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1 **Metallogeny of a Pan-African oceanic arc: VHMS and gold deposits in the Ariab-Arbaat**
2 **belt, Haya terrane, Red Sea Hills (Sudan).**

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36 16 **Abstract**

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39 17 In the Neoproterozoic Arabo-Nubian Shield, the Ariab gold deposits in the Red Sea Hills
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41 18 (Sudan) are associated with volcanogenic massive sulphide (VMS) deposits, formed in the
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43 19 context of a Tonian ensimatic arc of the Haiya terrane. The mined gold is in gossans, re-
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45 20 concentrating primary gold. All the primary gold occurrences of the Ariab belt, whatever their
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47 21 style (VHMS-, quartz lode- or shear-zone-hosted), appear to have been syn-to late-kinematic,
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49 22 i.e., formed in the context of the final Pan-African collision leading to the Arabian-Nubian
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51 23 Shield (D3 event). Gold concentration occurs as native gold or electrum deposited in late (syn-
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53 24 to post-metamorphic) tectonic features, usually in sets of (micro)cracks, mostly within earlier
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55 25 pyrite or arsenopyrite. It is associated with either galena or bismuth tellurides. Gold deposition
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26 occurred in relation with an intense fluid circulation of metamorphic CO₂-H₂O rich fluids which
27 underwent significant strong decompression from lithostatic to hydrostatic pressures.
28 Decompression, together with temperature decrease and dilution, are probably at the origin of
29 gold deposition, although no clear evidence of unmixing of the volatile was observed.
30 The late-D3 gold event in the Ariab belt has all the makings of an orogenic gold system, notably
31 similar to the Late Carboniferous gold event in the West European Variscan belt. Thus, it is
32 possible to propose that, at the end of the collision events, fluids released from the newly formed
33 terrane root were channelled towards the upper crust through the major D3 shear zones (the
34 Oko shear zone for the Ariab belt). The fluids outpouring from these major drains were then
35 conveyed through fault plays and damaged zones, up to specific chemical traps like massive
36 sulphide ore bodies or quartz veins systems. Thus, the Ariab gold district could be classified as
37 a representative of the shear-zone related orogenic gold deposits.

38
39 Keywords: VMS, ANS, gold, gossan, aqueous-carbonic fluids

40 41 42 **1 INTRODUCTION**

43 The Arabian-Nubian Shield (ANS), which is the largest tract of juvenile continental crust of
44 Neoproterozoic age (Patchett and Chase 2002), is known for its gold endowment since the
45 Pharaonic times (5,500 yr before present): the Egyptian part of the shield hosts more than 250
46 gold production sites (Fig. 1), and extensive gold production exists in the Sudanese part (with,
47 in particular, the Hassai gold mine), Sudan being the second-ranking gold producer in Africa
48 (Johnson et al. 2017, Tamer et al. 2018). The shield (and particularly the Arabian segment) is
49 currently considered favourable for further gold exploration. Gold in the ANS is mainly present
50 as (i) “stratabound”, in tight association with massive sulphide deposits in the Neoproterozoic

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51 arc (Bakheit and Matheis 1993; Abu Fatima 2006; Barrie et al. 2016; Johnson et al. 2017;
52 Volesky et al. 2017), Hassai, Jb Sayid and Bisha being the main occurrences (Fig. 1), and (ii)
53 vein-type “orogenic” gold, closely related to the shear zones, and in particular, to the Najd
54 system (Zoheir et al. 2019a), with notably the large Sukari and Madd adh Dhabab world-class
55 deposits (Fig. 1). A few occurrences associated with BIF in arcs have long been considered
56 synsedimentary (e.g. Botros 2004). Recently, however, Zoheir and Akawy (2010) have
57 demonstrated that they belong to the family of "orogenic" gold deposits (Zoheir and Akawy
58 2010). A large gold-porphyry deposit (Jb Ohier, Fig. 1) has recently been described (Bierlein
59 et al. 2016). Other occurrences were claimed to be also of the porphyry type (e.g., Sasmaz
60 2020), but are likely to be variant of the orogenic class of deposits.

61 The ANS is also the host for various rare metal deposits, including U, Nb-Ta, Sn and W
62 (Lehmann et al. 2017), most of them in relation with the highly fractionated post-tectonic
63 granites (Sami et al. 2017a,b, 2020). Thus, about 17 known Neoproterozoic (620-530 Ma) rare-
64 metal granitic bodies occur in the Sinai, Eastern Desert-Medyan and Hijaz terranes, including
65 the world-class Ta deposit of Abu Dabbab and Ghurayyah. Besides, a few pegmatites are
66 known, with the world-class Ta-pegmatite at Kenticha in Ethiopia (Küster et al. 2009). Very
67 locally, emerald was formed in the Negrus shear zone separating the Central and Southern
68 Eastern Desert terranes, in series including leucogranites, biotite schists, and serpentinites
69 (Abdalla and Mohamed 1999).

70 The ANS is equally the host for other raw materials, although as a minor component. There are
71 podiform chromitite occurrences in several of the ophiolitic sutures (Bonavia et al. 1993; Ali et
72 al. 2020), and Algoma-type BIF (possibly related with “snowball earth” episodes: Stern et al.
73 2006) are present in some arc terranes (Abd El-Rhaman et al. 2012; Khalil and El-Shazly 2012;
74 El-Shazly and Khalil 2014).

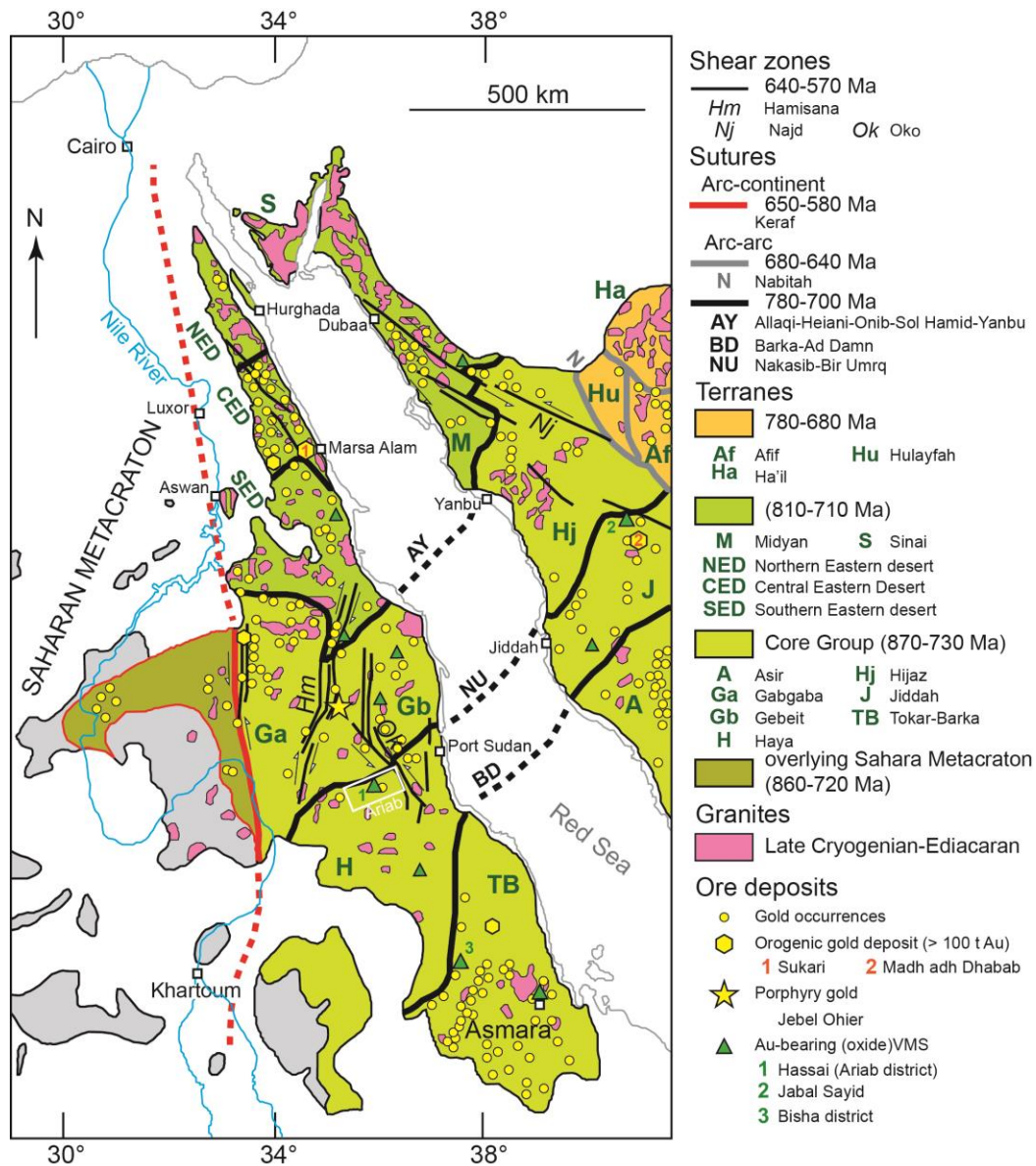


Figure 1. Schematic tectonic map of the Arabian-Nubian Shield, with the location of the Ariab belt (redrawn and adapted from Johnson et al., 2011, with complementary data from Barrie et al. 2016, Bierlein et al., 2016), and Zoheir et al., 2019)

The Hassai gold mine in the SW part of the central Red Sea Hills province, 250 km west of Port Sudan (Fig. 1), is the only mined deposit in Sudan. It was discovered as the result of an extensive exploration program in the Red Sea Hills, conducted in the '80s by a French-Sudanese

84 cooperation team, involving the French BRGM (Bureau de Recherches Géologiques et
85 Minières) and the Sudanese GRAS (Geological Research Authority of Sudan).

86 A total of about 20 gold and base metal deposits were finally discovered in the low-grade
87 metamorphic Ariab-Arbaat belt of the Haya terrane (Fig. 1). The Hassai mine was operated
88 since 1993 by the Ariab Mining Company, a GRAS-COMINOR (Areva) joint venture until
89 2015. After this date, the Sudanese Government has wholly controlled it. The main gold
90 mineralization at Hassai, hosted by a silica-barite rock in the close vicinity of a massive
91 sulphide body, was at first hand interpreted as a lateral hydrothermal exhalative of a VHMS
92 deposit (Cottard et al. 1986a). Recoché (1989) demonstrated the existence of supergene
93 enrichment in the oxidised part of the upper part of the massive sulphide, and it is indeed these
94 “gossan” deposits that are presently mined. The mine, which comprises three deposits (Hassai
95 S, Hadal Adawatib E, and Hadayamet), has produced ~ 85 t of gold until now.

96 In their review of the “Au-rich VMS” class of deposits, Dubé et al. (2007) erroneously ranked
97 Hassai as one of the giant representatives, quoting a 76 t gold content at a mean grade of 9.74
98 g.t⁻¹. They had indeed confused the Hassai reserves with the total bulk reserves of the gossan
99 ores in the whole Ariab camp. Nevertheless, it has since been recognized that the primary
100 sulphide deposits below the gossans are effectively members of the “Au-rich class”, with
101 estimated ~ 59 t of gold for Hassai S and ~ 75 t for Hadal Awatib E, at far lower Au contents,
102 however (1.47 and 1.1 g.t⁻¹ respectively).

103 There is no consensus on the origin of gold concentration in Au-rich VMS deposits. Gold is
104 thought to have been introduced in the massive sulphides either early (syngenetically) with
105 more or less pronounced tectonic-metamorphic remobilisation (e.g., Dubé et al. 2007, and
106 references therein) or later, as the result of fluid circulation in brittle fractures in association
107 with retrograde alterations (e.g., Taube 1986, Marquis et al. 1990a, Yeats and Groves 1998,
108 Gibson and Galley 2007). Besides, Dubé et al. (2007) list a series of characters specific to Au-

109 rich VMS deposits, including the nature of the felsic host, the existence or not of syn-volcanic
110 intrusives, the presence of advanced argillic alterations. On this basis, he suggested that this
111 class of deposits could be a submarine equivalent of the high sulphidation epithermal deposits
112 on land. The true VMS nature of some deposits included in the Au-rich VMS class by Dubé et
113 al. (2007) could, however, be debated (e.g., Boliden: Bergmann-Weihed et al. 1996; Mount
114 Morgan: Arnold and Sillitoe 1989).

115 On the other hand, fossil polymetallic VHMS deposits appear as relatively gold-poor in average
116 when compared with their modern equivalents (Glasby et al., 2000, Barrie and Hannington,
117 1999, Mercier-Langevin et al., 2010). Such low gold grades may indicate gold exportation out
118 from the re-crystallising massive sulphides during the metamorphic events. The observed gold
119 concentrations frequently occur in late brittle structures within the same VHMS bodies and
120 notably in their quartz-rich feeder stockworks prone to late faulting and fluid circulation. They
121 could, therefore, be considered as a particular occurrence of the orogenic class of gold deposits.
122 i.e., as the result of gold re-introduction by late orogenic fluids and trapping on the pre-existing
123 sulphides. Indeed, Marignac et al. (2003) reached this conclusion in their study of the gold
124 setting in the Carboniferous Iberian Pyrite Belt.

125 The present study of the Ariab field allows testing this hypothesis further. The extensive
126 exploration work done at the beginning of the 2000s in the Ariab field to evaluate the gold
127 potential of the unaltered roots of the deposits known by their gossans provided data that
128 allowed a full characterisation of this significant goldfield. Through the unravelling of the
129 tectonic setting of the massive sulphides, their relationships with felsic volcanic activity, and
130 the timing of gold introduction, the present work addresses the gold metallogeny in this key
131 area. We examined the possible roles of exhalative hydrothermal, granite-related, and orogenic
132 processes in the generation of primary gold concentrations before the supergene events in the
133 Red Sea Hills. Besides, the Ariab Au-field was compared with similar fields in the ANS.

134 **2. GEOLOGICAL SETTING**

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136 **2.1. The Arabian-Nubian Shield**

137 Around 25 Ma ago, the Red Sea opening separated the present-day Nubian and Arabian parts
138 of an initially continuous Neoproterozoic juvenile crust belt, the Arabo-Nubian Shield (ANS).
139 The ANS is the northern part an East African Orogen, extending southward to Madagascar. It
140 formed in relation with the closing of a Tonian Mozambique ocean and the formation of the
141 Gondwana supercontinent by the “ultimate collision of Neoproterozoic India with the Congo-
142 Tanzania-Bangweulu Block (in the south) and the Sahara (meta)craton (in the north)” (Johnson
143 et al., 2011).

144 Syntheses by Johnson et al. (2011) and Fritz et al. (2013) provided a robust framework for the
145 ANS structure and evolution, which has not been fundamentally altered since, although
146 clarified in several studies. The following presentation is mainly based on these authors.

147 According to Johnson et al. (2011), and Fritz et al. (2013), the ANS architecture (Fig. 1) may
148 be broadly described as the result of two successive major sequences of events. First, an
149 accretion stage, consisting in a series of arc/arc collisions, and ending with the amalgamation
150 of a proto-Arabian-Nubian Shield (p-ANS), encompassed the ca. 900-650/640 Ma interval
151 (Tonian-Cryogenian). It was followed by arc/continent collisions related to the squeezing of
152 the p-ANS between the Sahara (meta)craton and the Neoproterozoic India craton, during their
153 final convergence (East African Orogeny of Fritz et al. 2013, EAO) between ca. 635 and ca.
154 550 Ma (Late Cryogenian-Ediacaran).

155 The accretion stage may be further divided. Its initiation is the formation of a Core Group of
156 Tonian accreted terranes (Tabar-Barka-Arif, Haya-Jiddah, Gagbaba-Gebeit-Hijraz terranes, ca.
157 900-830 Ma) (Fig. 1), separated by ophiolitic sutures (Barka-Ad Damn, Nakasib-Biir Umq, Fig.
158 1) formed between 780 and 750 Ma. Then, occurred the Late Tonian-Early Cryogenian collision

159 of the Core Group with the northern ANS terranes (East Desert-Medyan terrane, ca. 750-715
1 160 Ma: Kosdroj et al., 2017, Hami et al. 2019; and Sinai terrane, ca. 820-740 Ma: Elisha et al.,
2 161 2019) along with the 710-690 Ma (Eyal et al. 2019) Allaqi-Heiani-Onib-Sol Hamid-Yanbu
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4 162 suture (Fig. 1). The final amalgamation of the p-ANS resulted from the ca. 650-640 Ma
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6 163 collision of the Afif terrane with the already assembled terranes. These events were associated
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8 164 with the first major metamorphic event (M1) in the ANS.
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10 165 The possible presence of Kibaran (c. 1.4-1.0 Ga: Fernandez-Alonso et al. 2012) nuclei,
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12 166 principally in some arcs of the northern terranes (East Desert-Medyan, Sinai), is documented
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14 167 by direct evidence (the 1.12-0.95 Ga Wadi Sa'al volcano-sedimentary complex in the Sinai
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16 168 terrane: Ali-Bik et al. 2017) or suggested by the sporadic occurrence of older than c.0.9 Ga
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18 169 zircons as either xenocrysts or cores in the magmatic arc rocks (Ali et al. 2016 and references
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20 170 therein), or detrital zircons in arc siliciclastic sediments (Li et al. 2018). Yet, it remains that
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22 171 the p-ANS, and in particular, the Core Group, are characterised by the pervasive emplacement
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24 172 of ensimatic island arc (Ali et al. 2020) and synorogenic granitoids at c. 845-c.700 Ma
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26 173 (Robinson et al. 2015) explaining a mostly juvenile lead isotope signature (Johnson et al. 2011
27
28 174 and references therein).
29
30 175 The second (EAO) sequence was marked by the late accreting events in both western and
31
32 176 eastern parts of the orogen. To the west, the collision of the p-ANS with the Saharan (meta)-
33
34 177 craton resulted in the Keraf and Atmar-Dalgado sutures (Fig. 1) at c. 620-600 Ma (Megerssa et
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36 178 al., 2020). To the east, the accretion of the Al Amar arc (690 Ma ophiolite) along the Nabitah
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38 179 suture (Fig. 1) occurred at 620-610 Ma (El-Bialy et al. 2020). Resulting from India's arrival, it
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40 180 was coeval with significant metamorphism (M2) and deformation in the Sinai terrane (Elisha
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42 181 et al. 2017, 2019). These events were accompanied/followed by the development of large intra-
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44 182 continental shear zones dilacerating the p-ANS, with NW-SE ductile sinistral shear zones and
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46 183 their conjugate N-S and NNE-SSW dextral structures (Fig. 1). The most important of these
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184 structures is the 2000 km-long crustal-scale NW-SE Najd system (Abdelsalam and Stern 1996),
185 with its numerous second and third-order splays in the East Desert terranes (Fig. 1). In NE
186 Sudan, the N140°E Oko sinistral shear zone (Abdelsalam 1984) is part of the Najd system. In
187 contrast, the dextral N-S Hamisana shear zone (Saeed et al. 2020) represents the conjugate set.
188 The activity of these shear zones extended from c. 630 to c. 580 Ma (Hassan et al. 2016 a, b)
189 or even 560-550 Ma (Hamisana, Oko, Keraf: Johnson et al. 2011). In the northern terranes, it
190 was associated with the exhumation of high-grade rocks of the arc roots, forming metamorphic
191 core complexes (Meyer et al. 2014). Unconformable “molassic” basins developed at the same
192 time (starting at c. 630 Ma, Bezenjani et al. 2014), as well as volcanism (Dokhan ignimbrite
193 610-590 Ma: Hamimi et al. 2019). Several magmatic pulses yield at that time to the syn-to late-
194 and post-kinematic granite intrusions (Lundmark et al. 2012). The basins history is interrupted
195 at c. 610 Ma, with low-grade metamorphism (M2) and deformation of the earliest sediments.
196 The development of an unconformity below the upper series (Bezenjani et al. 2014) during the
197 “first post-collisional cycle” defined by Eyal et al. (2019), recorded the final collision with the
198 Saharan metacraton. The succession of two types of granites is observed, with some overlap.
199 First came high-K calc-alkaline granites (635-590 Ma), followed by hypersolvus alkali feldspar
200 granites and peralkaline granites (608-580 Ma) (Be’eri-Shlevin et al. 2009). The granite plutons
201 are by far more frequent in the northern terranes (Fig. 1). Minor, but systematic mafic
202 magmatism (small intrusions and dykes) encompassed the late granite activity (Khalil et al.
203 2015). It continued until 545 Ma (Abbo et al. 2020), documenting extension in the ANS from
204 590 to 545 Ma (Fritz et al. 2013).

