

Decoupling ore formation, ductile fabric and hydrothermal upgrading in the giant Salobo IOCG deposit, Carajás Province, Brazil

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Supplementary Materials SM1 (deposit compilation) and SM3 (geochronological record of Salobo) are appended at the end of this file. SM2 (electron-microprobe data) is available as a separate supplementary file.

Comments and feedback are welcome and can be sent to the corresponding author.

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10 Abstract

11 The Salobo deposit is the largest Cu-Au deposit of the Carajás Mineral Province, northern
12 Brazil, and is hosted by altered volcanosedimentary schists of the Itacaiúnas Supergroup
13 along the Cinzento Lineament. Multiscale structural analysis, mineral chemistry and
14 published geochronology define four structural families: Pre-Dn, Dn, Dn+1 and Dn+2.
15 The Pre-Dn event comprises high-temperature K, K-Fe and Na-Ca hydrothermal
16 alteration, Fe-rich clinopyroxene, and chalcopyrite-magnetite mineralization associated
17 with Old Salobo-type quartz-feldspathic bodies and the ca. 2.57-2.55 Ga hydrothermal
18 system. Fe-rich clinopyroxene supports the high-temperature hydrothermal character of
19 this assemblage. Dn formed the penetrative NW-SE S_n schistosity, parallelizing and
20 deforming pre-existing ore, with localized mylonitic top-to-SW transport along the
21 southern wall-rock contact. Dn+1 produced oblique S_{n+1} foliation and shear bands with
22 top-to-NNW transport. Dn+2 comprises brittle, fluid-dominated structures with bornite,
23 chalcocite and minor late chalcopyrite in undeformed chlorite-altered veins. Chlorite
24 geothermometry, anchored on syn-Dn+2 vein chlorite, indicates ca. 250-290 °C for this
25 event, and chlorite chemistry supports involvement of a felsic fluid component. Together

with the spatial relation to Young Salobo granitic dykes, these data support correlation with the ca. 1.88 Ga granite-related system. Regional correlation favors linking Dn with the Transamazonian collision and Dn+1 with the Sereno event. Salobo is therefore interpreted as a Neoarchean IOCG system that was subsequently metamorphosed, structurally transposed, and locally upgraded during a Paleoproterozoic granite-related hydrothermal overprint.

Highlights

- Microtectonics separates ore, ductile fabric and hydrothermal upgrading at Salobo.
- Salobo is a Neoarchean IOCG deposit deformed and transposed in the Paleoproterozoic.
- The main Sn fabric is Transamazonian (2.10–2.05 Ga), not Neoarchean.
- Chalcocite Pb model age of 2.7 Ga records inherited, not syngenetic, lead.
- High-grade Cu is a ca. 1.88 Ga stoichiometric upgrade of Neoarchean ore.

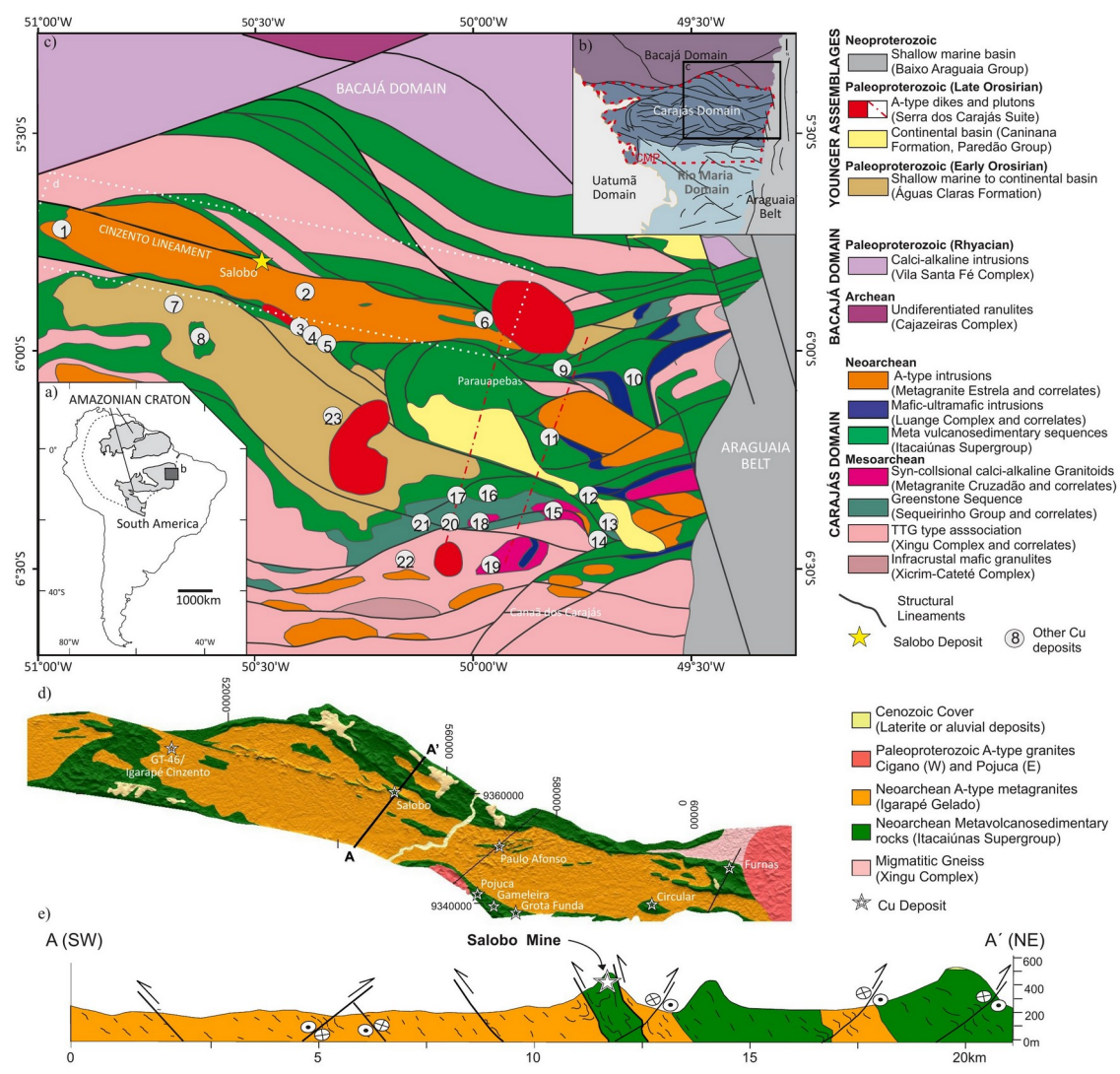
Keywords: Carajás, Salobo, IOCG, granite-related, metamorphosed deposit

Introduction

Major Precambrian IOCG provinces commonly form where deep structures focus high-temperature hydrothermal fluids through the crust, with examples ranging from syntectonic systems to post-orogenic extensional systems such as Olympic Dam (Reynolds, 2000; Groves et al., 2010). In polydeformed terranes, however, the structures that host ore, the fabrics that deform it and later fluid pathways may belong to different tectonic events, making the tectonic setting of ore formation difficult to resolve.

Salobo is a world-class Cu-Au deposit in the Carajás Mineral Province (Réquia and Fontboté, 2000; Burns et al., 2017), hosted by altered volcanosedimentary schists along the Cinzento Lineament, a polydeformed WNW-ESE corridor that also hosts several Cu-Au deposits (Costa and Siqueira, 1990; Tavares, 2015; Tavares et al., 2021; Fig. 1c, d). Its mineralization has long been considered structurally controlled, but the relative timing

51 between ore formation, metamorphism, ductile deformation and late hydrothermal
 52 overprint remains debated.



54 **Figure 1.** (a) Location of the Amazonian Craton in South America. (b) Location of the
 55 Carajás Mineral Province (CMP) in the southeastern Amazonian Craton, with the tectonic
 56 limits of the Carajás Domain; modified from Tavares et al. (2018). (c)
 57 Tectonostratigraphic map of northeastern Carajás; numbered circles are Cu and IOCG
 58 deposits listed in SM1 and the star marks Salobo. The dashed box on the Cinzento
 59 Lineament outlines the area of (d); modified from Tavares (2015). (d) Geological map of
 60 the Cinzento Lineament, showing Salobo and neighbouring Cu deposits and the trace of
 61 section A-A'; modified from Oliveira et al. (2016a, b, c). (e) Cross-section A-A' across
 62 the Cinzento Lineament through the Salobo mine (vertical exaggeration ~5x).

63 The deposit contains magnetite-, chalcopryite-, bornite- and chalcocite-bearing ore, with
 64 high-grade zones commonly associated with magnetitite bodies and late sulfide
 65 replacement. It still remains open whether present ore geometry reflects the original

66 Neoarchean hydrothermal system, later ductile transposition, or a younger brittle
67 hydrothermal overprint.

68 Most previous models link Salobo ore to syn-tectonic shear-zone reactivation (Machado
69 et al., 1991; Souza and Vieira, 2000; Réquia et al., 2003; Melo et al., 2017), whereas other
70 interpretations allow an earlier extensional event (Lindenmayer, 2003). Recent magnetite,
71 sulfur-isotope, apatite and geochronological studies further show that the deposit records
72 polyphase hydrothermal and metamorphic reworking (Campo-Rodríguez et al., 2022;
73 Diniz et al., 2023; Trunfull et al., 2020; Toledo et al., 2023), requiring a framework that
74 separates ore formation from subsequent fabric development and upgrading.

75 This study asks: (i) what structures occur at Salobo and how do they overprint one
76 another? (ii) what P-T conditions characterize the main mineral assemblages? and (iii)
77 was the main ore-forming event related to ductile deformation?

78 We combine fieldwork, drill-core logging, microtectonics, petrography and mineral
79 chemistry to establish the relative timing of ore emplacement with respect to deformation
80 and metamorphism. The resulting model reinterprets Salobo as a modified IOCG deposit
81 in which Neoarchean ore was deformed by younger ductile fabrics and locally upgraded
82 by a Paleoproterozoic granite-related hydrothermal event.

83 **Geological context**

84 **Carajás Mineral Province**

85 The Carajás Mineral Province (CMP), in the southeastern Amazonian Craton (Fig. 1a, b),
86 comprises Mesoarchean TTG gneisses, orthogranulites, greenstone fragments and
87 reworked granitoids overlain by Neoarchean volcano-sedimentary sequences of the
88 Itacaiúnas Supergroup (Grainger et al., 2008; Tavares et al., 2018). These include ca.
89 2.76-2.73 Ga bimodal volcanic rocks, BIFs, metacherts and exhalative units (DOCEGEO,

1988; Machado et al., 1991), later intruded by Neoproterozoic mafic-ultramafic bodies, A-type granitoids and smaller ca. 2.57 Ga quartz-feldspathic bodies traditionally grouped with the Old Salobo metagranite (Lindenmeyer, 1990; Machado et al., 1991; Melo et al., 2017). Paleoproterozoic sedimentation, deformation and ca. 1.88 Ga A-type magmatism overprinted this architecture (Tavares et al., 2018; Pollard et al., 2019). North of Salobo, the CMP is bounded by the Bacajá Domain; along this boundary, Rhyacian high-grade metamorphism and ca. 2.08-2.05 Ga fabric recrystallization provide an important regional reference for Paleoproterozoic reworking (Tavares et al., 2018; Tavares et al., 2026).

Archean-Paleoproterozoic tectonic evolution of the Carajás Mineral Province

The age of the CMP's main structural framework remains debated. One model links it to Archean oblique transpression/transtension and repeated reactivation between ca. 2.60 and 1.88 Ga (Pinheiro and Holdsworth, 2000; Pinheiro et al., 2013; Trunfull et al., 2020); another relates much of the penetrative fabric in northern Carajás to Paleoproterozoic reworking of older anisotropies (Tavares et al., 2018; Tavares et al., 2026). Tavares et al. (2018) correlate this framework with the Transamazonian Cycle at 2.10-2.05 Ga, during collision between the Carajás and Bacajá crustal blocks under NE-SW shortening and top-to-SW transport. Fabric-constrained $^{40}\text{Ar}/^{39}\text{Ar}$ data subsequently date recrystallization of the main regional fabric at ca. 2.08-2.05 Ga and the Sereno overprint at ca. 2.01-1.96 Ga (Tavares et al., 2026). Brittle to transtensional reactivation followed during ca. 1.88 Ga Uatumã magmatism.

Copper mineralization in the Carajás Mineral Province

Cu-Au mineralization in the CMP includes three components relevant to Salobo: Neoproterozoic volcanogenic/VMS systems formed during basin development, Neoproterozoic magnetite-rich IOCG systems, and Paleoproterozoic granite-related systems associated with 1.88 Ga A-type magmatism (Grainger et al., 2008; Xavier et al., 2012; Moreto et al.,

2015a; Pollard et al., 2019; Tavares et al., 2021; SM1). Deposits of both IOCG epochs share Fe metasomatism, low-sulfidation sulfides, breccia textures, LREE enrichment and structural control, but Neoarchean systems are generally magnetite-rich and deeper, whereas Orosirian systems are shallower, more silica-saturated and locally hematite-bearing. The tectonic regime of the older IOCG events remains contested, whereas the 1.88 Ga systems are more consistently attributed to brittle, high-fluid-pressure, extensional to transtensional settings.

Previous works on the Salobo Deposit

At Salobo, intense hydrothermal and metamorphic overprinting has led to contrasting interpretations of the protoliths, including metasedimentary, BIF-bearing volcanosedimentary, metavolcanic, orthogneissic and greenstone-belt affinities (Lindenmayer, 1990; Réquia and Fontboté, 2000; Lindenmayer, 2003; Melo et al., 2017). Previous work recognized amphibole-feldspar, garnet-biotite-quartz and grunerite-magnetite-almandine-fayalite-biotite parageneses, as well as widespread schistosity, granoblastic textures and locally mylonitic fabrics throughout the host and wall rocks (Lindenmayer, 1990; Machado et al., 1991; Réquia et al., 2003; Melo et al., 2017).

Published geochronology dates different components of a system that evolved for more than one billion years. U–Pb zircon data constrain Mesoarchean basement and ca. 2.76–2.73 Ga host rocks. Pb–Pb chalcocite gives 2705 ± 42 Ma, Re–Os molybdenite gives 2576 ± 8 and 2562 ± 8 Ma, and accessory minerals and altered zircon record ca. 2.59–2.43 Ga re-equilibration. Pb–Pb magnetite gives 2112 ± 12 Ma (Machado et al., 1991; Réquia et al., 2003; Tassinari et al., 2003; Melo et al., 2017; Toledo et al., 2023, 2026). This mixed record has supported syngenetic, syn-tectonic and multistage models.

Earlier models variously treat ore formation as syn-shear, post-metamorphic or multistage (Souza and Vieira, 2000; Réquia et al., 2003; Lindenmayer, 2003; Melo et al.,

2017). The 1880 Ma Young Salobo granite cuts the deposit and has usually been considered unrelated to ore formation, whereas regional work on the Cinzento belt interpreted Salobo as a pre-Dn ore system overprinted by Transamazonian metamorphism and later modified by Young Salobo-related fluids (Tavares et al., 2021). This study tests that model at deposit scale using systematic microtectonics and EPMA mineral chemistry.

Materials and methods

Structural analysis

We combined microtectonics, field observation and 1:10,000 geological mapping to reconstruct the overprinting sequence at Salobo. Direct observation and sampling were carried out in the open pit and surrounding area; geological cross-sections were built across the deposit; and 2000 m of drill core from nine Vale drill holes were logged and sampled. Structural measurements were reported using Clar notation, grouped by crosscutting relationships and tectonic style, and plotted with OpenStereo (Grohmann et al., 2010).

A set of 120 polished thin sections was prepared and described to identify mineral assemblages, microtextures, microstructural fabrics and the relative timing of mineral growth. Minerals were classified as pre-, syn-, late- or post-tectonic with respect to the recognized deformation events, following Passchier and Trouw (2005).

New geological and structural observations were integrated with Vale's lithological map in ArcGIS 10.2.2 to produce the geological-structural map presented here, and LeapfrogGEO software was used to build a 3-D model of the central part of the deposit (not shown), visualizing how the major structures affecting the host rocks at depth link to the regional structural framework of the Cinzento Lineament (Fig. 1d, e; Oliveira et al., 2016a, 2016b, 2016c).

Finally, we established a relative order of events and correlated it with previously published geochronological data (Machado et al., 1991; Tassinari et al., 2003; Réquia et al., 2003; Melo et al., 2017) to anchor the proposed deformation and hydrothermal events to an absolute age framework.

Electron probe microanalysis (EPMA)

Mineral chemistry analyses of the thin sections quantified major-element compositions and constrained the thermobarometric conditions of pyroxene and chlorite crystallization. Chlorite compositions were compared with those from other CMP Cu-Au deposits (Serra Verde, Estrela, Igarapé Bahia, Furnas, Sossego, Sequeirinho, GT-46 and Pojuca), and pyroxene compositions with those from the CMP mafic-ultramafic complex, the GT-34 and Lightning Creek deposits, and a charnockite from southern China; data sources for comparison datasets are listed in SM2 (Perring et al., 2000; Garcia, 2018). All analyses were carried out at the UFRJ Electron Microprobe Laboratory (LABSONDA).

