized under oxidizing conditions, with  $f\text{O}_2$  ranging from  $\text{NNO}$  to  $\text{NNO}+1$  (Sousa,  
255 2019). Dall'Agnol et al. (2017) described the textural and compositional variations in the  
256 magmatic epidote of the specific VJS varieties.

#### 257 5. TEXTURAL ASPECTS OF EPIDOTE IN THE STUDIED GRANITOIDS

258 Typical magmatic epidote crystals are hexagonal-euhedral and greenish-pink to pinkish-  
259 yellow or colorless, with oblique extinctions (Dawes and Evans, 1991; Schmidt and Poli,  
260 2004). In granitoid rocks, magmatic epidote crystals are usually subhedral and generally  
261 associated with biotite, hornblende and plagioclase (Pandit et al., 2014).

262 In the Archean granitoids selected for analysis, epidote is a common accessory mineral  
263 phase, sometimes present in appreciable modal proportions. In the Sapucaia TT, biotite is the  
264 main mafic phase, and the epidote modal content ranges from trace to 4%. The epidote  
265 exhibits two textural aspects (Santos et al., 2013b): (a) euhedral to subhedral crystals with  
266 well-defined faces, usually with automorphic contacts with biotite and irregular contacts with  
267 quartz and feldspars (Fig. 2a); and (b) crystals involving allanite cores, either metamictized or  
268 preserved (Fig. 2b). The sanukitoids of Ourilândia do Norte (Santos and Oliveira, 2016) have  
269 a high amphibole content, and epidote is a significant accessory mineral (ranging from trace  
270 to 3.9% and reaching 6.2% in the epidote-biotite-amphibole diorite variety). Mafic aggregates  
271 consisting of hornblende  $\pm$  actinolite, biotite, epidote, titanite and Fe-Ti oxides are common.  
272 Epidote occurs as zoned crystals, with or without allanite cores, or as euhedral to subhedral  
273 crystals that are partly or completely surrounded by biotite and/or hornblende (Fig. 2c).

274 In the VJS, oxidized ferroan biotite-hornblende monzogranite is the predominant  
275 variety, whose epidote has a modal content  $\leq 1.0\%$ . In reduced ferroan rocks of this variety,  
276 epidote occurs in a smaller proportion, is absent from most samples and is predominantly  
277 found as anhedral and rarely subhedral crystals with signs of intense corrosion. In the biotite  
278 monzogranite variety, epidote is only absent from one sample and exhibits variable modal  
279 content, albeit generally  $<1\%$  and peaking at 2.7%. The hornblende-biotite tonalite group has  
280 a smaller exposure area than that of other VJS varieties, and when present, the magmatic  
281 epidote content ranges from 0.2 to 0.4%.

282 In general, the epidote present in the study granitoids is commonly found in association  
283 with mafic aggregates consisting of biotite  $\pm$  hornblende  $\pm$  zircon  $\pm$  apatite  $\pm$  allanite  $\pm$  titanite  
284  $\pm$  opaque minerals. Based on textural aspects, the following types of epidote can be  
285 distinguished: (a) euhedral to subhedral crystals that are predominantly associated with biotite  
286 that shares straight contacts with epidote (Figs. 2a, 4c, 4d); (b) epidote in rims around allanite  
287 (Figs. 2b, 4c, 4f); and (c) anhedral to subhedral epidote with irregular contours with  
288 amphibole and/or feldspars (Figs. 2c, 2e, 2f). When in contact with biotite, the original shape  
289 of the epidote is generally preserved (Figs. 2a, 4c, 4f). In contrast, when in contacts with  
290 hornblende and felsic minerals, epidote crystals show signs of marked corrosion (Fig. 2e),  
291 which sometimes almost completely destroys the original facies. Allanite is also an accessory  
292 mineral commonly found in these granitoids, generally exhibiting a subhedral to anhedral  
293 habit. It is sometimes zoned and generally totally or partly surrounded by epidote (Figs. 2b,

294 4c). Subordinately, epidote is identified as a post-magmatic phase, forming irregular grains  
295 that are preferentially associated with altered plagioclase.

296 The textural aspects observed in the Archean granitoid epidotes of the CP are similar to  
297 those interpreted by Zen and Hammarstrom (1984), Bédard (2003) and Schmidt and Poli  
298 (2004), as diagnostic of magmatic epidote. However, Keane and Morrison (1997) stressed out  
299 that the shape of crystals is insufficient to differentiate the magmatic or subsolidus origin of  
300 epidotes because the absence or development of euhedral faces is not exclusive to direct  
301 magmatic crystallization. Accordingly, only the combination of textural, compositional and  
302 isotopic criteria can reliably indicate the origin of an epidote crystal.

## 303 **6.MINERAL CHEMISTRY**

304 Epidote crystals with textural evidence of magmatic origin were analyzed on a JEOL  
305 JXA-8230 electron probe microanalyzer (with five spectrometers in wavelength-dispersive  
306 spectroscopy (WDS) mode and one in energy-dispersive spectroscopy (EDS) mode) at the  
307 Microanalysis Laboratory, Institute of Geosciences (*Instituto de Geociências – IG*), Federal  
308 University of Pará (*Universidade Federal do Pará – UFPA*). The analytical procedure was  
309 performed on polished thin carbon coated sections under the following operating conditions:  
310 15 kV voltage acceleration; 20 nA current; 5  $\mu\text{m}$  beam diameter; and 20 second analysis times  
311 for F, Na, Al, Cr, K, Sr, Ni, Mn, Ba, Ti, Cl, V, Y, Ce, and La, and 10 second analysis times  
312 for Mg, Si, Ca, and Fe. The crystals used for the analyses were LIF for Fe, Mn, Ba, and Ti;  
313 PETJ for Ca and K; TAP for Na, Mg, Al, and Si; and LIFH for Ce and La. The standards used  
314 for calibration were orthoclase ( $\text{SiO}_2$  and  $\text{K}_2\text{O}$ ), rutile ( $\text{TiO}_2$ ), anorthite ( $\text{Al}_2\text{O}_3$ ), magnetite  
315 ( $\text{FeO}$ ), diopside ( $\text{MgO}$ ), rhodonite ( $\text{MnO}$ ), wollastonite ( $\text{CaO}$ ), sodalite ( $\text{Na}_2\text{O}$ ) and  $\text{CePO}_4$   
316 ( $\text{Ce}_2\text{O}_3$  and  $\text{La}_2\text{O}_3$ ).

317 In total, 105 analyses were performed (Table 1), with 41 in the Sapucaia TT (PFA-14  
318 and PFR-33A); 42 in the Ourilândia do Norte sanukitoids (NDP-47 and NDP-101); and 22 in  
319 the VJS granitoids (ADE-01D, AFD-11A, and AMP-77). The VJS analyses were added to  
320 those obtained by Dall’Agnol et al. (2017).