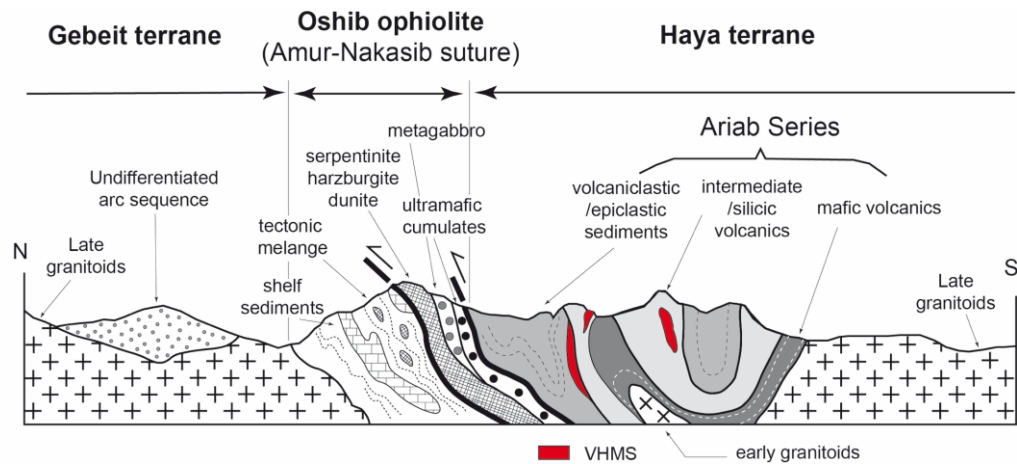
205 By the end of the Ediacaran, the entire ANS had been subjected to erosion and peneplanation,
206 forming a low-relief surface on which the Lower Cambrian sandstones were deposited (Avigad
207 et al. 2005).

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2 210 **2.2 The Ariab-Arbaat belt in the Haya terrane**

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4 211 The Ariab belt is located at the Haya terrane boundary with the Nakasib-Biir Umq ophiolitic
5 complex, separated from the complex by a major NW-verging thrust (Schandelmeyer et al.
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9 213 1994) (Fig. 2). Therefore, the Ariab belt history is intimately linked to the Core Group's
10 formation and its evolution.
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14 215 The time of formation of the northern Haya terrane, of which the Ariab belt is but the
15 northernmost unit, is bracketed between 887 ± 20 and 854 ± 9 Ma (Reischman et al. 1992). The
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19 217 history of the deformation of the Ariab arc related to the Haya/Gabgaba-Gebeit collision can
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21 218 be broken down into two events, D1 and D2 (Fig. 3). From D1 to D2, the shortening direction
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24 219 rotated from NE-SW to E-W, and the deformation evolved from compressive to transpressive,
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26 220 with dextral reactivation of the suture. According to Wipfler (1996), D1 thrusts and tight to
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29 221 isoclinal $N60^\circ-70^\circ E$ F1 folds, coeval with the development of the main regional S1 schistosity,
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31 222 generally close to S_0 in the Ariab belt, were steepened and refolded during D2. Folds F2 are
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34 223 of limited amplitude, a few metres to some ten metres, with axial plane S2 schistosity and strike
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36 224 $N0^\circ-30^\circ E$ upright. D1 and D2 occurred under the same low-grade (greenschist facies)
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39 225 metamorphic conditions.



227 Figure 2. Diagrammatic cross-section through the Gebeit-Oshib-Haya terranes, showing the
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2 228 Amur-Nakasib ophiolitic suture and the Ariab Series (redrawn from Abu Fatima, 2006, after
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4 229 Abdel Rahman, 1993 and Wipfler, 1996).
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9 231 After uplift and erosion, magmatic activity resumed in the amalgamated terranes, with the
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11 232 deposition in the Ariab area of the Arbaat volcano-sedimentary rocks, and the intrusion of the
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13 233 G2 granite suite. The ~790 Ma old Arbaat Group consists of tholeiitic, rift-related mafic
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15 234 volcanic rocks with subordinate felsic volcanic rocks and siliciclastic sedimentary rocks
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17 235 (Abdelsalam 2010). The intrusion age of the G2 suite is poorly constrained, between 750-740
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19 236 Ma and 710 Ma (Stern and Abdelsalam 1998). Final accretion of the proto-ANS onto the
20
21 237 Saharan (meta)craton was associated with a renewal of tectonic activity in the Haya terrane (D3
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23 238 event, Fig. 3). The D3 event was mainly transpressive, with an E-W shortening direction, at the
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25 239 origin of the F3 folds and S3 schistosity in the Ariab area. Small scale thrusts are also observable
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27 240 (see § III-5). D3 was coeval with the development of conjugated NW-SE to NNW-SSE sinistral
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29 241 and NNE-SSW dextral strike-slip faults under ductile, and then dominantly brittle, conditions
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31 242 at the shield scale (Fig. 1). Sinistral kinematics on the Oko shear zone occurred between ca.
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33 243 640 and ca. 560 Ma (Abdelsalam 1994).
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Cycle	Global event	Tectono-metamorphic event	Objects	Age
Pan-African	Juvenile arc in oceanic environment	Hydrothermal sub-sea floor metamorphism	Basalt to rhyolite submarine to subaerial volcanism (Ariab Series) and coeval plutonism (G1 suite) VHMS and barite deposits	ca. 900 Ma G1 granitoids at 888±3.7 Ma
	Arc accretion and thickening <i>unconformity</i>	D1 Obduction and oblique arc collision Low-grade metamorphism	Regional S1 schistosity Recumbent to upright regional N60-70°E F1 folds Transpressive SZ	ca. 770 Ma
		D2 Late shortening Low-grade metamorphism	Upright local N0-30°E F2 folds Regional S2 schistosity Dextral reactivation of D1 SZ	
	Evolved mature arc	Contact metamorphism	Awat-Aseriba calc-alkaline series Coeval to late plutonism (G2 suite)	730-720 Ma 720-680 Ma
	Final collision	D3 Transpression	Conjugated sinistral NW-SE to NNW-ESE and dextral NNE-SSW ductile to brittle SZ Faint local S3 schistosity	620-580 Ma
Gondwana break-up			Anorogenic plutonism	150 ± 4 Ma
Red Sea rift	Red Sea early opening	D4 Fracturation	NE-SW faults	Miocene

Fig. 3. Main events affecting the Haya terrane (see text for explanation and references).

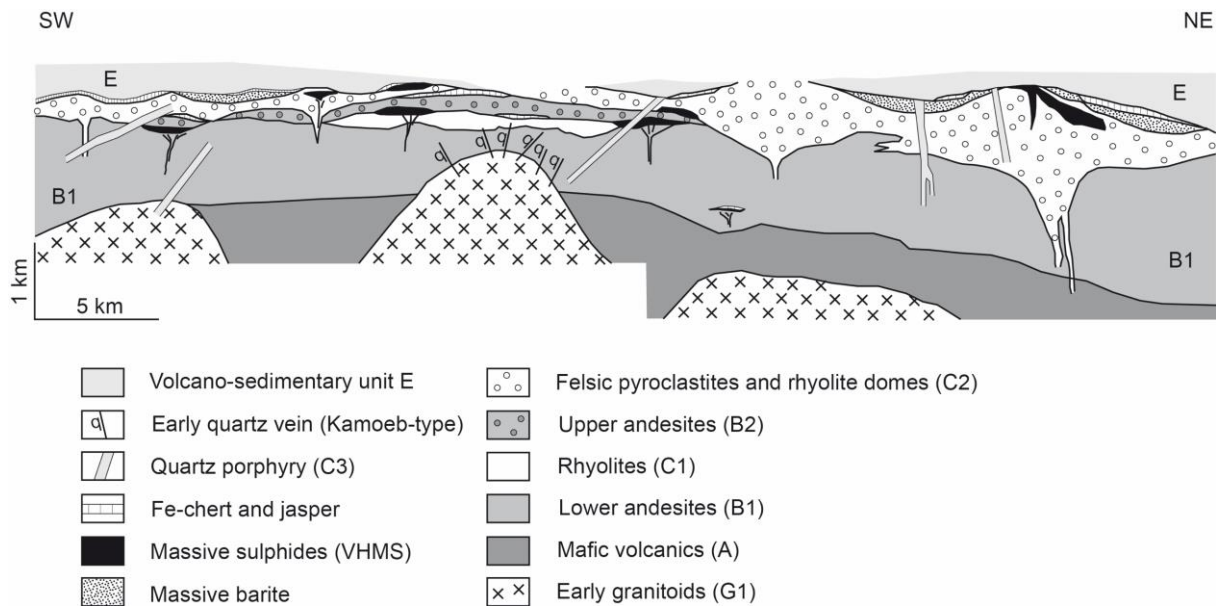
After that, the Haya Terrane remained stable, as the whole Nubian Shield in the Red Sea Hills.

The only recorded anorogenic magmatism associated with Gondwana break-up (Höhndorf et al. 1995), is the 150±4 Ma Jabal Tame complex in the Haya terrane (Kuster et al. 1993). The region then experienced NE-SW faulting related to the Red Sea's early opening (D4 event: Fig. 3).

256 **2.3 Lithostratigraphy of the Ariab belt**

257 **2.3.1. The Ariab series**

258 The Ariab series make the lowermost suite of the northern Haya terrane. The BRGM teams
 259 have subdivided the series into five litho-type units, labelled A to E, based on their main
 260 components, as a whole succeeding one to the other from bottom to top when seen in their pre-
 261 tectonic setting (Cottard et al., 1986b). However, there are complications, and it was necessary
 262 to introduce further subdivisions within the B and C units to reconstruct a “true”
 263 lithostratigraphic pile (Fig. 4).



266 Figure 4. Reconstruction of the Ariab series lithologic pile before the D1 deformation event
 267 (redrawn from Abu Fatima, 2006).

269 *Unit A: Tholeiitic series:* Basalt flows with minor andesitic basalt and microgabbro (the latter
 270 as sills and dykes) constitute a series of more than 1000 m thickness. Locally well-preserved
 271 pillow and vesicular textures testify for a submarine emplacement.

272 *Unit B: Intermediate volcanic unit B* at the outcrop is a thick volcanic unit with a thickness of
 273 about 1000 m, subdivided into two sub-units (B1 and B2, the latter varying from 10 to 200 m

274 in thickness) due to the local intercalation of a thin lenticular felsic sub-unit (C1, see below and
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2 275 Fig. 4). The B2 sub-unit is the D unit of Cottard et al. (1986b). Both B1 and B2 sub-units are
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4 276 mainly basaltic andesite to andesite in composition (Fig. 5). Most of the rocks are hyaloclastites,
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7 277 judging from the abundance of small vesicular angular fragments set in the chlorite schists that
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10 278 constitute the B unit's main rock-type. Yet, locally preserved pillow- and vesicular-lavas are
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12 279 evidence for subaquatic flows, and some massive fine-grained homogeneous facies may
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14 280 represent former lava flows and sills.
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285 *Unit C: Felsic volcanic unit*, rhyodacitic to rhyolitic in composition (Fig. 5), is subdivided into

286 three sub-units:

287 - C1 sub-unit is intercalated within the B unit (see above and Fig. 4) and consists of
288 small lava domes and flows.

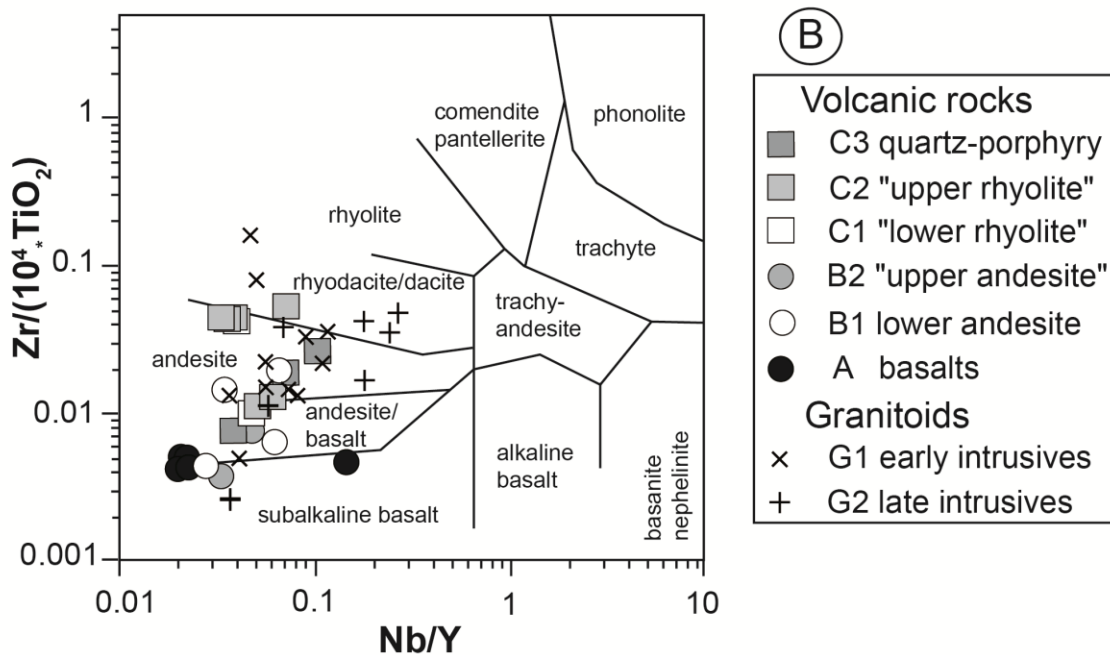
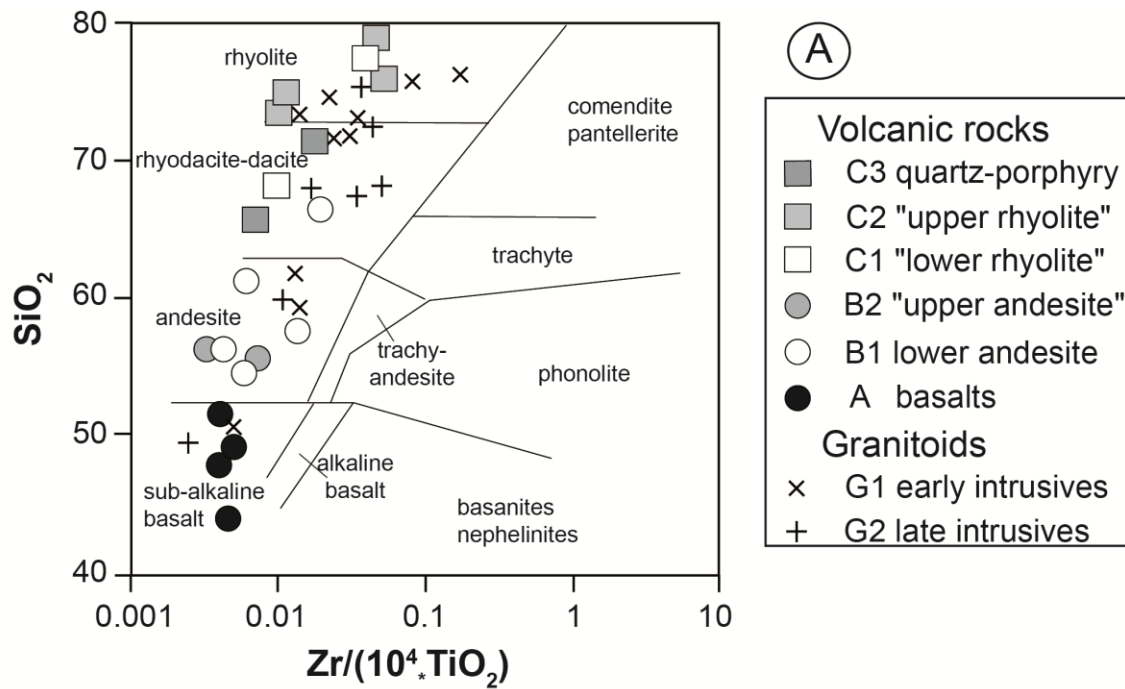
289 - C2 sub-unit, the main host for the massive sulphide and barite deposits, consists of
290 extrusive lavas and pyroclastics. The former exhibit homogeneous textures, with locally
291 preserved flow structures or autoclastic breccias. The rhyolitic facies contain quartz and K-
292 feldspar phenocrysts set in a fine-grained felsic matrix. Rhyodacitic facies are more porphyritic,
293 with quartz, plagioclase and hornblende phenocrysts. The latter display well-defined primary
294 bedding with alternating fine-grained (ashes? Pumice tuffs?) and coarse-grained (clastics)
295 horizons, in which fragments are so much transformed through hydrothermal alteration,
296 deformation, and metamorphism that their original features are usually obliterated.

297 Significant thickness variations (from a few metres up to several 100 m), correlated with
298 petrographic changes, suggest that the C2 sub-unit (and C1 sub-unit) is made of irregularly
299 distributed extrusive lava domes surrounded by pyroclastic aprons. Laterally, the C2 sub-unit
300 extends at the B sub-unit expense from the SW to the NE, with the total disappearance of the
301 B2 subunit (Fig. 4), suggesting some degree of interfingering between andesitic and more felsic
302 magmatism in the Ariab-Arbaat belt.

303 - C3 sub-unit is a set of rhyodacite dykes that crosscut all the Ariab series facies,
304 including C2 and the associated massive sulphide and barite deposits, to the notable exclusion
305 of the E unit. The dykes are quartz-porphyrries, with abundant quartz and feldspar phenocrysts
306 set in a fine-grained matrix with quartz, feldspars, and biotite. These dykes may be coeval with
307 the early granitoids (G1) (see below). Alternatively, the dykes could have been the feeders for
308 the explosive volcanism at the origin of the E unit.

309 *Unit E: Volcano-sedimentary unit E* is a thick (several 1000 m) sequence of volcano-
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2 310 sedimentary rocks, with a regional trend from dominantly volcano-clastic in the SW to more
3
4 311 silico-clastic in the NE. A thin and continuous layer of black siliceous metapelites with cherts
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6
7 312 and jaspers marks the unit's base and grade to layered tuffs, greywackes, and schists,
8
9 313 interbedded with fine-to coarse-grained epiclastic horizons, intermediate to silicic in
10
11 314 composition. The unit comprises greywackes and reworked ash-falls, alternating with marbles,
12
13 315 quartzites, and minor C-rich cherts and breccia-conglomerates of volcanic derivation.
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16 316 Greywackes are usually medium- to coarse-grained, with lithoclasts of various origin and
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18 317 rhyolitic quartz and feldspars as components.
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 320 Figure 5. Classification of the Ariab belt magmatic rocks in the SiO_2 vs Zr/TiO_2 (A) and
 321 Nb/Y vs Zr/TiO_2 (B) diagrams (Winchester and Floyd, 1977). The use of the "immobile" elements
 322 Zr and Ti partly alleviates the bias introduced by the intense hydrothermal alteration affecting
 323 most volcanic and many plutonic rocks (see text).

326 **2.3.2 Regional hydrothermal metamorphism of the Ariab Series**

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2 327 A propylitic assemblage (chlorite-albite-actinolite-epidote-pyrite) reveals a penetrative
3
4 328 hydrothermal alteration in the basalts and andesites of the A and B units. It is also found as
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7 329 porphyroclasts in the S1 schistosity. In the andesites, relictual chlorite-epidote-calcite
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10 330 amygdales may also be recognised.