Quantitative analyses used a JEOL JXA-8230 EPMA operated at 15 kV accelerating voltage, 20 nA beam current, and a 1 μ m beam diameter. Chlorite was analyzed in seven samples and clinopyroxene in one, with wavelength-dispersive spectroscopy (WDS) applied to both grain cores and rims against natural international standards; energy-dispersive spectroscopy (EDS) provided complementary qualitative and semi-quantitative data.

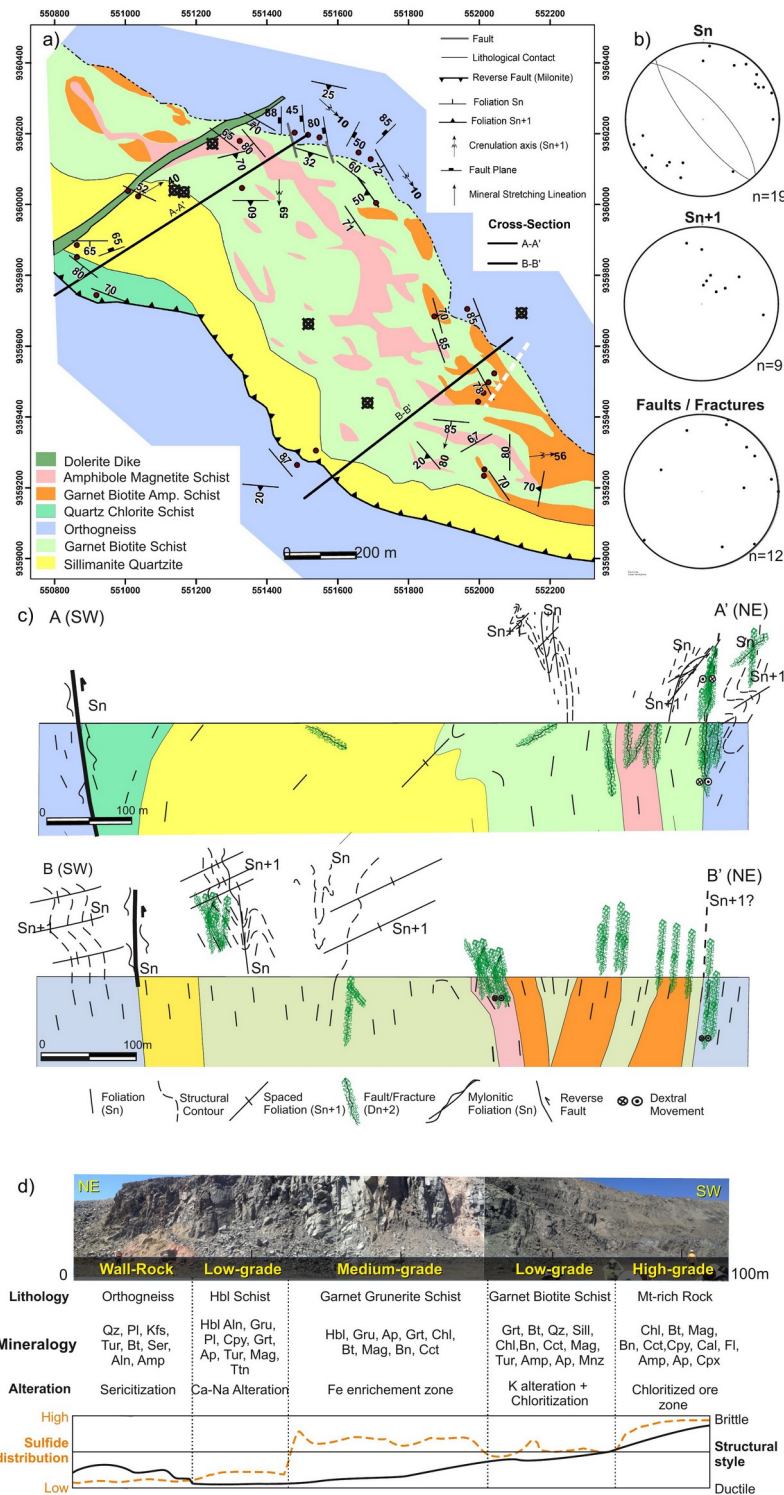
Chlorite data were recalculated with the WinCcac software (Yavuz et al., 2015) and pyroxene data with the WinPyrox software (Yavuz, 2013).

Results

Host and wall rocks of the Salobo deposit

The orthogneiss wall rocks form subvertical contacts with the ore-zone sequence along a narrow corridor: the southern wall dips 87° NE and the northern wall 85° SW (Fig. 2a, c).

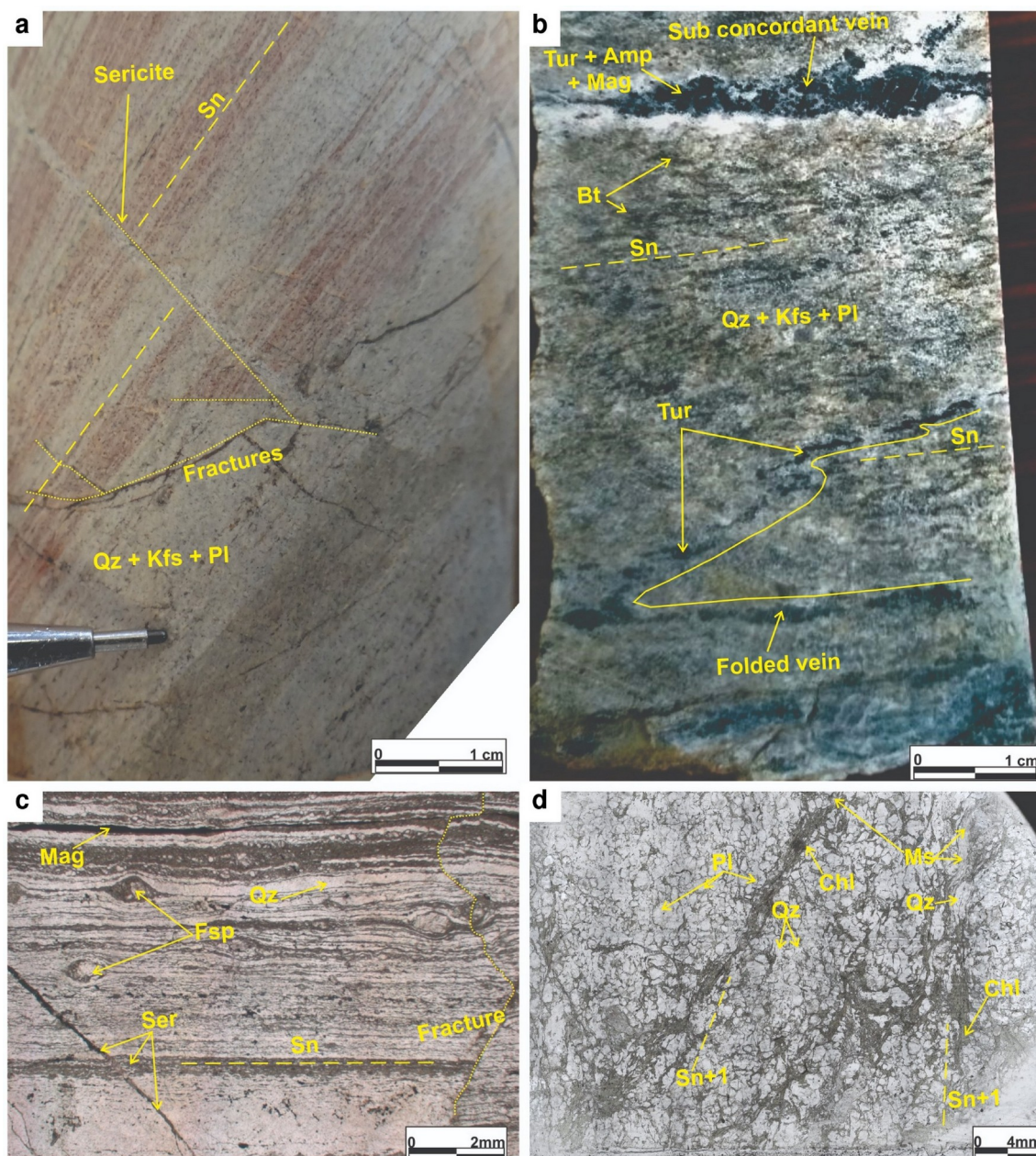
189 The southern wall is mylonitic, with alternating quartz-plagioclase and K-feldspar-rich
190 bands, feldspar porphyroclasts, down-dip lineation and fine recrystallized magnetite
191 parallel to the banding (Figs. 2a, 3a, c). Sericite partly replaces feldspar porphyroclasts
192 and bands along later brittle planes (Fig. 3c). A near-vertical reverse shear zone truncates
193 the contact with chlorite schists and quartzites. The northern orthogneiss is coarser
194 grained and less stretched, preserving a continuous S_n foliation that forms the axial plane
195 of Pre-Dn tourmaline-rich veins (Fig. 3b) and is locally cut by deformed biotite-rich veins
196 and by a spaced muscovite-chlorite S_{n+1} cleavage (Fig. 3d). Because its S_n -marking
197 minerals are far less stretched than in the south, the northern contact is not mylonitic and
198 preserves textures lost in the southern wall.



199

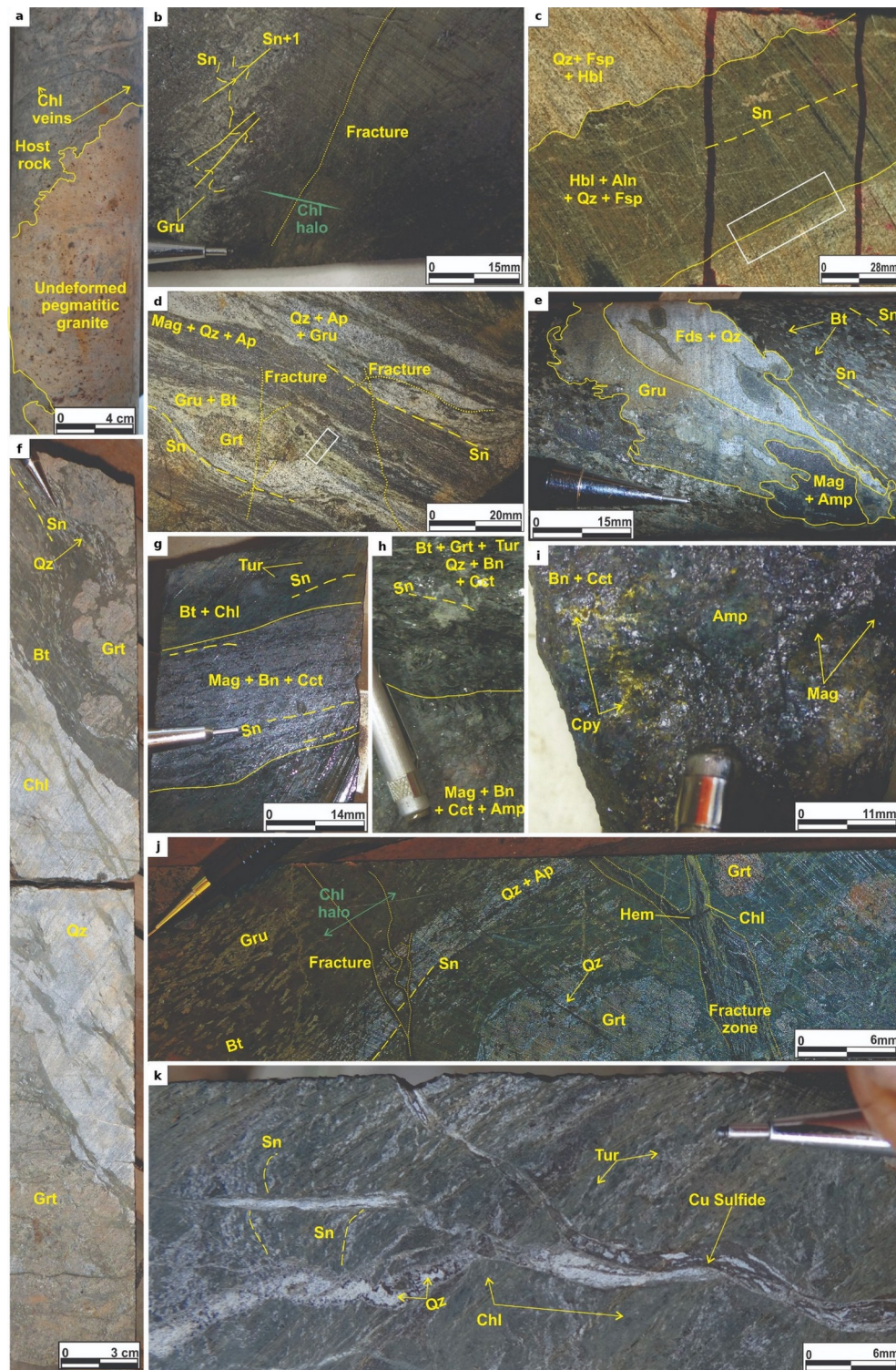
Figure 2. Structural framework of the Salobo deposit. (a) Geological-structural map of the deposit (level 250), produced from fieldwork integrated with Vale's lithological map and revised against published maps; the dashed white line locates the trace of section A-A' of Fig. 1d, e. (b) Equal-area, lower-hemisphere stereonets summarizing the Dn (Sn), Dn+1 (Sn+1) and Dn+2 (faults/fractures) structures. (c) Level-corrected, field-surveyed cross-sections A-A' and B-B' (horizontal and vertical scales equal). (d) Structure-grade panel: NE-SW sketch across the deposit, from the northern to the southern wall rock, showing lithology, mineralogy and alteration together with the distribution of ductile (Dn, Dn+1) versus brittle (Dn+2)

209 structural domains and sulfide distribution; note the coincidence of the high-grade
 210 zone with the brittle, chloritized domain.



211
 212 **Figure 3.** Core, hand-specimen and thin-section images of the orthogneiss wall rocks
 213 (Igarapé Gelado suite). (a) Drill core of the southern-wall banded orthogneiss: mylonitic
 214 Sn defined by alternating quartz-plagioclase and K-feldspar bands, crosscut by Dn+2
 215 fractures. (b) Northern-wall orthogneiss in hand sample: Pre-Dn tourmaline veins folded
 216 with Sn as axial plane; at the top, a tourmaline-amphibole-magnetite vein lies concordant
 217 with Sn. (c) Photomicrograph (plane-polarized light, PPL) of the southern-wall
 218 orthogneiss: feldspar porphyroclasts in a banded, finely recrystallized quartz-feldspar
 219 matrix with a magnetite veinlet parallel to Sn; sericite-filled microfaults crosscut the
 220 mylonitic fabric and sericite is disseminated over feldspar-rich bands. (d)
 221 Photomicrograph (PPL) of the northern-wall orthogneiss: plagioclase and quartz crosscut
 222 by the spaced Sn+1 cleavage marked by muscovite + chlorite. Note the contrast in
 223 stretching (a, b) and grain size (c, d) between the southern and northern walls.

224 The ore zone comprises five mappable units: sillimanite quartzite, chlorite schist, garnet-
225 biotite schist, garnet-biotite-amphibole schist and amphibole-magnetite schist (Fig. 2a),
226 plus deformed and undeformed granitic veins too small to map at deposit scale (Fig. 4a,
227 c). Contacts are gradual to sharp and most primary layering is strongly transposed parallel
228 to Sn.



229

230 **Figure 4.** Representative hand-sample and drill-core photographs of the Salobo ore zone.
 231 (a) Undeformed pegmatitic granite with chloritization halo. (b) Grunerite schist: Sn
 232 crenulated by Sn+1 and cut by chlorite-filled fractures. (c) Deformed amphibolite vein
 233 within Old Salobo-type metagranite; note the serrated, locally tectonized contacts. (d-f)
 234 Banded mineralized schists and chlorite veins; Sn bends slightly at vein-host contacts. (g,
 235 h) Amphibole-magnetite schist enveloped by biotite schist; magnetite-rich bodies locally
 236 crosscut the compositional banding. (i) Relict chalcopyrite islands rimmed by bornite-
 237 chalcocite in the high-grade zone. (j, k) Late chlorite-sulfide-quartz-calcite fractures
 238 cutting Sn, with jigsaw-puzzle fragments showing little rotation.

239 **Quartzite** – the impure quartzite thickens northwestward and contains quartz with minor
240 biotite, muscovite, sillimanite, garnet and secondary chlorite. Its contact with the schists
241 shows no shear fabric and is therefore interpreted as stratigraphic, though transposed
242 parallel to Sn; the contact with the orthogneiss is mylonitic, marked by complete quartz
243 recrystallization and grain-size reduction that lend the rock a saccharoidal aspect in hand
244 sample. It carries little Cu-Au mineralization.

245 **Garnet-biotite schist** – the most abundant host unit is a garnet- and biotite-rich schist
246 with granolepidoblastic texture, low magnetite content, granoblastic garnet
247 porphyroblasts up to 10 cm across with inclusion-rich cores, and local tourmaline-garnet-
248 biotite or quartz-garnet-apatite-magnetite layers. It commonly envelopes amphibole-
249 magnetite-rich schists and is overprinted by late bornite-chalcocite assemblages (Figs. 4h,
250 5c).

251 **Garnet-biotite-amphibole schist** – this unit contains two texturally distinct amphiboles –
252 dark-green, euhedral hornblende and light-green, acicular grunerite – together with
253 quartz, biotite, garnet and minor apatite, and locally preserves millimeter-scale
254 compositional banding between grunerite-biotite and magnetite-quartz layers.
255 Hornblende-rich varieties occur mainly near the northern ore zone and are spatially linked
256 to Old Salobo-type metagranitoid veins; grunerite commonly replaces hornblende,
257 forming partial coronas around it, especially in garnet-rich domains (Figs. 5d, 6g, h).