321 Tulloch (1979) and Vyhnal et al. (1991) evaluated the origin of epidote based on its  
322 chemical composition and concluded that pistacite contents ( $P_s$ : molar  $[\text{Fe}^{3+}/\text{Fe}^{3+}+\text{Al}]/100$ )  
323 ranging from 25 to 29% (molecular weight) correspond to magmatic epidotes, differing from  
324 those formed by subsolidus alteration of plagioclase ( $P_{s0-24}$ ) and biotite ( $P_{s36-48}$ ). Experimental

325 studies on synthetic rocks (Liou, 1973) also show that pistacite contents ranging from 25 to  
326 35% correspond to magmatic epidote.

327 In Sapucaia TT, the pistacite (Ps: molar  $[\text{Fe}^{3+}/\text{Fe}^{3+}+\text{Al}]/100$ ) content ranges from 26 to  
328 29 mol%, with values primarily clustering between 27 and 28 mol%, thus suggesting a  
329 magmatic origin of the crystals (Fig. 3; Table 1). In the Ourilândia do Norte sanukitoids, the  
330 Ps content of epidote ranges from 22 to 33 mol%, with values concentrating from 29 to 30  
331 mol%, that is predominantly in the magmatic origin field as well (Fig. 3; Table 1). The VJS  
332 varieties has Ps values ranging from 25 to 30 mol% (Fig. 3; Table 1), corroborating the  
333 findings of Dall'Agnol et al. (2017) and in line with a magmatic origin. Oliveira et al. (2010)  
334 found Ps values ranging from 26 to 33 mol% for the epidotes from the sanukitoid Rio Maria  
335 Suite, mostly overlapping with those of their analogues from Ourilândia do Norte.

336 In addition to the molar Ps content, previous empirical and experimental studies  
337 (Tulloch, 1979; Vyhnal et al., 1991; Liou, 1993; Sial et al., 1999; Oliveira et al., 2010;  
338 Dall'Agnol et al., 2017) suggest that a  $\text{TiO}_2$  content in epidote  $\leq 0.137$  wt. % is in line with its  
339 magmatic formation. The Sapucaia TT and VJS granitoid epidotes display  $\text{TiO}_2$  values always  
340 lower than 0.137 wt. % (Table 1), corroborating the textural and compositional assessments  
341 above indicative of magmatic origin. In the Ourilândia do Norte sanukitoids, the epidotes  
342 have  $\text{TiO}_2$  values predominantly  $\leq 0.137$  wt. % (Table 1).

343 Thus, the Sapucaia TT and Vila Jussara granitoid epidotes with textural evidence of  
344 magmatic origin exhibit compositional variations compatible with magmatic origin (Fig. 3;  
345 Table 1). Comparatively, despite predominantly plotting in the magmatic field, the Ourilândia  
346 do Norte sanukitoids epidotes have subordinate compositions compatible with their formation  
347 from altered plagioclase (Fig. 3). In this regard, these epidotes diverge from those of the Rio  
348 Maria Suite sanukitoids, which only show compositions within the range assigned to  
349 magmatic epidotes (Oliveira et al., 2010).

350 Studies conducted so far on Neoproterozoic granitoid from the Carajás Province (CP)  
351 corroborate the results of experimental researches showing that epidote formation depends on  
352  $f\text{O}_2$  and pressure, as shown by the Estrela Complex, Serra do Rabo and Planalto Suite  
353 granites, which formed under strongly reduced conditions and lack epidote. Conversely, the  
354 Villa Jussara Suite (VJS) (Dall'Agnol et al., 2017) and Vila União (Marangoanha et al., 2018)  
355 granitoids, which are mostly formed under oxidizing conditions, generally present epidote as  
356 an important accessory phase. The epidote crystals of the VJS show no considerable  
357 variations in  $\text{Fe}^{3+}$  content between the core and rims of the crystals and rarely display zoning,

358 thus suggesting that in the final stages of crystallization, the residual melt was not  
359 impoverished in  $\text{Fe}^{3+}$  and that  $f\text{O}_2$  remained constant, thereby favoring epidote stability.

## 360 **7. EPIDOTE DISSOLUTION AS A MAGMA ASCENT RATE INDICATOR**

361 In general, epidote crystallization occurs near the solidus temperature at relatively high  
362 pressure (Schmidt and Poli, 2004). However, epidote phenocrystals in dikes in which the  
363 matrix, representative of the residual melt, accounts for approximately two-thirds of the rock  
364 volume were described (Dawes and Evans, 1991). The presence of epidote in granitic rocks  
365 that finished its crystallization at relatively low-pressure conditions indicates that epidote  
366 likely started forming at greater depths and then was transported quickly enough to prevent its  
367 complete resorption during magma rise (Brandon et al., 1996). The absence of epidote may  
368 also result from its complete dissolution by residual magmatic melt. In turn, the presence of  
369 allanite cores surrounded by epidote may be due to the magma impoverishment in rare earth  
370 elements during crystallization, which would inhibit allanite crystallization and enable epidote  
371 epitaxial growth (Dawes and Evans, 1991).

372 Experimental evidence shows that the intensity of epidote crystal dissolution is directly  
373 related to changes in intensive crystallization parameters during magma rise. Accordingly,  
374 magma transport rates have been estimated based on the degree of epidote dissolution  
375 (Brandon et al., 1996; Sial et al., 2008; Brasilino et al., 2011; Pandit et al., 2014). This method  
376 was used to measure the ascent rate of Archean granitoid magmas from the CP, based on the  
377 methodological criteria detailed below:

- 378 (1) Euhedral and subhedral epidote crystals allow us to more accurately infer the original  
379 shape of the crystal and therefore are prioritized when estimating the degree of  
380 dissolution. Generally, these crystals are associated with biotite, amphibole and  
381 residual magmatic melt, which are currently represented by quartz and feldspar  
382 crystals;
- 383 (2) In the absence of isolated epidote crystals, epidote crystals involving allanite cores in  
384 epitaxial growth can be used;
- 385 (3) The pistacite content of the selected epidote crystals should fall within the magmatic  
386 field (Ps: 25-35 mol%; Liou, 1973);
- 387 (4) Calculating the crystal corrosion intensity is crucial, and these calculations must be  
388 based on maximum extent of dissolution identified in the petrographic microscope and  
389 always with measurements perpendicular to the crystal faces affected by corrosion;

390 (5) The duration of the corrosion process is calculated using the minimum apparent  
 391 diffusion coefficient between tonalitic magma and epidote ( $5 \times 10^{-17} \text{ m}^2 \text{ s}^{-1}$  for Si, Al,  
 392 Ca and F) at a temperature of  $750^\circ\text{C}$ , which can also be applied to other granitoid  
 393 magmas, according to criteria formulated by Brandon et al. (1996), using the equations  
 394 below:

$$395 \quad d_z = [(D_{\text{app}} \times t)^{1/2}]$$

$$396 \quad t = d_z^2 / (5 \times 10^{-17})$$

397 with

398  $d_z$  = dissolution zone width (m);

399  $D_{\text{app}}$  = apparent diffusion coefficient ( $5 \times 10^{-17} \text{ m}^2 \text{ s}^{-1}$ ); and

400  $t$  = partial epidote dissolution time (s)

401 (6) The magma emplacement depth is estimated using the Al-in-hornblende geobarometer  
 402 and is compared with estimates based on the geological and structural context of the  
 403 study rocks. The magma generation depth is generally deduced from petrological  
 404 criteria;

405 (7) The magma transport rate is calculated using the ratio between the path length  
 406 (difference between magma generation depth and emplacement depth) and the average  
 407 corrosion time of the epidote exposed to melt. In the original proposal, the path length  
 408 corresponds to the difference between 10 kbar, initial epidote crystallization pressure,  
 409 and emplacement depth (Brandon et al., 1996):

$$410 \quad \text{Tr} = L_r / t$$

411  $\text{Tr}$  = transport rate (m/yr)

412  $L_r$  = path length (m), defined by the equation:

$$413 \quad L_r = (10 - P_e) \cdot 10^4 / 3 \text{ (m)}$$

414  $P_e$  = emplacement pressure (kbar)

415  $t$  = partial dissolution time do epidote (yr)

416 The dimensions and length of the maximum dissolution of epidote crystals were  
 417 measured on thin sections representative of the Archean varieties of interest (Fig. 4).  
 418 Subsequently, the resulting values were clustered, and the arithmetic mean was assumed to be  
 419 representative of each sample, enabling calculations of average epidote dissolution and  
 420 transport rates of each magma (Table 2).



