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# Hydrothermal activity along a strike-slip fault zone and host units in the São Francisco Craton, Brazil – Implications for fluid flow in sedimentary basins

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#### 27 Abstract

This study combines multiscale analyses of geological, fault, fracture, and stable 28 isotope data to investigate strike-slip deformation and channeling of 29 hydrothermal fluids along the Cafarnaum fault and calcite veins at different 30 distances from the fault, which is a structure in the São Francisco Craton, 31 northeastern Brazil. Meteoric fluids with  $\delta D$  values near -45‰ and  $\delta^{18}O$  values 32 33 near -6.5‰ and temperatures at 40-70°C precipitated as calcite veins in the host carbonate units. The Cafarnaum fault, a N-S-striking vertical, ~170 km long fault 34 zone, juxtaposes Neoproterozoic carbonate rocks in the western block and 35 36 Mesoproterozoic siliciclastic rocks in the eastern block. A zone of restraining bends occurs at the central part of the fault, whereas termination zones of 37 horsetail geometry occur at both ends of the Cafarnaum fault. These zones are 38 marked by NW-SE-striking extensional faults that are oblique to the main N-S-39 striking fault zone, where hydrothermal deposits occur. The zone of influence of 40 the Cafarnaum fault is ~20 km wide around the main fault. The fault formed 41 during the Brasiliano orogeny (740-560 Ma) after Neoproterozoic carbonate 42 platform deposition. In contrast with the host units, fluids along the fault zone 43 44 originated in deeper levels of the crust and show much lower  $\delta^{18}$ O values, 45 indicating higher crystallization temperatures. These fluids caused brecciation in 46 the Neoproterozoic carbonate host rocks, whereas a subsequent decrease in fluid pressure and cooling near the surface resulted in the precipitation of a 47 hydrothermal paragenesis in veins, also affecting the host rock. 48

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Keywords: carbonate veins, hydrothermal fluid, strike-slip fault, Salitre
 Formation, São Francisco Craton

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#### 53 **1. Introduction**

Sedimentary basins display different fluid migration regimes depending on 54 the host rock, particularly in areas subjected to extensional tectonics. 55 Sedimentary stratified layers may allow basinal fluids to migrate laterally parallel 56 to bedding for hundreds of kilometers (Qing and Mountjoy, 1992). In most 57 instances, sedimentary basins display limited vertical fluid flow due to 58 impermeable layers; hence, fluid conduits cannot connect deep and shallow parts 59 of a basin. Pore water usually has a low-velocity flow regime, and its primary 60 geochemical values are therefore easily altered by reactions with host rock 61 minerals and mixing with other fluids (Bjorlykke and Egeberg, 1993). 62

Vertical fluid flow pathways in sedimentary basins are associated with fault 63 64 zones (Haneberg et al., 1999; Hardebeck and Hauksson, 1999). Depending on the structures and/or permeability properties, fault zones can either act as 65 hydraulic barriers or as preferential conduits for geofluid migration 66 (Gudmundsson, 2001; Rawling et al., 2001; Smeraglia et al., 2021; La Bruna et 67 al., 2021). These characteristics are also linked to the several 68 structural/diagenetic phases affecting the carbonate rocks from the earliest 69 diagenetic stages (e.g., La Bruna et al., 2020). In fact, selective cementation 70 and structural diagenetic processes are key factors in fault permeability control 71 (Hausegger et al., 2010; Ngwenya et al., 2000). 72

Strike-slip faults create complex and heterogeneous permeability anisotropy and strongly influence fluid migration in crustal fault zones (Caine et al., 2010; Bense et al., 2013; Arancibia et al., 2014). A wide variety of processes at various scales can occur during fault growth and lead to a large range of fault architectures and properties that influence fluid flow behavior (Wibberley and

Shipton, 2010). The activity of major and weak strike-slip fault systems is 78 influenced by fluid flow (e.g., Byerlee, 1990; Rice, 1992; Sleep and Blanpied, 79 1992; Evans and Chester, 1995; Zhang et al., 2001). The internal structure of 80 strike-slip faults is dominated by vein arrays and hydraulic breccias. These 81 features result from intense, deep-seated, and localized hydrothermal fluid flow 82 (Cox and Munroe, 2016). Fluid flow history can be investigated in exhumed faults 83 84 and fractures, which provide information about deformation mechanisms, fluidrock interactions, and bulk chemical redistributions (Arancibia et al., 2014; 85 Stevrer and Sturm, 2002). Among other features, synthetic faults, antithetic faults, 86 deformation bands, joints, stylolites, veins, and breccia have been recognized in 87 strike-slip fault zones affected by hydrothermal fluids (e.g., Fossen and Rotevatn, 88 2016; Choi et al., 2016; Liao et al., 2017; Peacock et al., 2017a, 2017b; Alsop et 89 al., 2020; Ostermeijer et al., 2020). However, there is a debate about which of 90 these structures, if any, exerts a primary influence on fluid flow and the role and 91 origin of fluids in strike-slip fault zones (e.g., Gudmundsson et al., 2002; 92 Gudmundsson, 2007). 93

Several studies show that hydrothermal deposits occur in the São Francisco Craton region, including those around fault zones, while the region remained tectonically stable during the Brasiliano/Panafrican orogenic cycle at 740-540 Ma (e.g., Almeida et al., 2000; Brito Neves et al., 2014). This study focuses on the Cafarnaum fault zone, which occurs as a lateral ramp (Fig. 1A, B, C, D). However, the relationship between the location and timing of hydrothermal deposit formation and fault geometry and evolution remains elusive.

101 This study is a multiscale and multidisciplinary approach that uses remote 102 sensing, aeromagnetic data, field observations, petrography, and stable isotope 103 geochemistry to compare the structural evolution of the hydrothermal system in 104 the siliciclastic and carbonate host rocks along the Cafarnaum fault zone and the inner basin. We present stable isotope analyses on carbonate host rocks, veins, 105 pockets, and fluid inclusions of the inner basin to reconstruct fluid-rock 106 interactions and build a model to predict the development of hydrothermal activity 107 on a regional scale, which can be used as a proxy for other basins elsewhere. 108 New stable isotope data of fluid inclusions and veins and previously published 109 110 data of sulfur in sulfides show that carbonate veins associated with fractures at the edge of the basin record much higher temperatures than those crystallized in 111 the central part of the basin. This study also describes and discusses the 112 kinematics, morphology, and magnetic characteristics of the fault zone and the 113 formation of various types of hydrothermal dilation breccias in the damage zone. 114 Finally, a comparison of the Precambrian Cafarnaum fault system and its host 115 116 rocks with other faults that affect other carbonate and siliciclastic units helps understand regional predictability. This study concludes that fluid flow occurs 117 mainly along extensional subsidiary faults and investigates the way they deform 118 the host rocks. The isotope data indicate that fluids are meteoric in origin and, 119 compared to the sedimentary basins, fluids that percolated the crystalline terrain 120 121 may have circulated into much deeper zones.

## 122 **2. Geological setting**

The study area is mainly composed of Mesoproterozoic rocks of the 123 Chapada Diamantina Group and Neoproterozoic units of the Una Group, 124 primarily the Salitre Formation (Fig. 1D, E). The groups contain distinct 125 formations and contrasting structural styles. Both major terrains are bounded by 126 a strike-slip fault that we name the Cafarnaum fault in this study. It acted as a 127 tectonic boundary between the aforementioned Mesoproterozoic 128 and 129 Neoproterozoic units (La Bruna et al., 2021).