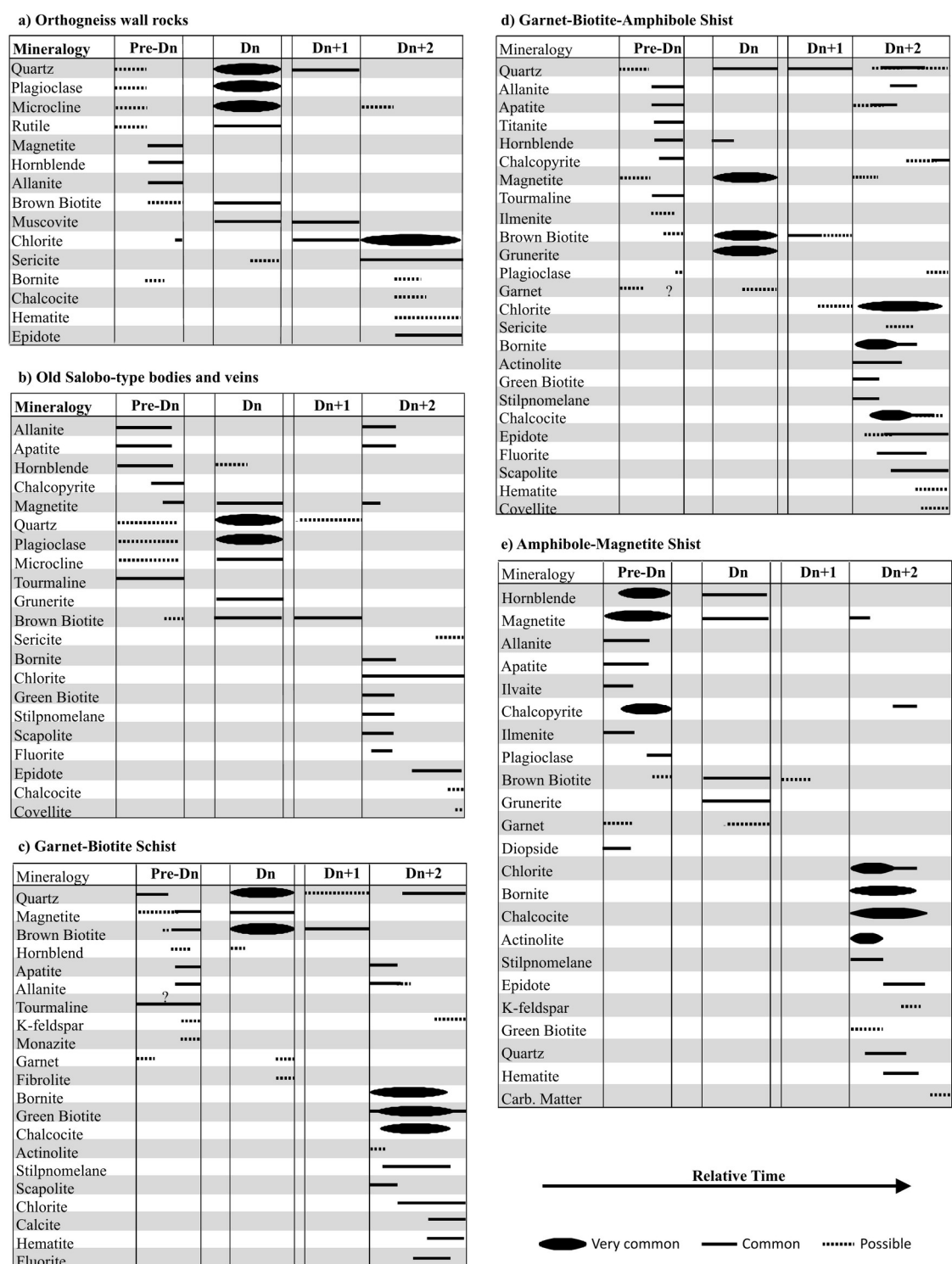


Figure 5. Relative crystallization order of minerals with respect to the structural families (Pre-Dn, Dn, Dn+1, Dn+2) for the main Salobo lithotypes: (a) orthogneiss wall rocks; (b) Old Salobo-type bodies and veins; (c) garnet-biotite schist; (d) garnet-biotite-amphibole schist; (e) amphibole-magnetite schist. Thick ellipses indicate very common minerals; solid lines, common; dotted lines, possible; question marks, uncertain timing. The diagrams show relative timing only and do not imply duration or absolute age.

265 **Chlorite schist** – is restricted to the southern deposit margin, between quartzite and
266 southern wall orthogneiss. It lacks the compositional banding typical of the other schists
267 and is dominated by oriented chlorite, quartz and copper sulfides.

268 **Magnetite schist (magnetitite) and sulfide breccia** – the main mineralized unit is a
269 magnetite-rich (25-90 vol.%), amphibole-bearing schist with minor clinopyroxene,
270 secondary chlorite, greenish biotite and variable bornite-chalcocite, commonly enveloped
271 by biotite-rich schists. Where magnetite exceeds ~50 vol.%, the rock is pseudo-massive
272 and Sn is defined only by submillimetric, amalgamated magnetite layers. Magnetite-rich
273 bodies are lentiform and elongated along Sn, but locally crosscut the compositional
274 banding of their host schists (Fig. 4g, h) – a discordance that identifies them as relict
275 hydrothermal veins rather than stratigraphic variants. They grade into sulfide-rich
276 breccias in which rounded to euhedral, millimetric to centimetric crystals of magnetite,
277 hornblende and clinopyroxene are set in a chalcopyrite-rich matrix (Figs. 4g-i, 6c).
278 Replacement of this matrix by bornite-chalcocite intensifies from the deposit margins
279 toward the high-grade core: at the periphery, chalcopyrite forms both the breccia matrix
280 and inclusions within magnetite, hornblende and clinopyroxene, whereas in the high-
281 grade zone chalcopyrite survives only as inclusions and as relict islands rimmed by
282 bornite-chalcocite (Fig. 4i), set in a bornite-chalcocite matrix in disequilibrium with the
283 Pre-Dn assemblage.

284 **Metagranitoids, granites and undeformed veins of the ore zone** – the Old Salobo-type
285 metagranitoid veins are grey, foliated, plagioclase-quartz-hornblende rocks with minor
286 magnetite – K-feldspar is notably absent – and serrated, intrusive, locally tectonized
287 contacts with the schists (Fig. 4c). Pink, post-tectonic granitic veins composed of K-
288 feldspar, quartz, biotite and chlorite cut all older rocks and are spatially associated with

289 chlorite halos, late fractures and calcite-epidote-fluorite-biotite-K-feldspar-allanite-
290 quartz-hematite-minor sulfide assemblages (Fig. 4a).

291 **Structural geology and relative crystallization order of minerals**

292 Four structural families affect the ore zone sequence and wall rocks: Pre-Dn, Dn, Dn+1
293 and Dn+2, each tied to a distinct mineral association or paragenesis. The Pre-Dn trend is
294 largely obliterated by later overprinting; Dn structures trend WNW-ESE to NW-SE with
295 subvertical dips to the NE; Dn+1 structures trend WSW-ENE to WNW-ESE with
296 moderate dips to the SSW or SSE; and Dn+2 structures are more variable in direction,
297 though a NW-SE trend with steep to subvertical dips to the SW is still recognizable.

298 **Pre-Dn structures and mineral association**

299 The oldest structural family, Pre-Dn, is recorded chiefly by deformed hydrothermal veins
300 and Old Salobo-type granitoid veins, most of them sub-parallel to the main foliation (Sn;
301 Figs. 4c, e, g, 6e).

302 These veins and zones are crosscut at a low angle by the Sn foliation and are locally
303 folded, with Sn parallel to their axial planes (Fig. 3b); even where concordant with Sn, the
304 foliation bends slightly at their contact with the host rock (Fig. 4d-h).

305 The same crosscutting criteria – truncation by Sn at a low angle, folding with Sn as axial
306 plane, and deflection of the foliation at vein contacts – apply to the hornblende-rich rocks,
307 Old Salobo-type metagranitoids, magnetite schists and sulfide breccias, which are
308 therefore treated as Pre-Dn vein-like bodies or hydrothermal alteration zones (Figs. 3b,
309 4g, h, 6b, e, f, h). Although most of these rocks preserve their Pre-Dn mineral association,
310 the minerals are usually re-equilibrated by later Dn, Dn+1 and/or Dn+2 events:
311 hornblende is partly replaced by coronas of grunerite (syn-Dn) or actinolite (syn-Dn+2)
312 (Fig. 6g, h), whereas clinopyroxene appears largely unaffected, showing only thin
313 amphibole coronas.

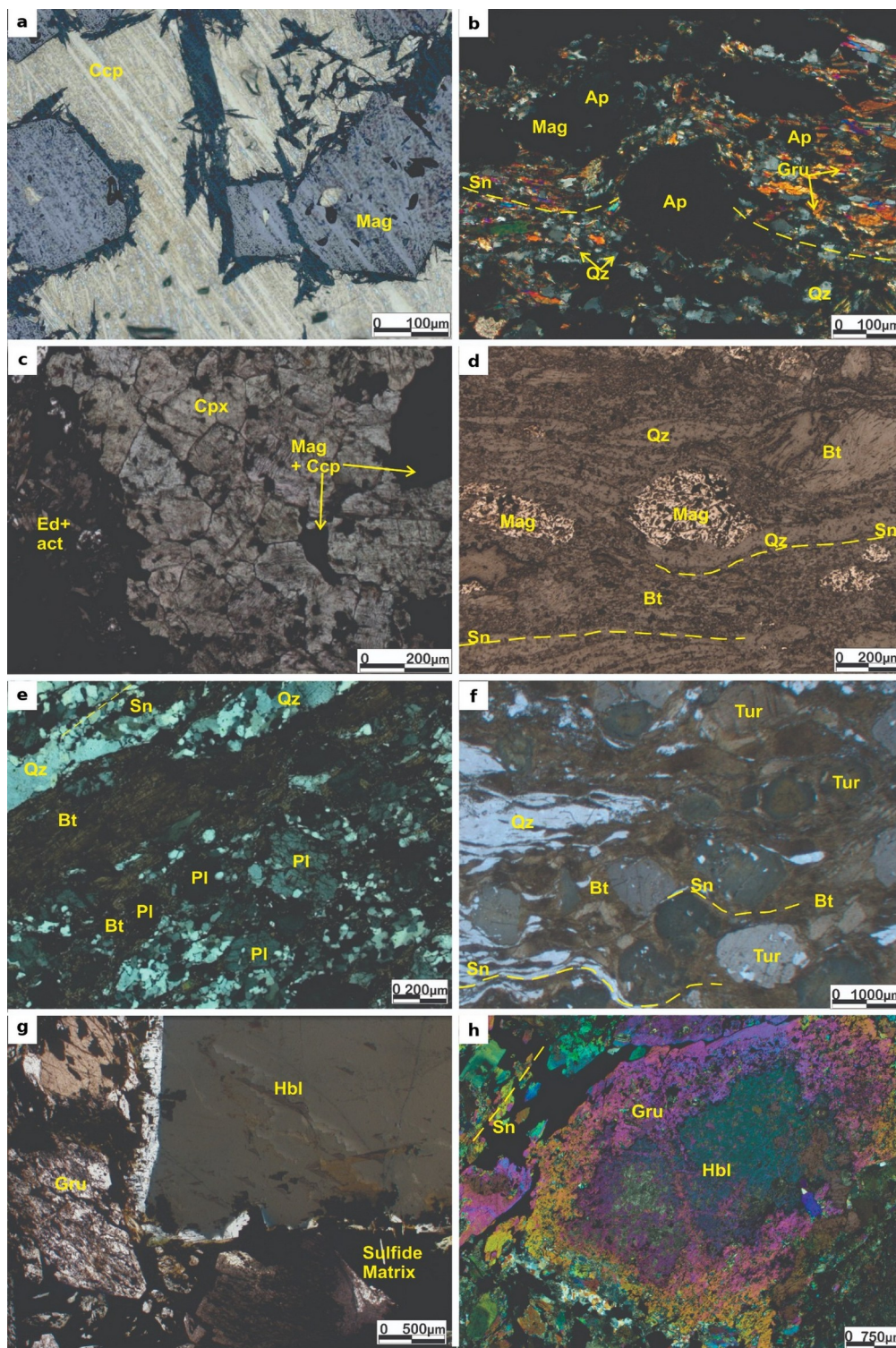


Figure 6. Photomicrographs of Pre-Dn and Dn relationships. (a) Reflected light: Pre-Dn magnetite in a chalcopyrite matrix cut by Dn+2 minerals. (b, d, f) Pre-Dn apatite, magnetite, biotite and tourmaline deflecting Sn around whole grains. (c) Clinopyroxene-hornblende paragenesis in a magnetite-chalcopyrite matrix; note chalcopyrite inclusions within the silicates. (e) Deformed biotite vein in orthogneiss, crosscut at a low angle by Sn. (g, h) Hornblende-bearing amphibole-magnetite schist with grunerite coronas replacing hornblende; host-rock pods between relict veins are reoriented parallel to Sn.

322 Where Sn is poorly developed, relict Pre-Dn veins survive as a slightly deformed,
323 stockwork-like network with a jigsaw-puzzle texture that forms Pre-Dn breccia fabrics in
324 thin section (Figs. 6e, g, 7b). In more strongly deformed areas, host-rock pods between
325 these veins are reoriented parallel to Sn (Figs. 6h, 7b), and the internal texture of the Pre-
326 Dn veins is largely obliterated by Dn. Individual veins are centimetric to metric in
327 thickness and centimetric to decametric in length; several preserve relict, deformed
328 syntaxial growth zoning. Restricted tourmalinite veins occur in association with the
329 hornblende-rich zones.

330 The Pre-Dn association includes magnetite, clinopyroxene, hornblende, allanite,
331 tourmaline, chalcopyrite, brown biotite, apatite, feldspars, ilmenite, titanite, monazite and
332 minor quartz (Figs. 5, 6a-d, f). In the sulfide breccias, xenomorphic garnet with magnetite
333 and hornblende is texturally distinct from the granoblastic garnet of the schists.
334 Magnetite, clinopyroxene and hornblende contain chalcopyrite inclusions even where the
335 surrounding matrix is replaced by bornite-chalcocite (Fig. 6a, c), implying that part of the
336 chalcopyrite is coeval with their crystallization. Some silicate and magnetite crystals of
337 the breccia framework are rounded, a shape consistent with chemical corrosion during
338 alteration or with abrasion during subsequent deformation.

339 This Pre-Dn association also occurs disseminated through the garnet-biotite-amphibole
340 schists and wall orthogneisses, where it deflects Sn around affected grains. Brown biotite
341 is largely confined to biotite-rich zones around magnetite schists and is reoriented or
342 recrystallized during later deformation.

343 Taken together, the relict jigsaw-puzzle textures, the stockwork-like geometry and the
344 low degree of fragment rotation in preserved pods are consistent with vein emplacement
345 during one or more pulses of brittle, fluid-saturated, largely extensional deformation with
346 limited open space (Jébrak, 1997), later partly obliterated by Dn.

347 **Dn structures and mineral association**

348 The Dn family records the deposit's main structural framework: compressive, ductile
349 structures affecting nearly every lithotype except the Young Salobo granites and dolerite
350 dyke. Its most prominent expression is the penetrative Sn schistosity (Fig. 7a, c, d), sub-
351 parallel to the ore sequence's compositional banding and defined by oriented syn-Dn
352 minerals (Figs. 7d, 8a) or recrystallized Pre-Dn minerals (Figs. 3c, 7b, e). Sn strike
353 controls the deposit's overall trend, from ENE-WSW to NNW-SSE with subvertical to
354 steep dips to the NW/SE (Fig. 2a, b), its wavy pattern reflecting late- to post-Dn refolding,
355 probably during Dn+1. A down-dip mineral lineation of biotite, amphibole and/or
356 magnetite is occasionally visible on Sn planes, and a few Dn folds affecting Pre-Dn veins
357 and banding were directly observed in the open pit (Fig. 7f, g).