421 The epidote dissolution and magma transport rates were calculated in the PFA-14  
422 sample for the Sapucaia TT (Figs. 4a, b; Table 2) and in the NDP-47 and NDP-101 samples  
423 for the sanukitoids from the Ourilândia do Norte region (Figs. 4c, d, e; Table 2). Granitoids  
424 from the VJS contain mineralogical and geochemical evidence of origin of the three study  
425 varieties from different magmas. Thus, the epidote crystal dissolution and magma transport  
426 rate were estimated separately in each variety. The biotite-hornblende monzogranite variety is  
427 represented by the MYF-77 (Fig. 4f) and PFR-16A (Fig. 4g) samples, biotite monzogranite is  
428 represented by the MDP-02A and LIF-04A samples (Figs. 4h, i, j) and biotite-hornblende  
429 tonalite is represented by the MDP-02E sample (Figs. 4k, l).

430 The emplacement pressure of the study granitoids is an important variable for  
431 calculating the magma transport rate. Dall'Agnol et al. (2017) proposed that the magmas that  
432 formed the VJS were generated under pressures between 800 and 1000 MPa, and  
433 emplacement pressures were estimated to range from 300 to 500 MPa based on Al-in-  
434 amphibole geobarometer and geological evidence; a value of 400 MPa is assumed here as the  
435 emplacement pressure of these granitoids. The Ourilândia do Norte sanukitoids exhibit ductile  
436 structures and evidence of penetrative deformation associated with mylonitic structures  
437 (Santos and Oliveira, 2016; Silva et al., 2018; Nascimento, 2020). Nascimento (2020), based  
438 on Al-in-hornblende geothermometer measurements, estimated an emplacement pressure of  
439 ~200 MPa for those sanukitoids. However, considering the structural features described in the  
440 present study, a higher emplacement pressure of ~350 MPa corresponding to the mesozone is  
441 estimated for the Ourilândia do Norte sanukitoids. The pressures during sanukitoids  
442 production in the Rio Maria region were estimated at 600 to 900 MPa by Oliveira et al.  
443 (2010), a range which we use in the present work. In regard to the Sapucaia TT, Santos et al.  
444 (2013b) demonstrated that these rocks show affinity to the TTG with high La/Yb and Sr/Y  
445 ratios of the CP, suggesting that they were derived from amphibolite garnet-based sources  
446 under high pressures (approximately 1500 MPa). Finally, based on the ductile deformational  
447 structures observed in these rocks, we estimate an emplacement pressure of ~500 MPa for the  
448 Sapucaia TT magma.

449 Using the equations proposed by Brandon et al. (1996), the average durations of  
450 epidote crystal dissolution of the study granitoids are estimated at ~4 years in the Sapucaia TT  
451 and ~6 years in the Ourilândia do Norte sanukitoids, whereas these periods are estimated at ~5  
452 years in bt-hb monzogranite, ~10 years in bt monzogranite and ~3 years in bt-hb tonalite for  
453 the VJS varieties (Table 2). The results show that the Sapucaia TT magma rose in the crust at

454 an ascent rate of ~4600 m/year, whereas the Ourilândia do Norte sanukitoid magma moved at  
 455 ~4300 m/year. Finally, in the VJS, the calculations show a comparatively higher magma  
 456 ascent rate for the three magmas (bt-hb monzogranite at ~8000 m/year; bt monzogranite at  
 457 ~7900 m/year; and bt-hb tonalite at ~8200 m/year; Table 2).

## 458 **8. DISCUSSION**

### 459 8.1 VISCOSITY AND LIQUIDUS TEMPERATURE IN THE STUDIED GRANITOID 460 MAGMAS

461 The emplacement of a granitoid body in the mid-to-upper crust is the end point of a  
 462 large-scale mass and energy exchange, which begins with magma generation and segregation  
 463 in the lower crust and is followed by magma ascent and the formation of plutons, batholiths  
 464 and stocks (Cruden, 1998). During magma ascent, the behavior of the silicate melt is  
 465 primarily determined by its viscosity, density, composition and volatile content, which  
 466 determine the magma ability to rise when combined with the geometric characteristics of the  
 467 conduit (see below). Among the physical properties, viscosity is one of the parameters with  
 468 the strongest effect on the production, transport and eruption of silicate magmas, varying in  
 469 the range  $10^{-1}$  to  $10^{14}$  Pa s (Giordano et al., 2008). This variation in viscosity by 15 orders of  
 470 magnitude determines the magmas' resistance to flow, which inversely affects its fluidity.

471 Experimental studies on the viscosity of silicate melts enabled Giordano et al. (2008)  
 472 to formulate a general model, which includes the effect of H<sub>2</sub>O and F, to predict the non-  
 473 Arrhenius Newtonian viscosity of silicate magmas, with the following equation:

$$474 \text{VFT } [\log \eta = A + B / (T (K) - C)],$$

475 where A is the upper temperature limit for the viscosity of all silica melts, which is -  
 476 4.55 (+0.2) (for example,  $\log \eta \sim 10^{-4.6}$  Pa s), and B and C are parameters that describe the  
 477 compositional control on viscosity (see Giordano et al., 2008). The rationale for the A  
 478 parameter being unique is that at super-liquidus temperatures, all silicate melts become highly  
 479 dissociated liquids regardless of their lower temperature structure and converge to lower  
 480 limiting value of viscosity, in accordance with experimental observations and theory. The B  
 481 and C terms embody 17 parameters (oxides content and various combinations of them,  
 482 including volatiles.) that allow to describe faithfully the viscous behavior of the vast majority  
 483 of terrestrial magma compositions, over fifteen log units of viscosity ( $10^{-1}$ - $10^{15}$  Pa s). The  
 484 model is regressed on more than 1770 experimental measurements, performed either at 1 bar  
 485 or at high pressure (ie with volatiles dissolved into the melt). Given that the compositions  
 486 dealt with here belong to common magma types, we believe that the application of Giordano

487 et al model retrieves the correct order of magnitude of melt viscosity, provided temperature  
488 and H<sub>2</sub>O content are well estimated, as discussed below.