The Chapada Diamantina Group is 1,000 m thick and includes the Morro do Chapéu Formation, which was deposited at ~1400-900 Ma (Pedreira et al., 1975; D'Angelo et al., 2020). This unit was affected by a first contractional inversion event that is mainly marked by symmetrical, N-S-trending, open folds (Danderfer et al., 2015; D'Angelo et al., 2020). High-angle fractures strike mostly N-S, NE-SW and NW-SE, and have a high degree of symmetry with the N-S regional folds (Danderfer et al., 2015).

The Salitre Formation is ~750 m thick in the central part of the Irecê Basin 137 138 (D'Angelo et al., 2020), which was deposited in a carbonate pelitic marine basin (Misi et al. 2005, 2011) at ~750 Ma (D'Angelo et al., 2020). This sequence was 139 deposited in the Irecê Basin, an asymmetric graben with an approximately N-S-140 141 oriented axis that plunges toward the north (Lagoeiro, 1990; Kuchenbecker et al., 2011; Brito Neves et al., 2012; D'Angelo et al., 2020). The Irecê Basin was 142 143 inverted in the Brasiliano orogeny, with a peak at ~600 Ma, which resulted in anomalous deformation concerning adjacent domains, with a series of south-144 verging fold and thrust systems (Lagoeiro, 1990; Teixeira et al., 2019; D'Angelo 145 et al., 2020). 146

The carbonate units of the Irecê Basin have similar Pb-Pb isochron ages and paleomagnetic poles, which fall close to ~520 Ma. This age is consistent with the Gondwana supercontinent's apparent polar wander path and indicates that isotopic and magnetic systems reset those of the Cambrian (Trindade et al., 2004). This event was related to regional-scale fluid migration and subsequent mineralization at the end of the Brasiliano orogenic cycle (Trindade et al., 2004).

An E-W-oriented magnetic telluric section across the Irecê and Morro do
 Chapéu Basins reveals lithospheric resistive blocks bounded by major conductive

deep zones, which are interpreted as faults. It shows that a lithospheric conductor, interpreted as a suture zone, occurs between the Neoproterozoic lrecê Basin and the Chapada Diamantina Group (Fig. 1D ). The high conductance zone is a combination of high porosity and high fluid salinity (Padilha et al., 2019).

Detailed geological mapping on both sides of the fault indicates a great number of occurrences of hydrothermal minerals associated with faults. These occurrences encompass metals such as Au, Pb-Zn, and Ba. These metals occur in sulfides associated with quartz veins in dolomite units close to the main faults. In a few cases, minerals such as barite also occur in the host carbonate rocks (Sampaio et al., 2001).

166 More detailed studies have also investigated hydrothermal silicification and dolomitization in a few karst systems. Several works, conducted in the São 167 Francisco Craton, have already investigated cave geometry, stratigraphy, 168 geochemistry, and mineralogy to indicate that fault and fracture systems were 169 used as conduits for deep-seated fluid flow (Klimchouk et al., 2016; Cazarin et 170 171 al., 2019; La Bruna et al., 2021; Pontes et al., 2021). In another case, approximately 100 km to the south of the study area, a N-S-striking, strike-slip 172 fault in the southern part of the Irecê Basin was the pathway for fluid flow confined 173 174 to the Salitre Formation during the Brasiliano orogeny. The first stage of Mgrich fluids caused extensive dolomitization in the Salitre Formation, which was 175 subsequently followed by Si-rich fluids that caused pervasive silicification in the 176 host units (Bertotti et al., 2020). 177

178

179 **3. Methods** 

180 The present study integrates (1) remote sensing and structural investigation, (2) Geophysical data and processing, (3) sampling, (4) mineralogy and petrography, 181 (5) stable isotope analysis of veins and host rock, and (6) isotope geochemistry 182 of fluid inclusions. We used the Shuttle Radar Topographic Mission (SRTM), 183 ALOS-PALSAR to map regional structures, and unmanned aerial vehicle imagery 184 for a detailed investigation of tectonic features (Fig. 2). The aeromagnetic data 185 186 used to map fault segments are from the Centro Norte Bahia Project is an airborne magnetic survey carried out by Companhia Baiana de Pesquisa Mineral 187 (CBPM) (Fig. 3). The petrography, mineralogy, isotope, and fluid inclusions are 188 based on the analysis of six outcrops in carbonate units of the Salite Formation 189 on both sides of the Cafarnaum fault. We present a complete description of data 190 and methods in the supporting material section (Methods – supporting material). 191

192

# 193 **4. Results**

#### 194 4.1 Qualitative field fracture analysis

195 Different fracture types were distinguished; joints display their peculiar plumose morphology (Pollard and Aydin, 1988), and are compart mentalized or not within 196 single beds (Fig. 4). For this reason, they are here named stratabound (SB) and 197 non-stratabound (NSB) fractures. Veins are also SB and NSB, but in some 198 199 outcrops they are subvertical and parallel to the tilted bed layers (Fig. 4C, D and 200 F). Clustered fracture and vein networks were in the proximity of fault zones (Fig. 5B,C, D and E). In particular, the high-resolution gualitative structural analysis 201 shows the following fracture sets: subvertical NW-SE fractures and veins (Fig. 202 4G, H and Fig.5B,C, D, E, F, G and H); minor subvertical NE-SW fractures and 203 204 veins (Fig. 4G and Fig.5B,C, D, E, F, G and H); minor subvertical N-S fractures

and veins (Fig.5C, D, and F); and subvertical E-W bed-parallel veins (Fig. 4C, D
and F).

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#### 208 4.2 Macroscale geometry and kinematics of the Cafarnaum fault

The compiled structural map presents the primary structural alignments in 209 an area of 38,000 km<sup>2</sup> (Figs. 2A, B, C, 3A, B). The tectonic structures were sorted 210 into strike-slip faults, normal faults, reverse faults, and fold hinges based upon 211 212 new original data interpretation and previous geological mapping (Cazarin et al., 2019; D'Angelo et al., 2019; Ennes-Silva et al., 2015; Souza et al., 2003). The 213 214 satellite alignments were divided into the following three fault sets: NNE-SSW, 215 NE-SW, and NW-SE. The NNE-SSW set is generally associated with strike-slip 216 left-lateral faults, as already presented by D'Angelo et al. (2019) and Danderfer Filho et al. (2015). A zone of restraining bends coincides with an uplifted area 217 218 (Fig. 2A). Additionally, drag folds occur on the west side of the Cafarnaum fault. The folded structures are E-W-striking thrusts of the Irecê Basin that bend where 219 220 they reach the Cafarnaum fault. Bed layers associated to these thrust faults were tilted to the subvertical position (Fig. 4A). In some case, the aforementioned 221 222 subvertical bed interfaces were affected by shearing as displayed by several 223 kinematics indicators (Fig. 4B). Many mineralized portions documented as bed parallel veins occur along bed interlayers (Fig. 4C, D and F). Some of the thrust 224 faults display drag folds as they approach the Cafarnaum fault. In contrast, the 225 226 NE-SW- and E-W-oriented sets are composed of reverse faults, as shown by D'Angelo et al. (2019) and Reis et al. (2013). 227

