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AUTHOR STATEMENT

**Cunha, I.RV:** 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.

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# 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 Neoarchean (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 Neoarchean 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 sill-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

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

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

In this context, it is relevant to determine variations of the viscosity of silicate magmas, because this parameter plays a key role in the transport dynamics, type of eruption and rates of physicochemical processes of natural magmas (Giordano et al., 2008), being crucial for a quantitative understanding of igneous processes (Persikov, 1991). The viscosity strongly depends on the composition of the melt, particularly its $SiO_2$ content, which determines its

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

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

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

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

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

paragenesis of natural rocks produced under different P-T-$fO_2$ conditions (Schmidt and Poli, 2004).

Experimental studies demonstrate the stability of epidote above 300 MPa; however, depending on the composition and oxygen fugacity, the minimum pressure for epidote crystallisation may vary (Schmidt and Poli, 2004). Crawford and Hollister (1982), based on the intersection of the solidus curve with that of epidote (Liou, 1973), proposed that in $H_2O$-saturated granites, magmatic epidote requires a minimum pressure of ~600 MPa. In granodioritic and tonalitic magmas, epidote is stable at pressures between 800 and 1000 MPa at temperatures near the solidus curve (Naney 1983; Liou 1993; Schmidt and Thompson, 1996). However, epidote may crystallise at pressures as low as ~300 MPa at $fO_2$ near the Hematite-Magnetite buffer, while at Ni-NiO, it is only stable at pressures ≥500 MPa (Naney, 1983; Schmidt and Thompson,1996). In general, epidote stability increases under more oxidizing $fO_2$ conditions (Liou, 1973; Schmidt and Thompson, 1996; Poli and Schmidt, 2004; Schmidt and Poli, 2004; Oliveira et al., 2010), and the pistacite contents characteristic of magmatic epidotes observed in experiments are mainly found between hematite-magnetite (HM) and nickel-NNO buffers (Liou, 1973). Oscillatory zones in epidote crystals may reflect changes in composition or in oxygen fugacity (Franz and Liebscher, 2004; Pandit et al., 2014). At low-pressures, outside the stability field of epidote, magnetite is usually the main $Fe^{3+}$-containing phase (Drinkwater et al., 1991) and its modal content increases in rocks without epidote crystallization (Hammarstrom and Zen 1992; Schmidt, 1993).

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

hornblende + K-feldspar + plagioclase + magnetite + $H_2O$ = epidote + biotite + quartz

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

plagioclase + amphibole + melt = biotite and epidote

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

Textural and compositional criteria used to distinguish between magmatic and secondary epidotes, include (1) euhedral to subhedral crystals with prismatic or hexagonal shapes; (2) mineralogical associations with primary hornblende and biotite; (3) epidote with oscillatory zoning, allanite core and irregular contact with quartz and plagioclase; (4) pistacite content ranging from 25 to 35%; and (5) $TiO_2$ content ≤ 0.137% (Liou, 1973; Tulloch, 1979; Zen and Hammarstrom, 1984; Zen, 1985; Evans and Vance, 1987; Dawes and Evans, 1991; Schmidt and Poli, 2004; Sial et al., 2008).

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

$$Fe^{2+} + OH^{-} \leftrightarrow Fe^{3+} + O^{2-} + 1/2\ H_2$$

Based on this reaction, Gieré and Sorensen (2004) suggested that the stability of allanite in magmas may be more dependent on T and $fH_2$ than on the content of rare earth elements.

## 3.REGIONAL GEOLOGY

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

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

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

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

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

### 4.1 TONALITE-TRONDHJEMITE ASSOCIATION OF SAPUCAIA

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

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

In mineralogical terms, the Sapucaia TT essentially consists of plagioclase and quartz, with highly subordinate microcline and with biotite as the main mafic phase, accompanied by epidote, opaques, allanite, titanite, apatite and zircon as accessory minerals. Geochemically, the granitoids are calc-alkaline with generally peraluminous compositions, no significant Sr and Eu anomalies, silica content ranging from 59 to 74 wt. %, relatively low ferromagnesian content ($Fe_2O_3$+MgO+MnO+$TiO_2$ < 5 wt. %), moderate Mg# (0.27-0.47) and high alumina values ($Al_2O_3$ > 15 wt. %) (Santos et al., 2013b). The samples selected for analysis derive from the eastern section of the SD (Fig. 1c) and belong to the Colorado Trondhjemite. Santos et al. (2013b) reported the presence of magmatic epidote in these rocks and conducted a preliminary mineralogical study based on scanning electron microscopy and EDS analyses.