489 The following representative samples were selected: the MYF-19A sample, containing  
490 69.3 wt. % SiO<sub>2</sub>, was chosen for the Sapucaia TT magma, whereas the viscosity of the  
491 Ourilândia do Norte sanukitoid magma was calculated based on the composition of the BRM-  
492 113B sample, with 65.7 wt. % SiO<sub>2</sub> (Table 3). For the VJS, the PFR-16, MDP-02A and MDP-  
493 02E samples with 71.4, 71.2 and 60 wt. % SiO<sub>2</sub> contents, respectively, were chosen as  
494 representatives of bt-hb monzogranite, bt monzogranite and bt-hb tonalite magmas,  
495 respectively (Table 3).

496 As the magmas of the studied granitoids were derived from different sources, the  
497 temperature ranges were estimated from experimental studies conducted on systems with  
498 similar compositions than those selected above, so as to constrain initial (liquidus) and final  
499 (solidus) crystallization temperatures. For the TTG, the most accepted mechanism for the  
500 origin of their parental magmas is the partial melting of hydrated metabasaltic rocks,  
501 including greenstones, amphibolites and eclogites, in different tectonic environments (Martin,  
502 1987; Winther, 1996; Condie, 2005; Clemens et al., 2006; Moyen and Martin, 2012). In  
503 particular, Clemens et al. (2006) proposed that TTG magmas are produced at temperatures  
504 between 900°C and 1000°C and that crystallization would start at ~880°C, which was the  
505 liquidus temperature estimated for Sapucaia TT magma. In turn, the absence of amphibole in  
506 these rocks suggests a low H<sub>2</sub>O content (Naney et al., 1983), which was estimated at 2%  
507 weight in the present study.

508 Sanukitoid suites are regarded as key components of late Archean terranes (2.95-2.54  
509 Ga), presenting geochemical characteristics similar to those of mantle- and crust-derived  
510 magmatic rocks (Moyen et al., 2003; Condie, 2005; Martin et al., 2005; Oliveira et al., 2010,  
511 2011) with higher temperatures compared to TTG. Oliveira et al. (2010) estimated a liquidus  
512 temperature of ~950°C and a H<sub>2</sub>O content of ~7% for the Rio Maria sanukitoids, which are  
513 used in this study for the Ourilândia do Norte sanukitoids.

514 The VJS varieties are derived from at least three magmas, two of monzogranitic  
515 composition (bt-hb monzogranite and bt monzogranite) and one tonalitic (bt-hb tonalite), but  
516 the temperatures estimated by Al-in-hornblende and zircon saturation thermometry  
517 (Dall'Agnol et al., 2017) indicate that the granitic and tonalitic magmas were formed at  
518 similar temperatures. Thus, a liquidus temperature of ~890°C and a solidus temperature of  
519 ~700°C was assumed for the three magmas that formed the VJS.

520 The calculated viscosity of each granitoid magma is listed in Table 4 and its variation as  
521 a function of temperature is shown on Fig. 5. The TT magma, with a relatively high content of  
522 silica and a low content of ferromagnesian components and water has a viscosity of  $\sim 10^{5.3}$  Pa  
523 s, which is the highest among all Archean magmas of this study (Fig. 5; Table 4). Conversely,  
524 the sanukitoid magma, which has a higher liquidus temperature ( $\sim 950^\circ\text{C}$ ), a lower silica  
525 content and a higher concentration of ferromagnesian constituents and water, shows a  
526 comparatively lower viscosity of  $\sim 10^{2.5}$  Pa s (Fig. 5; Table 4). Finally, both monzogranitic  
527 magmas of the VJS show a viscosity of  $\sim 10^{4.4}$  Pa s based on the liquidus temperature given  
528 their geochemical similarities, whereas the tonalitic magma of the same suite has a somewhat  
529 lower viscosity of  $\sim 10^{3.5}$  Pa s (Fig. 5; Table 4).

## 530 8.2 EFFECT OF PHYSICAL PROPERTIES ON MAGMA ASCENT AND EPIDOTE 531 DISSOLUTION

532 The compositional diversity of igneous rocks reflects the variation in processes and  
533 sources involved in magma generation (Pitcher, 1979; Huppert and Sparks, 1988; Whitney,  
534 1988; Patiño Douce, 1995; Brown, 2013; Scaillet et al., 2016; Cruden and Weinberg, 2018).  
535 Nevertheless, for decades, researchers have assumed that felsic magmas rose from deeper  
536 regions of the crust through diapirism essentially because the high viscosities assumed for  
537 these liquids seemingly precluded any form of rapid and channeled rise (Cruden and  
538 Weinberg, 2018), and also because of the broadly circular shape in map view of most granitic  
539 intrusions. However, field observations, experimental studies on the viscosity of magmatic  
540 liquids (Scaillet et al., 1996; Dingwell, 1999; Caricchi et al., 2007; Giordano et al., 2008) as  
541 well as theoretical considerations have demonstrated that felsic magmas can also quickly rise  
542 to different crustal levels through dikes, faults and shear zones (Clemens and Mawer, 1992;  
543 Petford et al., 1994, 2000; Oliveira et al., 2008; Cruden and Weinberg, 2018), in much the  
544 same way than more mafic magmas, such as basalts, do.

545 In this context, differences in physical properties and ascent rates between magmas may  
546 help to understand the textural differences observed in the epidotes of Archean granitoids of  
547 the CP. The millimetric dimension and euhedral character of the epidote grains suggest  
548 protracted periods of continuous granitic magma crystallization. Conversely, the textural  
549 evidence of corrosion and resorption indicates significant epidote dissolution during magma  
550 transport and crystallization in the crust due to changes in the physicochemical conditions  
551 (Pandit et al., 2014).

552 For the TT magma, with a generation pressure of 1500 MPa, a liquidus temperature  
 553 estimated at ~880°C, a low H<sub>2</sub>O content (~2%) and a relatively high silica content (70%  
 554 weight) markedly increase the viscosity, resulting in a magma rise rate of ~4600 m/year  
 555 (Table 2). However, such transport rates only correspond to the initial stages of magma rise  
 556 because during the process, the magma temperature may decrease, which will be reflected in  
 557 the increase in the degree of crystallinity and viscosity, thereby decreasing the magmatic flow.  
 558 In turn, sanukitoid magma, with an estimated generation pressure between 900 and 600 MPa  
 559 and emplacement in a syntectonic regime linked to transcurrent shear zones (Santos and  
 560 Oliveira, 2016; Silva et al., 2018; Nascimento, 2020), shows low viscosity (10<sup>2.5</sup> Pa s),  
 561 reflecting its high temperatures (Fig. 5; Table 4) and relatively high H<sub>2</sub>O content (7%, Table  
 562 3), with a relatively fast magma rise rate of ~4300 m/year. Moreover, sanukitoids have  
 563 epidote crystals with a more preserved shapes and a dissolution time estimated at ~6 years  
 564 (Table 2), whereas TT epidotes exhibit comparatively more intense corrosion and a shorter  
 565 dissolution time (~4 years; Table 2). In contrast, TT magma transport is fastest compared to  
 566 sanukitoid magmas analyzed in this study (Table 2). Therefore, for these two magmas, there is  
 567 no clear correlation between estimated magma ascent rates and mineral dissolution time.