The NW-SE fault set has been interpreted as composed of normal faults (D'Angelo et al., 2019). Termination zones of horsetail geometry occur at both

ends of the Cafarnaum fault (Fig. 2C). A few minor faults associated with veins 230 and breccia bodies also occur at the central part of the fault zone, as at the MAM 231 site (Fig. 5). These zones are marked by NW-SE-striking extensional faults 232 located at the extensional guadrant of the main N-S-striking fault zone (La Bruna 233 et al., 2021). There, hydrothermal deposits concentrate on a 20 km wide zone on 234 both sides of the central fault (Fig. 2A, C). Several hydrothermal minerals (e.g., 235 barite, galena) were documented in the Mam outcrop (Fig. 5B, C, D and E). In 236 these sites, complex vein/fracture networks were observed. Both veins and 237 fractures form a principal NW-SE striking set and a minor NE-SW set. 238

Both the NE-SW- and NW-SE-striking fault sets terminate against the NNE-SSW- to N-S-striking fault set. Eastward from the Cafarnaum fault, a larger folded zone was documented (Fig. 2B, C). Previous works have described how this sector is affected by several anticlines and synclines (Cazarin et al., 2019; D'Angelo et al., 2019; Danderfer Filho et al., 2015; Ennes-Silva et al., 2015; Souza et al., 2003). The fold hinges mainly trend along the NNE-SSW to N-S directions (Fig. 2B, C).

An area of ca. 14,000 km<sup>2</sup> was analyzed for structural magnetic lineament 246 map characterization (Fig. 3A, B). The 2,818 documented lineaments were sorted 247 into, I-order, II-order, and III order lineaments. The I-order lineaments are related 248 to regional-scale magnetic features and are composed of a major NE-SW set and 249 minor sets striking N-S and E-W, respectively. The II-order lineaments are 250 characterized by a major NE-SW lineament and a minor NW-SE to WNW-ESE 251 set associated with secondary magnetic anomalies. Both I- and II-order 252 lineaments are crosscut by the III order lineaments. 253

The III order lineaments exhibit a singular magnetic pattern, distinguished 254 by striking high-amplitude magnetic lineaments with the extension of dozens of 255 kilometers (Figs. 3A, B). They occur isolated or comprising sets with main NW-256 SE and minor NE-SW orientations. Sometimes, they are stepwise segmented (en 257 echelon) and shifted by hundreds of meters. They truncate other magnetic 258 lineaments in high to moderate angles, indicating a more recent geological event. 259 Due to high magnetization contrast with bedrocks, linear and extensive 260 waveform, and sparse spatial distribution, we associate these magnetic 261 lineaments with mafic dikes as many authors elsewhere (e.g., Schwarz et al., 262 263 1987; Demarco et al., 2020). In fact, the NNW-SSE trending Chapada Diamantina mafic dike swarm is intrusive into the Mesoproterozoic sedimentary sequences 264 of the Espinhaço Supergroup within the Paramirim Aulacogen (Brito, 2008; 265 266 Silveira et al., 2013). The magnetite-bearing dikes are fine to medium-grained diabases. Recently Pessano et al. (2021) associated NW-SE oriented magnetic 267 anomalies in the central portion of the São Francisco Craton with 268 Mesoproterozoic dikes of the Chapada Diamantina swarm. 269

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# 271 4.3 - Hydrothermal vein and breccia characterization

Two main types of structures related to hydrothermal activity occur in the Cafarnaum fault zone and its surroundings. The first type is hydrothermal breccias, which mainly occur in dilational jogs. The second is calcite veins, which are widespread in the fault zone and the host rocks away from the fault. Two sites of hydrothermal breccias occur along NW-SE striking extensional faults, which we describe below. In addition, we describe four sites with calcite veins at varying distances from the Cafarnaum fault. 279 Hydrothermal breccias are characterized by the interaction between rocks and hydrothermal solutions and are geometrically characterized by several 280 parameters, such as morphology, particle size distribution, fabric, and expansion 281 radius (Jébrak, 1997). Chemical and physical mechanisms can form these 282 breccia bodies: the first by selective dissolution and the second by the excess 283 tension exerted, which exceeds the tensile resistance of the material, or in some 284 285 combination. The analyzed bodies are classified as mosaic breccias that are formed by fluid-assisted breach (hydraulic fracturing) using the classification 286 287 given by Jebrák (1984b). Carbonate clasts are present, and clasts that are larger than 2 mm range between 60-75% of clasts and 75-100% of clasts. 288

Hydraulic breccias are rocks composed of angular to subangular fragments of dimensions ranging from 0.4 cm to 5 cm that are cut by several generations of fractures and veins. The fragments are present throughout the breccia bodies and are derived from adjacent rocks. Generally, they are monolithologic and represented by carbonate rocks corresponding to the Salitre Formation.

Carbonates also contain calcite and quartz geodes surrounded by intense 295 296 oxidation (limonitization). The interfragmentary filling is composed of iron oxide or quartz and calcite cement. The oxidized matrix is probably formed to replace 297 rich materials in iron from the cementing fluid. Quartz is associated with calcite, 298 299 galena, and malachite. There is intense veining by a network of veins with a branched structure that range from millimeters to centimeters thick and reach 300 30% of the total volume, and calcite veins of lesser thickness are related to quartz 301 302 veins.

The description of carbonates can be compared to descriptions made by 303 Souza et al. (1993), such as the association of intensely closed algal laminites 304 corresponding to the Nova América inferior subunit (transgressive cycle III) at the 305 MEL site, as well as area descriptions by Misi (1975) of fine dolomites with 306 307 ankerite, barite and galena and light dolomites with millimeter bands, calcite impregnations and microcrystals of pyrite and galena. Nevertheless, occurrences 308 309 only include sedimentary gaps along the Irecê Basin (Bonfim et al., 1985, and Souza et al., 1993). 310

The descriptions were made under transmitted light optical microscopy of the samples corresponding to MAM and MEL sites (Fig. 2A, C); the samples were divided into laminated carbonates with veins, laminated carbonates with a brittle aspect, and only the veins. The composition in these three divisions presents the same mineralogy in carbonates and the same mineralogy in the veins, differentiated by laminar and massive textural aspects in the case of carbonate alteration and fragmented textural aspects.

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#### 319 4.4 Mineralogy, texture, and isotope geochemistry

Sites from the central part of the basin (SOR, IRE, and ACH) and the eastern block (FAR) (Fig. 2C) are large carbonate pavements in which primary sedimentary features are observed. Except for the FAR site, these carbonates display vertical bedding and are crosscut by centimetric calcitic veins.

The limestones from the SOR site consist of microbial mats. The beds are oriented subvertically within tight N-S-trending folds (Fig. 2C). Two types of veins are observed at the SOR site. The first type of veins formed on transverse fault planes, which originate in E-W-striking low-angle thrust faults, predating folding to the present-day subvertical position (T16 and T19). The second type forms N-S-trending veins. These veins are associated with younger N-S shortening (T24). Isotope profiles across the veins reveal no significant difference in  $\delta^{13}$ C and  $\delta^{18}$ O concentration between the carbonate host rock and the transverse fault carbonate filling (Fig. 6A-D).