### 4.2 OURILÂNDIA DO NORTE SANUKITOID

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

In petrographic terms, the sanukitoid rocks exhibit well-preserved igneous textures, but are also marked by protomylonitic to mylonitic tectonic fabrics. The mineral assemblage essentially consists of quartz, plagioclase, various proportions of K-feldspar and mafic clusters formed by amphibole, biotite and epidote. The main accessory minerals consist of epidote (zoisite and clinozoisite), allanite, titanite, zircon, apatite, magnetite and ilmenite. Geochemically, sanukitoids and associated rocks exhibit a highly variable $SiO_2$ content, moderate $K_2O$ values and high Mg#, Cr and Ni values. They have a predominantly metaluminous character and a calc-alkaline to high-K calc-alkaline signature (Santos and Oliveira, 2016; Silva et al., 2018). Two textural types of magmatic epidote were distinguished: crystals with zoned or unzoned allanite cores and euhedral to subhedral crystals

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

### 4.3 VILA JUSSARA SUITE

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

Four main petrographic varieties were identified in the VJS: biotite-hornblende monzogranite (BHMzG), biotite-hornblende tonalite (BHTnl), biotite monzogranite (BMzG) and hornblende-biotite granodiorite (HBGd). The different rock varieties of the VJS are characterized by a similar mineralogy, from essential tectosilicates (quartz, plagioclase and potassic feldspar) to mafic minerals (hornblende and/or biotite), and by the main primary accessory minerals, ilmenite, magnetite, apatite and zircon, which can also include titanite, allanite and epidote. Geochemically, the VJS consists of ferroan granitoids (reduced and oxidized A-type granites) and mildly magnesian granites (cf. Dall'Agnol et al., 2017; Sousa, 2019). The four petrographic varieties of the VJS were formed under different conditions of oxygen fugacity. The reduced to moderately reduced BHMzG was formed under conditions varying from <FMQ to FMQ + 0.5. The oxidized BHMzG was generated under moderately oxidizing conditions (between NNO and NNO-0.5), whereas the magnesian granites were crystallized under oxidizing conditions, with $fO_2$ ranging from NNO to NNO+1 (Sousa, 2019). Dall'Agnol et al. (2017) described the textural and compositional variations in the magmatic epidote of the specific VJS varieties.

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

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

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

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

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

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

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

## 6.MINERAL CHEMISTRY

Epidote crystals with textural evidence of magmatic origin were analyzed on a JEOL JXA-8230 electron probe microanalyzer (with five spectrometers in wavelength-dispersive spectroscopy (WDS) mode and one in energy-dispersive spectroscopy (EDS) mode) at the Microanalysis Laboratory, Institute of Geosciences (*Instituto de Geociências* – IG), Federal University of Pará (*Universidade Federal do Pará* – UFPA). The analytical procedure was performed on polished thin carbon coated sections under the following operating conditions: 15 kV voltage acceleration; 20 nA current; 5 µm beam diameter; and 20 second analysis times for F, Na, Al, Cr, K, Sr, Ni, Mn, Ba, Ti, Cl, V, Y, Ce, and La, and 10 second analysis times for Mg, Si, Ca, and Fe. The crystals used for the analyses were LIF for Fe, Mn, Ba, and Ti; PETJ for Ca and K; TAP for Na, Mg, Al, and Si; and LIFH for Ce and La. The standards used for calibration were orthoclase ($SiO_2$ and $K_2O$), rutile ($TiO_2$), anorthite ($Al_2O_3$), magnetite (FeO), diopside (MgO), rhodonite (MnO), wollastonite (CaO), sodalite ($Na_2O$) and $CePO_4$ ($Ce_2O_3$ and $La_2O_3$).

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

Tulloch (1979) and Vyhnal et al. (1991) evaluated the origin of epidote based on its chemical composition and concluded that pistacite contents (Ps: molar $[Fe^{3+}/Fe^{3+}+Al]/100$) ranging from 25 to 29% (molecular weight) correspond to magmatic epidotes, differing from those formed by subsolidus alteration of plagioclase ($Ps_{0-24}$) and biotite ($Ps_{36-48}$). Experimental

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

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

In addition to the molar Ps content, previous empirical and experimental studies (Tulloch, 1979; Vyhnal et al., 1991; Liou, 1993; Sial et al., 1999; Oliveira et al., 2010; Dall'Agnol et al., 2017) suggest that a $TiO_2$ content in epidote ≤ 0.137 wt. % is in line with its magmatic formation. The Sapucaia TT and VJS granitoid epidotes display $TiO_2$ values always lower than 0.137 wt. % (Table 1), corroborating the textural and compositional assessments above indicative of magmatic origin. In the Ourilândia do Norte sanukitoids, the epidotes have $TiO_2$ values predominantly ≤ 0.137 wt. % (Table 1).