568 Finally, the magma viscosity calculated for the VJS indicates that near the liquidus  
 569 temperature, the two monzogranitic magmas exhibit a similar viscosity (~10<sup>4.4</sup> Pa s), which  
 570 subtly diverges during crystallization (Fig. 5). As the temperature decreases, the viscosity of  
 571 bt monzogranite becomes slightly higher than that of bt-hb monzogranite (Table 4, Fig.5),  
 572 reflecting the higher magma water content of the latter, which is necessary to stabilize  
 573 hornblende and capable of depolymerizing the magma, thus reducing its viscosity.  
 574 Accordingly, bt-hb monzogranite exhibits an estimated partial epidote dissolution time of ~5  
 575 years, whereas bt monzogranite has a dissolution time of ~10 years, both with inferred ascent  
 576 rates of ~7900 m/year (Table 2). In turn, the tonalitic magma has the lowest viscosity among  
 577 the suite granitoids (10<sup>3.5</sup>Pa s; Table 3) and therefore the fastest ascent rate (~8200 m/year)  
 578 and the lowest partial dissolution time of the epidote crystals (~3 years; Table 2).

579 It needs to be stressed that the ascent rates calculated above reflect not only the magma  
 580 viscosity but also the conduit width. This can be appreciated through the following classical  
 581 relationships which relates both parameters (Furbish, 1996):

$$w = \left(\frac{1}{3\eta}\right) * (\rho_c - \rho_l)gb^2$$

582 which gives the average rate of magma flow  $w$  ( $\text{m s}^{-1}$ ) in a dyke of width  $b$  (m) where  
 583  $\rho_c$  is the density of host rocks (about  $2700 \text{ kg m}^{-3}$ ) and  $\rho_l$  the density of melt (about  $2350 \text{ kg}$   
 584  $\text{m}^{-3}$ ),  $\eta$  the magma viscosity (Pa s), and  $g$  the gravitational acceleration ( $9.8 \text{ m}^2\text{s}^{-1}$ ). Given that  
 585 densities of both host rocks and melts are broadly similar, variations in  $w$  will mainly arise  
 586 from those of  $\eta$  and  $b$ . A somewhat wider dike width for TT magma could thus explain its  
 587 faster ascent rates relative to that of Sanukitoid, despite the latter having a lower viscosity  
 588 than the former.

589

### 590 8.3 EPIDOTE AS A PETROLOGICAL INDICATOR OF MAGMA EMPLACEMENT IN 591 THE ARCHEAN

592 Recent multidisciplinary studies involving high resolution geochronology and  
 593 numerical simulations suggest that most plutons are formed through the accumulation of  
 594 smaller bodies. Thus, plutons result from different pulses of magmatic activity. This factor  
 595 and the magma composition, emplacement rates of individual pulses, cooling rate and  
 596 emplacement geometry control the evolution of a pluton and its relationship with the tectonic  
 597 environment (Annen, 2011; de Saint Blanquat et al., 2011; Zibra et al., 2014). To analyze  
 598 magmatic transport, in addition to the distance travelled by the magma before solidifying, its  
 599 restite content and the physical properties of the host rocks, the physical and chemical  
 600 properties of the melt (T, P,  $f\text{O}_2$ ,  $\text{XH}_2\text{O}$ ) and the phases present in the system must also be  
 601 considered (Brown, 1994).

602 In this context, the epidote kinetics elaborated by Brandon et al. (1996) can be useful in  
 603 the research of emplacement mechanisms. Those authors have shown that the process of  
 604 epidote crystal dissolution to be fast: at  $700\text{-}800 \text{ }^\circ\text{C}$ , grains with sizes of  $0.2\text{-}3 \text{ mm}$  would only  
 605 need 2 to 2,000 years to dissolve (Schmidt and Poli, 2004). Another aspect is the size of  
 606 magmatic intrusions and the construction time of Archean plutons. To a first order, the  
 607 cooling time (s) of a magma sill of  $x$  (m) thickness can be estimated by applying the  
 608 following standard equation (e.g., Jaupart and Mareschal, 2010):

609

$$t = x^2 / D_{\text{heat}}$$

610 where ( $D_{\text{heat}}$ ) is the heat diffusion ( $10^{-7} \text{ m}^2/\text{s}$ ). A  $100 \text{ m}$  thick sill would fully crystallize  
 611 in about 1000 years. This time interval is long enough to dissolve an epidote crystal  $\sim 1.2 \text{ mm}$   
 612 (Brandon et al., 1996). Assuming this calculated value, the formation of an Archean pluton in  
 613 the CP, with an estimated thickness of  $\sim 4 \text{ km}$ , would require the juxtaposition of  $100\text{-m}$ -thick

614 sills fed by successive pulses and ~42,000 years of magmatic injection to form the pluton  
615 body. However, geologically, variable times can occur between intrusions, with shorter or  
616 longer periods depending fundamentally on the ability of the source to supply magma to the  
617 surface system. Altogether, the dissolution kinetics of the epidote, which formed grains of  
618 varying sizes in the CP, suggests that the mode of emplacement of Archean granitoid magmas  
619 was also dominated by a sequential process, i. e. incremental growth, similar to that inferred  
620 for their post-Archean counterparts, with crystallization of small individual bodies with an  
621 estimated thickness in the order of a hundred meters.

622 We note that in the case of the Ourilândia do Norte granitoids, Silva et al. (2018)  
623 suggested that such granitoids were formed through magma migration and accommodation of  
624 several mantle- and crust-derived pulses in a short period, corroborating the above hypothesis  
625 of progressive magma amalgamation via small batches. Similarly, the VJS was likely  
626 generated by different magmatic pulses from different sources (Dall'Agnol et al., 2017; Silva  
627 et al., 2020; Sousa, 2019). In detail, the TT magma appears to have been emplaced at  
628 somewhat deeper crustal pressures (~500 MPa; Table 5) relative to the sanukitoids (~350  
629 MPa; Table 5) and VJS (~400 MPa; Table 5) magmas. This would impart a comparatively  
630 slower crystallization rate and therefore a longer epidote reaction with the magmatic melt,  
631 allowing more intense corrosion of epidote crystals. In contrast, the shallower emplacement of  
632 the sanukitoid and Vila Jussara granitoid magmas allowed for a faster crystallization owing to  
633 slightly cooler host rocks. Accordingly, the process of epidote crystal dissolution was more  
634 quickly interrupted because their shielding was favored, resulting in crystals with less  
635 corrosion than those observed in the TT. We note also that experimental studies have  
636 demonstrated that high  $fO_2$  enhances epidote stability (Schmidt and Poli, 2004), which may  
637 have been an additional controlling factor of the extent of epidote dissolution (more  
638 pronounced at low  $fO_2$  relative to high  $fO_2$ , everything else being equal). Overall, as  
639 discussed above, a fast magma rise combined with a more superficial emplacement and a fast  
640 crystallization promote epidote preservation.

641 Thus, in addition to providing a more accurate characterization of the nature of magmas,  
642 the analysis of physical properties, when combined with experimental studies, epidote  
643 dissolution kinetics (Table 2) and viscosity estimates (Table 4), help to understand the magma  
644 rise, emplacement and crystallization processes of the Archean plutons in the CP.