The FAR site (Fig. 6E-H) consists of stromatolite bioherm colonies. Two 333 veins were analyzed. The first vein (T9) occurs in inter-stromatolite silt crosscut 334 by a sharp-edged 2.2 mm thick vein, filled with mosaic equidimensional 335 transparent calcite varying from 0.05 mm to 0.2 mm. The other vein (T10) 336 crosscuts a stromatolite unit. It has two different calcite fabrics: an 337 equidimensional mosaic with oriented sparry calcite crystals varying in size from 338 0.1 to 1 mm and a milky white area with no calcite crystals. Most calcite veins 339 340 from the FAR site display lower  $\delta^{13}$ C and  $\delta^{18}$ O values than the stromatolite host rock. Exceptions are a few carbonate samples from the vein that crosscut the 341 inter-stromatolite silt (Fig. 342 6G-H).

The ACH site consists of subvertical NNW-dipping dolostones. The 343 present-day NNE-SSW burial fault carries most veins and is more prominent than 344 345 its conjugate NW-SE counterpart (Fig. 7A-D). Veins at the ACH site lack signs of shear and are syntaxially filled with blocky crystals. In addition to veins, the 346 347 ACH site contains a pocket filled with crystal precipitates (T5) in which borders are vein wall remnants and dissolved host rock peloids (Fig. 7B-C). Since these 348 pockets dissolve burial-related fractures and faults, they formed later than the 349 burial fault set. In general, ACH samples show a higher porosity than other sites. 350 351 Isotope data across these veins reveal much lower  $\delta^{13}$ C and  $\delta^{18}$ O values than the isotope values of the host rock (Fig. 7C-D). 352

Carbonate rocks of the IRE site consist of thinly layered (0.2-20 cm) black 353 limestones with local slumps. The beds are vertically tilted and folded, with an 354 orientation of 355/86 (dip azimuth direction). Two types of veins are present (Fig. 355 97E-F): thinner (<1 cm) bedding-perpendicular and bedding-parallel pre- or 356 synfolding veins and thick veins (20 cm) filled with a mixture of calcite and barite 357 that irregularly crosscut folded bedding. They are syn- or post-folding veins. 358 359 Similar to the SOR site veins, IRE veins have isotope values similar to the host rock (Fig. 7E-F). 360

In contrast to sites in the central part of the basin, sites at the basin edges 361 (i.e., MEL and MAM) are strongly deformed and exhibit pervasive hydrothermal 362 features and hydraulic hydrothermal breccias (Fig. 8A-C). Samples from the MEL 363 site are mostly breccias with white calcite cement. Previous studies by Misi et al. 364 365 (2005) argue that the Salitre Formation stratigraphy controls the Pb-Zn concentrations. According to these authors, the Pb-Zn ore is associated with 366 silicified stromatolites that overlie a shallowing-upward sequence (Unit B1). The 367 MEL site is located in a flat area where sites are restricted to trenches dug by 368 mining operations. Samples may exhibit primary lamination similar to rocks of the 369 370 Salitre Formation. These features are obliterated by tectonic and hydrothermal processes toward the center of the brecciated zone at the outcrop scale (Fig. 8A). 371 372 Clasts of primary carbonate rocks may also occur in the hydrothermal breccia 373 cemented by white calcite devoid of laminations (Fig. 8B). Secondary vugulartype porosity that is less than 1% in the area is also observed under the 374 microscope. The limestones and breccias are crosscut by two generations of 375 376 veins: one measuring 4 mm thick and made of calcite and, quartz microcrystals and a second measuring approximately 0.3 mm thick and made of quartz. The 377 host carbonate and the cement display quite distinct mineralogy. The primary 378

Salitre carbonates consist mainly of calcite, dolomite, ankerite, siderite, iron 379 oxide, and limonite. SEM, XRD, and QUEMSCAN data indicate that the veins 380 and hydrothermal breccias display complex mineralogy, where the main minerals 381 are calcite, dolomite, galena, barite, quartz, sphalerite, illite, chlorite, zincite, 382 cerussite, malachite, magnesite, apatite, and chalcedony. Barite, apatite, chlorite, 383 and guartz are concentrated along fracture zones (Fig. 8D-G) and, in some 384 385 instances, may form larger aggregates. Thin section observations indicate that tectonic and hydrothermal processes were accompanied by the formation of 386 stylolites, dolomitization, silicification, limonitization, microfractures, folding, and 387 minerals with wavy extinction. 388

Samples (Figs.8D and E) at the MEL site display different facies of 389 carbonate breccia and isotope  $\delta^{13}$ C and  $\delta^{18}$ O analyses performed at specific 390 391 points in the samples. While the  $\delta^{13}$ C values range between 0.22‰ and -2.24‰, the  $\delta^{18}$ O values vary from -6.16‰ to -12.67‰. In most instances, the primary 392 limestone fabric was replaced by milky carbonate with large anhedral crystals. 393 Sample in Fig. 8C displays a cyclic succession of milky calcite and iron-rich 394 dolomite crosscut by veins containing galena. High-resolution  $\delta^{13}$ C and  $\delta^{18}$ O data 395 396 indicate that these samples may have an area with homogenous isotope values (Fig.8D-E), as well as areas with variable isotope values. 397

The MAM site is also located in a flat area in which sites are restricted to trenches of the mining operation. Centimetric veins of milky quartz occur in the carbonate host unit. As at the MEL site, XRD and SEM data at the MAM site indicate galena, zincite, ankerite, dolomite, cerussite, apatite, magnesite, anglesite, chlorite, and illite. However, in contrast to the MEL site, the MAM site exhibits strong silicification that may completely obliterate the primary carbonate texture. For instance, the QUEMSCAN images exhibit carbonate breccias replaced by silica in which ghost clasts can still be identified. Pores and laminated
illite and chlorite areas indicate that silica-rich fluids replaced mostly carbonate
minerals

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409 4.5 - Fluid inclusions

Table 01 presents the H and O isotopic compositions of fluid inclusions for 10 410 samples from the ACH, IRE, and SOR sites in the Irecê Basin and at the FAR 411 site in the eastern block of the Cafarnaum fault. The supplemented materials 412 detail linearity and memory effect corrections applied to the H and O fluid isotope 413 data. Table 01 also shows the average δ<sup>18</sup>O isotopic composition of the carbonate 414 associated with the fluid inclusions and the calculated temperature based on the 415 isotope fractionation between calcite and water. The calculated temperature 416 417 range is 40-74°C, with the highest values obtained in samples from the ACH and FAR sites 418

Except for one sample from the ACH site (ACH01.3Aa T5-1, left), all samples display negative  $\delta D_{SMOW}$  values. Fig. 9 plots the H and O isotope values with the global meteoric water line (GMWL) by Rozanski et al. (1993) and shows that the trending lines of samples from the ACH and FAR sites converge to a  $\delta D$  value of approximately -45‰ and a  $\delta^{18}O$  value of -6.5‰.