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12 331 The felsic rocks from the C unit occur as quartz-albite-sericite-pyrite ± chlorite ± tourmaline
13
14 332 schists. The latter result from the metamorphism and deformation of pre-kinematically altered
15
16
17 333 rocks (silicification, quartz-sericite-pyrite acid leaching), and exhibit in particular pyrite
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19 334 porphyroclasts rimmed by quartz pressure-shadows, and schistosed silicified fiammes. A
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22 335 chlorite ± quartz ± pyrite ± epidote pre-kinematic assemblage replaces former phenocrysts. At
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24 336 Hadayamet, hydrothermal leaching produced per-aluminous facies converted into dalmatianite
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27 337 during the subsequent tectonic and thermal events.

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29 338 These alterations were likely related to the massive sulphide deposition. They are most
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32 339 pronounced in the units hosting the VHMS deposits (B2, C1, C2) and attain their maximal
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34 340 development in the vicinity of the deposits themselves. Deformed pyrite-chlorite networks are
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36 341 thus found in B2 unit, and chlorite-quartz-sericite microfracture networks in C2, both features
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39 342 being likely paleo-stockworks.

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41 343 On the other hand, silica alteration is linked to the massive sulphide bodies, usually in the
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43
44 344 hanging-walls, and expressed as a pervasive replacement of other minerals by quartz.

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46 345 **2.3.3 Early granitoids (G1 suite)**

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48 346 Early granitoids make up to 40% of the Ariab-Arbaat belt's outcrops, generally occurring in the
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51 347 antiform cores. Along their margins, G1 intrusions display a penetrative foliation concordant
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53 348 with the regional S1 schistosity (see below). Remnants of the contact metamorphic aureolas
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56 349 occur as pre-kinematic assemblages of large actinolite-epidote-magnetite porphyroblastic
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58 350 assemblages (now, porphyroclasts) in the mafic to intermediate rocks, and of cordierite and
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351 andalusite in the felsic lavas of the Ariab Series close to the G1 margins. The G1 suite was
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2 352 therefore clearly emplaced pre-kinematically within the Ariab Series.
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4 353 The G1 suite is a typical I-suite consisting of biotite-hornblende granites to granodiorites,
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7 354 tonalite-trondjemites, and minor quartz-diorites (Fig. 5). Fine-, medium- and coarse-grained
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9 355 varieties (the latter, more abundant) coexist in the intrusions. The fine-grained facies
10
11 356 predominates and consists of a porphyritic biotite granodiorite, with showing plagioclase
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13 357 (oscillatory zoning around An₃₅ compositions), corroded quartz and rare K-feldspar
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16 358 phenocrysts. Only large rounded quartz phenocrysts are found in the coarse-grained variety.
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19 359 **2.3.4 Late (post-kinematic) granitoids (G2 suite)**

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21 360 Large plutons of undeformed equant granodiorite and granite (with minor tonalite) intrude, with
22
23 361 sharp contacts and chilled margins, all the Ariab series members, including the E-unit. They
24
25 362 produced contact metamorphism marked by post-kinematic biotite-chlorite assemblages. The
26
27 363 dominant facies in the G2 suite is a biotite porphyritic granodiorite, with mainly K-feldspar and
28
29 364 usually zoned An₃₀₋₄₀ plagioclase phenocrysts (Fig. 5).
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32 365 To the NE of the studied area, the G2 granitoids are known to be both intrusive and reworked
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34 366 in the Awat-Asoteriba series, meaning that they are partly coeval with this series, in agreement
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36 367 with the 723 ± 4 Rb-Sr age of the Awat-Asoteriba series and the 720-680 Ma ages of the G2
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38 368 suite.
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42 369 A faint deuteric alteration is present in the granitoids (K-feldspar sericitisation, biotite
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44 370 chloritisation) and could have been coeval with the widespread propylitisation observed within
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46 371 the Awat-Asoteriba series.
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51 373 **2.4. Geochemistry and geodynamic setting of the Ariab belt**

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53 374 Table 1 provides representative analyses of the main rock-types of the Ariab series and the two
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55 375 granitoid suites.
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376 **2.4.1. Volcanic rocks of the Ariab series**

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2 377 *Alteration effects:* As could be expected from the petrographical data, many analyzed samples
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5 378 display evidence of alteration effects, as, for example, high LOI values (up to 11%, Table 1).
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7 379 The A-B multi-cationic diagram of Debon & Le Fort (1988) was used to discriminate the effects
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10 380 of the alteration on the bulk rock compositions. In this diagram, the alumina excess (as $A = Al-$
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12 381 $Na-K-2Ca$, in milli-cations) is plotted vs the mafic charge (as $B = Fe+Mg+Ti$, in milli-cations).
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14 382 It is sensitive to the effects of either propylitisation or phyllic alteration. In this diagram, most
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17 383 volcanic rocks from the units A, B2 and C units display evidence of a significant alteration
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19 384 involving an increase of the peraluminous index and a rise in the mafic index (Fig. 6-A). From
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21
22 385 Fig. 6-A, a separation may be made between “unaltered” and “altered” rocks from the sample
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24 386 array (although consideration of the LOI does suggest that even “unaltered” rocks may have
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26
27 387 suffered limited alteration). Another alteration process, well seen in the field, is the silicification
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29 388 process, well displayed in the Q-P diagram of de La Roche (1990) in Fig. 6-C: with $Q = Si-$
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31 389 $(K+Na+2Ca)$ (quartz excess, as milli-cations) and $P = K-(Na+Ca)$ (feldspar balance, as milli-
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34 390 cations) this diagram (a combination of the classical Q-Ab-Or diagram with a Q-An-Or
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36 391 diagram), is well fitted for alterations involving quartz.

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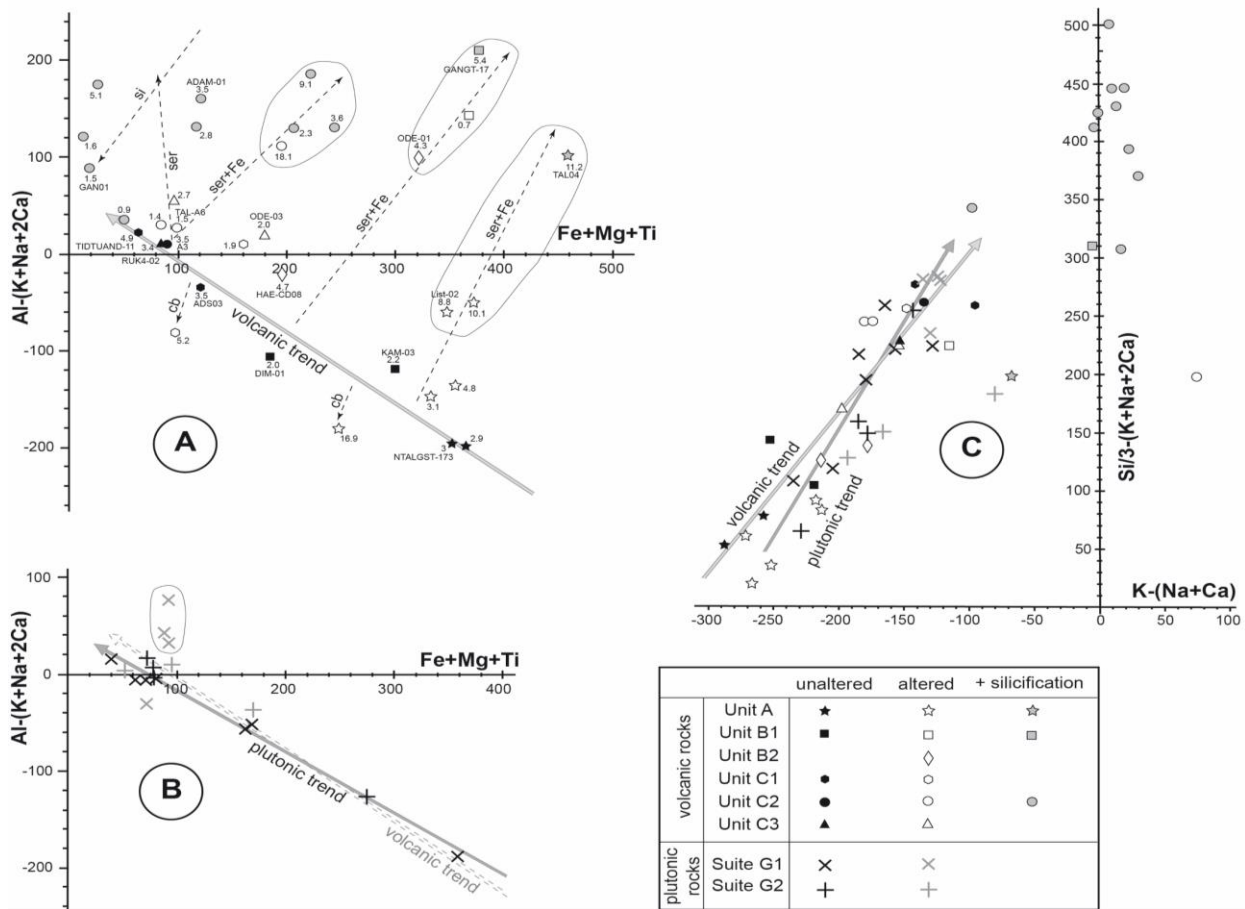


Figure 6. Projection of the Ariab belt magmatic rocks in adapted mineral-chemical diagrams. A. The A-B diagram of Debon and Le Fort (1988) for volcanic rocks of the Ariab Series. Numbers associated with rock symbols are the LOI. B. The A-B diagram of Debon and Le Fort (1988) for the Ariab-Arbaat belt plutonic G1 and G2 suites. C. The Q-P diagram of La Roche (198x) for all magmatic rocks of the Ariab-Arbaat belt. Although most of the volcanic and many plutonic rocks are affected by an intense alteration (involving an increase of the peraluminous index A), a differentiation trend, common to all the volcanic series and, at least, the G1 plutonic rocks, remains discernable in all diagrams (see text).

A quantitative assessment of these alteration effects would require a set of samples with identical initial compositions. This constraint was not satisfied with the sample array in this study. Nevertheless, a limited attempt was made, and isocons (Grant 1986) were drawn for

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406 couples of samples from the A, B1, and C2 units, respectively, selected based on their
407 relationships in the A-B diagram (Fig. 6-A). Assuming constant Al, practically all the elements
408 are more or less mobile, except possibly Si (Fig. 7). A problem arises from the Zr behaviour:
409 always well correlated with Hf, Zr is apparently either depleted or added in significant
410 quantities (in Fig. 7, the “zircon line” is often very far from the isocon line as defined by Al).
411 These apparent depletions or enrichments are anti-correlated with apparent enrichment or
412 depletion in Cr. They could reflect primary differences between each couple's samples related
413 to different partial melting ratios in the source. Thus, Cr is known to be enriched, whereas HFSE
414 and HREE are depleted when partial melting increases in the source. Also, Y and the HREE
415 usually plot on, or close to, the “zircon line” in Fig. 7.
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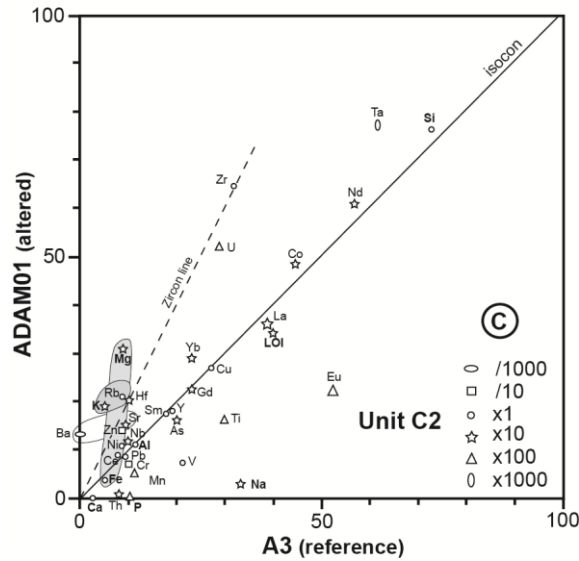
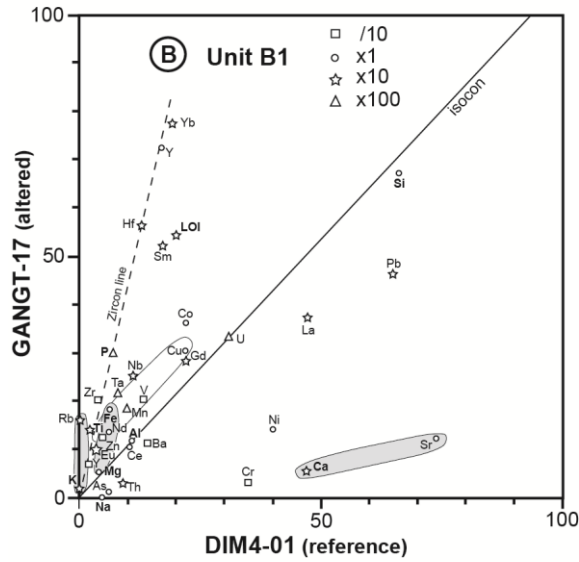
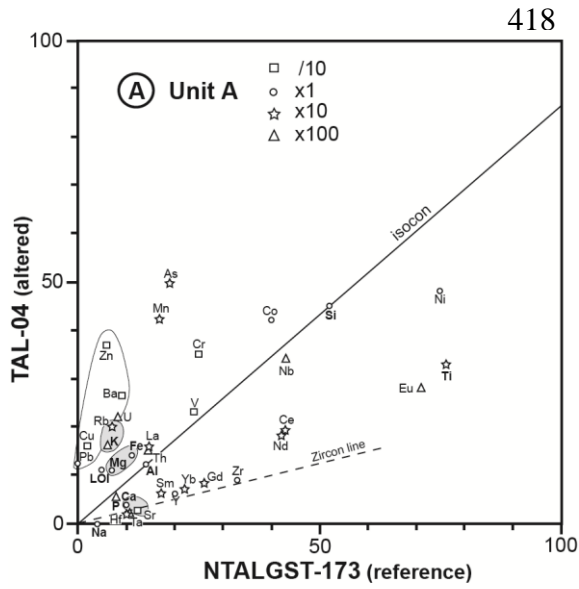


Figure 7. Isocon plots (Grant 1986) for selected couple of “unaltered”-altered rocks from (A): Unit A basalts; (B): Unit B1 andesites; and (C): Unit C2 rhyolites.

1 442 As a consequence, the apparent elemental mobility displayed in the isocon plots could be, at
2 443 least partly, due to primary differences between the compared rocks. Two different behaviours
3
4 444 are distinguished: i) some elements are either enriched/depleted in correlation with the apparent
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7 445 enrichment/depletion in Zr-Hf or Cr and can be considered as rather “immobile” elements, and
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9 446 ii) others behave independently and are considered as “mobile” elements.
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12 447 Among the “mobile” elements, there are those that behave very consistently in all the units: Fe
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14 448 and Mg are gained, as K and Rb, whereas Ca and Na are systematically depleted. Sr follows Ca
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17 449 excepted in the samples with a strong introduction of Ba, where Sr correlates to the latter (Fig.
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19 450 7-A, B, and C). Destructive alteration of plagioclase explains the Ca, Na, (Sr) depletions,
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21
22 451 whereas chloritisation is responsible for increasing the mafic elements. Mg input testifying for
23
24 452 seawater involvement and thus confirming the field evidence for submarine volcanism) and
25
26 453 sericitisation explains the behaviour of K and Rb. When present, Fe input is often higher than
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29 454 Mg input, in correlation with very high LOI, consistent with pyritisation. In the same way, Ba
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31 455 is often enriched through the alteration process, notably in the C unit, reflecting there the
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34 456 presence of primary deposits of massive barite (see below, § III-2). Other “mobile” elements
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36 457 behave more erratically: the base metals (Cu, Pb, Zn), U (mostly added, however), P and Ti.
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39 458 Certain elements, as Ce, are often immobile, anyway. Eu represents a particular. Thus, Eu is
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41 459 “mobile”, being often depleted, except in the basalts, where it is added. It could reflect fractional
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44 460 crystallization in the andesite-rhyolites series or cumulative processes in basalts. However,
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46 461 considering the REE profiles in Fig. 8, the intensity of the negative Eu anomaly is correlated to
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48
49 462 the intensity of alteration, particularly in the felsic metavolcanics, leading to consider Eu as a
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51 463 “mobile” element, and leached together with Ca and Sr during plagioclase destructive
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54 464 alteration. Nevertheless, as pointed by Leshner and Campbell (1987), it is likely that alteration
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56 465 did not create the Eu anomaly (when present) and was only responsible for its enhancement.
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58 466 These findings are consistent with the current knowledge on the mobility during alteration of
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1 467 elements such as Sb, Cs, Pb, Rb, Ba, K, U and P (Utsunomiya et al. 2006, and references
2 468 therein), and, indeed, with the somewhat erratic behaviour of most LILE in the spidergrams in
3
4 469 Fig. 8 to 11.

6
7 470 Among the “immobile “ elements, a distinction arises between those that correlate positively
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9 471 with Cr (i.e., increasing with partial melting in the source): La, Th, As; and those that correlate
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11 472 negatively with Cr (i.e., following zircon): Y, Yb, Sm, Gd, Nb, Ta.

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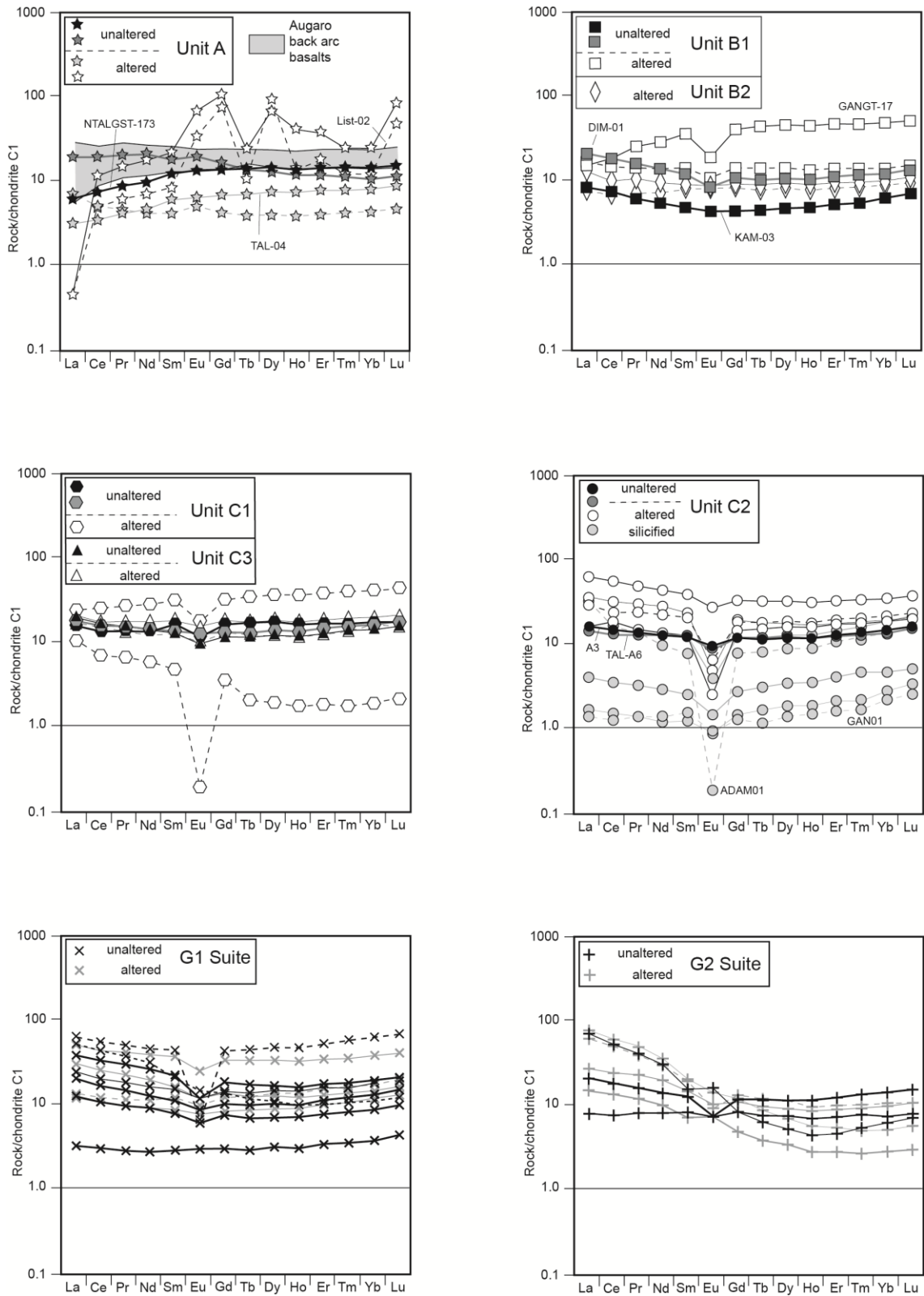


Figure 8. Chondrite-normalized REE plots for the Ariab belt magmatic rocks (C1 chondrite normalizing values from Anders and Grevesse 1989). Data from the Augaro back-arc basalts in Eritrea (Teklay et al. 2003) have been added to the Unit A plot for comparison (see text).