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

Studies conducted so far on Neoarchean granitoid from the Carajás Province (CP) corroborate the results of experimental researches showing that epidote formation depends on $fO_2$ and pressure, as shown by the Estrela Complex, Serra do Rabo and Planalto Suite granites, which formed under strongly reduced conditions and lack epidote. Conversely, the Villa Jussara Suite (VJS) (Dall'Agnol et al., 2017) and Vila União (Marangoanha et al., 2018) granitoids, which are mostly formed under oxidizing conditions, generally present epidote as an important accessory phase. The epidote crystals of the VJS show no considerable variations in $Fe^{3+}$ content between the core and rims of the crystals and rarely display zoning,

thus suggesting that in the final stages of crystallization, the residual melt was not impoverished in $Fe^{3+}$ and that $fO_2$ remained constant, thereby favoring epidote stability.

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

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

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

(1) Euhedral and subhedral epidote crystals allow us to more accurately infer the original shape of the crystal and therefore are prioritized when estimating the degree of dissolution. Generally, these crystals are associated with biotite, amphibole and residual magmatic melt, which are currently represented by quartz and feldspar crystals;

(2) In the absence of isolated epidote crystals, epidote crystals involving allanite cores in epitaxial growth can be used;

(3) The pistacite content of the selected epidote crystals should fall within the magmatic field (Ps: 25-35 mol%; Liou, 1973);

(4) Calculating the crystal corrosion intensity is crucial, and these calculations must be based on maximum extent of dissolution identified in the petrographic microscope and always with measurements perpendicular to the crystal faces affected by corrosion;

(5) The duration of the corrosion process is calculated using the minimum apparent diffusion coefficient between tonalitic magma and epidote ($5x10^{-17}$ $m^2$ $s^{-1}$ for Si, Al, Ca and F) at a temperature of 750°C, which can also be applied to other granitoid magmas, according to criteria formulated by Brandon et al. (1996), using the equations below:

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

$$t = d_z^2/(5x10^{-17})$$

with

$d_z$ = dissolution zone width (m);

$D_{app}$ = apparent diffusion coefficient ($5\times10^{-17}$ $m^2$ $s^{-1}$); and

t = partial epidote dissolution time (s)

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

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

$$Tr = L_r/t$$

Tr = transport rate (m/yr)

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

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

$P_e$ = emplacement pressure (kbar)

t = partial dissolution time do epidote (yr)

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

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

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

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

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

## 8. DISCUSSION

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

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

Experimental studies on the viscosity of silicate melts enabled Giordano et al. (2008) to formulate a general model, which includes the effect of $H_2O$ and F, to predict the non-Arrhenius Newtonian viscosity of silicate magmas, with the following equation:

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

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

et al model retrieves the correct order of magnitude of melt viscosity, provided temperature and $H_2O$ content are well estimated, as discussed below.

The following representative samples were selected: the MYF-19A sample, containing 69.3 wt. % $SiO_2$, was chosen for the Sapucaia TT magma, whereas the viscosity of the Ourilândia do Norte sanukitoid magma was calculated based on the composition of the BRM-113B sample, with 65.7 wt. % $SiO_2$ (Table 3). For the VJS, the PFR-16, MDP-02A and MDP-02E samples with 71.4, 71.2 and 60 wt. % $SiO_2$ contents, respectively, were chosen as representatives of bt-hb monzogranite, bt monzogranite and bt-hb tonalite magmas, respectively (Table 3).

As the magmas of the studied granitoids were derived from different sources, the temperature ranges were estimated from experimental studies conducted on systems with similar compositions than those selected above, so as to constrain initial (liquidus) and final (solidus) crystallization temperatures. For the TTG, the most accepted mechanism for the origin of their parental magmas is the partial melting of hydrated metabasaltic rocks, including greenstones, amphibolites and eclogites, in different tectonic environments (Martin, 1987; Winther, 1996; Condie, 2005; Clemens et al., 2006; Moyen and Martin, 2012). In particular, Clemens et al. (2006) proposed that TTG magmas are produced at temperatures between 900°C and 1000°C and that crystallization would start at ~880°C, which was the liquidus temperature estimated for Sapucaia TT magma. In turn, the absence of amphibole in these rocks suggests a low $H_2O$ content (Naney et al., 1983), which was estimated at 2% weight in the present study.