## 645 9. CONCLUSION

646 Magmatic epidote is commonly found in various Archean granitoids of the CP. By  
647 combining a detailed study of the textural aspects with the determination of magma physico-  
648 chemical properties, epidote dissolution kinetics and viscosity, we reached the following  
649 conclusions:

- 650 (1) The epidote of the TT and VJS varieties has pistacite contents indicative of a magmatic  
651 origin, whereas that of the Ourilândia do Norte sanukitoid has a Ps content  
652 predominantly compatible with primary formation and to a lesser extent with formation  
653 through plagioclase and biotite alteration. In addition, the TiO<sub>2</sub> content ( $\leq 0.137$  wt. %) in  
654 the epidote of these granitoids also indicates magmatic origin.
- 655 (2) The kinetics showed that the epidote dissolution time of tonalite-trondhjemite was  
656 shorter ( $\sim 4$  years) than that of sanukitoid ( $\sim 6$  years). In turn, at the beginning of the  
657 magma rise, the TT magma transport speed was faster ( $\sim 4600$  m/years) than that of  
658 sanukitoid ( $\sim 4300$  m/years). Conversely, in the VJS, monzogranitic magmas exhibited  
659 epidote dissolution times of  $\sim 5$  years for bt-hb monzogranite and  $\sim 10$  years for bt  
660 monzogranite, even though the initial magma transport rates were similar ( $\sim 7900$   
661 m/year and  $\sim 7906$  m/year for bt-hb monzogranite and bt monzogranite, respectively). In  
662 turn, the VJS tonalitic magma had the shortest epidote dissolution time ( $\sim 3$  years) and  
663 the highest magma rise rate ( $\sim 8210$  m/year).
- 664 (3) The viscosity of the CP Archean magmas increases from sanukitoid to the tonalitic  
665 variety of the VJS, followed by monzogranitic magmas of the same suite and peaks in  
666 tonalite-trondhjemite. These variations in viscosity, in the case of tonalite-trondhjemite  
667 and sanukitoid magmas, show no clear match with magma ascent rates, these magmas  
668 having similar ascent rates but different viscosities ( $10^{5.3}$  and  $10^{2.5}$  Pa s, respectively).  
669 This probably reflects variations in conduit width that may have modulated magma  
670 ascent rates. In contrast, in the VJS, the correlation between viscosity and magma rise  
671 rate seems more coherent. The monzogranitic magmas show similar viscosity values  
672 ( $\sim 10^{4.4}$  Pa s) and very similar magma rise rates. They contrast with the tonalitic magma,  
673 which shows the fastest rise among all studied granitoids and a viscosity ( $10^{3.5}$  Pa s)  
674 lower than that of the VJS magmas.
- 675 (4) The sanukitoid and Vila Jussara granitoid magmas are emplaced closer to the surface  
676 and therefore tend to crystallize more rapidly than the TT magma, which displays  
677 epidote crystals more intensely corroded than those of the first two magmas.



678 (5) The short epidote crystal dissolution time in the CP suggests that these Archean  
679 magmas were emplaced through a mechanism involving incremental stacking of sill-  
680 like bodies with a thickness of about a hundred meters, each cooling in approximately  
681 1000 years. In such a view, approximately 42,000 years of magmatic feeding would  
682 have been required through several pulses to form ~ 4-km-thick plutonic bodies,  
683 assuming no time interval between each injection.

684 (6) Magmatic epidote is generally considered to be a phase absent from granitoids of  
685 different Archean cratons worldwide (Schmidt and Poli, 2004), despite being relatively  
686 common in the CP. This study shows that magmatic epidote crystals can be preserved,  
687 albeit rarely, in Archean granitoids under fast magma rise and crystallization conditions.

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## 1029 **FIGURE CAPTIONS**

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1031 Figure 1. (a) Tectonic provinces of the Amazonian Craton with the location of the Carajás  
 1032 Province shown (based on Santos et al., 2003); (b) Simplified geological map of the Carajás  
 1033 Province highlighting the division into the Rio Maria, Sapucaia, Canaã dos Carajás domains  
 1034 and Carajás Basin (Source: Silva et al., 2018, modified); (c) Geological map of the Sapucaia  
 1035 Domain and its transition into the Canaã dos Carajás and Rio Maria Domains, with the  
 1036 location of the studied samples from the Vila Jussara Suite granitoids, the tonalite-Colorado  
 1037 Trondhjemitic association of Sapucaia and the Ourilândia do Norte Suite sanukitoids.

1038 Figure 2. Textural relationship of the magmatic epidotes in the CP Archean granitoids.  
 1039 Trondhjemitic-tonalites of Sapucaia: (a) subhedral epidote partly surrounded by biotite and  
 1040 hornblende, exhibiting an irregular shape in contact with alkali-feldspar; (b) allanite crystal  
 1041 with an epidote rim, showing generally regular contacts with hornblende and irregular  
 1042 contacts with quartz; Ourilândia sanukitoids: (c) epidote with an irregular shape and signs of  
 1043 resorption through residual magmatic fluid in contact with hornblende, quartz and  
 1044 plagioclase; dissolution is apparently more intense when in contact with felsic minerals; (d)  
 1045 mafic aggregate of epidote, hornblende, biotite, titanite opaque minerals and apatite; (e)  
 1046 epidote crystal with irregular contacts with hornblende, showing more intense dissolution in  
 1047 contacts with plagioclase and alkali-feldspar; and (f) anhedral epidote surrounded by  
 1048 hornblende, with irregular contacts resulting from partial resorption. Abbreviations according  
 1049 to Kretz (1983).

1050 Figure 3. Histogram of compositional variation in the pistacite component (Ps: molar  
 1051  $[\text{Fe}^{3+}/\text{Fe}^{3+}+\text{Al}]/100$ ), in molecular percentage (mol.%), in Sapucaia TT epidotes, Ourilândia  
 1052 do Norte sanukitoids and Vila Jussara granitoids. Ranges of compositional variation in Ps in  
 1053 magmatic and secondary epidotes formed by plagioclase and biotite alteration are also  
 1054 indicated.



1055 Figure 4. Textural aspects of thin sections used in the dissolution kinetic study of epidote in  
1056 the Archean granitoids from the CP. The original shapes of the corroded epidote crystals are  
1057 indicated by black lines; dotted lines indicate the largest dimension of the crystal, and red  
1058 lines delimit the current contour of the corroded crystals. Line segments delimited by arrows  
1059 show where were defined the maximum dissolution values for each crystal. The PFA-14  
1060 sample represents the Sapucaia tonalite-trondhjemite; the NDP-47 and NDP-101 samples  
1061 represent the Ourilândia do Norte sanukitoid; and the MYF-77 sample represents the VJS  
1062 biotite-hornblende monzogranite variety. (a) and (b) epidote crystals are best preserved when  
1063 in contact with biotite and partly reabsorbed by the residual magmatic fluid currently  
1064 represented by feldspars; (c) epidote involving an allanite euhedral core partly surrounded by  
1065 biotite and exhibiting maximum dissolution in contact with felsic minerals; (d) subhedral  
1066 epidote crystal heavily corroded when in contact with feldspars; (e) epidote with marked  
1067 resorption at the edges, which is less marked in contacts with biotite; and (f) allanite with an  
1068 epidote rim, preserved when in contact with biotite and corroded along those with felsic  
1069 minerals that formed from residual melt.