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479 Most of the REE and HFSE, currently used for constraining magmatic rocks' origin and
480 geological setting, may be taken as practically immobile through the alteration (e.g., Leshner et
481 al. 1986; Leshner and Campbell 1987). The element ratios are also not significantly affected by
482 alteration (e.g., Condie, 2005). This contention may be extended to the Ariab series, as indeed
483 may be seen in Fig. 8, 9, 10 (and 11), in which alteration has no significant impact on the HREE
484 and HFSE profiles.

485 *Geochemical characterisation:* The less altered rocks of the Ariab Series are metaluminous to
486 weakly peraluminous, low in K₂O (from less than 0.1% in the Unit A basalts to up to 1.34% in
487 the Unit C rhyolites) and rather iron-rich (Fe# [FeO*/FeO*+MgO] from 0.5-0.6 in the basalts
488 to 0.8-0.9 in the rhyolites). They display in the A-B and Q-P diagrams a typical calc-alkaline
489 trend (Fig. 6-A and C).

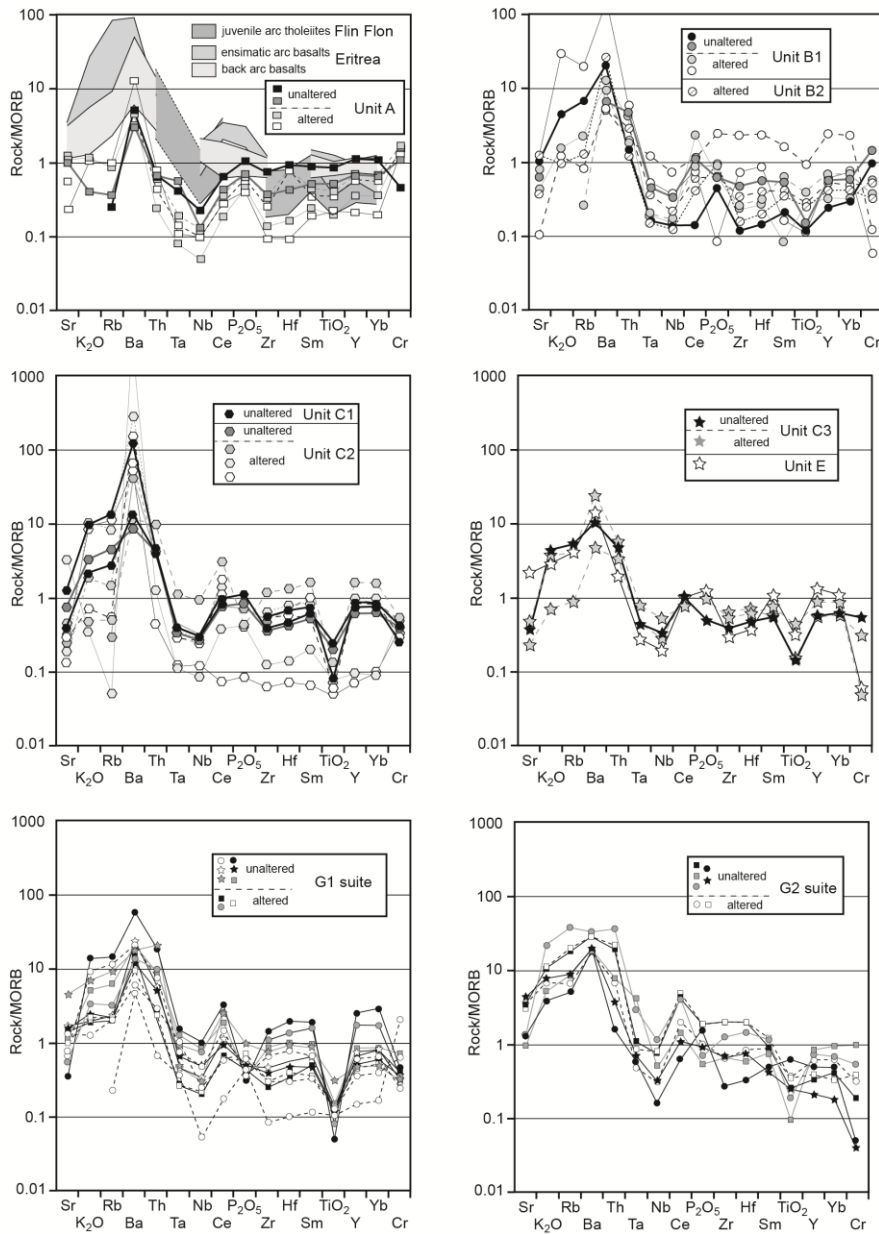
490 All lithologies from these three units display similar characteristics:

- 491 - The chondrite-normalised REE profiles (Fig. 8) are flat, with very low [La/Yb]_N (from
492 0.4 to 1.4, with a slight increase in the C Unit felsic lavas) and [Gd/Yb]_N (from 0.7 to 1.5)
493 ratios. The Eu negative anomalies in the less altered rocks are either absent (or even positive),
494 as in Unit A basalts and andesitic basalts (Eu/Eu* between 1 and 1.19), or moderate as in Unit
495 B basaltic andesites and andesites (Eu/Eu* between 1 and 0.73) and Unit C rhyodacites and
496 rhyolites (Eu/Eu* between 0.72 and 0.96). More pronounced negative anomalies are present,
497 mainly in the felsic rocks, but this is due to the alteration effects discussed in the previous
498 section. Either significant plagioclase fractionation (e.g., Leshner et al. 1986) or a plagioclase
499 restitic phase in the source (e.g., Hart et al., 2004) may explain the negative Eu anomalies in
500 similar metafelsic rocks. In the case of the Ariab series, we must conclude that these processes
501 were of far lesser importance in their genesis. Silicification effects are particularly evident in

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2 502 the REE profiles, acting as a diluting effect (obviously superimposed upon the plagioclase-
3 503 destructive alteration responsible for most of the Eu anomalies) (Fig. 8).

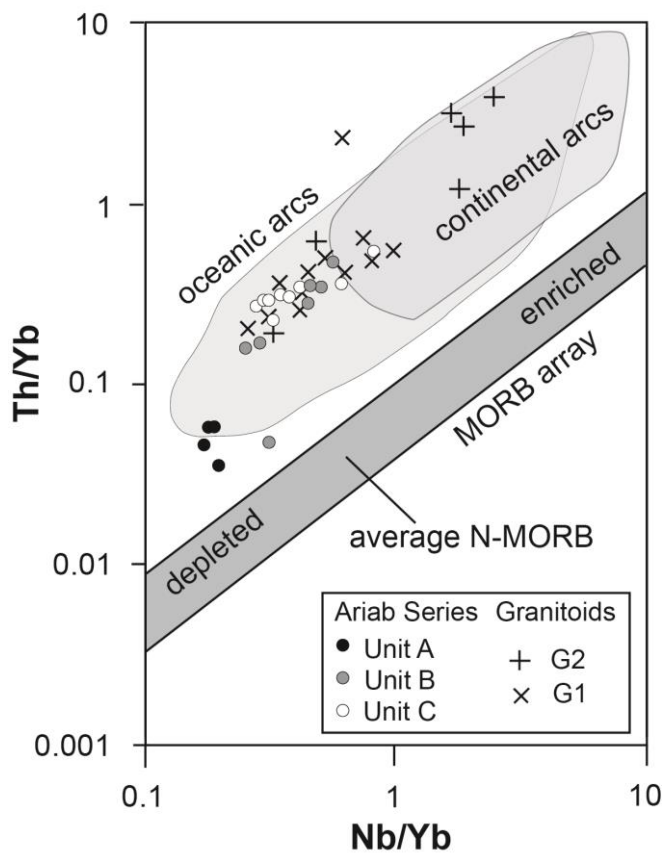
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5 504 - The MORB-normalized spider diagrams (Fig. 9) show LILE enrichment and HFSE
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7 505 depletion.

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9 506 Except for the Unit A basalts, there is consistently a positive anomaly in K-Rb-Ba, with a clear
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11 507 Ba spike contrasting with a constant Sr depletion, accentuated in the altered rocks. The Ba
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13 508 enrichment is amplified with alteration, particularly in the C Unit felsic rocks. It may be
14
15 509 suspected that K-Rb depletion in the basalts is due to an early “spilitisation” on the seafloor
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17 510 with K to Na exchange and Mg enrichment (e.g., Lentz 1998). Such alteration
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19 511 (“keratophyrisation” in felsic rocks) could have been present in all Ariab lithologies and then
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21 512 masked by subsequent K-Rb hydrothermal enrichment associated with sericitisation. Yet, it
22
23 513 seems complicated to assign the low-K nature of the Ariab series to alteration processes only:
24
25 514 it seems likely that whatever alteration processes were in operation, they only accentuated the
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27 515 initial geochemical signature. Thorium, always relatively enriched compared to niobium, is
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29 516 either depleted (Unit A) or enriched (Units B and C).
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 518 Figure 9. MORB-normalized plots for the Ariab-Arbaat belt magmatic rocks, using the
 519 normalization and ordering scheme of Pearce (1983) (in Rollinson 1993), with LILE on the left
 520 and HFSE on the right, and incompatibility growing outward from Ba-Th (save the Cr addition
 521 on the right). Data for comparison with Ariab basalts: Flin Flon arc tholeiites, redrawn from
 522 Syme et al. (1998); Eritrea arc and back-arc basalts, redrawn from Teklay et al. (2003) (see
 523 text).

526 Among the HFSE, Ti and Ta-Nb display a strong negative anomaly, whereas Ce-P and Hf are
 527 always slightly enriched relatively to Zr. The Ti/V ratios are still ≤ 20 in the basalts and
 528 andesites (Units A and B), contrasting with elevated values (generally > 100) in the felsic rocks,
 529 due to a dramatic lowering in the V contents (see Table 1).
 530 Chromium is usually depleted, with the only exception of the B1 Unit andesites. This depletion
 531 may be related to either higher partial melting ratios in the sources or fractional crystallisation
 532 processes in a magmatic chamber.



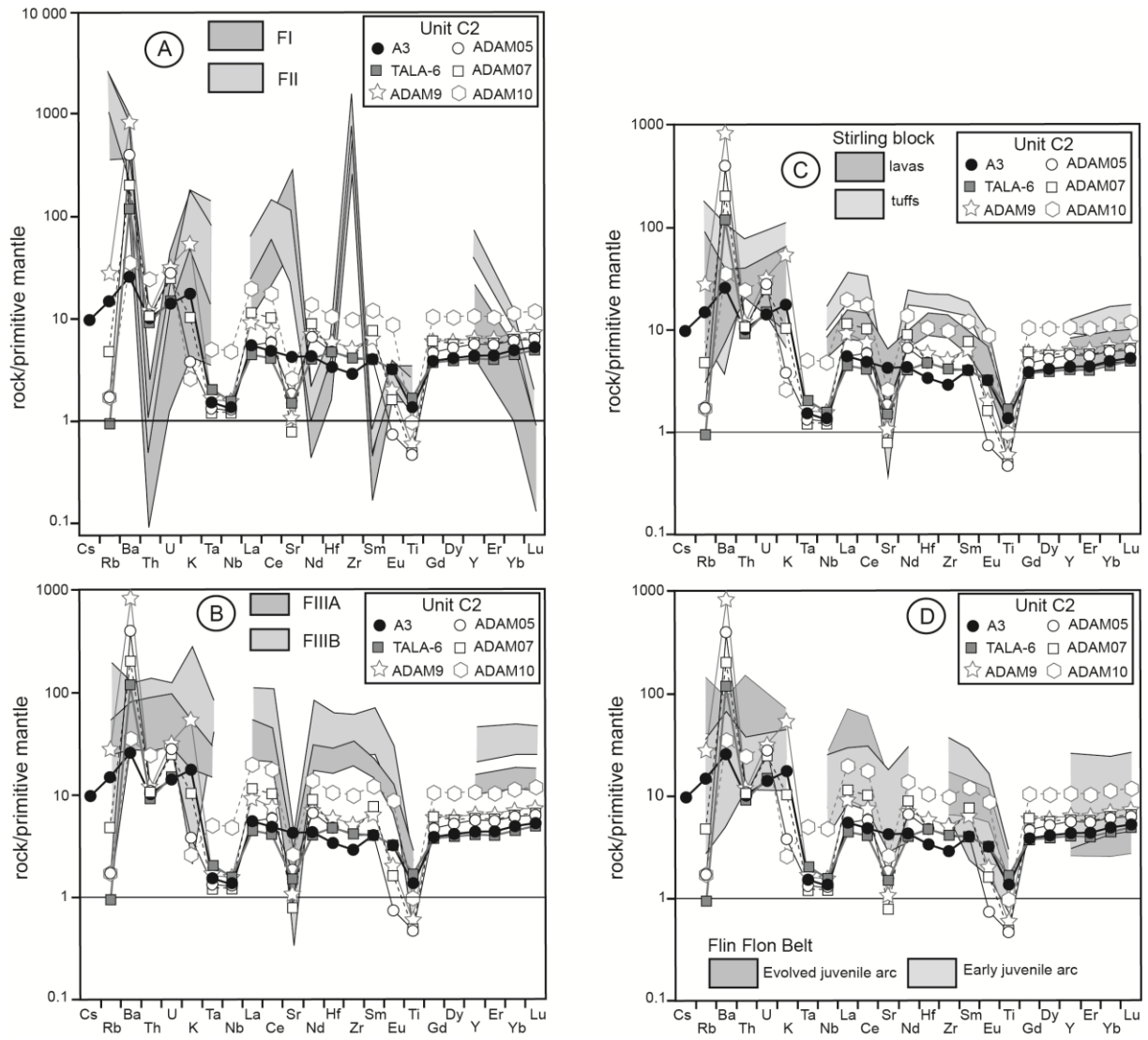
533
 534 Figure 10 Projection of the Ariab-Arbaat belt volcanic rocks in the Th/Yb vs Nb/Yb diagram
 535 of Pearce and Peate (1995), showing clearly their oceanic island-arc affinity. G1 and G2 suites
 536 have also been projected for comparison.

537
 538 All these features are consistent with a subduction-related setting (e.g., Rollinson 1993).
 539 Moreover, the B and B Units' basalts and andesites display characteristics that are most

540 suggestive of an ensimatic juvenile arc environment, the Unit A basalts, particularly with their
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2 541 low-K and very low $[La/Yb]_N$ signatures, being arc tholeiites. A back-arc environment for the
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4 542 mafic and intermediate lavas of the Ariab series may be excluded based on the low Ti/V ratios
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7 543 (back-arc basalts displaying Ti/V consistently >20 , Rollinson 1993) and direct comparison with
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10 544 reference series. In Fig. 9, basalts from the A Unit are compared with juvenile arc tholeiites
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12 545 from the Paleoproterozoic Flin Flon Belt (Canada) and with both ensimatic arc and back-arc
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14 546 basalts from the nearby Neoproterozoic greenstones in Eritrea. It is apparent that the mafic
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17 547 rocks from the Ariab series display MORB-normalised profiles more similar to those of the arc
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19 548 basalts than to back-arc ones. The latter lacks the strong Nb anomaly and the Ce-P relative peak
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22 549 that typify the arc tholeiites, a feature seen on the Ariab series REE profiles (Fig. 8). Finally,
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24 550 the Th/Yb vs Nb/Yb relationships (Pearce and Peate, 1995) equally strongly point towards an
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27 551 oceanic island-arc setting for the mafic, intermediate, and felsic rocks (Fig. 10).
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29 552 Unit C's felsic rocks (especially those of Unit C2) are discussed in more detail as they host most
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32 553 of the Ariab VMS and massive barite deposits. There is indeed a special relationship between
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34 554 felsic lava's characteristics and their capacity to host such deposits (Leshner et al. 1986; Lentz
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36 555 1998; Hart et al., 2004). Hart et al. (2004) define four classes of felsic lavas, FI to FIV: (i) FI
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38
39 556 dacites and rhyodacites, displaying steep REE profiles ($[La/Yb]_N$ 6-34) high Sr and Zr/Y ratio
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41 557 (9-31), low HFSE contents, and negative to moderately positive Eu anomalies (Eu/Eu^* 0.87-
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44 558 2.0), (ii) FII rhyodacites and rhyolites displaying gentle sloping REE profiles ($[La/Yb]_N$ 2-6),
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46 559 moderate Zr/Y ratio (6-11), intermediate Sr and HFSE contents and variable Eu anomalies
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49 560 (Eu/Eu^* 0.35-1.4), (iii) FIII rhyolites displaying flat REE profiles ($[La/Yb]_N$ 1-4) and
50
51 561 subdivided into FIIIa (low Zr/Y 4-7, intermediate HFSE, variable Eu anomaly: Eu/Eu^* 0.37-
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53 562 0.94) and FIIIb (low Zr/Y 2-6, high HFSE, low Sr, pronounced negative Eu anomaly: Eu/Eu^*
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56 563 0.2-0.61) subtypes, and (iv) FIV rhyolites, similar to FIIIa subtype but displaying very flat to
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58 564 slightly LREE-depleted REE profiles, low REE and low HFSE contents. While FI felsic rocks
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565 are always barren, some FII and many FIII (mainly Archean) and FIV groups are associated
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2 566 with VHMS districts. According to Hart et al. (2004), FIII-IV magmas are the products of
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4 567 repetitive partial melting of a single “hydrated basalt source” with an only minor contribution
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7 568 (if any) of a crystal fractionation process. The Unit C felsic rocks of the Ariab Series display
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10 569 all the FIIIa and FIV groups' characteristics, including a very low Zr/Y ratio, between 1.6 and
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12 570 3.6. Fig. 11 compares felsic rocks from the C2 sub-unit with FI to FIII references (Leshner et al.
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14 571 1986) and with two VHMS-hosting felsic rocks from FIV group (Lentz 1998): the
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17 572 Paleoproterozoic Flin Flon Belt, Canada (Syme et al. 1988; Whalen et al. 1998) and the
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19 573 Neoproterozoic Stirling lavas from Cape Breton Island, Nova Scotia (Dostal et al. 1992). Both
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21
22 574 are members of arc tholeiite suites in juvenile oceanic arcs, displaying a characteristic “bi-
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24 575 modal” arrangement with a few andesites intermediate between basalts and andesi-basalts, on
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27 576 the one hand, and rhyodacites and rhyolites, on the other hand. This bi-modality is a feature
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29 577 equally visible in the Ariab Belt.