424

#### 425 **5. Discussion**

426 5.1 – Fault evolution and hydrothermal fluids

427 Cratons are composed of thick and cool lithospheric keels with high resistivities 428 and low porosities (e.g., Ferguson et al., 2012; Selway, 2014). However, several 429 studies have increasingly indicated that cratons present low-resistivity zones in 430 the lithosphere that behave as weakness zones prone to deformation, such as

ductile shear zones and faults (e.g., Pinto et al., 2010; Thiel and Heinson, 2013; 431 Dong et al., 2015). These shear zones/faults provide a high permeability and are 432 pathways for deep-seated fluids to ascend through the whole lithosphere (Caine 433 et al., 2010; Bense et al., 2013). The study of hydrothermal fluids in fault zones 434 in cratons may explain the permeability of fault zones and host units and deep 435 geothermal exploration constraints (Taillefer et al., 2017). The boundary between 436 437 the Irecê Basin and the Chapada Diamantina Group is marked by the Cafarnaum fault (Figs. 1, 2) (La Bruna et al., 2021), which coincides with a high conductor 438 mapped by a magnetotelluric survey (Fig.10). This zone was prone to 439 hydrothermal activity and is consistent with high porosity-permeability, high-fluid 440 salinity, and sulfide emplacement. In a few cases, it served as a conduit for mafic 441 volcanism (Teixeira et al., 2017; Padilha et al. 2019). 442

Faults behave as high permeability conduits that facilitate fluid flow in the 443 Earth's crust (Cox and Munroe, 2016). Deep-seated hydrothermal fluids 444 445 precipitate minerals that form veins and breccias and decrease the permeability of the lithospheric fault zones (Sibson et al., 1988, 1990). The precipitation of 446 hydrothermal minerals in fault zones and host rocks is caused by decompression 447 and cooler conditions (Calvin et al., 2015). Therefore, based on the 448 characteristics presented here, the Cafarnaum fault was a structure prone to 449 hydrothermal activity. It controlled the upward hot fluid flow, indicated by the 450 relationship between fault geometry and hydrothermal deposits (Fig. 10). 451

452 . Hydrothermal fluids channelized along faults can affect thousands of 453 square kilometers, even in nonmagmatic settings (Nabavi et al., 2020). The 454 hydrothermal activity in faults with shallow crustal levels under fluid overpressure 455 is controlled by the geometry of the crustal-scale fault zone (Bellot, 2008). 456 Dilational jogs flanking continental-scale strike-slip faults, for example, are locations prone to hydrothermal boiling and implosive brecciation (Sibson, 1987).
Active examples of hydrothermal activity in dilational jogs occur in compressional
settings such as New Zealand (Brathwaite et al., 1986 in 24) and the French
Pyrenees (Taillefer et al., 2017).

We interpret the N-W-striking faults that arrest against the main N-Strending Cafarnaum fault as dilational jogs that facilitate hydrothermal fluid ascension and flow (Fig. 10). Hydrothermal minerals are concentrated in dilational jogs on both sides flanking the Cafarnaum fault, either on carbonate units in the western block or mainly on siliciclastic units in the eastern block (Fig. 2). This indicates that the structure, rather than the lithology, controls the fluid flow along the fault.

Hydrothermal breccias occur in dilational jogs on both sides of the main Cafarnaum fault (Fig. 2A, C). These breccias form when fluid migration becomes explosive (e.g., Jébrak, 1997). Subsequent precipitation of hydrothermal minerals forms breccias and heals the fault, decreasing their permeability (Katz et al., 2006; Taillefer et al., 2017). Fluid flow and subsequent precipitation have been interpreted with the seismic cycle and fault-valve behavior, influencing breccia occurrence (Taillefer et al., 2017).

The results shown in this study document how strike-slip faults such as the 475 Cafarnaum fault are efficient pathways for fluids and that these fluids caused the 476 widespread silicification and the precipitation of Ca-bearing minerals. A recently 477 performed in the Morro Vermelho Karst System, located some 150km to the S of 478 the Cafarnaum fault and developed within the same carbonate succession, 479 provides more insights into the temporal succession of fluid circulation and 480 481 precipitation/dissolution (Bertotti et al., 2020). During the first stage, fluids flowed along two main aquifers, the Chapada Diamantina guartz arenites and the 482

overlying Salitre Formation carbonates, separated by the Bebedouro Formation 483 glacial sediments aguitard. The flow was associated with pervasive dolomitization 484 of a 100s of m wide body overlying a deep-seated strike-slip fault in the 485 carbonates. With increasing displacement, the strike-slip fault grew upward (e.g., 486 Dooley and Schreurs, 2012), thereby first affecting the Chapada Diamantina 487 aquifer and eventually reaching the Salitre carbonates. With establishing a 488 through-going fracture zone, Si-rich fluids previously confined to the Chapada 489 Diamantina aquifer invaded the Salitre aquifer, causing widespread dissolution 490 and karst formation and precipitation of Si both in the host rock as a silica-crust 491 coating the caves. We suggest that such a temporal succession could also be 492 applicable to the mineralizations of the Cafarnaum fault zone. 493

494

#### 495 5.2 – Geochemistry of hydrothermal fluids, fluid pathways and tectonics

Fluid flow in the crust is a powerful mechanism to remobilize chemical 496 elements and concentrate metals of economic interest (Heinrich and Candela, 497 498 2014, Yardley and Bodnar, 2014).. The efficiency of heat transfer and chemical remobilization depends on heat and chemical gradients, fluid-rock interactions, 499 and tectonic settings. Many studies have addressed the question of how deep 500 fluid penetrates the crust (e.g., Nesbitt and Muehlenbachs, 1989; Fricke, 501 Wickham et al., 1992; Haines, Lynch et al., 2016). Most authors agree that fluid 502 503 may penetrate as deep as 10 to 15 km into the crust, mainly in crystalline terrains submitted to an extensional tectonic regime. 504

The isotope data presented here show that fluid sources in the central part and at the Irecê Basin edges had the same origin but underwent distinct pathways. The new data presented in this study allow a discussion of the source of these fluids, their primary isotopic composition, their interaction with the sedimentary rocks on both sides of the Cafarnaum fault, and how deep they may
have penetrated each kind of terrain.

511 Different generations of carbonate veins and breccias crosscut the 512 sedimentary rocks of the Irecê Basin. Figs. 6 and 7 show that carbonate 513 veins from the central part of the basin do not display significant carbon and 514 oxygen isotope differences relative to the carbonate host rock. More significant 515 isotopic differences occur in samples from the ACH and FAR sites, where the 516 veins present more negative isotopic values than the host rock (Fig.6).

Data in this study suggest that the  $\delta^{18}$ O value of the carbonate veins may 517 be explained by a meteoric fluid source (Fig.11), by a higher temperature of 518 crystallization (Table 02) of these carbonates, or by the combination of both 519 processes. The oxygen isotope fractionation between calcite and water varies 520 521 from 28‰ to 7‰ in the temperature range of 25-250°C (Chacko et al. 1991; Kim, O'Neil et al., 2007; Chacko and Deines, 2008). In contrast, the low  $\delta^{13}$ C values 522 observed in the carbonate may only be explained by an external source of 523 carbon, since the carbon isotope fractionation values between calcite and HCO3-524 and calcite and CO<sub>2</sub> are much less than 4‰ at temperatures below 200°C 525 526 (Deines et al., 1974; Chacko et al., 1991; Chacko and Deines 2008). Samples from the MEL site, which are located at the edge of the basin, display even more 527 528 negative oxygen isotope values. As shown in Fig. 11, isotope data from this site display a narrow range of  $\delta^{13}$ C values and a wide range of  $\delta^{18}$ O values. Compared 529 to the veins from the central part of the Irecê Basin, the lower δ<sup>18</sup>O values of their 530 carbonates indicate interactions with more <sup>18</sup>O-depleted fluids or higher 531 crystallization temperatures. 532