Sanukitoid suites are regarded as key components of late Archean terranes (2.95-2.54 Ga), presenting geochemical characteristics similar to those of mantle- and crust-derived magmatic rocks (Moyen et al., 2003; Condie, 2005; Martin et al., 2005; Oliveira et al., 2010, 2011) with higher temperatures compared to TTG. Oliveira et al. (2010) estimated a liquidus temperature of ~950°C and a $H_2O$ content of ~7% for the Rio Maria sanukitoids, which are used in this study for the Ourilândia do Norte sanukitoids.

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

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

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

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

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

For the TT magma, with a generation pressure of 1500 MPa, a liquidus temperature estimated at ~880°C, a low $H_2O$ content (~2%) and a relatively high silica content (70% weight) markedly increase the viscosity, resulting in a magma rise rate of ~4600 m/year (Table 2). However, such transport rates only correspond to the initial stages of magma rise because during the process, the magma temperature may decrease, which will be reflected in the increase in the degree of crystallinity and viscosity, thereby decreasing the magmatic flow. In turn, sanukitoid magma, with an estimated generation pressure between 900 and 600 MPa and emplacement in a syntectonic regime linked to transcurrent shear zones (Santos and Oliveira, 2016; Silva et al., 2018; Nascimento, 2020), shows low viscosity ($10^{2.5}$ Pa s), reflecting its high temperatures (Fig. 5; Table 4) and relatively high $H_2O$ content (7%, Table 3), with a relatively fast magma rise rate of ~4300 m/year. Moreover, sanukitoids have epidote crystals with a more preserved shapes and a dissolution time estimated at ~6 years (Table 2), whereas TT epidotes exhibit comparatively more intense corrosion and a shorter dissolution time (~4 years; Table 2). In contrast, TT magma transport is fastest compared to sanukitoid magmas analyzed in this study (Table 2). Therefore, for these two magmas, there is no clear correlation between estimated magma ascent rates and mineral dissolution time.

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

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

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

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

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

Recent multidisciplinary studies involving high resolution geochronology and numerical simulations suggest that most plutons are formed through the accumulation of smaller bodies. Thus, plutons result from different pulses of magmatic activity. This factor and the magma composition, emplacement rates of individual pulses, cooling rate and emplacement geometry control the evolution of a pluton and its relationship with the tectonic environment (Annen, 2011; de Saint Blanquat et al., 2011; Zibra et al., 2014). To analyze magmatic transport, in addition to the distance travelled by the magma before solidifying, its restite content and the physical properties of the host rocks, the physical and chemical properties of the melt (T, P, $fO_2$, $XH_2O$) and the phases present in the system must also be considered (Brown, 1994).

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

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

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

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

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

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

## 9. CONCLUSION

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

(1) The epidote of the TT and VJS varieties has pistacite contents indicative of a magmatic origin, whereas that of the Ourilândia do Norte sanukitoid has a Ps content predominantly compatible with primary formation and to a lesser extent with formation through plagioclase and biotite alteration. In addition, the $TiO_2$ content ($\leq$ 0.137 wt. %) in the epidote of these granitoids also indicates magmatic origin.

(2) The kinetics showed that the epidote dissolution time of tonalite-trondhjemite was shorter (~4 years) than that of sanukitoid (~6 years). In turn, at the beginning of the magma rise, the TT magma transport speed was faster (~4600 m/years) than that of sanukitoid (~4300 m/years). Conversely, in the VJS, monzogranitic magmas exhibited epidote dissolution times of ~5 years for bt-hb monzogranite and ~10 years for bt monzogranite, even though the initial magma transport rates were similar (~7900 m/year and ~7906 m/year for bt-hb monzogranite and bt monzogranite, respectively). In turn, the VJS tonalitic magma had the shortest epidote dissolution time (~3 years) and the highest magma rise rate (~8210 m/year).

(3) The viscosity of the CP Archean magmas increases from sanukitoid to the tonalitic variety of the VJS, followed by monzogranitic magmas of the same suite and peaks in tonalite-trondhjemite. These variations in viscosity, in the case of tonalite-trondhjemite and sanukitoid magmas, show no clear match with magma ascent rates, these magmas having similar ascent rates but different viscosities ($10^{5.3}$ and $10^{2.5}$ Pa s, respectively). This probably reflects variations in conduit width that may have modulated magma ascent rates. In contrast, in the VJS, the correlation between viscosity and magma rise rate seems more coherent. The monzogranitic magmas show similar viscosity values (~$10^{4.4}$ Pa s) and very similar magma rise rates. They contrast with the tonalitic magma, which shows the fastest rise among all studied granitoids and a viscosity ($10^{3.5}$ Pa s) lower than that of the VJS magmas.