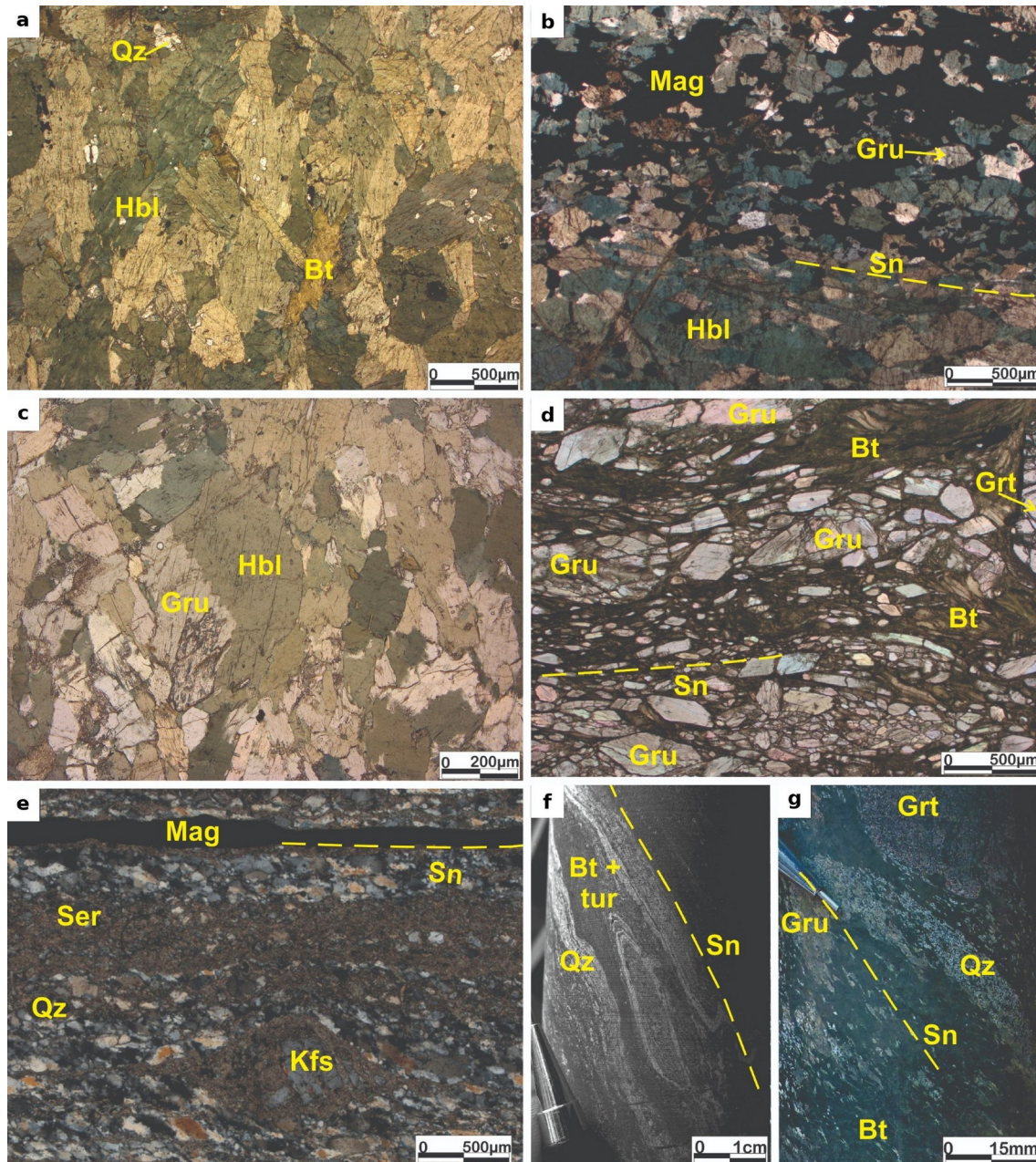


Figure 7. Dn microstructures and core-scale folds. (a-d) Polygonal hornblende, grunerite and biotite with straight grain boundaries defining or overprinting Sn (granonematoblastic to granolepidoblastic fabrics); in (b), a relict Pre-Dn jigsaw-puzzle texture is partially transposed by Sn. (e) Southern-wall orthogneiss with mylonitic Sn, stretched magnetite and later sericite. (f, g) Tight to rootless folds with Sn as axial plane.

Sn becomes mylonitic only near the southern contact between the ore zone and the orthogneiss, where a finely recrystallized quartz-feldspar matrix hosts K-feldspar and plagioclase porphyroclasts with wavy extinction, mantle-core structures and subgrain rotation - microstructures consistent with medium-temperature (500-650°C) mylonitic fabrics (Trouw et al., 2010). Down-dip stretching lineation and S/C fabrics in this

southern shear zone indicate reverse, top-to-SW shear, consistent with a NE-SW shortening vector.

Minerals crystallized along Sn define the Dn mineralogy. Grunerite (early- to syn-Dn), sillimanite and muscovite (syn-Dn) are exclusive to this event, as is most garnet (syn- to late-Dn), except the xenomorphic ore-zone garnet. Hornblende, magnetite (Figs. 6b, d, 7b, e), quartz, plagioclase, K-feldspar and brownish biotite are considered Pre-Dn minerals recrystallized during Dn (Fig. 5a-e), though part of the biotite/hornblende may have formed anew.

Sn-oriented grunerite, hornblende and biotite form straight grain boundaries and polygonal fabrics, consistent with high-temperature recrystallization after deformation (Fig. 7a, c, d). Pre-Dn hornblende occurs in equilibrium with biotite or is partly replaced by grunerite. Garnet has granoblastic textures and records two growth patterns: cores crowded with inclusions of quartz, biotite, tourmaline and accessory allanite and monazite, surrounded by inclusion-poor, limpid rims. No spiral or snowball inclusion patterns were observed; internal trails are straight to gently curved, locally oblique to the external foliation, and Sn deflects around whole grains, core and rim alike (Fig. 8a). These relations constrain the limpid rims as syn- to late-Dn overgrowths, whereas the inclusion-rich cores may be older, inherited porphyroblast domains (see Discussion).

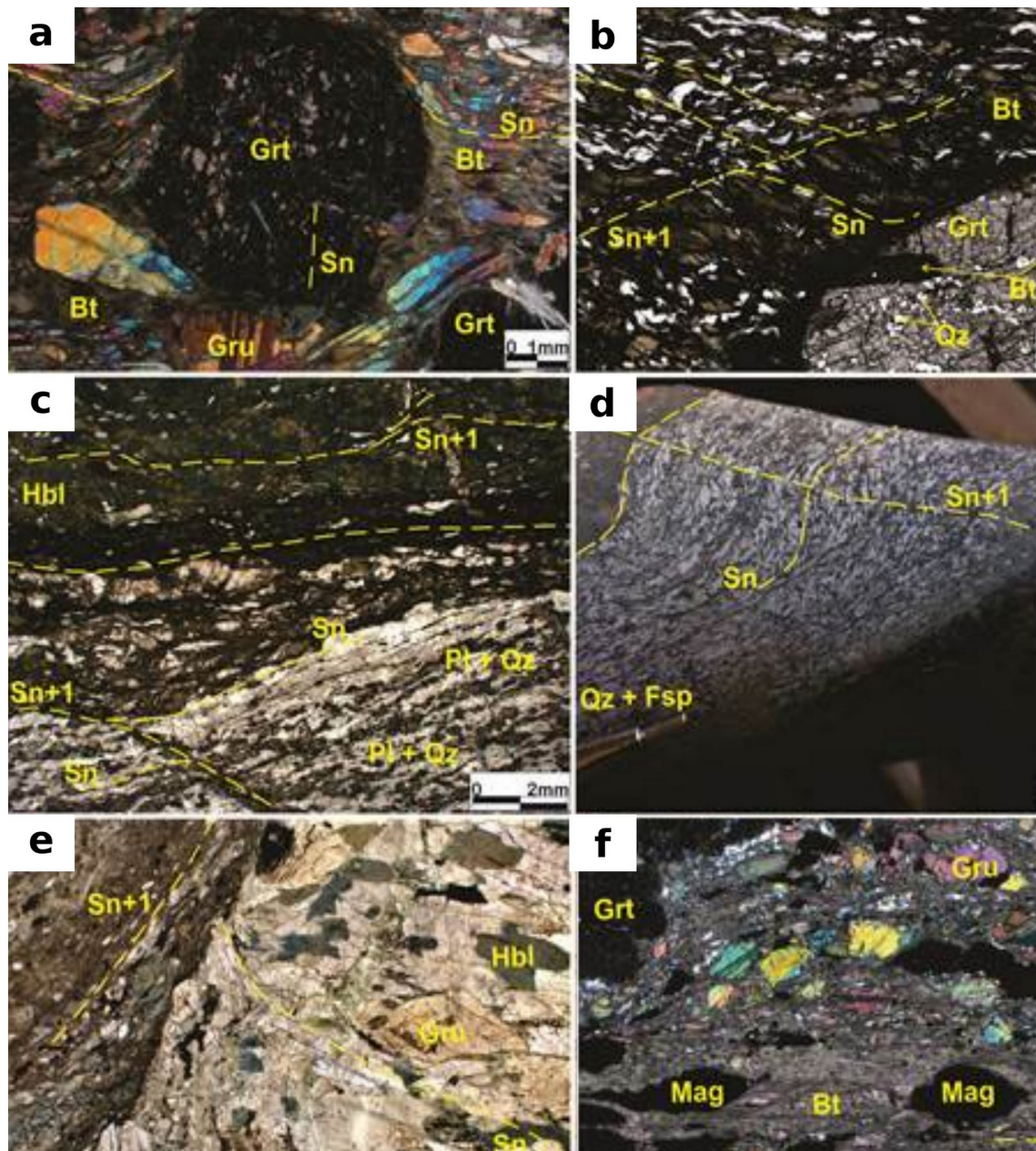


Figure 8. Dn and Dn+1 microstructures. (a) Granoblastic garnet with inclusion-rich core and limpid rim; Sn deflects around the whole grain. (b) Garnet-biotite schist in which Sn is dragged into an oblique Sn+1 shear band (Dn+1 drag fold); note garnet with quartz preserved in the low-strain domain. (c, e) Sn truncated by low-temperature Sn+1 mylonitic planes and shear bands with sharp contacts against the host fabric. (d) Asymmetric Dn+1 folds with sub-horizontal axes. (f) Fine-grained Sn+1 ultramylonite (over 75% comminuted matrix) with brittle magnetite and grunerite porphyroclasts and chlorite-muscovite-quartz strain shadows.

Dn+1 structures and mineral association

Dn+1 structures are ductile-brittle, oblique to the Dn trend, and less penetrative, formed under a compressive to transpressive system; they cluster along discrete, spaced

discontinuities where S_{n+1} foliation, shear bands and folds occur together. Like D_n , this family affects every rock type in the deposit except the Young Salobo granites and the dolerite dyke.

S_{n+1} is a spaced, largely dissolution-controlled cleavage that locally forms crenulation axial planes or develops near D_{n+1} shear bands (Figs. 3d, 4b, 8c). It hosts fine chlorite, muscovite and biotite – syn- D_{n+1} biotite is markedly thinner than its syn- D_n counterpart – and the associated shear bands form thin, submillimetric ultramylonite layers with sharp contacts against the host fabric and top-to-NNW transport, commonly nucleated on earlier lithological or structural discontinuities (Fig. 8c, e). The ultramylonites contain over 75% finely comminuted matrix overgrown by biotite and chlorite, with brittle magnetite and feldspar porphyroclasts, hornblende and grunerite porphyroclasts, biotite fish, and chlorite-muscovite-quartz strain shadows (Fig. 8f) – microstructures indicating relatively low-temperature (250-500 °C) mylonitization, consistent with greenschist-facies conditions (Trouw et al., 2010).

D_{n+1} folds are either drag folds linked to the shear bands (Fig. 8b) or asymmetric, open to tight folds (Fig. 8d) with sub-horizontal axes and a top-to-N/NNW vergence, more common in the sillimanite quartzite unit and possibly related to its greater thickness shown in cross-section B-B' (Fig. 2c). Together, fold vergence and shear-band transport point to a N-S to NNW-SSE shortening vector for D_{n+1} .

D_{n+2} structures and mineral association

D_{n+2} comprises brittle, subvertical fractures and faults associated with a substantial volume of undeformed, hydrothermally altered veins. These structures preferentially reactivate earlier discontinuities, especially D_n and D_{n+1} fabrics, though they can also cut discordantly across them (Fig. 9a, b), with a weak but recognizable NW-SE main orientation, parallel to S_n , along planes that are subvertical or dip steeply to the SW.

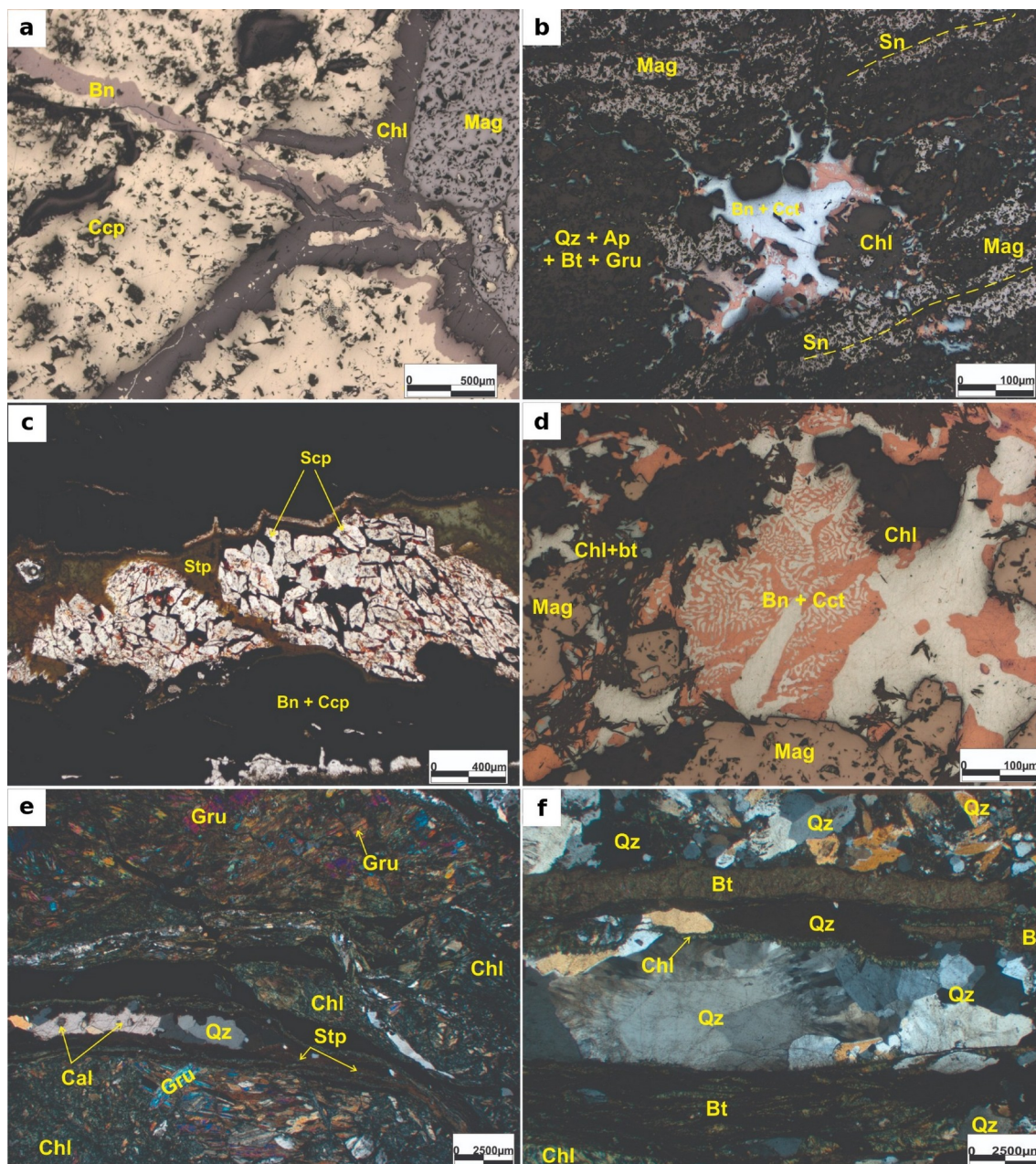
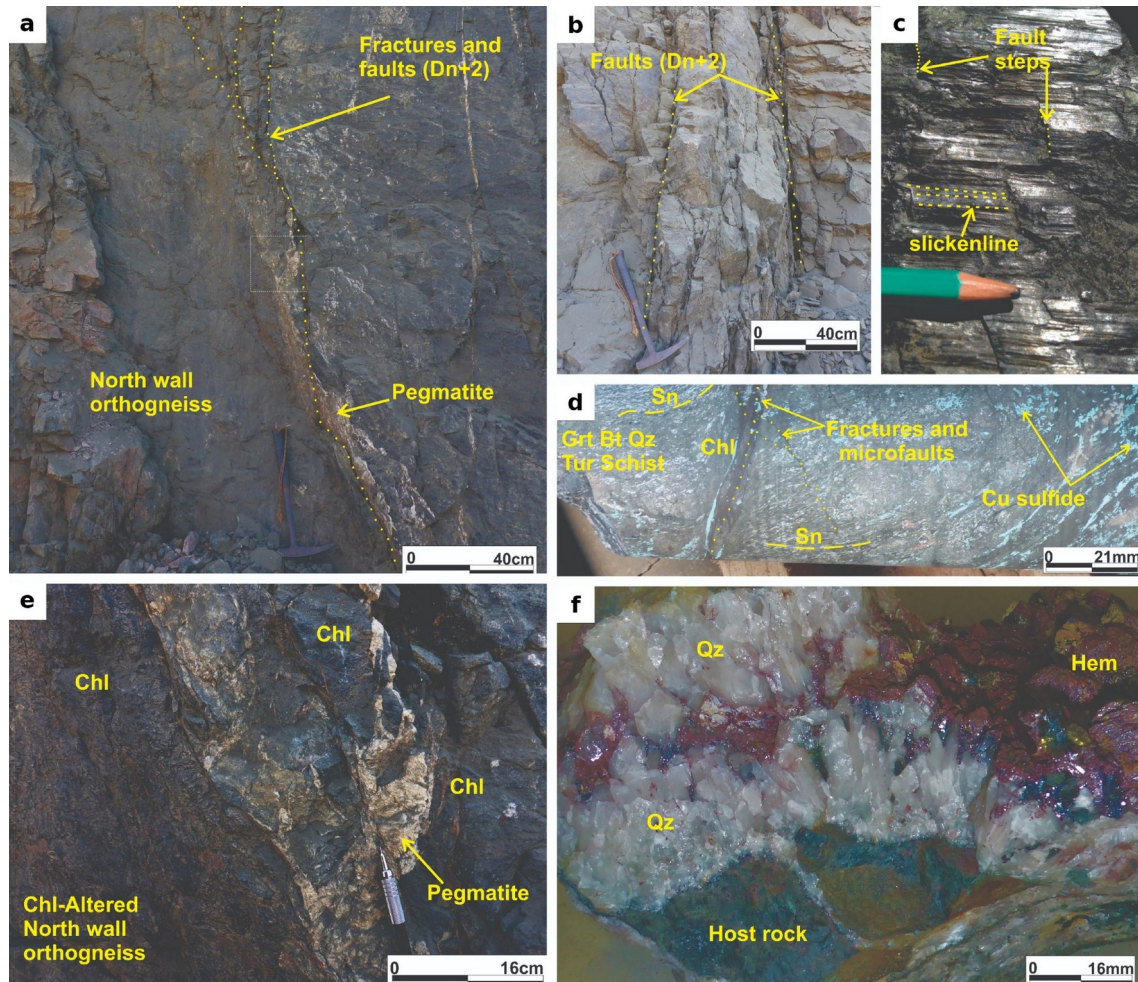


Figure 9. Dn+2 veins and sulfide replacement. (a, b) Reflected light: Pre-Dn magnetite-chalcocopyrite and Sn foliation cut by bornite-chalcocite-bearing fractures; bornite veinlets crosscut chalcocopyrite, indicating that chalcocopyrite was no longer stable during Dn+2. (c) Banded scapolite-stilpnomelane-bornite vein. (d) Symplectitic bornite-chalcocite intergrowth around magnetite. (e, f) Undeformed chlorite-quartz-calcite-stilpnomelane veins cutting grunerite-rich rocks; euhedral quartz lacks wavy extinction except near the rims of larger crystals (f).