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 580 Figure 11. Comparison of C2 felsic lavas with reference series in primitive mantle-normalized
 581 LILE-HFSE plots (normalizing values from Ringwood 1991). A: Comparison with the FI and
 582 FII Archean felsic metavolcanic rocks (redrawn from Lesher et al. 1986). B: Comparison with
 583 the FIIIA and FIIIB Archean felsic metavolcanic rocks (redrawn from Lesher et al. 1986). C:
 584 Comparison with the Neoproterozoic Cape Breton Island felsic metavolcanic rocks from an
 585 ensimatic arc (redrawn from Dostal et al. 1972). D: Comparison with the Paleoproterozoic Flin
 586 Flon Belt felsic metavolcanic rocks from a primitive ensimatic arc (redrawn from Syme et al.
 587 1998).

590 **2.4.2. Early granitoids (G1 suite)**

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2 591 Although affected to some extent by hydrothermal alteration, the G1 granitoids display in the
3
4 592 A-B and Q-P diagrams a typical calc-alkaline trend, identical to the volcanic trend (Fig. 6-B
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7 593 and C). They are metaluminous, K-poor ($K_2O \leq 2.04$), and rather Fe-rich (Fe # from 0.5 in the
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10 594 more mafic term to 0.8). They display flat REE profiles (Fig. 8), similar to those of the A unit
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12 595 for the more mafic facies and the C unit for the more felsic rocks. Compared to the C unit
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14 596 profiles, however, they are slightly more fractionated ($[La/Yb]_N$ 0.9-2.1), the HREE
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16 597 fractionation and the deepening of the negative Eu anomaly (Eu/Eu^* 0.57-1.04) being
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18
19 598 correlated. The MORB-normalised plots again compare well with the C unit plots (Fig. 9),
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21
22 599 although there is only a Ce positive anomaly, P being relatively depleted. These features are
23
24 600 consistent with an oceanic arc-setting (Fig. 11). As the G1 granitoids are very close in
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26 601 composition to the C unit's felsic volcanics, a genetic connection is possible. Thus, the G1
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29 602 plutons could have been the feeders for all or a part of the C unit felsic extrusives (in particular,
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31 603 C3 sub-unit), or the reworked volcanoclastics of the E unit.

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34 604 **2.4.3. Late granitoids (G2 suite)**

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36 605 The G2 granitoids are calc-alkaline metaluminous rocks (Fig. 6-B), close in composition to
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39 606 granitoids of the G1 suite. They are K-poor ($K_2O \leq 1.76$) and rather Fe-rich (Fe# from 0.7 to
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41 607 0.8). Compared to the G1 suite, however, they display steeper REE profiles (except for the
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43
44 608 more mafic terms), with $[La/Yb]_N$ between 1.1 and 15.4, and the Eu anomalies are more
45
46 609 variable (Eu/Eu^* 0.58-1.44) (Fig. 8). Besides, in the MORB-normalised plots, the Ti negative
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49 610 anomaly is smoothed or absent (Fig. 9). These characteristics refer the G2 suite to the FII or
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51 611 even the FI groups of Hart et al. (2004), favouring an arc-setting, however more akin to a
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53 612 continental arc than the G1 suite (Fig. 10).

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615 **2.4.4. Summary: the geodynamic setting of the Ariab belt**

616 In conclusion, the petrographic and geochemical evidence points to the lower suites of the Haya
617 terrane representing the remains of a mature island-arc, with volcanic activity shifting from an
618 early submarine mafic (immature) stage (A unit) to a final shallow marine to subaerial felsic
619 (mature) stage (C and E units, G1 suite). By contrast, felsic intrusions of the G2 suite testify for
620 a mature continental arc setting.

621

622 **3. GOLD MINERALIZATIONS**

623

624 Until recent times, gold from the Ariab area mainly came from Miocene supergene enrichments
625 of earlier gold (pre)concentrations in massive sulphide deposits, as gossans – the so-called
626 “silica barite rocks (SBR)” (Recoché 1989). However, it appears that primary gold is not
627 restricted to massive sulphide deposits as three primary gold-bearing associations are now
628 recognised: (i) gold in VHMS deposits, ii) related massive barite and exhalative hematite-
629 magnetite schists, (iii) gold hosted by pre-orogenic quartz veins (Kamoeb), and (iv) gold within
630 carbonate-pyrite-quartz alteration zones related to D3 SZ (Derbikwan), and v) supergene gold
631 enrichments (mantos).

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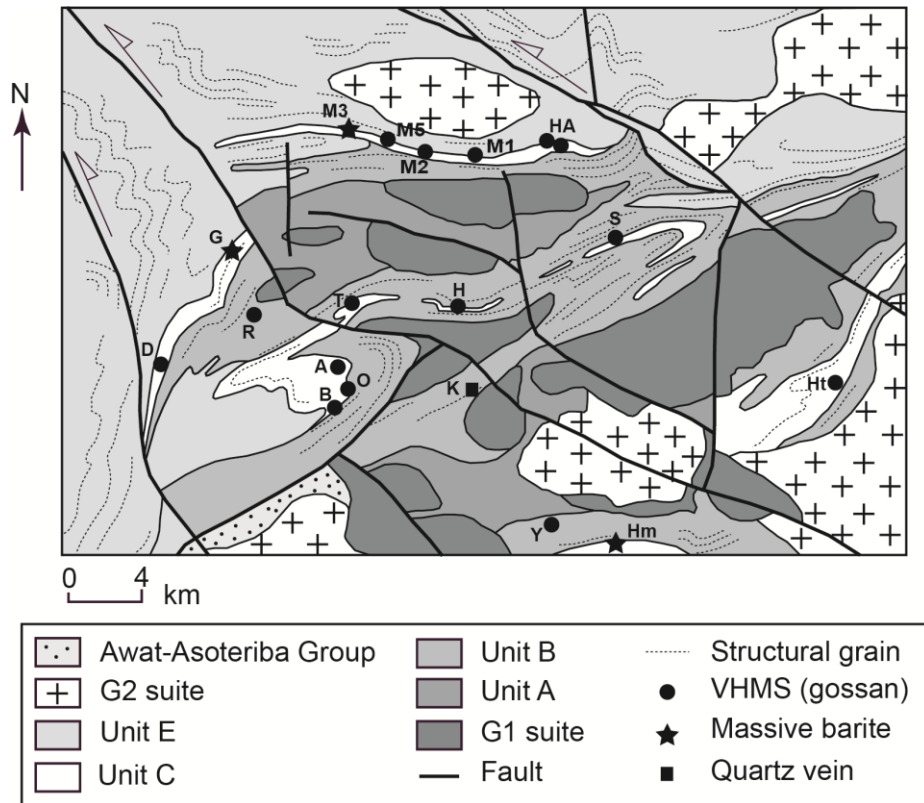
633 **3.1. Gold-bearing massive sulphide (VHMS) deposits**

634 **3.1.1. General characteristics**

635 More than 20 lenses (or lenses clusters) of massive sulphide (≥ 60 % pyrite) are known in the
636 Ariab belt (Fig. 12). They are mostly hosted by the felsic members of the Ariab Series (C1 and
637 C2 sub-units). They were generally strongly deformed during the D1 event, and many primary
638 features may have been obliterated. Nevertheless, it is possible to reconstruct gross pre-
639 kinematic features. The massive sulphide bodies appear to have been concordant layered lenses,

640 with a high aspect ratio (2-25 m x 1,500 m). The presence of a pipe-like stockwork cannot be
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2 641 demonstrated. The footwall sulphide disseminations and diffuse stringer stockworks (down to
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4 642 30 m) within strongly altered volcanic rocks (quartz-sericite-chlorite-pyrite), seems to have
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7 643 been evenly distributed beneath the massive sulphides. Clear zoning is expressed, with a central
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10 644 chlorite-rich zone, grading outward in a quartz-sericite zone. Pyrite is ubiquitous, as early
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12 645 dissemination or stringers (Fig. 13-a). Chlorite is a ripidolite. It may be associated with a
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14 646 tourmaline (hydroxy-feruvite), for example, at Medadip. The sericite is usually a Ba-rich
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17 647 phengite, up to 4.18 BaO wt. % (Fe-rich when shielded in early pyrite), apparently responsible
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19 648 for the systematic Ba enrichment associated with alteration (see the previous section).
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22 649 The primary mineral association was pyrite ± chalcopyrite ± sphalerite, with a noticeable lack
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24 650 in base metals (1.3 % Cu, 2 % Zn on average) and a very low Pb content, in good agreement
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26 651 with the oceanic setting of the Ariab arc. The gold grade varies from 0.6 to 1.2 g.t⁻¹ as an
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29 652 average. The Ariab massive sulphide deposits may be classified as polymetallic VHMS of the
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31 653 Zn-Cu type (Franklin 1993).

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656 Figure 12. Geologic map of the central part of the Ariab area, with the main gold occurrences
657 (redrawn from Abu Fatima 2006).

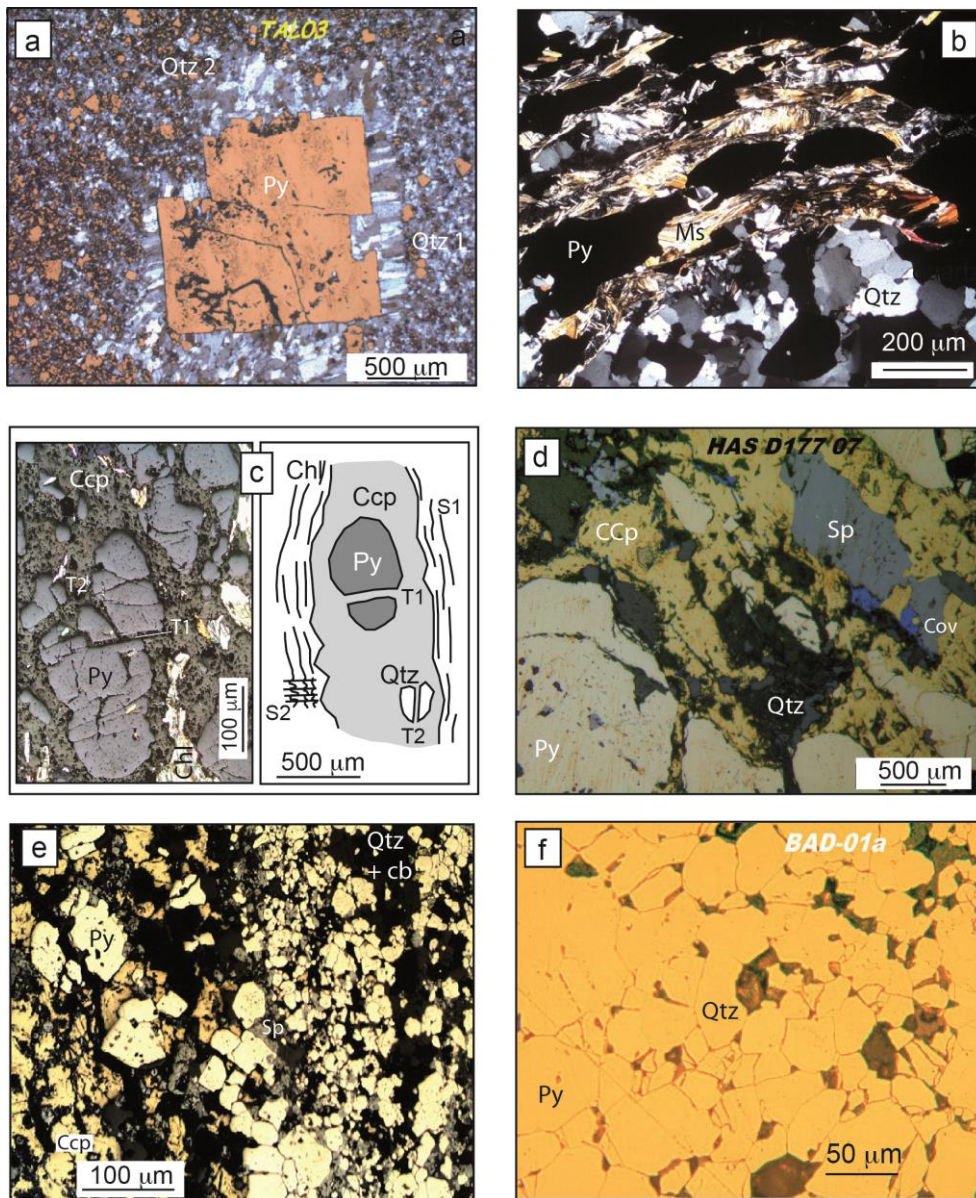
658 A Adasselakh B Baderuk D Dimdim G Ganaet H Hassai HA Hadal Aouatib Hm Hamim
659 Ht Hadayamet K Kamoeb M1-M5 Medadip 1 to 5 O Oderuk R Rukab S Shidimam T
660 Talaidirut Y Yonim

662 3.1.2. Post-depositional history and the timing of gold deposition

663 Based on a detailed study of the mineral/deformation relationships, Abu Fatima (2006)
664 reconstructed the complicated history following the syngenetic deposit of primary sulphides
665 and footwall alteration (Fig. 14).

666 (i) Syn-G1 recrystallisation: the transformation of pyrite into pre-kinematic magnetite 1
667 is currently observed in the massive sulphides, and seems spatially correlated with the relicts
668 of syn-G1 contact metamorphism in the surrounding volcanic rocks.

669 (ii) Syn-metamorphic recrystallisation: chlorite and muscovite currently define the main
 670 S1 schistosity, with numerous porphyroclastic remnants of the pre-kinematic assemblages (Fig.
 671 13-b). The massive sulphides may be strongly reworked, with superimposition of a syn-D1
 672 metamorphic banding, as either layers with ductile elongation in S1 chalcopyrite (Fig. 13-c, d)
 673 or cataclastic bands of microcrystalline pyrite (py1) wrapping pyrite porphyroclasts or both
 674 (Fig. 13-e).

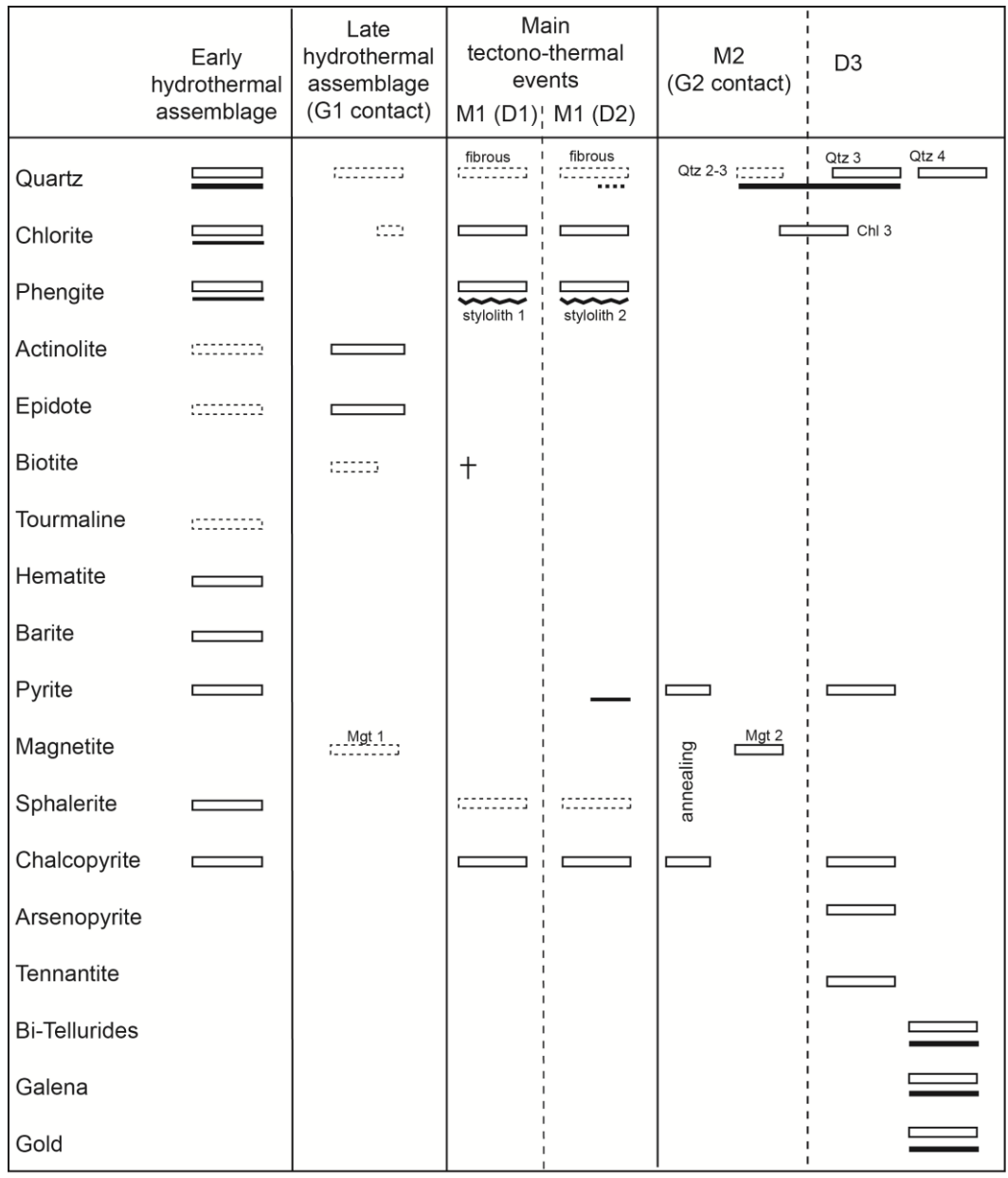


675
 676 Figure 13 Post-depositional transformations in the VHMS deposits. **a.** Porphyroclastic pyrite
 677 (py) in a quartz-sericite matrix (diffuse stockwork, Talaidrut) Qz1 and 2: two generations of
 678 syn-kinematic fibrous quartz (“pressure-shadows”). **b.** Porphyroclastic muscovite (mus) and

679 pyrite (py) in a schistosed stockwork (Medadip 1) quartz: quartz amygdale. **c.** Differential
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2 680 behaviour of massive sulphides during deformation (Yonim): fractured pyrite (py) and ductile
3
4 681 chalcopyrite (cp) (left); sketch of relationships in the same thin section (right) chl: chlorite; qtz:
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6
7 682 porphyroclastic quartz; S1, T1, and S2, T2: two generations of schistosity and related tension
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9
10 683 gashes. **d.** Differential behaviour of massive sulphides (Hassai): porphyroclastic pyrite (py) and
11
12 684 sphalerite (sp) fractured and enclosed in ductile chalcopyrite (cp) elongated in S1; Qz: quartz;
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14 685 cv: supergene covellite. **e.** Two pyrite generations in the massive sulphides (Medadip 5):
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16
17 686 cataclastic pyrite (py1), quartz (qz), and chalcopyrite (cp) banding (S0+S1) and porphyroblastic
18
19 687 euhedral pyrite (py2). **f.** Annealed pyrite (py) (Baderuk) Qz: quartz.

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22 688 Sphalerite may be present as porphyroclasts, embedded into remobilized chalcopyrite,
23
24 689 sometimes with chalcopyrite disease (Fig. 13-d) or be itself remobilized, being often associated
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26 690 with chalcopyrite in “pressure-shadows” around pyrite. Most often, massive pyrite aggregates
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29 691 are fractured, with fibrous quartz infillings, in two grossly orthogonal directions, reflecting the
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31 692 changing shortening directions from D1 to D2 (see above, section II-4).