Fig. 11compares the isotopic composition of the carbonates studied here with previous isotope data reported for the same area. The diagram shows that

the data in this study have a wide range of  $\delta^{18}O$  (-13.0‰ to 1.8‰) and  $\delta^{13}C$  (-535 10‰ to 10‰) values. However, most of our samples have a narrower range of 536  $\delta^{13}$ C (-5‰ to 1‰). The exceptions are samples from the IRE site that exhibit 537 higher  $\delta^{13}$ C values and a few samples from the ACH, FAR, and SOR sites that 538 present  $\delta^{13}$ C values below -5‰. Samples from the IRE site are associated with 539 <sup>13</sup>C-enriched carbonates from the upper section of the Irecê Basin and plotted as 540 541 squares in Fig. 9 . These primary high  $\delta^{13}$ C carbonates have been reported in both the Irecê Basin (Misi 1988, Misi and Kyle 1994, Borges, Balsamo et al. 2016, 542 Caird, Pufahl et al. 2017) and other Neoproterozoic basins (Santos, Alvarenga et 543 al. 2000). Carbonate veins with  $\delta^{13}$ C values below -5‰ are probably related to 544 the same fluid that is responsible for the carbonates that formed the calcretes 545 previously described in the basin (Borges, Balsamo et al. 2016, Caird, Pufahl et 546 al. 2017). Published isotope data of these carbonates, plotted as "stars" in Fig.11, 547 also present low  $\delta^{13}$ C values. 548

The source of fluids related to the carbonate veins from the central part of 549 550 the Irecê Basin may be further constrained by the isotopic composition of fluid inclusions trapped in the carbonate veins. Based on the oxygen isotopic 551 composition of these fluid inclusions and the associated carbonate, we estimate 552 the temperature of carbonate crystallization to be between 39 and 67°C (Table 553 02). These temperature estimates were based on Kim and O'Neil (1997) oxygen 554 isotope fractionation equation between calcite and water. Since calcite and fluid-555 inclusions may be affected by post-entrapment isotope exchange during 556 exhumation (Nooitgedacht et al., 2021), these results should indicate minimum 557 temperature conditions. Assuming an average thermal gradient of 30 C/km, these 558 fluids circulated at depths reaching 1000 m within the crust. Fig. 9 displays the 559 mean global meteoric water line by Rozanski et al. (1993) and the hydrogen and 560

oxygen isotope values of these fluid inclusions. It also shows trending lines for 561 fluid inclusions from the ACH and FAR sites, indicating that they converge to a 562  $\delta D$  value near -45‰ and a  $\delta^{18}O$  value near -6.5‰ along the mean meteoric water 563 line (Fig. 9). We argue that these isotopic values represent the local meteoric 564 water, which upon interaction with the host rock changed its isotopic composition 565 along the mixing lines. A similar process has been described in active 566 567 hydrothermal systems (Criss and Taylor Jr, 1986), in which there is also a more extensive range of  $\delta^{18}$ O values compared to  $\delta$ D values. 568

The isotopic variation observed in Fig. 11 may be explained by different 569 geological scenarios. Arrows I and II represent veins formed by mixing meteoric 570 571 fluids and carbonate host rocks of the inner part of the basin at low-temperature conditions (between 30 and 40°C). Arrow I indicates mixing between these fluids 572 573 and carbonates from the lower section. In contrast, arrow II represents the mixing between these fluids and the <sup>13</sup>C-enriched carbonates from the upper section. 574 Arrow III represents carbonates formed by the same fluids as those of the 575 calcretes, which have more negative  $\delta^{13}$ C values. Samples from the MEL site, 576 represented by arrow IV, fall within the same range of  $\delta^{13}$ C values for most 577 samples from the central part of the basin. However, they also have more 578 579 negative  $\delta^{18}$ O values, indicating that isotope exchange between the meteoric fluids and carbonates alone may not be reconciled with the observed data. We 580 argue that carbonates from the MEL site crystallized from the same meteoric fluid 581 but at higher temperatures. These fluids percolated through conduits that allowed 582 them to reach deeper parts of the crust and return to shallow crustal levels without 583 losing much heat. Compared to thrust systems alone, thrust followed by strike-584 slip and extensional faulting may provide deep fluid conduits. Based on hydrogen 585 isotopes, Nesbitt and Muehlenbachs (1989) concluded that the tectonic regime 586

might drastically control the depth of fluid interaction in the crust. This interpretation also agrees with previous studies based on fluid inclusions and sulfur isotope geothermometry performed in the MEL site area (Misi and Kyle 1994, Misi, Iyer et al. 1999, Misi, Iyer et al. 2005), Fluid-rock interactions may also explain the presence of base metals in these high-temperature veins, suggesting that these chemical elements were scavenged from deeper crustal levels.

The role of deep crustal fluids at the MEL and MAM sites is further 594 suggested by the petrographic and mineralogical features observed at these sites 595 (Fig. 8). In addition to the high concentration of base metals (e.g., Pb, Zn, Fe), 596 these sites display a pervasive replacement of the primary carbonates by silica. 597 598 Silicification events are recognized as a diagenetic process in which Si-rich fluids affect a host rock, modifying its texture and mineralogy (Menezes et al., 2019). 599 For example, Haldar and Tislier (2014) documented a silicification process where 600 601 opal/chalcedony/low-temperature quartz replaces calcite/aragonite/dolomite. 602 The percolation of meteoric fluids at the deep crustal level provided the required conditions to mobilize silica at the MEL and MAM sites. 603

604 Deep meteoric fluid circulation along the Carfarnaun fault is comparable to other worldwide geological examples in which surface waters penetrate along 605 606 tectonict structures. For instance, based on the Friedman and O'Neil (1977) isotope fractionation diagram, Mozafari et al. (2015) reconstructed the 607 608 composition of the parent fluids that were in equilibrium with the vein infill in the 609 Jabal Qusaybah Anticline (Adam Foothills, North Oman). The authors discussed 610 a paleofluid evolutionary model related to the deformation front of the foreland 611 fold and thrust belt. The strike-slip tectonic regime introduces a different type of fluid flow linked to the fault zone. The shifts in the stable isotope values forming 612

613 different fields with comparable values (Fig. 12) possibly originate from different phases of strike-slip movement. A similar spread in the fluid inclusion isotope 614 ratios was discussed by de Graaf et al. (2020) after analyzing the wide variety of 615 vein-type mineralization caused by deep-seated brines in the Hartz Mountains, 616 617 Germany. Other studies examined the complex recrystallization processes along dolomitization fronts causing dissolution of the host limestones and precipitation 618 619 of dolomite crystal structures depending on the porosity-permeability properties of the sediments present in the fault systems (Koeshidayatullah et al., 2020). 620

Similar to the Cafarnaun fault, the interplay between tectonics and fluid 621 circulation has played a major role in the evolution of the inner Northern 622 Apennines. As shown by Brogi et al. (2020), active transfer zones associated 623 with an extensional tectonism control the deposition of travertine deposits by 624 enhancing fluid circulation. Other studies in the same area have shown that 625 626 these trending faults have also controlled the development of magmatic activities 627 (Brogi et al., 2021) and Hg-Sb ore deposits (Brogi et al., 2011), further suggesting that these structures may connect different geological systems and 628 play a major role in the remobilization of chemical constituents. They also 629 630 indicate that competition between crustal stretching and surface uplift continuously switches the local intermediate stress axis, thus promoting quick 631 changes in the direction of the maximum permeability (Liotta and Brogi, 2020). 632 These changes further promote lateral and vertical migration of fluids within the 633 634 system.