1070 Figure 4 continuation. Textural aspects of epidote in the VJS granitoids. Sample PFR-16A is  
1071 representative of the biotite-hornblende monzogranite variety; samples MDP-02A and LIF-  
1072 04A are representative of the biotite monzogranite variety and sample MDP-02E is  
1073 representative of biotite-hornblende tonalite. (g) anhedral epidote in irregular contact with  
1074 hornblende showing a more pronounced resorption in contacts with feldspars and quartz; (h)  
1075 subhedral epidote crystal with intensely corroded portions; (i) epidote with rectilinear contacts  
1076 with biotite and irregular contacts with opaque minerals and feldspars; (j) allanite surrounded  
1077 by epidote and in association with opaque minerals, corroded in contacts with feldspars; and  
1078 (k) and (l) heavily corroded epidote crystals in contact with hornblende and felsic minerals.  
1079 Abbreviations according to Kretz (1983).

1080 Figure 5. Diagram showing the variation in the calculated magma viscosity of TT, sanukitoid  
1081 and the three varieties from the VJS from the CP as a function of temperature. Viscosity was  
1082 calculated based on the model by Giordano et al. (2008). The red circle at the beginning of the  
1083 curve represents the estimated liquidus temperature of each magma.

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Table 1. Representative electron microprobe analyses of Archean granitoid epidotes from the Carajás Province.

Sample	Sapucaia tonalite-trondhjemite						Ourilândia do Norte Sanukitoid						Vila Jussara Suite										
	PFA-14			PFR-33A			NDP-47			NDP-101			ADE-01D		AFD-11A		AMP-77						
Analyses	C1.1-4	C1.1-5	C1.1-6	C1.1-1	C2.1-3	C2.1-4	C2.1-2	C2.1-3	C3.1-1	C3.1-3	C3.1-4	C3.1-5	C3.1-6	C3.1-4	C3.1-6	C2.1-8	C3.1-1	C3.1-2	C2.1-1	C2.1-3	C2.1-6	C2.1-7	C2.1-8
SiO <sub>2</sub>	37.74	38.23	37.82	37.62	38.2,503	38.05	38.21	38.16	37.93	37.53	37.61	37.93	37.51	37.60	37.86	37.92	37.59	37.76	37.55	37.82	37.96	37.82	37.59
TiO <sub>2</sub>	0.10	0.09	0.07	0.10	0.04	0.05	0.10	0.11	0.09	0.05	0.04	0.09	0.13	0.09	0.04	0.08	0.01	0.10	0.09	0.09	0.06	0.00	0.07
Al <sub>2</sub> O <sub>3</sub>	23.10	23.74	23.01	22.63	23.18	23.22	24.90	24.48	23.84	21.37	21.28	24.31	20.64	22.04	22.24	23.28	23.04	22.68	23.45	23.28	24.08	23.96	22.34
Fe <sub>2</sub> O <sub>3</sub>	13.49	12.98	13.92	12.83	13.50	13.49	9.73	10.19	12.72	15.39	15.61	11.78	15.71	13.60	14.84	13.41	14.04	14.64	12.37	13.83	12.76	13.13	15.12
MnO	0.19	0.19	0.25	0.38	0.36	0.37	0.18	0.16	0.10	0.13	0.16	0.16	0.14	0.10	0.16	0.21	0.18	0.20	0.19	0.23	0.21	0.22	0.18
CaO	23.70	23.53	23.45	23.23	23.13	23.16	23.48	23.52	23.42	23.28	22.45	23.35	22.69	23.25	23.23	23.67	23.36	23.04	23.27	23.31	23.48	23.29	23.08
Na <sub>2</sub> O	0.00	0.00	0.00	0.00	0.00	0.02	0.01	0.01	0.01	0.02	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.01	0.01	0.01
K <sub>2</sub> O	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.02	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.02	0.00	0.01
Ce <sub>2</sub> O <sub>3</sub>	0.05	0.02	0.01	0.00	0.00	0.08	0.03	0.03	0.01	0.00	0.00	0.06	0.00	0.01	0.04	0.04	0.00	0.01	0.03	0.00	0.00	0.05	0.01
La <sub>2</sub> O <sub>3</sub>	0.00	0.04	0.04	0.00	0.00	0.02	0.03	0.00	0.01	0.02	0.00	0.03	0.02	0.00	0.01	0.00	0.05	0.07	0.04	0.00	0.00	0.00	0.00
Total	98.37	98.83	98.59	96.80	98.23	98.47	96.67	96.66	98.14	97.79	97.15	97.72	96.86	96.70	98.42	98.61	98.30	98.51	97.01	98.54	98.58	98.47	98.43
Number of cations on the basis of 13 oxygens																							
Si	3.120	3.134	3.122	3.121	3.140	3.138	3.141	3.142	3.127	3.140	3.160	3.132	3.169	3.126	3.137	3.124	3.113	3.124	3.103	3.118	3.117	3.112	3.118
Ti	0.006	0.006	0.005	0.006	0.002	0.003	0.006	0.007	0.006	0.003	0.002	0.005	0.008	0.006	0.003	0.005	0.001	0.006	0.006	0.005	0.004	0.000	0.004
Al	2.251	2.294	2.239	2.213	2.256	2.258	2.413	2.376	2.317	2.107	2.108	2.366	2.055	2.159	2.172	2.260	2.249	2.211	2.283	2.262	2.330	2.324	2.183
Fe <sup>+3</sup>	0.839	0.800	0.865	0.890	0.839	0.837	0.669	0.702	0.789	0.969	0.987	0.732	0.999	0.946	0.926	0.832	0.875	0.911	0.855	0.858	0.788	0.813	0.944
Mn	0.013	0.013	0.018	0.026	0.025	0.026	0.013	0.011	0.007	0.009	0.011	0.011	0.010	0.007	0.011	0.014	0.013	0.014	0.014	0.016	0.015	0.015	0.013
Ca	2.099	2.067	2.073	2.064	2.047	2.047	2.068	2.075	2.068	2.087	2.021	2.066	2.054	2.071	2.063	2.089	2.073	2.042	2.060	2.059	2.066	2.053	2.051
Na	0.000	0.000	0.000	0.000	0.000	0.004	0.002	0.002	0.001	0.003	0.000	0.000	0.001	0.000	0.000	0.000	0.003	0.000	0.003	0.000	0.001	0.002	0.002
K	0.000	0.001	0.001	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.002	0.000	0.000	0.000	0.000	0.001	0.001	0.000	0.002	0.000	0.001
Pistacite	27	26	28	29	27	27	22	23	25	32	32	23	33	30	30	27	28	29	27	28	25	26	30

Table 2. Estimation of the transport rate of magmas based on the kinetic dissolution study of the Archean granitoid epidotes from the Carajás Province.