Faults trending in multiple directions carry mostly sub-horizontal, subordinately subvertical, slickenlines on chloritized planes, with dextral or normal steps and displacements of at most a few meters (Fig. 10b, c). Fractures show jigsaw-puzzle textures with little clast rotation (Figs. 4j, k, 9e, 10d), commonly radiating from a central

436 fracture, and some Dn+2 structures are spatially associated with undeformed Young
 437 Salobo-type pegmatitic veins (Fig. 10a, e). Both fractures and faults are more common
 438 within the high-grade ore, where intense Dn+2 chloritization can obliterate the pre-
 439 existing ductile fabric (Fig. 2d).



440
 441 **Figure 10.** Dn+2 brittle structures. (a) Pegmatitic granite spatially associated with Dn+2
 442 faults. (b) Set of subvertical Dn+2 faults in the pit wall. (c) Chloritized fault plane with
 443 sub-horizontal slickenlines and dextral steps. (d) High-grade ore cut by crackle-breccia
 444 fractures with jigsaw-puzzle texture and minimal clast rotation. (e) Chlorite-altered north-
 445 wall orthogneiss cut by an undeformed Young Salobo-type pegmatitic vein spatially
 446 associated with Dn+2 fractures. (f) Syntaxial vein texture with euhedral quartz and
 447 hematite against the pervasively chlorite-replaced host rock.

448 Veins are mainly syntaxial, locally zoned, with crystals grown from the walls inward and
 449 locally with radial habit; laminated veins and subordinate ataxial veins also occur (Figs.
 450 9c, e, f, 10f). They are commonly enveloped by chlorite-green biotite halos up to 15 m

wide, in which chlorite and green biotite replace earlier garnet and biotite, locally up to complete obliteration of the ductile fabric. Syn-Dn+2 veins and alteration zones contain undeformed crystals of chlorite, green biotite, quartz, allanite, bornite, chalcocite, stilpnomelane, calcite, K-feldspar, epidote, fluorite, hematite, scapolite, actinolite, sericite, chalcopyrite, covellite, digenite, uraninite and minor magnetite.

The dominant bornite-chalcocite assemblage is linked to Dn+2 replacement and veining, locally as symplectitic intergrowths (Fig. 9d), either replacing Pre-Dn chalcopyrite (Fig. 9a) or forming new mineralized veins and nodules at fracture intersections (Fig. 9b); minor earlier bornite cannot be ruled out, and a minor, late-Dn+2 chalcopyrite generation grows with calcite and quartz in the cores of zoned syntaxial veins, later than bornite, and in late veinlets crosscutting the bornite-chalcocite assemblage.

Quartz crystals are mostly euhedral and lack wavy extinction, which appears only near the rims of larger crystals (Fig. 9e, f); no other syn-Dn+2 mineral shows evidence of ductile strain.

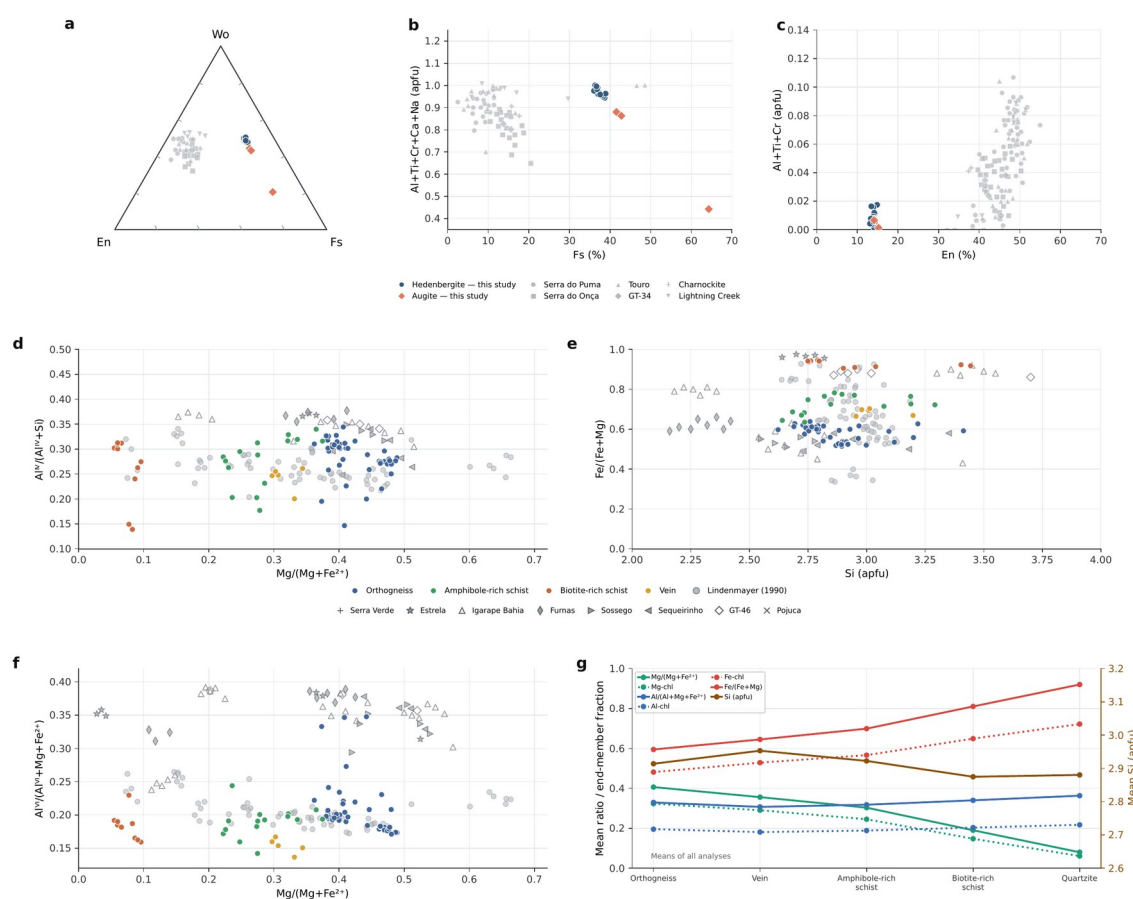
Mineral chemistry

Mineral chemistry was used as an independent test of the structural-paragenetic framework, targeting one mineral from the Pre-Dn high-temperature assemblage and one from the Dn+2 low-temperature assemblage. Clinopyroxene and chlorite were chosen because, at Salobo, each is restricted to one of these assemblages and, in IOCG systems, they respectively mark high-temperature K-Fe-Ca alteration and low-temperature, shallow alteration (Williams et al., 2010).

Clinopyroxene mineral chemistry

We analyzed 21 spots by wavelength-dispersive spectroscopy across cores and rims of Pre-Dn clinopyroxene from a single thin section (SA-QUATRO; Supplementary Material 2, SM2), chosen because this mineral appears largely unaffected by later Dn, Dn+1 and

476 Dn+2 deformation, metamorphism or alteration. Structural formulae were normalized to 4
 477 cations and 6 oxygen atoms, with ferric iron estimated following Papike et al. (1974).
 478 The resulting structural formulae give $Al = 0.005\text{--}0.029$, $Cr \leq 0.001$, $Ti \leq 0.002$, $Na =$
 479 $0.011\text{--}0.041$, $Ca = 0.405\text{--}0.971$, $FeT = 0.738\text{--}1.272$, $Mg = 0.254\text{--}0.301$, and $Mn = 0.007\text{--}$
 480 0.014 . On Rock's (1990) Wollastonite (Wo)-Enstatite (En)-Ferrosilite (Fs) diagram (Fig.
 481 11a), most analyses plot as hedenbergite (average $Wo_{49}En_{14}Fs_{37}$), with three crystal
 482 cores plotting as augite (average $Wo_{35}En_{15}Fs_{50}$); overall, the Salobo clinopyroxene has
 483 $En < 20\%$ and $Al+Ti+Cr < 0.02$ (Fig. 11b, c). Rims are systematically richer in total Fe
 484 than cores.



485
 486 **Figure 11.** Mineral chemistry. (a) Clinopyroxene classification (Wo-En-Fs; Rock, 1990):
 487 Salobo analyses (black crosses) against comparison datasets (sources in SM2); most
 488 analyses plot as hedenbergite and three crystal cores as augite. (b) $Al+Ti+Cr+Ca+Na$
 489 (apfu) versus Fs . (c) $Al+Ti+Cr$ (apfu) versus En ; note $En < 20\%$ and $Al+Ti+Cr < 0.02$. (d)
 490 Chlorite classification (Moazzen, 2004) by host-rock group, with comparison deposits.

491 (e) $\text{Fe}/(\text{Fe}+\text{Mg})$ versus Si (apfu). (f) $\text{AlIV}/(\text{AlIV}+\text{Mg}+\text{Fe}^{2+})$ versus $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$
492 parent-rock discrimination; Salobo vein chlorite plots near fields typical of felsic parent
493 rocks. (g) Mean chlorite ratios by rock-type group, showing progressive Fe enrichment
494 from orthogneiss through veins and schists to quartzite.

495 Compared with other IOCG occurrences, Salobo clinopyroxene has unusually high
496 ferrosilite and low enstatite contents, with low $\text{Ti}+\text{Cr}+\text{Al}$ values. This is more consistent
497 with crystallization from an aqueous hydrothermal fluid than with igneous or
498 metamorphic clinopyroxene. The empirical intracrystalline-equilibrium thermometer of
499 Dal Negro et al. (1982) yields an indicative temperature of $\sim 595^\circ\text{C}$ for hedenbergite. A
500 single augite core gives 4.5 kbar with the Nimis (1999) geobarometer; because this
501 calibration was developed for magmatic clinopyroxene, the estimate is treated as
502 indicative only and is not used to constrain emplacement depth.

503 **Chlorite mineral chemistry**

504 We performed 75 WDS analyses of major elements in Dn+2 chlorite (samples EB-11B,
505 EB-13B, EB-7A, EB-3A, EB-2A, EB-4 and 38817), compared with the data of
506 Lindenmayer (1990; see SM2). Cationic compositions were recalculated on 14 oxygen
507 atoms with estimated Fe^{+3} (Yavuz et al., 2015) and grouped by host rock (SM2): (i)
508 orthogneiss, (ii) amphibole-rich schist, (iii) biotite-rich schist, (iv) quartzite (data from
509 Lindenmayer, 1990), and (v) syn-Dn+2 veins. In groups (i)-(iv), chlorite occurs as very
510 fine, euhedral crystals disseminated through the host rock.

511 Chlorite data were compared with those from other CMP Cu-Au deposits using the
512 $\text{AlIV}/(\text{AlIV}+\text{Si})$ - $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ classification diagram of Moazzen (2004) (Fig. 11d).
513 Salobo chlorite mostly plots as chamosite to magnesian chamosite, with minor
514 clinocllore compositions depending on host rock.

515 Recalculated Si contents range from 2.600 to 3.450. $\text{Fe}/(\text{Mg}+\text{Fe})$ is higher in the ore zone
516 than in orthogneiss or veins, and the quartzite (Lindenmayer, 1990 data) carries the most

Fe-rich chlorite (Fig. 11e). Group averages show progressive Fe enrichment from orthogneiss through veins and schists to quartzite (SM2; Fig. 11g).

Compared with chlorite from other CMP Cu-Au deposits, Salobo vein chlorite is distinct because its composition is weakly controlled by host lithology. On the $Mg/(Mg+Fe)$ versus $Al/(Al+Mg+Fe)$ diagram it plots close to fields typical of felsic parent rocks (Fig. 11f).

Chlorite formation temperatures were estimated with two geothermometers: Kranidiotis and MacLean (1987), suited to chlorite in aluminous hydrothermal sulfide deposits, and Cathelineau (1988), more broadly applicable across diagenetic, hydrothermal and metamorphic settings. The former runs systematically higher, likely reflecting differing sensitivity to Salobo's Fe-rich compositions; we favor Cathelineau (1988) as more broadly applicable. Individual estimates are widely dispersed (ca. 120-380 °C, Cathelineau, 1988 calibration) owing to host-rock buffering and local intergrowths. Because disseminated chlorite is partly buffered by host-rock composition, the temperature of the Dn+2 event is anchored on the syn-Dn+2 vein chlorite (group v), which yields Cathelineau (1988) means of ca. 250 °C (this study) and ca. 285 °C (data of Lindenmayer, 1990), i.e. ca. 250-290 °C; disseminated chlorite from the host-rock groups gives consistent means of ca. 270-310 °C (group statistics in SM2). Analyses with elevated K₂O (>0.5 wt.%) likely reflect biotite intergrowths; they are flagged in SM2 and do not affect the vein-based estimate.

Discussion

Our microtectonic study identifies distinct P–T–t conditions for each structural family and requires time between successive events. We provide no new absolute ages, but calibrate the relative sequence against published geochronology. The four structural families

record at least four tectonothermal episodes (Table 1). Pre-Dn is itself composite and may preserve several Neoarchean to Siderian hydrothermal pulses.

Table 1. Structural synthesis and geological interpretation of the four structural families recognized at Salobo (Pre-Dn, Dn, Dn+1, Dn+2).

Structural Family	Geological Evidence	Mineralogy	Main Lithotypes observed	Tectonic Regime	Geological Interpretation	Inferred age
Pre-Dn	Relict jigsaw-puzzle textures and open-space-filling, recorded only in foliated hydrothermal alteration zones discordant to the compositional banding of the schists and gneiss.	Amphibole, clinopyroxene, (fayallite), chalcopyrite, magnetite, allanite, apatite, ilmenite, plagioclase, tourmaline, monazite	Wall-rock orthogneiss; mineralized breccias; amphibole-magnetite schists	Extension under high T, low P	Hydraulic breccias and hydrothermal zones spatially linked to Old Salobo-type quartz-feldspathic bodies; main Fe, K, Na-Ca and ore alteration.	~2.57-2.45 Ga
Dn	Early stage: Sn foliation, with a down-dip Ln mineral stretching lineation; Late stage: progressive deformation with closed folds, a parasitic 'Z' limb pattern and SW vergence, associated with reverse faults and medium-temperature mylonites.	Varies with the early mineral composition of the host lithotype: biotite, grunerite, bluish amphiboles, plagioclase, garnet, quartz	Grt-bt schist, orthogneiss, Grt-bt-Amp-schist, Amp-Mt-schist, Old Salobo veins	Ductile, compressive	Formation of Salobo's main structural framework; deformation and amphibolite-facies metamorphism	~2.10-2.05 Ga (favored correlation)
Dn+1	Spaced Sn+1 cleavage oblique to the main foliation, with asymmetric crenulation folds of subhorizontal axis (NNW top-vergence), and WNW-ESE / ENE-WSW low-temperature mylonitic zones.	Muscovite, chlorite, biotite, albite	Garnet-biotite schist, orthogneiss, garnet-amphibole-biotite schist	Brittle-ductile, compressive to transpressive	Structural overprint on the Salobo framework under greenschist-facies conditions, probably related to a late tectonic reactivation pulse.	~2.01-1.96 Ga (favored correlation)
Dn+2	Undeformed stockwork veins with open-space-filling textures, closely associated with quartz, chlorite and hematite, crosscutting all previously formed structures; high-angle stepping planes with subhorizontal lineation, mostly dextral displacement.	Chlorite, chalcopyrite, bornite, chalcocite, allanite, calcite, epidote, biotite, stilpnomelane, feldspar, fluorite, quartz, actinolite, hematite, minor magnetite, carbonaceous matter, scapolite	All lithotypes (excluding dolerites)	Extension with transtensional component	Low-T hydrothermal alteration along fractures and faults; associated with undeformed pegmatoid granitoid dykes and veins; correlated with the ca. 1.88 Ga system.	~1.88 Ga (correlation)

Host-rock protoliths

The host-rock mineralogy was overprinted by Pre-Dn hydrothermal alteration, Dn-Dn+1 metamorphism and Dn+2 hydrothermal alteration. The compositional banding is best interpreted as primary stratification transposed by deformation.