(4) The sanukitoid and Vila Jussara granitoid magmas are emplaced closer to the surface and therefore tend to crystallize more rapidly than the TT magma, which displays epidote crystals more intensely corroded than those of the first two magmas.

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

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

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## FIGURE CAPTIONS

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

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

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

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

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

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

| Sapucaia tonalite-trondhjemite | | | | | | | Ourilândia do Norte Sanukitoid | | | | | | | | | | Vila Jussara Suite | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Sample | 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 |
| $SiO_2$ | 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 |
| $TiO_2$ | 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 |
| $Al_2O_3$ | 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 |
| $Fe_2O_3$ | 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 |
| $Na_2O$ | 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 |
| $K_2O$ | 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 |
| $Ce_2O_3$ | 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 |
| $La_2O_3$ | 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 |
| $Fe^{+3}$ | 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 |

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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 | 519 | 80 | 4.1 | 4109 |
| Spot 2 | | 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 | 440 | 85 | 4.6 | 4732 |
| Spot 2 | | 79 | 4.0 | 5478 |
| Spot 3 | | 99 | 6.2 | 3488 |
| Spot 4 | 498 | 68 | 2.9 | 7393 |
| Spot 5 | | 122 | 9.4 | 2297 |
| Spot 6 | | 88 | 4.9 | 4415 |
| NDP-101 | | | | |
| Spot 1 | 609 | 121 | 9.3 | 2335 |
| Spot 2 | | 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 | 415 | 65 | 3 | 7469 |
| Spot 2 | | 63 | 3 | 7951 |
| PFR-16A | | | | |
| Spot 1 | 253 | 47 | 1 | 14286 |
| Spot 2 | | 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 | 812 | 110 | 8 | 2608 |
| Spot 2 | | 279 | 49 | 405 |
| LIF-04A | | | | |
| Spot 1 | 464 | 132 | 11 | 1811 |
| Spot 2 | | 107 | 7 | 2756 |
| Spot 3 | | 65 | 3 | 7469 |
| Spot 4 | 600 | 60 | 2 | 8766 |
| Spot 5 | | 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 | 541 | 74 | 3 | 5763 |
| Spot 2 | | 96 | 6 | 3424 |
| Spot 3 | 337 | 74 | 3 | 5763 |
| Spot 4 | | 42 | 1 | 17890 |
| Average | | | 3 | 8210 |

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

| Lithology | TT | Sanukitoid | Vila Jussara Suite | | |
|---|---|---|---|---|---|
| Sample | MYF-19A | BRM-113B | PFR-16A | MDP-02A | MDP-02E |
| $SiO_2$ | 69.30 | 65.68 | 71.44 | 71.20 | 60.00 |
| $TiO_2$ | 0.30 | 0.41 | 0.36 | 0.49 | 1.06 |
| $Al_2O_3$ | 16.20 | 15.52 | 12.89 | 13.90 | 13.90 |
| $FeO_t$ | 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 |
| $Na_2O$ | 5.20 | 4.90 | 3.46 | 3.42 | 3.38 |
| $K_2O$ | 1.50 | 2.30 | 3.93 | 3.91 | 1.75 |
| $P_2O_5$ | 0.09 | 0.16 | 0.07 | 0.13 | 0.31 |
| $H_2O$* | 2.0 | 7.0 | 4.0 | 4.0 | 4.0 |

* $H_2O$ wt. % estimate based on rock mineralogy and experimental data. Data sources: [(1)]Santos et al.(2013b); [(2)]Santos and Oliveira (2016); [(3)]Dall'Agnol et al. (2017) and Silva (2019).

Table 4. Viscosity[(1)] of Archean granitoid magmas from the Carajás Province at different liquidus[(2)] to solidus temperatures

| Melt | TT | Sanukitoid | Vila Jussara Suite | | |
|---|---|---|---|---|---|
| Sample | MYF-19A | BRM-113B | PFR-16A | MDP-02A | MDP-02E |
| Temperature (°C) | Viscosity ($10^x$ 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 |

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(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[(1)] | Ourilândia do Norte Sanukitoid[(2)] | Vila Jussara Suite[(3)] | | |
|---|---|---|---|---|---|
| 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 |
| $XH_2O$ (wt. %) | 2 | 7 | 4 | 4 | 4 |
| $fO_2$ | 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.

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## 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.

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**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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