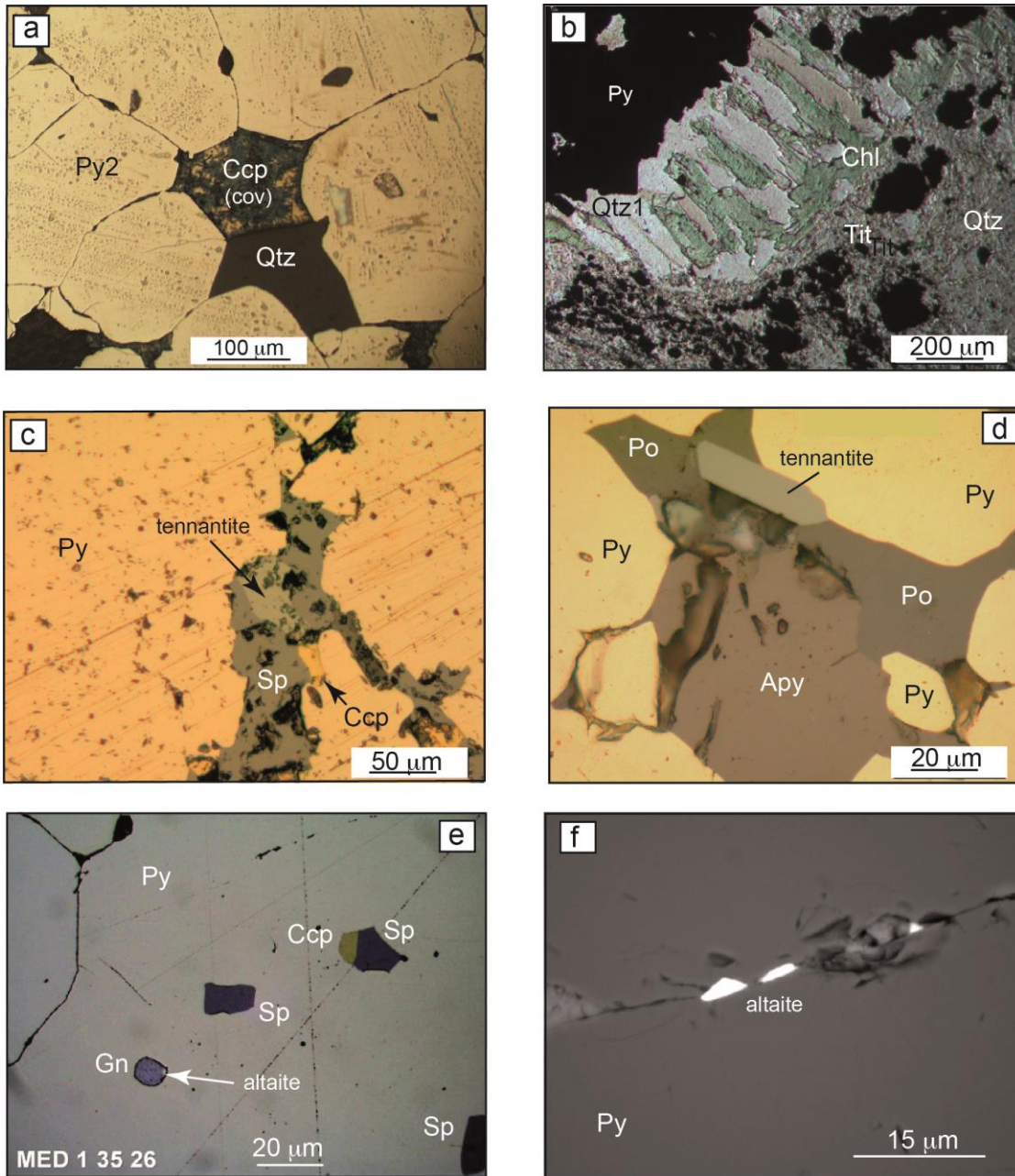
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VHMS Major phase
 VHMS Minor phase
 Kamoeb

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Figure 14. Paragenetic evolution in the VHMS and Kamoeb deposits.

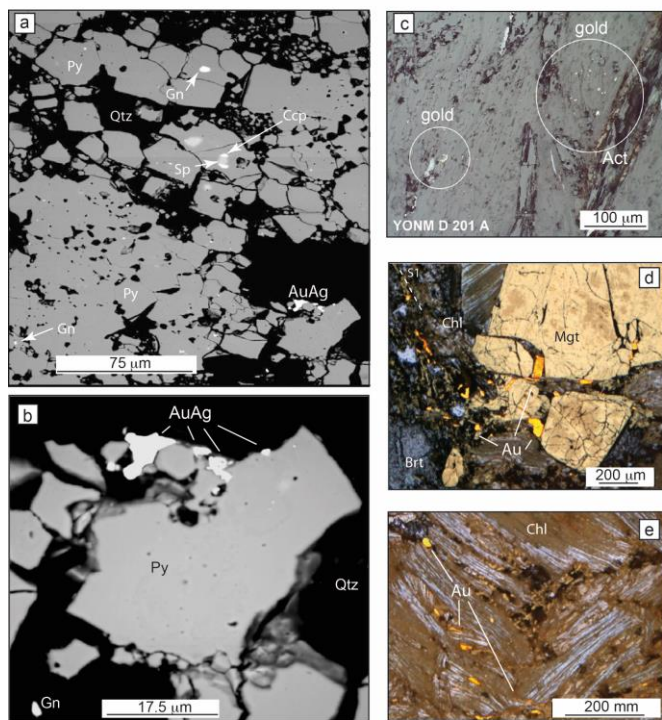


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 41 697
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 43 698 Figure 15. Late events in the massive sulphides. **a.** Quartz (Qtz) and chalcopyrite (Ccp) possibly
 44 recrystallised with pyrite (Py) annealing (Medadip 5). **b.** Replacement of syn-kinematic fibrous
 45 quartz (Qtz1) by post-kinematic chlorite (Chl) (Barut) Qtz: quartz; Tit: titanite. **c.** Fracture-
 46 699 dissolution in pyrite (Py), filled with chalcopyrite (Ccp), sphalerite (Sp), and tennantite (Hadal
 47 700 Ouatib). **d.** Dissolution cavity in pyrite (Py), filled by pyrrhotite (Po), arsenopyrite (Apy), and
 48 701 tennantite (Hassai). **e.** Microvugs inside annealed pyrite (Py), filled with sphalerite (Sp),
 49 702 chalcopyrite (Ccp), galena (Gn), and altaite (SEM-BSE, Medadip 1); note that the microvugs
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705 developed independently from the pyrite joints. **f.** Microcrack in pyrite (Py) with altaite sealings
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2 706 (SEM-BSE, Hadal Aouatib).

3
4 707 (iii) Development of post-kinematic euhedral pyrite (py2) (Fig. 13-e) and pyrite
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6 708 annealing, without conspicuous BMS accumulations at the triple junctions except for a few
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9 709 chalcopyrite grains (Fig. 13-f and 15-a), are often (although not systematically) observed. The
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11 710 timing of this event is obscure, as it may record either the peak temperature of metamorphism
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14 711 associated with the relaxation following the collision events or the thermal effect of G2
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16 712 emplacement.

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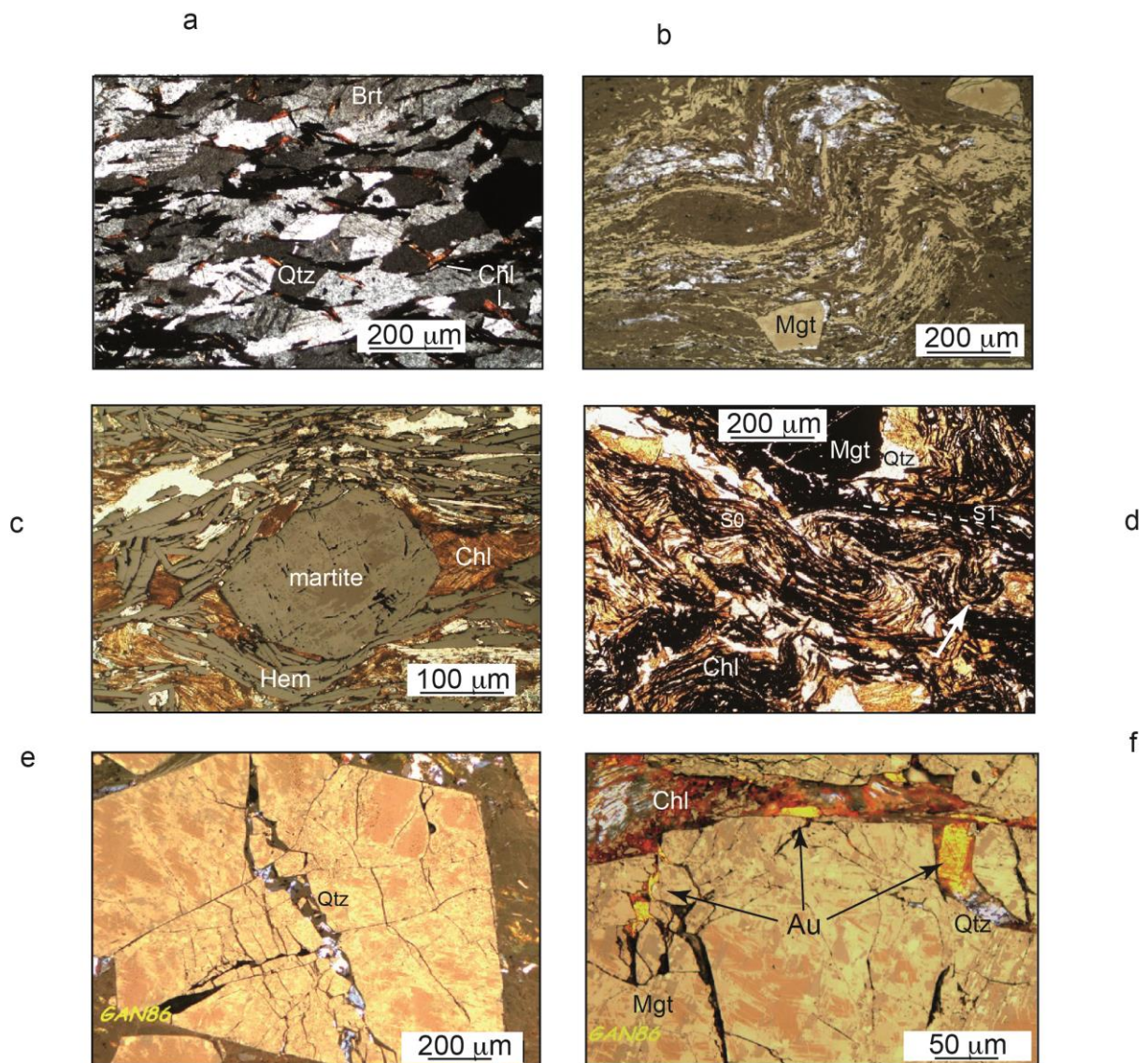


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50 715 Figure 16. Gold setting in late microstructures. **a.** Brecciated microcrystalline pyrite (pyrite 1)
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52 716 with quartz filling (SEM-BSE, Adassedakh); Gn: galena; Sp: sphalerite; Ccp: chalcopyrite;
53
54 717 AuAg: electrum. **b.** Detail of a: gold is associated with recrystallised pyrite in relation with the
55
56 718 brecciation. **c.** Native gold along actinolite (Act) cleavages (SEM-BSE, Yonim). **d.** Gold (Au)
57
58 719 in brittle fractures affecting magnetite (Mgt) and in re-activated S1 schistosity marked by
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720 chlorite (Chl) (Ganaet); Brt: barite; Q2: fibrous (syn-D2) quartz (“pressure-shadow” around the
 1
 2 721 magnetite). e. Gold (Au) in kinked chlorite (Chl) (D2 structure, Ganaet); note that gold is either
 3
 4 722 along cleavage or cross-cutting them, demonstrating a post-D2 introduction.
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10 724 (iv) Syn- G2 recrystallisation: apart from pyrite annealing, some interkinematic D2-D3
 11
 12 725 recrystallisation is observed, which may be tentatively attributed to the thermal effects of the
 13
 14 726 G2 suite intrusion. These include pyrite transformation into magnetite 2, clearly distinct from
 15
 16 727 the syn-G1 magnetite 1, the latter being consistently coated by syn-kinematic fibrous quartz),
 17
 18 728 chlorite recrystallisation (Fig. 15-b), and quartz healing.
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730 Figure 17. Massive barite and associated hematite-chlorite schists. **a.** Porphyroclastic barite
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2 731 (Brt) (Hamim) Qtz: quartz; Chl: chlorite. **b.** F2 microfold affecting the S1 hematite schistosity
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4 732 (Ganaet) Mgt: magnetite. **c.** Early magnetite (Mgt) wrapped in the S1 hematite (Hem)
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6
7 733 schistosity (Medadip 3); Chl: iron-stained chlorite. **d.** Early (pre-S1) microfolds (slumps?)
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9
10 734 (white arrow) (Hamim) S0: primary bedding; S1: main schistosity; Chl: iron-stained chlorite;
11
12 735 Mgt: early magnetite; Qtz: quartz in “pressure-shadow”. **e.** Late D3 microcracks and quartz
13
14 736 (Qtz) infillings in magnetite (Ganaet). **f.** Gold association with late D3 microcracks in magnetite
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16
17 737 (Mgt) (Ganaet).
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19 738
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22 739 (v) Syn-D3 microfracturation and hydrothermal recrystallisation: microcracks with
23
24 740 polymetallic mineralization crosscut all the preceding parageneses, including annealed pyrite
25
26 741 and magnetite 2 (Fig. 15-c). Pyrite is often dissolved, with polymetallic mineralization coating
27
28 742 the resulting micro-vugs (Fig. 15-d, e). Polymetallic mineralization includes pyrrhotite,
29
30 743 chalcopyrite, sphalerite, tennantite, arsenopyrite, galena, and altaite (PbTe), but not gold. Brittle
31
32 744 fracturing with galena-altaite sealing (Fig. 15-f) or fracturing and brecciation (hydraulic
33
34 745 fracturing), with mainly quartz cement (Fig. 16-a), appears as slightly later.
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39 746 (vi) Gold is always found as tiny particles of either native gold or electrum. It is most often
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41 747 associated with the latest syn-D3 manifestations (Fig. 16-a, b). More or less disseminated in the
42
43 748 mafic silicates (chlorite, actinolite), it is systematically intercalated along the cleavages, as, for
44
45 749 example, at Yonim (Fig. 16-c), and Ganaet (Fig. 16 d,e,f).
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49 750 (vii) Post-dating gold, the latest hydrothermal events are carbonate vein fillings (mainly,
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51 751 calcite), usually preceding supergene alteration, and likely related to the D4 fracture event.
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53 752 Within the massive sulphide bodies, supergene alteration mostly affects chalcopyrite, with
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55 753 covellite as the main product, and to a lesser extent, pyrite.
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755 **3.1.3. Synthesis**

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2 756 Quantitatively, most of the BMS presently seen in the Ariab VHMS deposits (mainly,
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4 757 sphalerite) are expressed in the D3 hydrothermal event. Yet, sphalerite and chalcopyrite were
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7 758 present since the early deposition of massive sulphides and were remobilised throughout the
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10 759 G1 intrusive and D1-D2 tectonic-metamorphic events. Therefore, it is likely that most of the
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12 760 zinc and copper, now present as late-D3 BMS, was remobilised from the primary stock. The
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14 761 case for As is more problematic since all the observed arsenopyrite and tennantite are part of
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16
17 762 the D3 paragenesis. There is no evidence, however, of significant As contents in the primary
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19 763 pyrites. The timing of Ag introduction in the Ariab massive sulphides (with 18 g.t⁻¹ on average)
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21
22 764 also remains difficult to explain, because tennantite is the only silver-hosting mineral in these
23
24 765 deposits.

25
26 766 On the other hand, gold is only seen in late-D3 micro-structures, as also it is the case for the
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29 767 galena-altaite association. Contrary to zinc and copper, there is no evidence of an early
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32 768 introduction of gold and lead. However, early galena would have been easily reworked in S1
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34 769 structure, as usually in metamorphosed Pb-bearing VHMS worldwide. Early gold, if present,
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36 770 could have been dissimulated into primary sulphides, such as pyrite. But it would have been
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39 771 liberated and reconcentrated through the D1-D2 tectonic-metamorphic events, or even more
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41 772 likely, during the strong remobilisation of base metals at the onset of the syn-D3 hydrothermal
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43
44 773 event. That it was not observed puts a strong argument against the hypothesis of gold
45
46 774 remobilization from a primary stock at the end of the D3 event (although this hypothesis may
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48 775 be realistic for silver).

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50
51 776 There thus seems to exist a severe decoupling between gold and the main BM, the latter being
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53 777 part of the VHMS depositing event. In contrast, gold, together with lead, altogether rare in the
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56 778 Ariab deposits, were introduced later in the system, as tellurium and possibly silver.

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780 **3.2. Gold-bearing massive barite and hematite schists**

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2 781 Several small barite lenses (about 10mx100m) are found close to the top of the C2 unit (Fig.
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4 782 4). Individual lenses contain no more than 150 to 300 kg Au, and the global resource in gold of
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7 783 the barite-hosted deposits is only 0.5 t @ 7.1 g.t⁻¹ Au. The barite lenses are usually topped by a
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9
10 784 thin layer of Mn-rich chert (1.9 to 3.1wt. % Mn) and black shales, possibly an exhalative
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12 785 horizon.

13
14 786 **3.2.1. General characters**

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17 787 *Description:* The barite bodies are characterised by a conspicuous banding, with alternating
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19 788 barite-quartz and hematite-chlorite-barite-quartz layers (10 to 20 cm thick).

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21
22 789 In the barite-quartz bands, barite is pre-kinematic, displaying a porphyroclastic texture (Fig.
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24 790 17-a), with more or less pronounced annealing, not unlike the primary pyrite in the VHMS
25
26 791 bodies. At the same time, quartz is remobilized as ribbon quartz.

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28
29 792 The iron-rich bands display a laminated structure, with alternating barite (porphyroclastic)-
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31 793 quartz and hematite-chlorite-quartz laminas (50 to 100 µm?). The latter, platy hematite and
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33
34 794 chlorite (a ripidolite with Mg # ~ 0.5) defines S0-1 (Fig. 17-b). There are preserved amygdales
35
36 795 showing a pre-kinematic quartz-magnetite-chlorite association, and some porphyroclastic
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38
39 796 hematised euhedral magnetites (martite) may be observed (Fig. 17-c). Locally, a pre-S1
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41 797 complex folding of the platy hematite ribbon structure is well expressed, as, for example, at
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43
44 798 Medadip 3 and Hamim (Fig. 17-d).

45
46 799 *Interpretation:* The quartz-chlorite-hematite-(magnetite) layers are likely to represent
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48
49 800 metamorphosed chert-iron hydroxide hydrothermal deposits on the sea-floor, with magnetite as
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51 801 an early diagenetic product. Early (pre-S1) folds may be interpreted as slumps structures. These
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53 802 observations suggest, in turn, that the barite-quartz layers could have been of exhalative origin
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55
56 803 as well.