Hedenbergite, Fe-rich chlorite and thin quartz-magnetite layers support iron formation and mafic rocks as important protoliths. We interpret the recognizable protoliths as

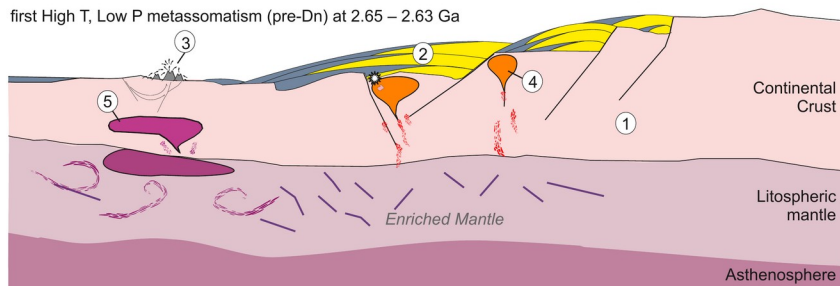
volcanic, clastic and exhalative sedimentary rocks of the Itacaiúnas Supergroup; chlorite schists were probably mafic rocks, quartzites arenites to Fe-rich chemical sediments, and magnetite-quartz layers relicts of oxide- and silicate-facies BIF. Hornblende schists, magnetite schists and sulfide breccias are instead treated as Pre-Dn hydrothermal alteration zones whose protoliths were largely obliterated.

This interpretation agrees with previous recognition of a volcanosedimentary substrate (DOCEGEO, 1988; Lindenmayer, 1990) but differs from models that treat hornblende- and magnetite-rich rocks as simple stratigraphic BIF variants (Lindenmayer, 2003; Melo et al., 2017). Although volcanogenic or syn-sedimentary features have been proposed for north Carajás IOCG-class deposits more broadly (Martini and Dreher, 2024), the discordant, vein- and breccia-hosted character of the Pre-Dn mineralized zones at Salobo favors the epigenetic hydrothermal origin adopted here.

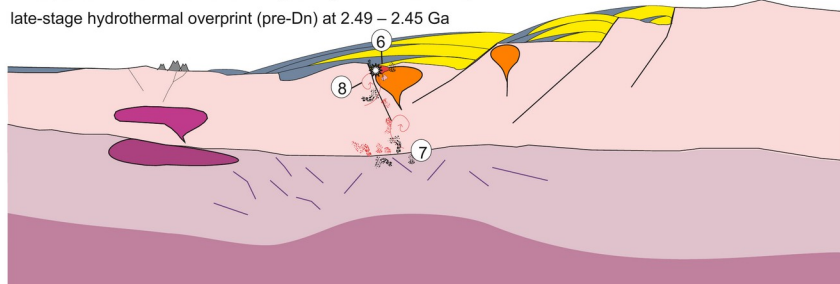
The first mineralizing event

The Pre-Dn event, possibly grouping several closely spaced pulses, involved high-temperature metasomatism and the earliest recognizable IOCG-type ore, characterized by chalcopyrite + magnetite (Fig. 12, Stages I and II). Old Salobo-type quartz-feldspathic bodies are spatially and texturally associated with this system, and the ore-forming stage is broadly correlated with the 2.57-2.56 Ga Re-Os molybdenite age of Réquia et al. (2003). Fe-rich clinopyroxene supports a high-temperature, hydrothermal and moderately reducing assemblage consistent with hedenbergite stability (Gustafson, 1974). The augite-core-to-hedenbergite-rim zoning of the clinopyroxene further records progressive Fe enrichment of the fluid during crystal growth, an independent chemical expression of this Fe metasomatism.

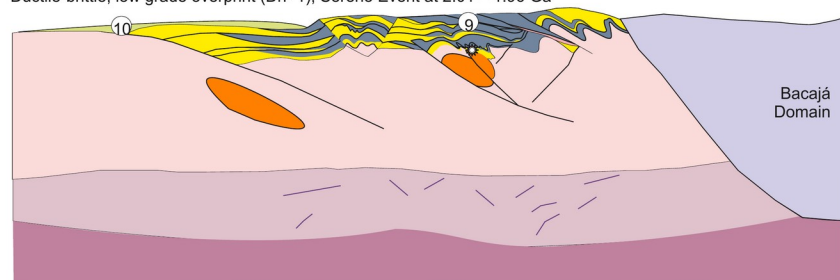
Formation of Salobo host rocks at 2.76 – 2.73 Ga; first Cu input at ca. 2.70 Ga (min. age);
first High T, Low P metassomatism (pre-Dn) at 2.65 – 2.63 Ga



Principal IOCG mineralization event (pre-Dn) at 2.57 – 2.55 Ga;
late-stage hydrothermal overprint (pre-Dn) at 2.49 – 2.45 Ga



Ductile deformation and metamorphism (Dn), Transamazonian Cycle (Carajás – Bacajá collision at 2.10 – 2.05 Ga);
Ductile-brittle, low grade overprint (Dn+1), Sereno Event at 2.01 – 1.96 Ga



Brittle, extensional, shallow fluid-dominated deformation (Dn+2); granite-related hydrothermalism and ore upgrading during the ca. 1.88 Young Salobo magmatism

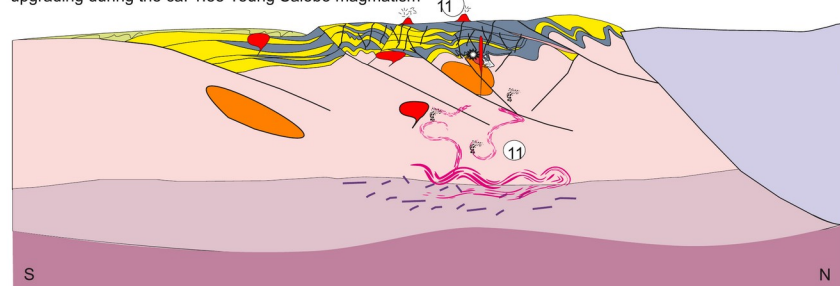


Figure 12. Schematic Neoproterozoic evolution of the Salobo deposit. Stage I (Neoproterozoic, 2.76-2.65 Ga): formation of the host-rock sequence, first Cu input and the earliest high-T, low-P Pre-Dn metasomatism. Stage II (Neoproterozoic to Siderian, 2.57-2.45 Ga): principal Pre-Dn IOCG event, with extensional, fluid-saturated veining and high-temperature alteration with chalcopyrite-magnetite ore at 2.57-2.55 Ga, followed by a late Pre-Dn hydrothermal overprint at 2.49-2.45 Ga. Stage III (Rhyacian to early Orosirian, 2.10-1.96 Ga): Dn compressive ductile deformation and amphibolite-facies metamorphism transposing the mineralized zones (correlated with the Transamazonian Cycle, 2.10-2.05 Ga), followed by the Dn+1 ductile-brittle, greenschist-

585 facies overprint (correlated with the Sereno Event, 2.01-1.96 Ga). Stage IV (late
586 Orosirian, ca. 1.88 Ga): Dn+2 brittle, fluid-dominated fracturing, chloritization and
587 bornite-chalcocite mineralization associated with Young Salobo granite emplacement.
588 Numbered circles key the individual features discussed in the text.

589 The clinopyroxene data are not used as a standalone genetic discriminator, but as a
590 chemical constraint on the Pre-Dn paragenesis already defined by crosscutting
591 relationships and ore textures. Whether the ca. 2.57 Ga age dates pristine crystallization of
592 Old Salobo-type granite remains uncertain, because hydrothermally altered quartz-
593 feldspathic bodies elsewhere along the Cinzento Shear Zone may record partial resetting
594 or mixed zircon behavior (Toledo et al., 2023). Therefore, Old Salobo-type bodies are
595 treated here as components of the Pre-Dn hydrothermal-structural architecture, not as a
596 magmatic source.

597 The magnetite-chalcopyrite-clinopyroxene-amphibole-garnet paragenesis records high-
598 temperature alteration. Fe-rich clinopyroxene with amphibole supports pyroxene-hornfels
599 conditions above 600 °C, followed by amphibolite-facies recrystallization during Dn.
600 Comparison with GT-34 and Lightning Creek reinforces a hot metasomatic origin
601 (Garcia, 2018; Perring et al., 2000). Titanite and garnet petrochronology independently
602 resolve the composite Pre-Dn history. A U–Pb titanite age of 2653 ± 13 Ma and a Lu–Hf
603 garnet age of 2635 ± 18 Ma record a high-temperature precursor before the principal ore
604 event. A Sm–Nd garnet age of 2573 ± 58 Ma records re-equilibration by hot fluids under
605 low-pressure conditions (Toledo et al., 2026). This setting is consistent with the
606 unfoliated Pre-Dn assemblages and hydraulic breccias. Accessory-mineral inclusions
607 contributed Nd and Hf to the garnet concentrates, so the bulk isochrons may date the
608 thermal-hydrothermal system as a whole. The xenomorphic garnet in sulfide breccias
609 remains undated.

610 K, K–Fe and Na–Ca metasomatism match the main alteration families of IOCG systems
611 (Williams et al., 2010). Their present assemblages are composite and metamorphosed.
612 Biotite, K-feldspar, magnetite, tourmaline, plagioclase, titanite and apatite include Pre-Dn
613 components, while sillimanite and at least part of the garnet rims formed or recrystallized
614 during Dn. Metamorphism erased much of the primary halo geometry, so the current
615 mineral distribution cannot be read as a simple alteration profile.

616 Relict open-space fillings and jigsaw-puzzle textures place Pre-Dn alteration in an
617 extensional, fluid-saturated setting (Jébrak, 1997). Boron isotopes in tourmaline along the
618 Cinzento belt require hybrid fluids that leached older Itacaiúnas exhalative and
619 volcanosedimentary rocks (Melo et al., 2021). This inherited reservoir provides an
620 independent basis for the early contribution recorded by the 2705 ± 42 Ma chalcocite
621 model age. Comparable alteration extends at least 700 m below the current Salobo pit
622 (Carvalho et al., 2025). Dn erased many early overprinting relations, so individual Pre-Dn
623 pulses cannot be separated at deposit scale.

624 **Tectonic controls of earlier mineralization**

625 Several authors have interpreted Archean ore formation at Salobo as synchronous with
626 shear-zone reactivation along the Cinzento Lineament at ca. 2.57 Ga (Machado et al.,
627 1991; Souza and Vieira, 2000; Réquia et al., 2003; Melo et al., 2017; Toledo et al., 2026).
628 In contrast, our microtectonic data indicate that Dn parallelized and deformed pre-existing
629 hydrothermal zones that more likely formed under extension.

630 Classic syn-tectonic IOCG deposits such as Ernest Henry and Monakoff contain
631 mylonitic fabrics around ore, ore precipitation in compressional dilational sites and clear
632 synchronicity with regional metamorphism (Davidson et al., 2002; Holwell et al., 2014;
633 Williams et al., 2010). These relations are absent from the first Salobo event. The ore zone

is mostly schistose, and Sn stretches and parallelizes pre-existing magnetite-chalcopyrite bodies. The sulfides do not occupy Dn dilational jogs.

Pre-Dn and Dn assemblages resemble alteration in some Australian syn-tectonic deposits (Davidson et al., 2002; Holwell et al., 2014), but the similarity reflects comparable P–T conditions during Salobo's main metamorphic deformation. It does not establish syn-Dn ore formation. Quartz-biotite temperatures of 565 ± 50 °C may record Dn re-equilibration of K- and Fe-alteration assemblages, instead of peak Pre-Dn pyroxene-hornfels metasomatism (Melo et al., 2019). Dn and Dn+1 still control the present orebody geometry by folding, transposing and reorienting older mineralized zones.

The second mineralization event

Two mineralizing stages have long been recognized at Salobo, but they are commonly treated as parts of a single evolving system (Souza and Vieira, 2000; Réquia et al., 2003; Melo et al., 2017). We instead constrain the second stage to the Dn+2 fabric and alteration, separated from the first by two intervening tectono-metamorphic events. The syn-Dn+2 alteration is spatially tied to Young Salobo granitoid dykes; the alteration shares the brittle emplacement style of the dykes, and vein chlorite chemistry supports a felsic fluid component; together, these relations are consistent with correlation to the ca. 1.88 Ga granite-related system (Fig. 12, Stage IV).

Chlorite geothermometry gives ca. 250-290 °C for syn-Dn+2 alteration, far cooler than the Pre-Dn event. Our textural framework does not require assigning the ca. 2.45 Ga monazite age of Melo et al. (2017) to Dn+2, because monazite is not paragenetic with the Dn+2 assemblage; it may instead record a Siderian hydrothermal or recrystallization episode that predates Dn.

Dn+2 alteration is pervasive and strongest in the high-grade ore zone, where chloritized fractures and faults are abundant (Fig. 2d). Replacement of Pre-Dn chalcopyrite by syn-

Dn+2 bornite and chalcocite increased the Cu tenor. Chalcopyrite carries about 34.6 wt.% Cu, compared with 63.3 wt.% in bornite and 79.9 wt.% in chalcocite. Replacement can therefore raise contained Cu by a factor of about 1.8–2.3 per unit mass and 2.2–3.1 per unit volume without external Cu input. This bornite-chalcocite domain corresponds to the high-grade ore mined at Salobo.

This distinction also avoids conflating two different Paleoproterozoic processes. Granoblastic, foliated magnetite is best interpreted as a record of ductile metamorphic reworking of older magnetite, whereas the low-temperature bornite-chalcocite-digenite-fluorite-calcite assemblage records brittle fluid flow and sulfide remobilization (Campo-Rodríguez et al., 2022). The former is compatible with the ca. 2.1 Ga magnetite age of Tassinari et al. (2003); the latter is better correlated with the ca. 1.88 Ga granite-related event.

The Dn+2 assemblage of chlorite, quartz, biotite, fluorite, hematite and carbonate resembles 1.88 Ga Cu-Au systems elsewhere in Carajás (Pollard et al., 2019). Campo-Rodríguez et al. (2022) support the same separation between two Paleoproterozoic processes. Mgt III records metamorphic reworking of older magnetite and may correspond to the 2112 ± 12 Ma Pb–Pb age. Their low-temperature bornite-chalcocite-digenite-fluorite-calcite assemblage matches Dn+2. Whether Dn+2 introduced new copper remains unresolved, but the local U, LREE, Au and high-grade Cu enrichment is consistent with remobilization and upgrading.

The mixed isotopic record: inherited lead, chronometer retentivity and Paleoproterozoic disturbance

The published isotopic record of Salobo spans nearly 1.6 Gyr and includes almost every method applied to Precambrian ore systems (Table 2; for the complete geochronological record, see the Supplementary Material 3). It falls into two families. Robust chronometers

(Re-Os in molybdenite, Lu-Hf in garnet, U-Pb in titanite, monazite and least-altered zircon) date mineral growth and resist later reworking. Disturbance-prone systems (stepwise-leached Pb-Pb model ages, whole-rock Rb-Sr and K-Ar) record the last recrystallization or fluid interaction experienced by each phase. Treated in isolation, the second family yields the "mixed ages" that have long clouded the geochronology of the deposit (cf. Trunfull et al. 2020). Tied to the structural families defined here, the same record resolves into a coherent sequence: host-rock formation, an early hydrothermal contribution, the principal ore event, Siderian re-equilibration, Paleoproterozoic deformation and granite-related upgrading (Table 2).

Table 2. Event-based synthesis of published geochronological constraints at Salobo and their relationship with the structural and hydrothermal sequence proposed in this study.

Event and structural position	Representative constraints (Ma)	Dated material, method and source	Main published interpretation	Integrated interpretation in this study
Host rocks and inheritance <i>Pre-Dn host record</i>	ca. 2950–2731	Zircon U–Pb, orthogneiss, amphibolite and Igarapé Gelado gneiss (Machado et al., 1991; Melo et al., 2017; Toledo et al., 2023)	Crystallization and metamorphism of basement and host-rock protoliths.	Formation and reworking of the host rocks. These ages do not date Cu mineralization.
Early metal contribution <i>Early Pre-Dn</i>	2705 ± 42; 2701 ± 30	Chalcocite Pb–Pb and zircon U–Pb (Tassinari et al., 2003; Melo et al., 2017)	Primary syngenetic Cu mineralization or early fluid-assisted resetting.	Older Cu-bearing or exhalative contribution, consistent with tourmaline B isotopes (Melo et al., 2021). It does not date the orebody.
High-temperature metasomatism <i>Early Pre-Dn</i>	2653 ± 13; 2635 ± 18	Titanite U–Pb and garnet Lu–Hf (Toledo et al., 2026)	Dynamic metamorphism and high-temperature Fe-silicate growth.	High-temperature, low-pressure, fluid-driven re-equilibration before the principal ore event.
Principal IOCG event <i>Principal Pre-Dn ore event</i>	2576 ± 8; 2562 ± 8; 2594 to 2535 (broader array)	Molybdenite Re–Os, sulfide Pb–Pb, accessory-mineral U–Pb and garnet Sm–Nd (Machado et al., 1991; Réquia et al., 2003; Tassinari et al., 2003; Melo et al., 2017; Toledo et al., 2023, 2026)	Ore formation, metamorphism, granite crystallization or hydrothermal resetting.	The main mineralizing event occurred at 2.57–2.55 Ga. Re–Os directly dates ore; the wider array records mineral growth and fluid-assisted re-equilibration.
Siderian re-equilibration <i>Pre-Dn, post-ore overprint</i>	2497 ± 5; 2452 ± 14; 2427 ± 130	Titanite and monazite U–Pb; chalcopyrite Pb–Pb (Machado et al., 1991; Tassinari et al., 2003; Melo et al., 2017)	Late metamorphism, hydrothermal activity or renewed sulfide mineralization.	Post-ore recrystallization, Pb loss and open-system behavior. No separate principal Cu pulse is required.
Dn reworking <i>Syn-Dn</i>	2112 ± 12; 2101 ± 130	Magnetite Pb–Pb and whole-rock Rb–Sr (Tassinari et al., 2003)	Paleoproterozoic hydrothermal magnetite growth and alteration.	Magnetite re-equilibration is compatible with Dn at ca. 2.10–2.05 Ga. The Rb–Sr result is an imprecise open-system constraint.