635

## 636 **6. Conclusion**

This study in the São Francisco Craton focuses on the hydrothermal activity in the Cafarnaum fault and its host rock and yields the following conclusions. The São Francisco Craton is a cold and thick block preserved from
deformation in the Brasiliano orogeny (740-580 Ma). However, a few tectonothermal events affected the Craton along its boundary in Neoproterozoic times.
One of these events was repeated hydrothermal activity along the Cafarnaum
fault, a N-S-striking, 170 km long, strike-slip fault that juxtaposes Neoproterozoic
carbonate units and Mesoproterozoic siliciclastic-carbonate units in the northern
part of the Craton.

Hydrothermal boiling and implosive brecciation occurred along the fault. 646 Decompression and cooler conditions induced precipitation of hydrothermal 647 minerals in N-W-striking dilational jogs, mainly flanking the northern and southern 648 fault terminations. The hydrothermal fluid structures are composed of 649 hydrothermal breccias close to the main fault zones and along dilational jogs. In 650 addition, calcite veins in the host units away from the fault are also part of the 651 hydrothermal system. Therefore, the geometry of faults at shallow crustal levels 652 653 influences the location of hydrothermal deposits.

Based on isotopic geochemistry, we show that meteoric water was the 654 655 main fluid source that percolated the sedimentary rocks of the Irecê Basin and the Cafarnaum fault zone. Fluid inclusions in carbonate veins from the central 656 part of the basin indicate a meteoric fluid with a  $\delta D$  value near -45‰ and a  $\delta^{18}O$ 657 658 value near -6.5%. Temperature estimates based on the oxygen isotopic fractionation between the carbonate veins from the central part of the basin and 659 the trapped fluid inclusions indicate temperatures ranging between 40 and 70°C. 660 These temperature conditions agree with the lower  $\delta^{18}$ O values of veins 661 compared to the carbonate host rock. A similar  $\delta^{18}$ O fluid value (-6.5‰) is 662 obtained based on the interaction between these fluids and carbonates from the 663 lower and upper parts of the succession. 664

In contrast, carbonates at the front edges of the basin associated with the Cafarnaum fault exhibit much lower  $\delta^{18}$ O values, indicating higher crystallization temperatures. These carbonates are also associated with base metals and silicarich fluids, suggesting that the fault behaved as a conduit for deeper fluid circulation in the basement. The mineral paragenesis (e.g., galena, sphalerite, barite, chlorite, illite, and quartz) and brecciated features associated with the veins and fault support this interpretation.

672

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Figure 2. (A) Simplified geological map modified after Levantamento Aerogeofisico da Área de Centro Norte Bahia - CBPM, 2011/12 Geologia - Mapa Geológico do Estado da Bahia - CBPM/CPRM, 2003. (B) Structural framework of the study area superposed on ALOS PALSAR imagery. Dashed lines represent the inferred faults. Continuous white lines represent both anticline and syncline axes. (C) Line drawing of the structural framework presented in B. (D) Lower hemisphere equal-area projections of great circles representing the attitude of the documented faults and equal-area projection/density contour plots of the poles of the measured faults. (E) Lower hemisphere equal-area projections of fold hinges and relative density contour plots. (the map location can be found in Fig. Geological map).



Figure 3. Behavior of magnetic lineaments in the study area: (A) Map of first derivate. (B) Lineaments Line drawing of the map presented in A. (C-D-E) Rose diagram representing the attitude of the documented first, second and third order lineaments.



Figure 4. (A) Outcrop view of subvertical bed layers in the Achado outcrop; (B) Close up view of a bed parallel slip surface. Kinematic indicators, such as striae are visible along the slip surface. (C) Close up view of a large bed parallel vein and bed perpendicular fractures; (D) Close up view of bed parallel veins and bed perpendicular fractures; (D) Close up view of bed parallel veins and bed perpendicular fractures; (D) Close up view of bed parallel veins and bed perpendicular fractures and veins. (E) Rose diagramm of the documented fractures in the Ach outcrop. (F) Lower hemisphere equal area projection of veins measured in the Ach outcrop. (G) and (H) Rose diagramm of the documented fractures and veins in the Far and Sor outcrops. See Fig. 2 for outcrops location.



Figure 5. (A) Digital elevation model (DEM) of the Mamonas outcrop (outcrop localization in fig. 2. Mam). Red lines are associated to the structural lineaments; (B)(C) Outcrop view of veins and fractures. (D) Outcrop view of a large vein located in a fault damage zone. (E) Rose diagramm of the identified structural features from the DEM in Fig. 5a; (F) Rose diagramm of the documented veins and fractures displayed in Fig. 5b; (G) Rose diagramm of the documented veins and fractures displayed in fig. 5c; (H) Rose diagramm of the documented veins and fractures displayed in fig. 5d. See Fig. 3 for outcrops location.



Figure 6.A) On the left, thin section SOR01.2Aa. On the center graphical representation of SOR01.2Aa with veins in green and fault breckla in multicolor. On the right, zoom in on sheeted character of vein T16. B) On the left, thin section SOR01.2Ab. On the center graphical representation of SOR01.2Ab. On the right zoom in on bedding parallel styloites and NNE-SSW fractures. C) On the left, thin section SOR01.3Aa. On the center Micro map of all present features. On the right, crosscuting of shear fracture T19 and N-S vein T24. D) Stable isotope transects of T16.T19 and T24. E) Thin section FAR02.3Bb. On the center, graphic mpresentation of thin section. On the left, zoom in on transect T10 across two calcite phases including crystalization F) Transect T10. G) Thin section FAR014D, and zoom in on vein T9 with sampling location including crystalization. H) Transect T9.



Figure 7. A) Thin section ACH01.4Ba with T8 location and two zones and, zoom in on zone boundary between (1) no host rock remnants and (2) with host rock remnants 8) Thin section ACH01.3Aa and, zoom in on transect T5. C) Transect T5. The red sites mark a change in 5 "C-inu and a new growth plane. D) Stable lootope transect of T6. Note the difference in 5"O. E) Thin section IRE01.2Aa, and, zoom in on vein infill with needle shaped aregonite remnants. F) Stable lootope transects of vein T11, T13 and T14 from IRE. Host rock and vein infill have an almost identical lootope composition.



Figure 8. Samples from Melancias area showing the effects of hydrothermal alteration and hydraulic fracturing. A) Carbonate that shows partially preserved primary features grading to a brecciated zone. In the lower part of the sample, the primary features were obliterated by the hydrothermal alteration. B)Vein of carbonate by sucessive precipitation of carbonate and Fe-rich zones. C) Hydralic breccia with clasts of the primary carbonate. D) and E) Thin section from Melancias in which calcite is the main mineral. The diagram also presents  $\delta^{13}$ C and  $\delta^{18}$ O across profiles. F) and G) QemScan image of samples from Melancias showing the hydrothermal mineralogy made of barite, quartz, illite, and chlorite. Note that these mineras concentrate along fractures.



Figure 10. Fluid inclusion data GMWL based on Rozanski et al. (1993). Regions of marine and meteoric waters based on Moore (1989). Sample plot in both meteoric and marine domain. Trend lines of ACG and FAR cross the GMWL on the same point (~-6,5‰ for  $\delta^{18}O_{ysMOW}$ ; -45‰ for  $\delta D_{ysMOW}$ ).