804 As a whole, the barite-iron oxide lenses are reminiscent of the external zones of many kuroko-
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2 805 type VHMS deposits, marking the vanishing of the exhalative seafloor hydrothermal systems.
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4 806 **3.2.2. Gold setting**

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7 807 Gold is present in the barite lenses, with values up to 7.0 g.t⁻¹ (Ganaet) or even 8.0 g.t⁻¹
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9 808 (Hamim), generally associated with a few silver (up to 12 g.t⁻¹), likely as electrum. Bulk
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11 809 analyses demonstrate that gold is concentrated in the thick barite-quartz layers, the iron-rich
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13 810 layers being practically barren. Yet, likely due to polishing problems, gold was rarely seen in
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15 811 thin sections from barite-quartz layers, as small xenomorphic grains of native gold at barite
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17 812 boundaries. Inversely, gold was readily observed in thin sections from the hematite-chlorite
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19 813 schists, as tiny (10-15 µm) grains of native gold, either associated with late-D3 microcracks,
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21 814 that are particularly well expressed in the magnetite porphyroclasts (Fig. 15-d, 16-e, f, and 17-
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23 815 e,f), or dispersed in chlorite cleavage, clearly post-dating D1 and D2 deformation (Fig. 15-d,
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25 816 e).
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31 817 **3.2.3. Synthesis**

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34 818 Gold concentration in massive barite-quartz layers is of difficult interpretation. On the one
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36 819 hand, it could point to an early (syngenetic) enrichment in the exhalative barite. It is known
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38 820 from present-day barite-rich polymetallic VMS on the sea-floor that gold is effectively
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40 821 concentrated in Ba-rich environments (e.g., Herzig et al. 1993). Alternately, barite deposition
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42 822 could simply be the corollary of the H₂S-oxidation process, which by itself leads to gold
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44 823 precipitation (Ihle et al. 2005), in particular in barite lenses capping the massive sulphides
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46 824 (Huston et al. 1992). Thermodynamic calculations (McClenaghan et al., 2009) suggest that
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48 825 barite and gold co-precipitation is a distinct possibility. The systematically late-D3 introduction
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50 826 of gold in the iron-rich layers could, in this case, be interpreted as a remobilisation halo. Early
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52 827 gold enrichment can be expected in the other exhalative levels, such as the iron-rich layers and
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54 828 Mn-rich layer topping the barite horizon, with remobilisation beginning with the metamorphic
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829 episode. As such, a process is not observed, the inverse relationship may be suspected, i.e., that
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2 830 gold was introduced during D3 in the barite lenses, particularly the massive barite-quartz layers.

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7 832 **3.3. Gold-bearing quartz veins: the Kamoeb deposits**

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9 833 The Kamoeb deposits (Fig. 18) are the unique occurrence in the Ariab area of the quartz-gold

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12 834 lode-type, otherwise frequent in the Red Sea Hills (see Fig. 1). The Kamoeb deposits consist of

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15 835 a set of pre-kinematic (folded, boudinaged, and schistosed) quartz veins, 0.5 m to 5m thick, up

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17 836 to 3 km along strike (Fig. 18). Rocks mainly from the C unit host them, where they are mostly

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19 837 sub-concordant due to tectonic accordance. They are also present at the G1-C contacts, with

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21
22 838 high discordance, or even inside the G1 granitoids. Four veins display significant gold values

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24 839 (Fig. 18): KN (4.3 g.t⁻¹), KS1 and 2 (4 g.t⁻¹), and KE (6 g.t⁻¹), but KN, with about 1 t gold, is

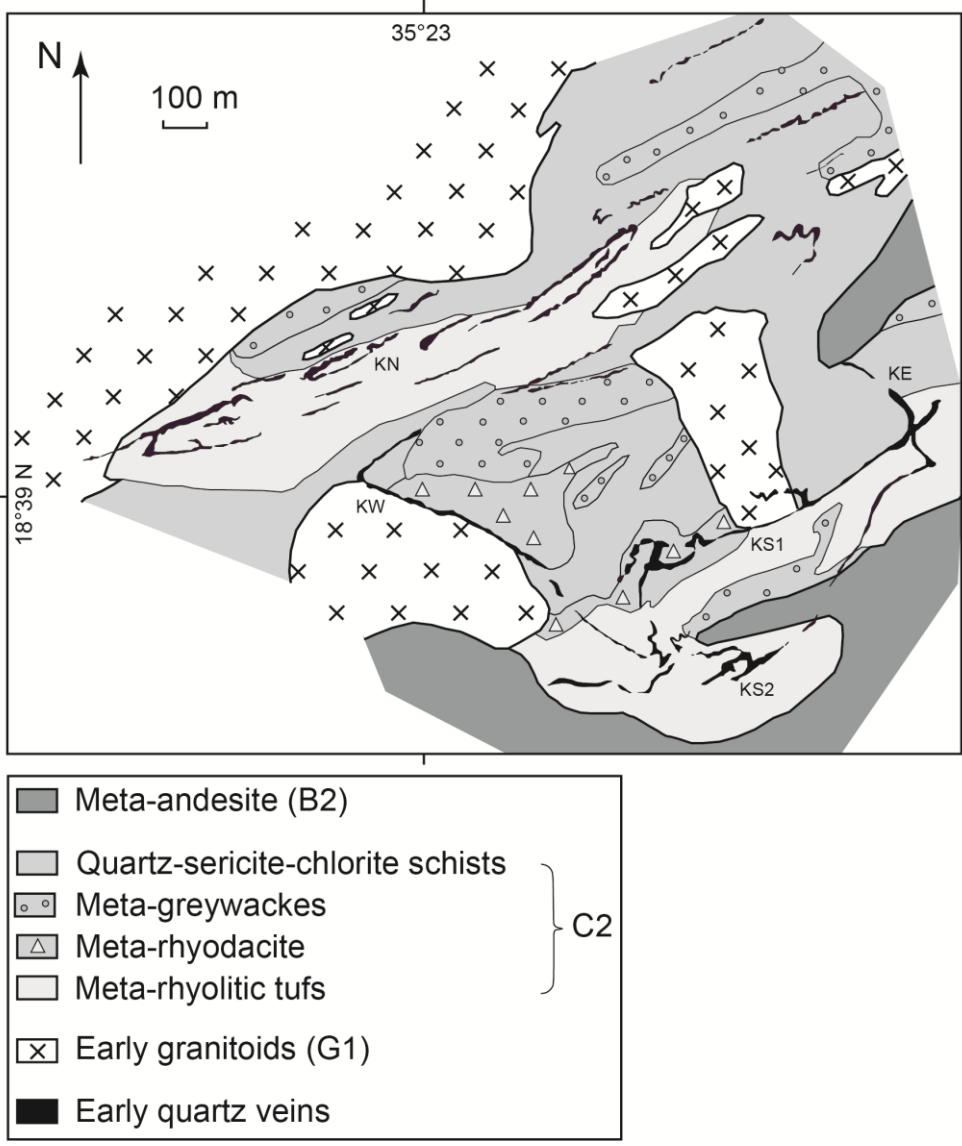
25
26 840 the only significant reserve.

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29 841 The vein system is associated with zoned alteration haloes, up to 10 m thick, consisting of

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32 842 proximal sericite (a Ba-phengite, up to 1.35 BaO wt.%) -pyrite (“bleaching”) zone and a distal

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34 843 Fe-carbonate-chlorite assemblage.

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844
845 Figure 18. Geological map of the Kamoeb area, displaying the Kamoeb quartz vein setting
846 (redrawn from Abu Fatima 2006, after Cortial et al., 1985). KE: Kamoeb east; KN: Kamoeb
847 north; KS1, KS2: Kamoeb south, vein 1 and 2; KW: Kamoeb west.

848
849 **3.3.1. Quartz microstructures**

850 At the microscopic scale, all the Kamoeb quartz display ductile deformation microstructures
851 testifying for D1 and D2 deformation stages, confirming that quartz deposition at Kamoeb was
852 a pre-kinematic event. Minerals from the alteration haloes, being equally affected by the D1
853 and D2 events, were synchronous of quartz deposition.

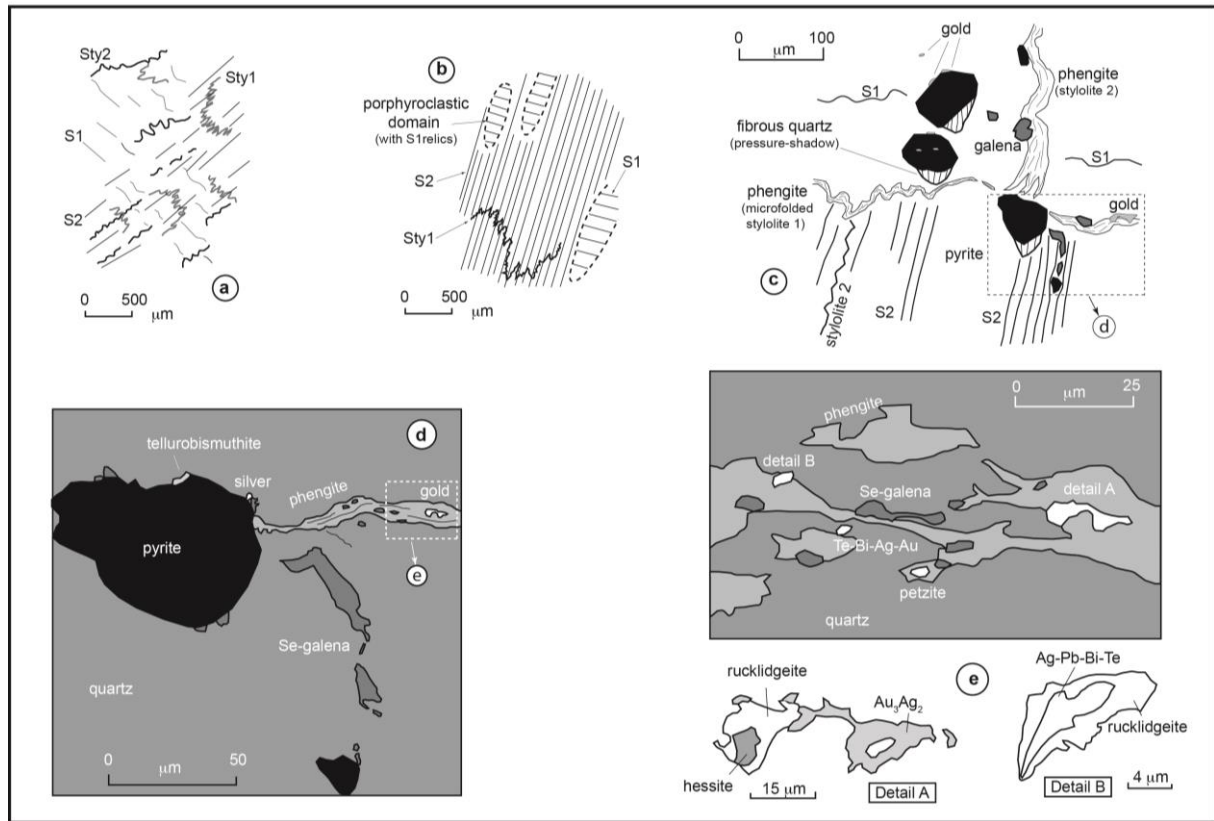


Figure 19. Microstructures in the deformed Kamoeb quartz and the gold and sulphides settings.

a. Existence of two successive flattening surfaces (S1, S2) materialised by microcrystalline quartz, and the associated stylolites (Sty 1, Sty 2) expressed as either indented grain contacts or (more often) as Ba-phengite accumulations. Stylolites 1 are micro-folded with S2 as axial planes. Sample KAM 9209-3A. **b.** Schematic relationships in mylonitic quartz (Sample KAM 9309-123), with penetrative S2 and relicts of the pre-existing D1 microstructures, as S1 preserved within porphyroclastic domains and intensely microfolded stylolite 1. **c.** Sulphide setting in deformed quartz (Sample KAM9209-3C), with D1 relicts (mainly early phengite stylolite 1) and penetrative D2 features (S2, phengite stylolite 2). Pyrite is syn-kinematic for D2 (quartz pressure-shadows), whereas galena and gold overprint D2 features. **d.** Detail of **c** (redrawn from SEM-BSE imaging). **e.** Aspect of **d** (redrawn from SEM-BSE imaging).

1 867 The D1 and D2 events are expressed as two sets of more or less orthogonal planar features: (i)
2 868 schistosities, marked by parallel bands of microcrystalline quartz (sometimes partly annealed),
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4 869 and (ii) stylolites, either as indented crystal boundaries or, more often, as accumulations of a
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7 870 Ba-phengite. Mutual relationships indicate that one set (S1, stylolite 1) is earlier than the other
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10 871 (S2, stylolite 2). In particular, early stylolites (Sty 1) are usually strongly microfolded with S2
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12 872 as an axial plane (F2 microfolds) (Fig. 19-a to c).

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14 873 Two other types of microstructures overprint the D1 and D2 features: (i) diffuse bands of
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17 874 recrystallised quartz with healed microcracks, still exhibiting a weak plastic deformation, and
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19 875 (ii) clear quartz veinlets corresponding to a macroscopic veining perceptible at the outcrop
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22 876 scale. These late features are interpreted as possible syn- to late- D3 events of hydrothermal
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24 877 circulation within the quartz veins.

25 26 878 **3.3.2. Mineralization**

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29 879 Quartz veins host pyrite, galena, native gold and electrum, and several tellurides. The
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32 880 crystallization sequence of these minerals may be evaluated with the relative chronology of the
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34 881 above-described quartz microstructures (Fig. 14).

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36 882 *Pyrite* is late-kinematic for D2 and occurs as sub-euhedral porphyroblasts (~ 100 μ m in size)
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39 883 overprinting most D1 to D2 plastic deformation features, including stylolites 2. It still exhibits
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41 884 “pressure-shadows” textures with fibrous quartz, consistent with the D2 shortening direction
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44 885 (Fig. 19-c).

45
46 886 *Galena* (a selenium-rich variety, up to 7,000 ppm Se) as small anhedral crystals (~ 10 to 50
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49 887 μ m) is typically later, apparently following sealed (D3?) microcracks in the quartz and
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51 888 crosscutting pyrite. It is mainly found in close association with both generations of phengite
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54 889 stylolites (Fig. 19-c to e).

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56 890 As tiny blebs a few μ m in size, gold and tellurides are distinctly coeval and, such as galena, are
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58
59 891 late in the system. They are intimately associated with the phengite stylolites or the quartz-

892 pyrite boundaries (Fig. 19-c to e). Gold is present either as native gold or electrum (Au_3Ag_2) or
893 gold-bearing tellurides. Based on SEM-EDS quantitative analyses, several tellurides were
894 found, including calaverite (AuTe_2), petzite (Ag_3AuTe_2), hessite (Ag_2Te), and altaite (PbTe),
895 but also tellurobismuthite (Bi_2Te_3), rucklidgeite ($\text{Bi, Pb}_3\text{Te}_4$) and volynskite (AgBiTe_2), without
896 any preferential distribution.

3.3.3. *Synthesis*

From the preceding, main conclusions are as follows:

(i) The Kamoeb quartz vein systems are spatially associated with the G1 granitoids.

They may be interpreted as peribatholithic quartz vein swarms reworked during the tectonic-thermal events affecting the Ariab area.

(ii) Gold, Te, Bi, Pb, and Se were introduced after the main D1-D2 events, likely at the late-kinematic D3 stage. Given the apparent absence of pre-kinematic gold-bearing association, it is difficult to envisage a hypothetical syn-quartz gold stock's late remobilisation.

(iii) As far as gold is considered, there is an evident parallel between the mineralization sequence at Kamoeb and in the VHMS (and barite) deposits. Gold is associated with Te, Bi, and Pb in the two types of deposits, and coeval galena is Se-enriched in both. These data reinforce the idea that gold was introduced during the late D3 stage in the Kamoeb quartz and emphasise the late introduction of silver in both types of deposits.

(iv) The dispersion and the variety of Au- and Ag-bearing minerals may be taken as evidence for a series of very local chemical equilibria, suggesting, in turn, low fluid-rock ratios at the time of gold deposition. The lack of gold-related quartz deposition also confirms it. Alternatively, the variety of Bi-Te minerals observed at Kamoeb could result from fractional crystallisation of a Bi-Te melt, and the gold enrichment is due to gold scavenging by this melt (Wagner 2007). Such melts do not exist, however, at low temperatures ($< 300^\circ\text{C}$). The systematic association of the gold, tellurides, and galena within phengite stylolites suggests that

1 917 the latter acted as preferential fluid pathways. However, a kind of topochemical effect cannot
2 918 be dismissed. The preferential association of gold- and Te-bearing minerals with the quartz-
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4 919 pyrite boundaries, on the other hand, is there strongly suggestive of a topochemical effect.
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9 921 **3.4. Gold mineralization close to the Nukasib shear zone: the Derbikwan showing**

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11 922 In the Derbikwan area, close to the Nukasib suture, swarms of poorly outcropping quartz veins
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14 923 hosted in carbonate-pyrite-rich mafic schists are spatially associated with a colluvial gold
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16 924 prospect. Although the quartz veins are barren, gold was worked in old times in discrete bodies
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19 925 (1-8m wide, 100-300 m long) striking parallel to the quartz veins on an overall distance of ~ 9
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22 926 km. The deposit is at present poorly known, with only a few exploratory trenches dug at the
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24 927 time of our field study (2007). The following developments are a preliminary statement on the
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26 928 nature and origin of this type of gold occurrence in the Ariab area.
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29 929 **3.4.1 Quartz veins**

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31 930 Exploratory trenches favoured field observations of the quartz vein system. Conjugated flat-
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34 931 lying fault veins with sheared schists at both the footwall and the hanging wall are associated
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36 932 with sub-horizontal tension gashes (0.3 to 1 metre wide), consistent with the D3 shortening
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39 933 direction. Thus, the quartz vein array may be interpreted as an inverse fault splay associated
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41 934 with the D3 reactivation of the early (D1) Nukasib shear zone.
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43 935 **3.4.2 Carbonate-pyrite-rich mafic schists**

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46 936 Unfortunately, no sample from the close vicinity of the quartz veins could be studied. Yet, in
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48 937 one exploratory trench, the significant gold content was found in sheared pyrite-schists at the
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50
51 938 quartz footwall. The quartz-carbonate-chlorite-sericite-pyrite schists appear as pre-D1 altered
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53 939 mafic rocks. The large (mm-sized) pyrite cubes, as well as chlorite-quartz-epidote assemblages
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55
56 940 replacing an unknown mafic mineral (clinopyroxene?), are all pre-kinematic, with spectacular
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58 941 “pressure-shadows” (quartz + Fe-carbonate) recording both the D1 and the D2 events.
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942 **Interpretation:** Although poorly constrained, the Derbikwan occurrence would appear as syn-
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2 943 D3 mineralization. Gold could have been conveyed through the quartz veins pathway and be
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4 944 trapped in the pyrite-rich surrounding schists. The noticeable absence of any sulphides inside
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7 945 the quartz veins would explain their barren character since no mineral trap such as pyrite could
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10 946 have favoured gold deposition in the quartz.