Event and structural position	Representative constraints (Ma)	Dated material, method and source	Main published interpretation	Integrated interpretation in this study
Dn+1 and post-Dn cooling <i>Late Dn to post-Dn+1</i>	2022 ± 32 to 1903 ± 35	Muscovite, biotite and amphibole K–Ar (Cordani et al., 1984)	Transamazonian cooling.	Mineral cooling or recrystallization after Dn. These values do not record renewed ore formation.
Dn+2 and younger disturbance <i>Dn+2 to post-Dn+2</i>	1880 ± 80; 1800	Young Salobo Granite Rb–Sr; whole-rock and mineral Rb–Sr/K–Ar (Cordani, 1981; Cordani et al., 1984)	Granite emplacement, late resetting.	The ca. 1.88 Ga granite constrains chlorite alteration and Cu upgrading.

Note: only representative ages are shown; the complete dataset and individual original interpretations are provided in Supplementary Table Sx.

The principal mineralizing component is the most securely dated. Re-Os molybdenite ages of 2576 ± 8 and 2562 ± 8 Ma date ore-related sulfide crystallization directly and provide the strongest anchors for the main IOCG event (Réquia et al. 2003). Around this anchor, Pb-Pb ages of Pre-Dn tourmaline (2587 ± 150 Ma; Tassinari et al. 2003) and bornite-chalcopyrite (2579 ± 71 Ma; Réquia et al. 2003), U-Pb ages of titanite (2581 ± 5 Ma) and monazite (2551 ± 2 Ma; Machado et al. 1991), garnet Sm-Nd re-equilibration at 2573 ± 58 Ma (Toledo et al. 2026) and the 2594-2535 Ma cluster of variably reset zircon from Old Salobo-type bodies and high-grade ore (Machado et al. 1991; Melo et al. 2017; Toledo et al. 2023) define a single ca. 2.59-2.54 Ga high-temperature interval. The internal spread reflects contrasting closure behaviour, inherited components and fluid-assisted re-equilibration among chronometers.

Chalcocite provides the clearest example of inheritance. It is texturally the youngest ore mineral at Salobo, undeformed, vein- and fracture-hosted, replacing chalcopyrite within the Dn+2 assemblage correlated with the ca. 1.88 Ga Young Salobo system. Yet it yields the oldest Pb-Pb age of the deposit, 2705 ± 42 Ma, originally interpreted as dating syngenetic primary mineralization (Tassinari et al. 2003). Both observations are reconciled if the result is read as a model age of the lead source: the ca. 1.88 Ga chalcocite inherited, essentially unmodified, the lead of the Neoproterozoic sulfides it replaced. The

markedly uraniferous character of the deposit, shown by its anomalous U_3O_8 and ThO_2 contents (Tassinari et al. 2003) and uraninite in the Dn+2 assemblage, reinforces this reading, since in a U-rich system, strongly radiogenic lead accumulates in proportion to the integrated U-Pb history of the source volume. The ca. 2.7 Ga slope requires that the U-Pb fractionation sampled by the chalcocite is Neoproterozoic; lead generated during, or uraninite formed by, the ca. 1.88 Ga event would drag the leachates toward a younger or geologically meaningless mixed slope. The inherited age is geologically informative. It converges with the ca. 2.76 Ga slope of the complete Pb dataset of Tassinari et al. (2003), with ca. 2.76-2.74 Ga sulfide Pb at Igarapé Bahia and Pojuca (Tassinari et al. 2003 and references therein), with the 2701 ± 30 Ma resetting record of the Igarapé Gelado mylonitic gneiss (Melo et al. 2017) and with boron-isotope evidence that the ore fluids interacted with the Neoproterozoic exhalative pile (Melo et al. 2021). An early, ca. 2.7 Ga Cu-bearing or exhalative contribution to the metal budget is therefore plausible and is preserved in the lead inventory. This does not make the present orebody syngenetic. Together with the pre-ore Fe-silicate record at 2653 ± 13 Ma (titanite) and 2635 ± 18 Ma (garnet Lu-Hf; Toledo et al. 2026), it gives explicit anatomy to the composite Pre-Dn history proposed above, in which basin-stage metal was absorbed into the present mineralized zones by the principal ca. 2.57-2.55 Ga event. A second consequence follows directly: the Dn+2 upgrading operated as a closed system for lead - had the granite-related fluid introduced significant external metal, the chalcocite would carry Orosirian or mixed lead. This is independent isotopic confirmation of the stoichiometric upgrading argument developed above.

The younger disturbed ages complete the ladder in order of increasing susceptibility. Titanite at 2497 ± 5 Ma (Machado et al. 1991), hydrothermal monazite at 2452 ± 14 Ma (Melo et al. 2017), texturally secondary, microfracture-filling chalcopyrite at 2427 ± 130

741 Ma and a 2430 ± 37 Ma Sm-Nd mineral isochron on garnet-bearing schists (Tassinari et
742 al. 2003) converge on a discrete Siderian episode of recrystallization and Pb loss within
743 the composite Pre-Dn record. Late euhedral magnetite at 2112 ± 12 Ma and a whole-rock
744 Rb-Sr errorchron of 2101 ± 130 Ma on chloritized gneisses (Tassinari et al. 2003) are
745 compatible with Rhyacian re-equilibration during Dn, but neither is treated here as a
746 direct deformation age; the textural link is instead provided by the recrystallized, Sn-
747 parallel magnetite fabric described above. Open-system Rb-Sr model ages of ca. 2.24-
748 2.17 Ga (Cordani et al. 1984) are mixed records of the same broad Paleoproterozoic
749 disturbance. In a deposit as U-rich as Salobo, stepwise-leaching ages of intermediate
750 position are the most vulnerable to mixing between inherited Neoarchean lead and
751 younger in-situ radiogenic lead hosted in U-bearing micro-inclusions. We therefore attach
752 interpretive weight only to intervals supported by at least two independent isotopic
753 systems, as is the case for each interval discussed here.

754 The K-Ar record closes the ladder at its low-retentivity end. No K-bearing phase so far
755 dated in the Salobo schists retains Archean argon: muscovite gives 2022 ± 32 and $1988 \pm$
756 47 Ma, amphibole 1940 ± 45 Ma, biotite 1971 ± 27 to 1903 ± 35 Ma, and plagioclase 1740
757 ± 27 and 1358 ± 31 Ma (Cordani et al. 1984 and references therein; Table 2). These are
758 total-gas ages and cannot be assigned individually to events. The muscovite, amphibole
759 and biotite values cluster around the regional Sereno-to-Uatumã interval, whereas
760 plagioclase records still younger Ar loss. This pattern mirrors the fabric-constrained
761 $^{40}\text{Ar}/^{39}\text{Ar}$ record elsewhere in northeastern Carajás, where Sereno-related mica
762 reservoirs coexist with discrete ca. 1.88 Ga Uatumã-related components (Tavares et al.
763 2026). Two ensemble conclusions are robust. First, the absence of retained Archean K-Ar
764 ages at deposit scale shows pervasive post-Archean resetting within the deforming, fluid-
765 affected Cinzento corridor. Second, the mica-amphibole record is consistent with

superposed Sereno and Uatumã disturbance, matching the fine-grained syn-Dn+1 biotite and muscovite and the pervasive Dn+2 chlorite-sericite alteration described above.

The ca. 2.1 Ga Rb-Sr "chloritization" age does not conflict with the ca. 1.88 Ga age proposed here for the Dn+2 chlorite. Two chlorite generations coexist at Salobo: disseminated retrograde chlorite in the host rocks, formed during Dn-Dn+1 greenschist-facies reworking, and undeformed vein-related chlorite of the Dn+2 assemblage. The Rb-Sr errorchron was obtained on pervasively chloritized wall-rock gneisses, the host-buffered material deliberately excluded from our vein-anchored thermometry, and is best read as dating the older, retrograde generation, within a Rhyacian to Orosirian window consistent with its large uncertainty.

Read in this way, the three-stage model of Tassinari et al. (2003) – primary mineralization at ca. 2.7 Ga, re-equilibration at ca. 2.5 Ga and greenschist-facies overprint at ca. 2.1 Ga – anticipated much of the sequence proposed here. What it lacked was the structural framework that assigns each disturbance to a mappable event, the textural evidence that reassigns the chalcocite itself to the ca. 1.88 Ga stage, and the Re-Os anchor that fixes the principal ore event at ca. 2.57-2.55 Ga.

Regional correlation and implications

The absolute framework of the deposit now has an explicit hierarchy. The composite Pre-Dn record comprises an early, ca. 2.70 Ga metal contribution preserved in inherited sulfide lead, a ca. 2.65-2.64 Ga pre-ore thermal precursor (Toledo et al. 2026) and the principal IOCG event at ca. 2.57-2.55 Ga, anchored by Re-Os molybdenite (Réquia et al. 2003) and followed by Siderian re-equilibration at ca. 2.50-2.43 Ga. At the young end, Dn+2 is correlated with the ca. 1.88 Ga Young Salobo system (Machado et al. 1991) through crosscutting relations, mineralogy, chlorite chemistry and regional comparison

790 (Pollard et al. 2019; Tavares et al. 2021). Dn and Dn+1 must therefore lie between these
791 brackets.

792 Two ages have been proposed for the main deformation: syn-ore ca. 2.57 Ga reactivation
793 or Paleoproterozoic Transamazonian deformation (Pinheiro and Holdsworth 2000;
794 Pinheiro et al. 2013; Tavares et al. 2018; Trunfull et al. 2020). At Salobo, the ca. 2.57 Ga
795 record is better explained by Pre-Dn extension and hydrothermal activity than by shear-
796 zone reactivation. The 2112 ± 12 Ma magnetite age is compatible with Rhyacian re-
797 equilibration during Dn, although it is not treated as a direct deformation age and inherited
798 or mixed Pb components cannot be excluded (Tassinari et al. 2003). The top-to-SW
799 transport, NE-SW shortening, tectonic style and amphibolite-facies grade match the
800 Transamazonian Cycle near the Carajás-Bacajá boundary, favoring correlation of Dn with
801 the 2.10-2.05 Ga event (Fig. 12, Stage III). This correlation is independently supported at
802 regional scale by fabric-hosted hornblende (2078 ± 15 Ma) and biotite (2075 ± 14 Ma)
803 ages that date recrystallization and closure within the penetrative fabric of northern
804 Carajás (Tavares et al. 2018; Tavares et al. 2026).

805 If the penetrative Sn fabric at Salobo were Neoarchean, the ca. 2.57-2.55 Ga ore event and
806 the main deformation would be broadly coeval, as proposed in earlier models. Three
807 observations argue against this. The ore-hosting veins and breccias are truncated, folded
808 and transposed by Sn rather than occupying Sn-related dilational sites; the regional fabric
809 that Sn matches in style, kinematics and metamorphic grade is independently dated as
810 Rhyacian by fabric-hosted hornblende and biotite elsewhere in northern Carajás (Tavares
811 et al. 2026); and the survival of Neoarchean isotopic records at Salobo is compatible with
812 recrystallization-controlled argon systematics, because ca. 2.52 Ga hydrothermal
813 amphibole reservoirs survive regionally where later re-equilibration was incomplete

(Tavares et al. 2026). The Neoarchean age of the ore and the Paleoproterozoic age of the fabric are therefore not in tension.

Dn+1 shares NNW-SSE shortening, top-to-NNW transport and low metamorphic grade with the regional Sereno Event, anchored regionally at 2009 ± 19 Ma and 1965 ± 11 Ma (Tavares et al. 2026), but no direct age constraint exists at Salobo; the K-Ar mixtures discussed above are consistent with, though not diagnostic of, Dn+1 recrystallization. Figure 12 (Stages I-IV) summarizes the proposed evolution.

We nevertheless treat these age assignments as correlations. The model explains the observed overprinting sequence and fits regional tectonometamorphic patterns, but it does not exclude earlier 2.60-2.45 Ga reactivation elsewhere in Carajás or locally cryptic Siderian events that were later transposed by Dn. The garnet record along the Cinzento belt is consistent with this hierarchy. Lu-Hf data indicate that at least part of the garnet, most plausibly the inclusion-rich cores described above, is Neoarchean (2635 ± 18 Ma; Toledo et al. 2026), and Sn deflects around whole garnet grains. Because Dn temperatures at Salobo (500-650 °C) lie below the Sm-Nd closure temperature of coarse garnet (~700-750 °C) and isotopic resetting appears to require hot fluids rather than temperature alone (Toledo et al. 2026), a recrystallization-dominated, fluid-poor Dn would leave the Neoarchean garnet record intact, as observed. We therefore regard the limpid garnet rims as candidate Transamazonian overgrowths on inherited Neoarchean cores, though pre- to syn-Dn growth or recrystallization of the rims cannot be excluded.

Conclusions

Integration of structural analysis, mineral chemistry and published geochronology leads to the following conclusions.

1. Four overprinted structural families are recognized at Salobo. Pre-Dn, Dn, Dn+1 and Dn+2 carry distinct hydrothermal or metamorphic assemblages. Their

contrasting P–T conditions and crosscutting relations require at least four tectonothermal episodes, and Pre-Dn may itself combine several pulses.

2. The overall geometry of the Neoproterozoic mineralized zones is controlled by the younger Dn ductile and Dn+1 ductile-brittle structures.

3. The deposit's main structural framework likely formed during the Transamazonian Cycle (2.10–2.05 Ga) and was later overprinted by a second, compressive-transpressive system correlated with the Sereno Event (2.01–1.96 Ga; Tavares et al., 2018; Tavares et al., 2026); no evidence was identified that either event formed a major ore stage or introduced substantial external metal.

4. The second hydrothermal event is best correlated with ca. 1.88 Ga magmatic-hydrothermal activity, almost 700 m.y. younger than the ca. 2.55 Ga ages previously assigned to this stage, in a shallow, brittle, fluid-dominated, extensional-transensional setting correlated with emplacement of the Young Salobo granite, with a felsic fluid component recorded by vein chlorite chemistry. Its main sulfides, bornite and chalcocite with minor late-stage chalcopyrite, are concentrated mostly in the high-grade central zone of the deposit.

5. The anomalously high copper grades in this central zone were likely reconcentrated during this Paleoproterozoic hydrothermal event.

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CRediT authorship contribution statement

Eric Bernard Gonçalves Prado: Investigation (fieldwork, drill-core logging, petrography, microtectonics and mineral chemistry), Writing – original draft. Felipe Mattos Tavares: Conceptualization, Investigation (fieldwork and structural analysis), Formal analysis (regional correlation), Writing – original draft, Writing – review & editing. José Carlos Sícoli Seoane: Resources, Writing – review & editing. Julio Cezar Mendes: Investigation (electron-microprobe analyses and mineral chemistry), Writing – review & editing. Junny Kyley Mastop de Oliveira: Investigation (regional mapping), Writing – review & editing.

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Declaration of competing interest

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. Access to the Salobo mine, drill core and unpublished company maps was granted by Vale S.A. for the purposes of this academic study; Vale S.A. had no role in the structural interpretation, data analysis, or the decision to publish.

Data availability

The deposit compilation (SM1), electron microprobe data supporting the mineral chemistry results of this study (clinopyroxene and chlorite analyses; SM2), and the full geochronological record of Salobo (SM3) are provided as Supplementary Materials. Vale assay data underlying the structure-grade panel of Fig. 2 are proprietary and were used with permission; the panel is schematic.

Declaration of generative AI and AI-assisted technologies in the manuscript preparation process

During the preparation of this work, the authors used OpenAI's ChatGPT and Anthropic's Claude to assist with language editing, the organization and development of the manuscript text, data processing and consistency checks, and the preparation and refinement of figures and tables. All AI-assisted outputs, including calculations, interpretations, figures and tables, were critically reviewed, verified and edited by the authors. No generative AI tool was used to generate the primary analytical data. The authors take full responsibility for the content of the published article.

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Supplementary Material 1 (SM1)

Mineral resources, host rocks, mineralisation style and structural controls of IOCG and other Cu-Au deposits in the Carajás Province shown in Figure 1.