Sapucaia tonalite-trondhjemite				
Sample	Grain width (µm)	Maximum width of dissolution (µm)	Partial dissolution time (years)	Transport rate (m/year)
PFA-14				
Spot 1		80	4.1	4109
Spot 2	519	73	3.4	4935
Spot 3	259	82	4.3	4693
Average			4	4579
Ourilândia do Norte Sanukitoid				
Sample	Grain width (µm)	Maximum width of dissolution (µm)	Partial dissolution time (years)	Transport rate (m/year)
NDP-47				
Spot 1		85	4.6	4732
Spot 2	440	79	4.0	5478
Spot 3		99	6.2	3488
Spot 4		68	2.9	7393
Spot 5	498	122	9.4	2297
Spot 6		88	4.9	4415
NDP-101				
Spot 1		121	9.3	2335
Spot 2	609	81	4.2	5211
Spot 3		98	6.1	3560
Average			6	4323
Granitoids of the Vila Jussara Suite				
Sample	Grain width (µm)	Maximum width of dissolution (µm)	Partial dissolution time (years)	Transport rate (m/year)
MYF-77				
Spot 1		65	3	7469
Spot 2	415	63	3	7951
PFR-16A				
Spot 1		47	1	14286
Spot 2	253	155	15	1314
Spot 3		60	2	8766
Average			5	7957
Sample	Grain width (µm)	Maximum width of dissolution (µm)	Partial dissolution time (years)	Transport rate (m/year)
MDP-02A				
Spot 1		110	8	2608
Spot 2	812	279	49	405
LIF-04A				
Spot 1		132	11	1811
Spot 2	464	107	7	2756
Spot 3		65	3	7469
Spot 4		60	2	8766
Spot 5	600	34	1	27299
Spot 6		51	2	12133
Average			10	7906
Sample	Grain width (µm)	Maximum width of dissolution (µm)	Partial dissolution time (years)	Transport rate (m/year)
MDP-02E				
Spot 1		74	3	5763
Spot 2	541	96	6	3424
Spot 3		74	3	5763
Spot 4	337	42	1	17890
Average			3	8210

Table 3. Whole rock geochemical composition used to calculate the viscosity of the Sapucaia tonalite-trondhjemite<sup>(1)</sup>, Ourilândia do Norte sanukitoid<sup>(2)</sup> and Vila Jussara Suite<sup>(3)</sup> magmas.

Lithology Sample	TT	Sanukitoid	Vila Jussara Suite		
	MYF-19A	BRM-113B	PFR-16A	MDP-02A	MDP-02E
SiO <sub>2</sub>	69.30	65.68	71.44	71.20	60.00
TiO <sub>2</sub>	0.30	0.41	0.36	0.49	1.06
Al <sub>2</sub> O <sub>3</sub>	16.20	15.52	12.89	13.90	13.90
FeO <sub>t</sub>	2.79	3.49	4.28	2.97	8.18
MnO	0.04	0.04	0.06	0.03	0.12
MgO	0.77	2.07	0.30	0.80	2.78
CaO	2.85	3.43	1.91	2.23	5.56
Na <sub>2</sub> O	5.20	4.90	3.46	3.42	3.38
K <sub>2</sub> O	1.50	2.30	3.93	3.91	1.75
P <sub>2</sub> O <sub>5</sub>	0.09	0.16	0.07	0.13	0.31
H <sub>2</sub> O*	2.0	7.0	4.0	4.0	4.0

\* H<sub>2</sub>O wt. % estimate based on rock mineralogy and experimental data. Data sources: <sup>(1)</sup>Santos et al.(2013b); <sup>(2)</sup>Santos and Oliveira (2016); <sup>(3)</sup>Dall’Agnol et al. (2017) and Silva (2019).

Table 4. Viscosity<sup>(1)</sup> of Archean granitoid magmas from the Carajás Province at different liquidus<sup>(2)</sup> to solidus temperatures

Melt Sample Temperature (°C)	TT	Sanukitoid	Vila Jussara Suite		
	MYF-19A	BRM-113B	PFR-16A	MDP-02A	MDP-02E
Viscosity (10 <sup>x</sup> Pa s)					
950		2.5			
940		2.6			
930		2.7			
920		2.7			
910		2.8			
900		2.9			
890		2.9	4.4	4.4	3.5
880	5.3	3.0	4.5	4.5	3.6
870	5.4	3.1	4.6	4.6	3.7
860	5.6	3.2	4.7	4.7	3.8
850	5.7	3.2	4.8	4.8	3.9
840	5.8	3.3	4.8	4.9	4.0
830	5.9	3.4	4.9	5.0	4.1
820	6.0	3.5	5.0	5.1	4.2
810	6.1	3.6	5.1	5.2	4.3
800	6.2	3.6	5.2	5.3	4.4
790	6.4	3.7	5.3	5.4	4.6
780	6.5	3.8	5.4	5.5	4.7
770	6.6	3.9	5.5	5.6	4.8
760	6.8	4.0	5.7	5.7	4.9
750	6.9	4.1	5.8	5.8	5.0
740	7.0	4.2	5.9	5.9	5.2
730	7.2	4.3	6.0	6.0	5.3
720	7.3	4.4	6.1	6.2	5.4
710	7.5	4.5	6.2	6.3	5.6
700	7.6	4.6	6.4	6.4	5.7

(1) Viscosities calculated on the basis of the model proposed by Giordano et al. (2008); (2) The liquidus temperatures of the various magmas are given on Table 5.

Table 5. Estimates\* of intensive crystallization parameters of representative samples of Sapucaia tonalite-trondhjemite, Ourilândia do Norte sanukitoid and Vila Jussara Suite magmas

Lithology	Sapucaia TT <sup>(1)</sup>	Ourilândia do Norte Sanukitoid <sup>(2)</sup>	Vila Jussara Suite <sup>(3)</sup>		
Sample	MYF-19A	BRM-113B	PFR-16A	MDP-02A	MDP-02E
Generation Pressure (P MPa)	1500	600-900	800-1000	800-1000	800-1000
Emplacement P (MPa)	500	350	400	400	400
Liquidus Temperature (T°C)	880	950	890	890	890
Solidus T (°C)	700	700	700	700	700
Saturation in Zr T (°C)	794	750	856	888	778
XH <sub>2</sub> O (wt. %)	2	7	4	4	4
<i>f</i> O <sub>2</sub>	NNO±0.5	NNO±0.5 and NNO±2.5	FMQ±0.5	NNO±0.5	NNO±1

\*Estimates based on (1) Clemens et al. (2006) and Santos et al. (2013b); (2) Oliveira et al. (2010) and Santos and Oliveira (2016); (3) Dall'Agnol et al. (2017); cf. text.

**HIGHLIGHTS**

- Epidote is relatively common in the Archean granitoids of the Carajás Province.
- The variations in the viscosity of the Archean magmas do not show a clear match only with magma ascent rates.
- The epidote crystal dissolution time in the Carajás Province suggests that Archean magmas were emplaced through a mechanism involving incremental growth.
- A fast magma rise combined with a more superficial emplacement and a fast crystallization promote epidote preservation.
- Epidote dissolution kinetics and viscosity estimates, help to understand the magma rise, emplacement, and crystallization processes of the Archean plutons.

**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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