Figure 10. Conceptual model proposed for the study area. Cartoon displaying the current regional scale configuration and the connection with the deep magnetic lineaments.





Figure 11. Carbon and oxygen isotopes of samples from the central part of the basin (IRE, FAR. ACH, and SOR) and samples from the Melancias outcrop compared to other published isotope data from the basin. This plot shows different scenarios of isotope evolution: I. Interaction between meteoric fluid and carbonates of the lower stratigraphic section; II. Interaction between meteoric fluid and carbonates of the upper stratigraphic section; III. Low d<sup>13</sup>C carbonates related to the same fluids responsible for calcrete formation in the area; IV. Carbonates formed by high-temperature fluids that interacted with deep-crustal levels. Hostrock data (Misi and Kyle, 1994; Misi and Veizer, 1998; Borges et al., 2016; Caird et al., 2017); Calcrete data (Borges et al., 2016; Caird et al, 2017).

Table 1: Isotopic composition of carbonates from central part of the basin. IRE; ACH; SOR; FAR. V refers to vein, HR to host-rock, and BREC to breccia.

Achado - ACH			Fazenda Arrecife - FAR				Irecê - IRE				Soares - SOR				
Sample	δ <sup>13</sup> C	δ <sup>18</sup> Ο	V/HR	Sample	δ¹³C	δ <sup>18</sup> Ο	V/HR	Sample	δ¹³C	δ <sup>18</sup> Ο	V/HR	Sample	δ <sup>13</sup> C	δ <sup>18</sup> Ο	V/HR
Т3	0.1	-5.6	HR	T9.1	-3.2	-5.2	HR	T11.3	9.8	-4.2	V	T16.1	-0.9	-6.8	HR
T3.2	-4.5	-8.2	V	T9.3	-7.7	-7.1	V	T11.5	9.9	-4.1	V	T16.3	-1.4	-7.4	V
T3.3	-5.2	-7.5	V	T9.5	-7.2	-6.8	V	T11.7	9.8	-4.2	V	T16.5	-0.8	-7.1	V
T3.4	-5.6	-8.7	V	Т9.7	-3.4	-5.8	HR	T11.9	9.5	-4.4	HR	T16.7	-0.9	-7.4	V
T3.5	-4.6	-8.6	V	T10.1	-2.5	-7.4	HR	T13.1	9.2	-4.3	HR	T16.9	-0.7	-6.9	V
T3.6	0.1	-5.8	V	T10.3	-3.0	-7.6	V	T13.2	9.4	-4.3	V	T16.11	-0.8	-7.1	V
T4.1	-0.3	-5.2	HR	T10.5	-2.6	-7.6	V	T13.3	9.3	-4.3	HR	T16.12	-0.4	-7.3	HR
T4.2	-0.3	-6.4	HR	T10.7	-2.6	-6.9	V	T13.4	9.0	-4.3	HR	T17.1	0.4	-7.1	V
T4.3	-3.9	-7.2	HR	T10.9	-3.4	-7.5	V	T13.5	9.1	-4.4	V	T17.3	-2.4	-8.2	v
T4.4	-3.4	-7.5	HR	T10.11	-3.5	-7.5	V	T14.1	7.3	-4.6	V	T17.5	-0.1	-6.2	V
T4.5	-0.3	-10.9	V	T10.13	-4.0	-7.7	v	T14.3	7.5	-4.0	v	T19.1	-0.5	-6.6	HR
T4.6	0.4	-10.0	v	T10.15	-2.7	-7.2	HR	T14.5	7.5	-4.6	v	T19.2	-0.5	-6.1	V
T4.7	-0.8	-9.0	v		,	<i>·</i> · <b>_</b>		T14.7	7.1	-4.0	v	T19.3	-0.4	-6.4	v
T4 8	-0.4	-11 7	v					T14 9	61	-4 7	v	T19.4	-0.3	-6.4	v
T4 9	0.5	-10.4	v					T14 11	7.6	-4 1	v	T19 5	-0.3	-6.3	v
T4 10	0.3	-10.4	v					T14.11	5.2	-5.0	v	T19.5	-0.5	-6.9	v
T4 12	0.3	-10.0	v					T14.15	75	-4 1	v	T19.7	-0.5	-6.7	v
T/ 12	0.5	_8.2	v					T1/ 15	7.0	-4.0	v	T10.0	0.3	-7.0	ЧR
T5 1	-0.3	_2 2	ЧR					T14.15	7.5	-3.0	v	T2/ 1	-2.1	-1.6	V N
T5.1	-0.5	-2.J	1 IIX 1/1					T14.17	7.7	-3.5	V	124.1	-3.1	-1.0	v
T5.2	-5.1	-0.2	V					T12 1	7.7 9.7	-4.0	v				
	-0.1	-0.4	v					112.1	0.7	-4.5					
15.5 TE 6	-0.2	-0.7	V												
	-0.1	-9.1	V												
	-9.9	-8.4	V												
	-9.1	-7.7	V												
15.9	-0.0	-2.1													
10.1 TC 2	2.2	-2.0													
16.2 TC 2	-4.7	-8.4	V												
16.3	-5.1	-8.6	V												
16.5 TC C	-2.7	-7.0	BREC												
	1.2	-3.0	BREC												
16.7	-2.4	-7.1	BREC												
16.8	2.2	-2.4	BREC												
	T.P	-3.0	нк												
18.3	-4.0	-7.0	V												
18.5	-6.3	-6.8	V												
18.7	-4.6	-6.8	V												
18.8	-3.0	-5.7	V												
18.9	-2.5	-7.1	V												
18.11	-3.0	-6.9	V												
18.12	-4.2	-7.1	V												
18.13	-5./	-6.4	V												
18.1/	-3.2	-8.3	V												
18.19	-2.4	-8.2	V												
18.21	-1.9	-7.9	V												
18.22	-1.5	-8.0	V												
T8.24	-2.0	-7.6	V												
T8.26	2.1	-2.8	HR												

Table 02			

Sample	Location	Fluid δD <sub>smow</sub>	Fluid δ <sup>18</sup> O <sub>SMOW</sub>	Calcite $\delta^{18}O_{PDB}$	Calcite $\delta^{18}O_{SMOW}$	T in C
ACH01.2Bb T4-1	ACH	-22.7	-2.5	-7.4	23.3	39
ACH01.2Bb T4-2	ACH	-24.0	-3.8	-10.1	20.5	47
ACH01.3Aa T5-1 (left)	ACH	1.9	0.8	-8.6	22.1	66
ACH01.4BaT8-1 (grey)	ACH	-15.7	-1.6	-8.0	22.7	48
IRE01.2Aa T11	IRE	-32.0	1.4	-4.2	26.6	43
IRE01.3Ba T14	IRE	-27.9	0.3	-4.1	26.6	37
SOR01.3Aa T19	SOR	-36.0	-2.5	-6.5	24.3	35
FAR01.4D T9	FAR	-10.5	0.9	-7.0	23.7	56
FAR02.3Bb T10-1 (milky, right)	FAR	-8.2	2.0	-7.6	23.1	67
FAR02.3Bb T10-2 (transparant, left)	FAR	-19.0	-0.7	-7.5	23.1	51

Table 2: Isotope data from fluid inclusions of carbonates from the central part of the basin. The temperature was calculated based on the oxygen isotopic composition of calcite and fluid, according to Kim & O'Neil (1997).