Symbol	Deposit	Mineral Resources	Mineralisation age	Host rocks	Mineralisation	Mineralisation Control	Age reference	Resources reference	Mineralisation reference
Star	Salobo	1178 Mt @ 0.63% Cu, 0.4 g/t Au	2576 Ma	Gneiss and metavolcanosedimentary rocks	Bornite, chalcocite, magnetite ± minor chalcopyrite	Structural control, mineralisation along foliation plane of the Cinzento Shear Zone	Requia et al. (2003)	Vale (2012)	Requia et al. (2003)
1	GT-46 / Igarapé Cinzento	-	2600 (1858) Ma	Orthogneiss	Chalcopyrite, bornite, covellite, chalcocite, magnetite	Structural control, mineralisation along foliation plane of the Cinzento Shear Zone	Silva et al. (2006)	-	Silva et al. (2006)
2	Paulo Afonso	330 Mt @ 0.95% Cu, 0.04 g/t Au	-	Metavolcanosedimentary rocks	-	-	-	Centaurus Metals Report (2017)	-
3	Pojuca	58.2 Mt @ 0.78-0.89 wt% Cu and 8.4 Mt @ 0.49-1.11 wt% Zn	~2.76 Ga	Metavolcanosedimentary rocks	Chalcocite, magnetite, sphalerite, pyrrhotite, chalcopyrite	Strata bound	DOCEGEO (1988)	Carvalho (2004)	Dreher et al. (2018)
4	Gameleira	535 Mt @ 0.57% Cu, 0.12 g/t Au	2614 (1839) Ma	Metavolcanosedimentary rocks	Chalcopyrite, bornite, pyrite, magnetite	Stockwork	Marschik et al. (2005) / Pimentel et al. (2003)	Centaurus Metals Report (2017)	Marschik et al. (2005) / Pimentel et al. (2003)
5	Grota Funda	-	2531 Ma	Metavolcanosedimentary rocks	Chalcopyrite, magnetite, pyrrhotite, sphalerite, minor Ti-hematite	Structural control within shear zones	Hunger (2018)	-	Hunger (2018)
6	Furnas	550 Mt @ 0.71% Cu, 0.3 g/t Au	~2.76 Ga	Metavolcanosedimentary rocks	Bornite, chalcopyrite, chalcocite, magnetite	Strata bound	Dreher et al. (2018)	Centaurus Metals Report (2017)	Dreher et al. (2018)
7	Breves	50 Mt @ 1.22% Cu, 0.75 g/t Au	1880 Ma	Metasediments, Águas Claras Formation	Chalcopyrite, pyrite, pyrrhotite	Greisen	Tallarico et al. (2004)	Tallarico et al. (2004)	Tallarico et al. (2004)
8	Igarapé Bahia-Alemão	230 Mt @ 1.26% Cu, 0.83 g/t Au	2.76 Ga (2575 Ma)	Metavolcanosedimentary rocks	Chalcopyrite, bornite, magnetite	Strata bound - NE-SW trend with steep dip	Galarza et al. (2008) / Tallarico et al. (2005)	Centaurus Metals Report (2017)	Tallarico (2003) - Ronzê et al. (2000)
9	Serra Verde	-	2609 Ma	Metavolcanosedimentary rocks	Chalcopyrite, pyrrhotite, pyrite and minor magnetite	Structural control, NW-SE brittle fault	Marschik et al. (2005)	-	Villas and Santos (2001)
10	Maravaia	2.14 Mt @ 4.2% Cu, 0.66 g/t Au	-	Gneiss, metavolcanosedimentary rocks	Native copper, cuprite, chalcocite, chalcopyrite, pyrite, magnetite	Tension gashes related to NW-SE shear zones	-	Ni 43-101 (2014)	Ni 43-101 (2015)
11	Antas Norte	6.38 Mt @ 2.38% Cu, 0.50 g/t Au	-	Metavolcanosedimentary rocks	Chalcopyrite, pyrrhotite	Breccia pipe	-	JORC May 2014	JORC May 2014
12	Estrela	230 Mt @ 0.50% Cu, 0.01 g/t Au	1.88 Ga?	Metavolcanosedimentary rocks, metagranitoids and metagabbros	Chalcopyrite + pyrrhotite + pyrite + magnetite	Stockwork	Volp (2005)	Centaurus Resources report	Volp (2005)
13	Cristalino	482 Mt @ 0.65% Cu and 0.06 g/t Au	2700 Ma	Gneiss	Chalcopyrite, pyrite, magnetite	Stockwork	Soares et al. (2001)	NCL Brasil (2005)	Huhn et al. (1999a)
14	Borrachudos	-	2011 Ma	Metavolcanics and metagabbro	Chalcopyrite, pyrite, magnetite	Lithological	Calabrez (2015)	-	Previato (2016)
15	Bacuri	-	2758 Ma	Gneiss and metagabbro	Chalcopyrite, pyrite, magnetite, hematite	Disseminated in mylonites and deformed quartz veins	Moreto et al. (2015, Miner. Dep.)	-	Moreto et al. (2015)
16	Castanha	-	1.88 Ga?	Acid meta-subvolcanics	Chalcopyrite, pyrite, magnetite, pyrrhotite	Veins, veinlets, stockwork and breccias with foliated clasts	Pestilho (2011)	-	Pestilho (2011)
17	Jatobá	-	-	Metavolcanics and metagabbro	Co-chalcopyrite, Ni-pyrrhotite, Co-pyrite, Co-magnetite, marcasite	Disseminated along foliation coeval to hydraulic breccias, late fractures and veins	Veloso et al. (2016)	-	Veloso et al. (2016)
18	Bacaba	-	2681 (2060) Ma	Gneiss and metagabbro	Chalcopyrite, bornite, covellite, chalcocite, magnetite, hematite	Veins and replacement zones ruled by mylonitic foliation	Moreto et al. (2015, Miner. Dep.)	-	Moreto et al. (2015)
19	Visconde	-	-	Gneiss, metavolcanics, and metagabbro	Chalcopyrite, bornite, pyrite, magnetite	Shear-zone controlled and late breccias and veins	-	-	Craveiro et al. (2012)
20	Sossego	85% of 335 Mt @ 1% Cu, 0.28 g/t Au	1897 Ma	Metagranite	Chalcopyrite, magnetite, pyrite	Subvertical breccia pipes and veins	Moreto et al. (2015)	Lancaster, Oliveira et al. (2000)	Monteiro et al. (2008)
20	Sequeirinho	15% of 335 Mt @ 1% Cu, 0.28 g/t Au	2685 Ma	Metagranite, metagabbro, metavolcanics and gneiss	Chalcopyrite, magnetite, pyrrhotite, pyrite	Breccia, disseminations along mylonitic foliation, veins and stockwork breccias	Moreto et al. (2015)	Lancaster, Oliveira et al. (2000)	Monteiro et al. (2008)
21	Alvo 118	170 Mt @ 1.0% Cu, 0.3 g/t Au	1869 Ma	Metabasalt, metagranite, metagabbro, porphyry dykes	Chalcopyrite, bornite, rare magnetite, hematite	Stockwork	Tallarico (2003)	Rigon et al. (2000)	Tallarico (2003)
22	GT-34	-	2724 Ma	Gneiss	Pentlandite, magnetite, hematite	Shear-zone control with rounded fragmental ore breccias	Garcia (2018)	-	Garcia (2018)
23	Águas Claras	9.5 Mt @ 2.43 g/t Au	-	Granite and greenschist-facies metasedimentary rocks	Chalcopyrite, pyrite, pyrrhotite	Supergenic	-	Silva and Villas (1997)	Soares et al. (1994) / Silva and Villas (1997)

Supplementary Material 3

Published geochronological constraints from the Salobo deposit and immediate host-rock system, grouped by their role in the proposed multistage framework. Original interpretations follow the cited studies. The relative position and revised interpretation integrate mineral context, isotopic behavior and the structural sequence proposed here.

Dated mineral, phase or rock	Age (Ma)	Method and published source	Original interpretation	Position relative to this study	Revised interpretation and observations
Host rocks and inherited record					
Xingu Complex orthogneiss	2950 ± 25	U–Pb zircon; Melo et al. (2017)	Magmatic crystallization of the basement host.	Pre-Dn host	
Salobo schists	ca. 2900	Rb–Sr whole-rock reference isochron; Cordani et al. (1984)	Archean formation or metamorphism of the Salobo sequence.	Pre-Dn host or inherited record	Broad Mesoarchean inheritance in an open system. It is consistent with old basement components.
Strongly altered Xingu host rock	2857 ± 6.7	U–Pb zircon; Melo et al. (2017)	Metamorphism and deformation of the basement.	Pre-Dn host	
Igarapé Gelado gneiss	2763 ± 4.4	U–Pb zircon; Melo et al. (2017)	Magmatic crystallization.	Pre-Dn host	
Foliated amphibolite	2761 ± 3	U–Pb zircon; Machado et al. (1991)	Crystallization of the amphibolitic protolith.	Pre-Dn host	
Granitic vein	2758	U–Pb zircon; Machado et al. (1991)	Crystallization of a granitic vein.	Pre-Dn host	
Igarapé Gelado gneiss	2744 ± 4; 2731 ± 7	U–Pb zircon LA–ICP–MS; Toledo et al. (2023)	Magmatic crystallization of the gneiss.	Pre-Dn host	
Early exhalative and pre-ore record					
Chalcocite leachates	2705 ± 42	Pb–Pb step leaching; Tassinari et al. (2003)	Primary syngenetic Cu mineralization.	Early Pre-Dn	Model age for an early Cu-bearing contribution.
Igarapé Gelado mylonitic gneiss	2701 ± 30	U–Pb zircon; Melo et al. (2017)	Tectonic or hydrothermal resetting during mylonitization.	Early Pre-Dn	Early fluid-assisted zircon disturbance.
Amphibolite, titanite	2653 ± 13	U–Pb titanite; Toledo et al. (2026)	Dynamic metamorphism of the amphibolitic host.	Early Pre-Dn	High-temperature, low pressure, fluid-driven pre-ore re-equilibration of iron-rich protholiths.
Garnet-amphibole schist	2635 ± 18	Lu–Hf garnet; Toledo et al. (2026)	High-temperature Fe-silicate growth before the main IOCG event.	Early Pre-Dn	
Principal Pre-Dn IOCG event					
Old Salobo quartz-feldspathic body, FD405-DA02	2594 ± 6	U–Pb zircon LA–ICP–MS, projected upper intercept; Toledo et al. (2023)	Hydrothermal disturbance or growth in altered zircon.	Pre-Dn re-equilibrated zircon	
Tourmaline leachates	2587 ± 150	Pb–Pb step leaching; Tassinari et al. (2003)	Hydrothermal tourmaline related to tectonic reactivation and ore formation.	Pre-Dn alteration	Low-precision value that overlaps the main event.
Amphibole-rich rock, titanite	2581 ± 5	U–Pb titanite; Machado et al. (1991)	Metamorphic or hydrothermal titanite growth.	Pre-Dn high-T alteration	High-temperature hydrothermal accessory-mineral growth at the onset of the main IOCG event.
Bornite-chalcopyrite leachates	2579 ± 71	Pb–Pb step leaching; Réquia et al. (2003)	Cu-sulfide mineralization.	Pre-Dn ore	
Molybdenite	2576 ± 8	Re–Os; Réquia et al. (2003)	Ore-stage molybdenite crystallization and IOCG mineralization.	Pre-Dn ore	
Old Salobo quartz-feldspathic body	2573 ± 2	U–Pb zircon ID–TIMS, two grains; Machado et al. (1991)	Magmatic crystallization of the Old Salobo Granite and temporal link with ore.	Pre-Dn re-equilibrated zircon	Mixed or hydrothermally reset zircon is plausible.

Dated mineral, phase or rock	Age (Ma)	Method and published source	Original interpretation	Position relative to this study	Revised interpretation and observations
Garnet-amphibole schist	2573 ± 58	Sm–Nd garnet; Toledo et al. (2026)	Second thermal event related to IOCG formation and hot fluids.	Pre-Dn	Legacy open-system value. Its numerical agreement with the main event is suggestive.
Salobo schist	2570	Rb–Sr whole-rock model age; Cordani et al. (1984)	Archean event recorded by an individual whole-rock model age.	Pre-Dn, open system	
Molybdenite	2562 ± 8	Re–Os; Réquia et al. (2003)	Ore-stage molybdenite crystallization and IOCG mineralization.	Pre-Dn ore	Hydrothermal zircon re-equilibration within the principal high-temperature event.
Amphibole-rich rock, zircon	2555 +4/–3	U–Pb zircon; Machado et al. (1991)	Metamorphism and deformation of the Salobo sequence.	Pre-Dn re-equilibrated zircon	
Old Salobo quartz-feldspathic body	2554 ± 3	U–Pb zircon LA–ICP–MS, upper intercept; Toledo et al. (2023)	Hydrothermal resetting or new zircon growth.	Pre-Dn re-equilibrated zircon	Accessory-mineral growth or re-equilibration during the main IOCG event.
Magnetite-rich rock, monazite	2551 ± 2	U–Pb monazite; Machado et al. (1991)	Metamorphic or hydrothermal monazite growth.	Pre-Dn	
Old Salobo quartz-feldspathic body	2549 ± 7	U–Pb zircon LA–ICP–MS, concordia; Toledo et al. (2023)	Hydrothermal zircon growth or re-equilibration.	Pre-Dn re-equilibrated zircon	Hydrothermal resetting within the main event.
Old Salobo quartz-feldspathic body	2547 ± 5.3	U–Pb zircon SHRIMP; Melo et al. (2017)	Isotopic resetting associated with intense deformation and fluid flow.	Pre-Dn re-equilibrated zircon	
High-grade Cu–Au ore, zircon	2535 ± 8.4	U–Pb zircon SHRIMP; Melo et al. (2017)	Isotopic resetting in ore affected by intense deformation and fluid flow.	Pre-Dn re-equilibrated zircon	Young tail of fluid-assisted zircon re-equilibration.
Siderian re-equilibration					
Amphibole-rich rock, titanite	2497 ± 5	U–Pb titanite; Machado et al. (1991)	Late metamorphic or hydrothermal titanite growth.	Pre-Dn post-ore re-equilibration	Siderian recrystallization or Pb loss within the composite Pre-Dn record.
Cu–Au ore, monazite	2452 ± 14	U–Pb monazite; Melo et al. (2017)	Late hydrothermal monazite crystallization in the ore zone.	Pre-Dn post-ore re-equilibration	
Chalcopyrite leachates	2427 ± 130	Pb–Pb step leaching; Tassinari et al. (2003)	Late Cu-sulfide mineralization or tectonic reactivation.	Pre-Dn post-ore re-equilibration	Very imprecise open-system Pb re-equilibration.
Dn–Dn+1 reworking and cooling					
Salobo schist	2240; 2170	Rb–Sr whole-rock model ages; Cordani et al. (1984)	Paleoproterozoic disturbance of the Salobo sequence.	Pre- to syn-Dn, open system	Mixed open-system values that anticipate the Transamazonian overprint. They are not deformation ages.
Ore-zone magnetite leachates	2112 ± 12	Pb–Pb step leaching; Tassinari et al. (2003)	Paleoproterozoic hydrothermal magnetite crystallization.	Syn-Dn magnetite re-equilibration	Magnetite recrystallization or isotopic re-equilibration during the main deformation event
Chloritized gneiss	2101 ± 130	Rb–Sr whole rock; Tassinari et al. (2003)	Paleoproterozoic hydrothermal overprint.	Syn-Dn	Low-precision open-system record broadly compatible with Dn re-equilibration.
Salobo Formation schist, muscovite	2022 ± 32	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn / Dn+1	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.
Salobo Formation schist, muscovite	1988 ± 47	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn / Dn+1	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.
Salobo Formation schist, biotite	1971 ± 27	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.
Salobo Formation schist, biotite	1948 ± 32	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.
Salobo Formation schist, amphibole	1940 ± 45	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn / Dn+2	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.

Dated mineral, phase or rock	Age (Ma)	Method and published source	Original interpretation	Position relative to this study	Revised interpretation and observations
Salobo Formation schist, biotite	1920 ± 53	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.
Salobo Formation schist, biotite	1903 ± 35	K–Ar, Tassinari et al. (1982)	Transamazonian cooling.	Syn-Dn	Cooling or recrystallization during Dn / Dn+1, with possible Dn+2 disturbance.
Dn+2 and younger disturbance					
Young Salobo Granite	1880 ± 80	Rb–Sr whole rock; Cordani (1981)	Crystallization of the Young Salobo Granite.	Syn-Dn+2 magmatic emplacement	Regional bracket for Dn+2 granite-related chlorite alteration and Cu upgrading.
Salobo schist or intrusive granite	1800	Rb–Sr whole-rock model or reference age; Cordani et al. (1984)	Late resetting and granite emplacement.	Dn+2 to post-Dn+2 resetting	Open-system value with no independent significance.
Salobo Formation schist, plagioclase	1740 ± 27	K–Ar; Tassinari et al. (1982)	Minimum age from Ar loss.	Post-Dn+2 disturbed record	Incomplete Ar retention.
Salobo schist	1700	Rb–Sr whole-rock model age; Cordani et al. (1984)	Late isotopic resetting.	Post-Dn+2 disturbed record	Open-system value with no independent significance.
Salobo Formation schist, plagioclase	1358 ± 31	K–Ar; Cordani et al. (1984), data of Tassinari et al. (1982)	Minimum age from Ar loss.	Post-Dn+2 disturbed record	Incomplete Ar retention.