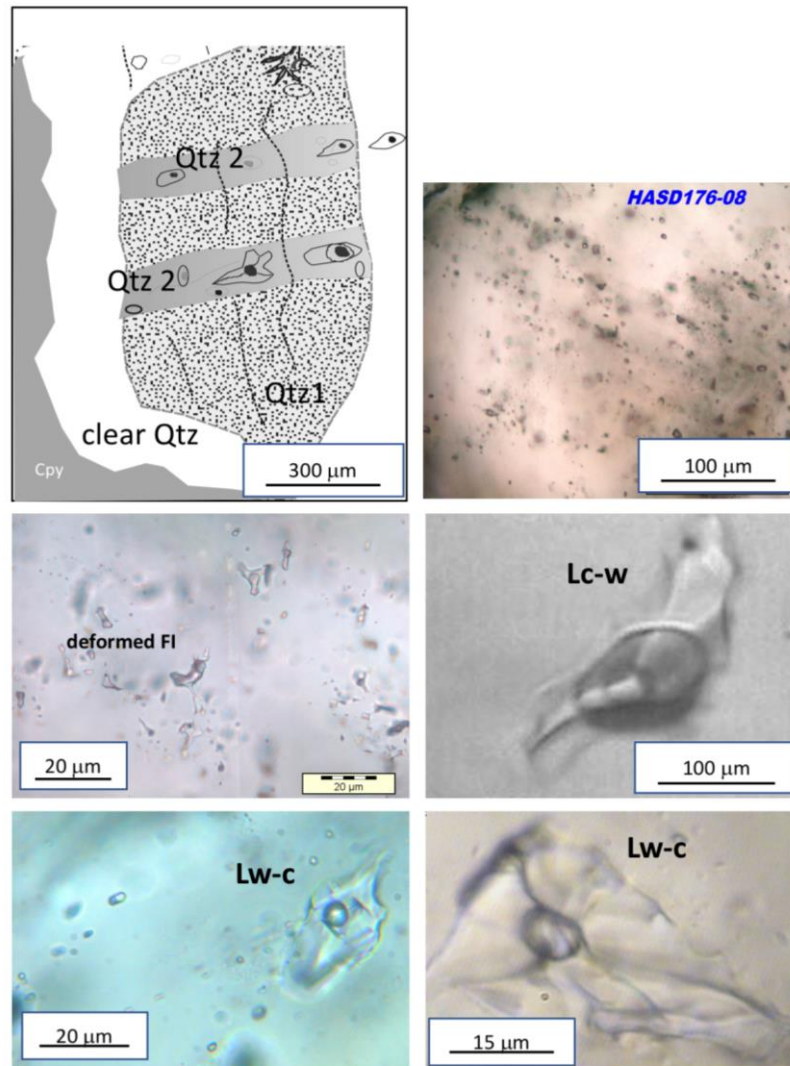
14 948 **3.5. P-T conditions of gold deposition in the VHMS and Kamoeb deposits: a fluid inclusion** 15 16 17 949 **study**

19 950 Early quartz in either the VHMS or the Kamoeb deposits mainly contains decrepitated fluid
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21 951 inclusions (FI), with stellate or cusped textures and peripheral clouds of tiny FI. Such features
22
23
24 952 suggest implosion during the tectonic-thermal events, as it could have been expected (Marignac
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26 953 et al. 2003; Marignac and Cathelineau 2006). It is only in late undeformed quartz (inter D2-D3
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28
29 954 Q₂₋₃ in the VHMS, and above all late-D3 Q₃ in both) that preserved FI may be studied. In the
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31 955 VHMS, these are often very small (< 5 µm) FI, only amenable to microthermometric
32
33
34 956 measurements using x800 magnification. The Kamoeb quartz display, as a rule, larger FI of
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36 957 more comfortable study.

39 958 **3.5.1 FI typology and petrography**

41 959 The FI nomenclature follows Boiron et al. (1992). Briefly, the fluid inclusion terminology is
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43 960 based on their final homogenisation mode as either liquid (noted L) or vapour (noted V).
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46 961 According to their components' nature, indicated by a subscript, c for CO₂-dominated volatile
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49 962 species, w for water. For example, 3-phase FI at room temperature that finally homogenises in
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51 963 the liquid phase is noted as Lc-w. In contrast, those inclusions homogenising to the liquid that
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53 964 remain 2-phase even at low temperature, but displaying volatile species (more commonly as
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56 965 clathrates) are noted Lw-c. The volumetric fraction of the volatile phase (fvp) was estimated by
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966 reference to the volumetric chart of Roedder (1979). FI assemblages correspond to fluid
 967 inclusions considered as following synchronous nomenclature of Bodnar (2003).



968
 969 Fig. 20: Main types of fluid inclusions in quartz from Kamoeb and VHMS. a: quartz textures
 970 and simplified types of occurrences of fluid inclusions. b: microphotograph showing fluid
 971 inclusion FI assemblages in Q₂₋₃, c: deformed fluid inclusions in Q₁, d: aqueous-carbonic
 972 three-phase Lc-w inclusion, e-f: Lw-c aqueous-carbonic two-phase fluid inclusions

973
 974 The following types were observed in the present study (Fig. 20):

975 Lc-w FIA, mostly seen in the Kamoeb samples, have a large size (20-30 µm) and an fvp
 976 ranging between 0.6 and 0.9. They are also frequent in D2-3 quartz from the VHMS, where

977 most of them are 3-phase FI at room temperature. They are mainly present as isolated 3-D
 978 clusters (i.e., possibly primary, or at least, early), elongated as microdomain Q₂₋₃ within Q₁.

979 Lw-c FIAs are absent in Kamoeb but present in Haddal Aouatib and Yonim. Lw-c FIs
 980 are 2-phases FI at room temperature and have a fvp ranging from 0.2 to 0.3. Based on a few
 981 intersecting relationships of difficult interpretation, they seem to post-date the Lc-w. They are
 982 mostly found as FIPs, and in the D3 VHMS quartz.

FI type	Tm CO ₂ (°C)	Tm ice (°C)	Tm clat (°C)	Th CO ₂ (°C) (L)	Th (°C)
Lc-w					
Kamoeb (n=66)	-58.3 to -56.8	-5.6 to -2.2	2.6 to 7.9	16.3 to 26.0	220 to 310
VHMS (n=118)	-59.7 to -57.2	-7.6 to -2.5	1.8 to 8.3	22.8 to 29.7	237 to 278
Lw-c					
VHMS (n=10)		-7.1 to -2.2	1.9 to 8.0		260 to 355

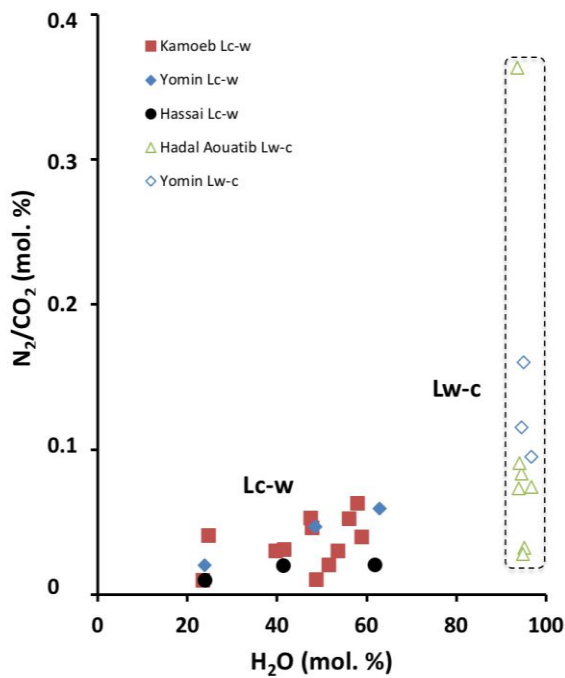
984
 985
 986 Table 2. Summary of the microthermometric measurements in the Kamoeb and VHMS (Yonim,
 987 Hadal Ouatib, and Hassai) quartz. Tm CO₂: melting temperature of the solid CO₂ Tm ice: final
 988 melting temperature of ice. Tm clat: final melting temperature of clathrates. Th CO₂:
 989 homogenisation temperature of the volatile phase. Th: bulk homogenisation temperature. L:
 990 homogenisation in the liquid phase, respectively.

991 3.5.2. Microthermometry and Raman results

992
 993 A summary of the microthermometric results is given in Table 2. In the Lc-w FI, melting
 994 temperatures of CO₂ range from -59.7° to to 56.8°C. They are thus lower than for pure CO₂ (-
 995 56.6°C), indicating the presence of other volatiles. Raman microanalysis (Table 3) confirms the
 996 presence of small amounts of N₂, and the significant absence of a methane component. Th CO₂
 997 ranges from 16 to 30 °C in the liquid phase.. Tm ice is ranging from -7.6 to -2.2°C and Tm clat

998 is observed between 1.8 and 8.3 °C. Bulk homogenisation occurs on a wide range of
 1
 2 999 temperatures, mostly between 220° and 310°C.

3
 4
 5 1000 In Lw-c FI, Tm clat attesting the presence of gas species is ranging from 1.9 to 8.0°C. Tm ice
 6
 7 1001 is measured between -7.1 to -2.2°C. Bulk homogenization occurs between 260 and 355°C.
 8
 9 1002 The nitrogen concentration, mainly in the range 10 to 35 mol.% in the Lw-c FI, is higher than
 10
 11
 12 1003 in the Lc-w FI. (1 to 6 mol.%).



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 39 1004
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 41 1005 Figure 21. Calculated bulk compositions (in molar fractions) of representative FI s from the
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 43 1006 quartz at Kamoeb and several VHMS (Yonim, Haddal Aouatib, Hassai) in an N₂/CO₂-H₂O
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 45
 46 1007 diagram.

50 51 1009 3.5.3. Bulk compositions

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 53 1010 Calculated bulk compositions (Table 3) are reported in the N₂/CO₂ versus H₂O diagram (Fig.
 54
 55 1011 21). The same volatile-rich fluids (Lc-w) are present in both the Kamoeb and the VHMS quartz.
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 57
 58 1012 In WHMS quartz, a second fluid stage is recorded as Lw-c FIs which are characterised by (i) a
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 60 1013 water content higher than 94 mol.%, (ii) an increase of the N₂ /CO₂ ratio compared to Lc-w FI,
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1014 N₂ being always a minor component, the volatile fraction remaining CO₂-dominated, with no
 1
 21015 contribution of any CH₄ component.
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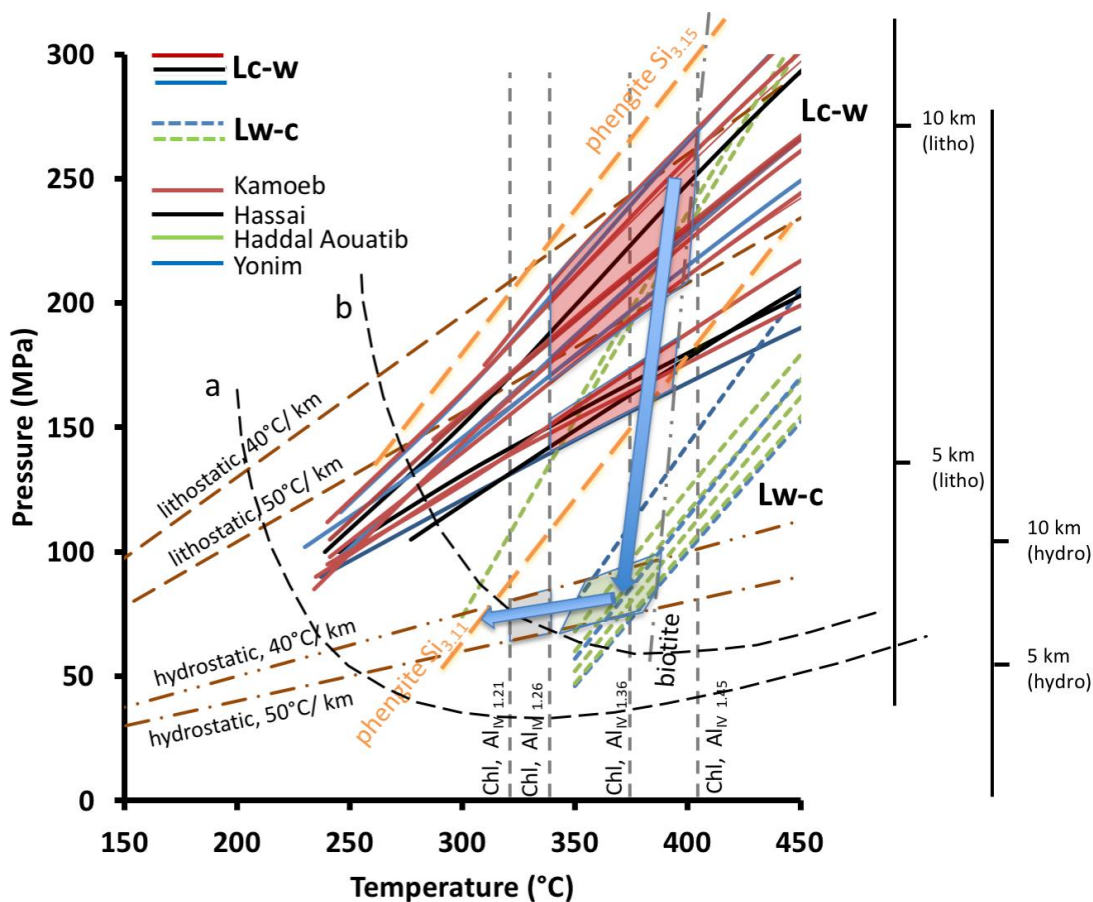
	Microthermometry (°C)					Raman data (mol.%)			Bulk composition (mol. %)			
	Tm CO ₂	Tm ice	Tm clat	Th CO ₂ (L)	Th	fvp (%)	ZCO ₂	ZN ₂	XH ₂ O	XCO ₂	XN ₂	XNaCl
Lc-w												
Kamoeb												
KAME02-11	-57.1	-4.2	7.6	23.2	241.5	75	97.0	3.0	48.8	49.3	0.5	1.4
KAME02-12	-57.3	-4.3	3.3	20.1	248.3	65	95.0	5.0	56.2	38.5	2.0	3.3
KAME02-13	-57.0	-5.3	2.6	23.6	234.1	70	98.0	2.0	51.7	44.3	0.9	3.1
KAME02-14	-58.1	-2.2	4.8	21.8	190*	80	97.0	3.0	39.7	56.8	1.7	1.8
KAME02-16	-57.1	-3.4	7.1	26.0	240.3	80	98.0	2.0	41.7	54.8	1.7	1.8
KAME02-18	-57.0	-3.0	7.1	23.6	235.7	90	96.0	4.0	24.7	71.6	2.9	0.8
KAME02-110	-57.2	-4.1	6.3	22.6	220.5	90	99.0	1.0	23.5	74.9	0.7	0.9
KAME02-111	-57.0	-5.1	7.2	20.6	241.3	75	95.5	4.5	47.9	48.3	2.2	1.6
KAME02-114	-57.8	-4.2	4.5	22.2	240.4	65	94.0	6.0	58.0	36.7	2.3	3.0
KAME02-23	-57.2	-5.6	7.5	19.8	310.1	65	96.0	4.0	58.9	37.8	1.5	1.8
KAME02-25	-57.0	-4.4	5.5	22.2	287.6	75	95.0	5.0	47.6	47.8	2.5	2.1
KAME02-26	-56.8	-3.1	7.9	22.1	281.9	70	97.0	3.0	53.6	43.7	1.3	1.4
Yonim												
YONM-202-NB1	-56.9	-2.5	4.3	25.4	237.1	90	98.0	2.0	23.9	73.5	1.5	1.1
YONM-202-NB2	-57.2	-4.7	6.4	25.4	230.3	70	96.5	3.5	48.5	47.3	2.2	2.0
YONM-202-NB3	-57.2	-5.5	5.8	25.4	243.2	75	95.5	4.5	48.5	47.3	2.2	2.0
YONM-202-NB4	-58.3		4.4	23.6	246.2	60	94.0	6.0	62.8	32.0	1.9	3.3
Hassai												
HASD176-081	-57.3	-3.6	4.2	23.3	239.5	60	98.0	2.0	62.0	34.3	0.7	3.0
HASD176-084	-56.8		7.7	22.8	245.3	90	99.0	1.0	24.0	74.7	0.7	0.6
HASD176-088	-57.3	-2.7	7.6	29.7	277.7	80	98.0	2.0	41.5	55.4	1.1	2.0
Lw-c												
Hadal Ouatib												
HDND77-111		-5.1	4.2		285*	30	89.0	11.0	95.2	3.1	0.1	1.6
HDND77-112		-5.0	8.0		341.9	20	89.5	10.5	94.9	3.6	0.1	1.4
HDND77-113		-5.6	7.6		260.5	30	86.0	14.0	94.0	4.1	0.3	1.6
HDND77-114		-5.1	6.5		354.7	30	83.0	17.0	94.6	3.6	0.3	1.5
HDND77-117		-2.3	4.3		275.7	30	84.5	15.5	96.7	2.7	0.2	0.4
HDND77-118		-7.1	6.3		278.3	20	80.0	20.0	94.1	3.3	0.3	2.3
HDND77-119		-2.2	6.6		345.9	20	50.0	50.0	93.7	3.3	1.2	1.8
Yonim												
YONM 202-N2-1	-56.8	-7.1			250 *	30	65.0	35.0	95.1	2.5	0.4	2.0
YONM 202-N2-2	-56.6	-3.6	1.9		240 *	35	71.5	28.5	94.6	2.6	0.3	2.5
YONM 202-N2-3	-56.6		1.7		220*	25	70.5	29.5	96.7	2.1	0.2	1.0

1017
 1018 Table 3. Bulk compositions from Raman microanalysis of selected FI from both the Kamoeb
 1019 and the VHMS (Yonim, Haddal Aouatib, and Hassai) quartz. Tm CO₂: melting temperature of
 1020 the solid “CO₂”. Tm ice: final melting temperature of ice. Tm clat: final melting temperature of
 1021 clathrates. Th CO₂(L): homogenisation temperature of the volatile phase in the liquid state. Th:
 1022 bulk homogenisation temperature, *: decrepitated under heating.
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 1024
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1026 **3.5.4. P-T-t evolution**

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21027 Isochores were calculated for a set of FI of known bulk compositions, and are drawn in Fig.
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4
51028 22. They record a pressure drop from two series of Lc-w, and then the Lw-c trapping conditions
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71029 with two sets of isochores.
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9
101030 The absence of biotite in the regional package of minerals means that regional temperature was
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121031 never higher than ca. 420°C at ca. 350 MPa (e.g., Yardley 1989). More precise temperature and
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141032 pressure estimations may be achieved using the Si-in-Muscovite geo-barometer of Massonne
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16
171033 and Schreyer (1987) (assuming that Ba substitution to K does not significantly alter the
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191034 equilibrium) and the Al-in-Chlorite geothermometer of Cathelineau (1988). The latter is known
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21
221035 to yield relevant results, given that the chlorite grew in an Al-saturated medium, and no
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241036 interlayered chlorite-other phyllite phase is used (Frimmel 1997, in Gartz and Frimmel 1999),
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271037 two conditions respected in the present study. However, the problem is to find the relevant
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291038 compositions, since the Ariab-Arbaat belt experienced a multistage thermal history, starting
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311039 from the hydrothermal stage of metamorphism.
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33
341040 An electron microprobe survey of chlorite and phengite compositions in the belt revealed a
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361041 clear bi-modal distribution of Al_{iv} p.f.u (in chlorite) or Si p.f.u (in phengite), with Al_{iv}-in-
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38
391042 chlorite usually comprised between either 1.60 and 1.36 (most of them around 1.45) or 1.26
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411043 and 1.21 (with exceptional values as low as 1.07), and Si-in-phengite clustering either between
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43
441044 3.19 and 3.24 (up to 3.30) or 3.11 and 3.15 (exceptionally as low as 3.07). Two observations
45
461045 allow an interpretation of these distributions. First, kinked syn-S1 or syn-S2 phengites (i.e.,
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491046 likely un-recrystallized phengites having escaped later re-equilibration) display Si p.f.u. typical
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511047 of the high-pressure range of compositions, whereas late overgrowths on early phengites yield
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531048 Si p.f.u. in the lower (low P) range of compositions. Second, in a Barut sample, an early quartz-
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561049 chlorite-phengite hydrothermal association preserved in an undeformed vug provided the
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581050 higher Al_{iv}-in-chlorite found in our survey. In contrast, a late (possibly, syn-D3) chlorite
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1051 overprinting fibrous syn-kinematic D2 quartz yielded $Al_{iv} = 1.26$, in the low (low-T) range of
 1052 compositions. The highest temperatures recorded by the Al-in-chlorite geo-thermometer of
 1053 Cathelineau (1988) and the highest pressures recorded by the Si-in-phengite geo-barometer of
 1054 Massonne and Schreyer (1987) provide a maximal P-T estimate. The latter is likely those of the
 1055 first tectonic-metamorphic events (D1 and D2). The lowest values record more or less
 1056 overprinting during a second thermal event, during D3 or later. When reported in the P-T
 1057 diagram in Fig. 22, the two P-T pairs appear to be entirely consistent with the isochore sets,
 1058 allowing the construction of a P-T-t path, shown in Fig. 20, characterised by a sub-isothermal
 1059 pressure decrease from lithostatic to hydrostatic pressures. The isochores of a few FI from
 1060 Haddal Ouatib are shifted to lower temperatures. However, it is difficult to know if they are
 1061 trapped under litho, infra-litho or hydrostatic pressures.



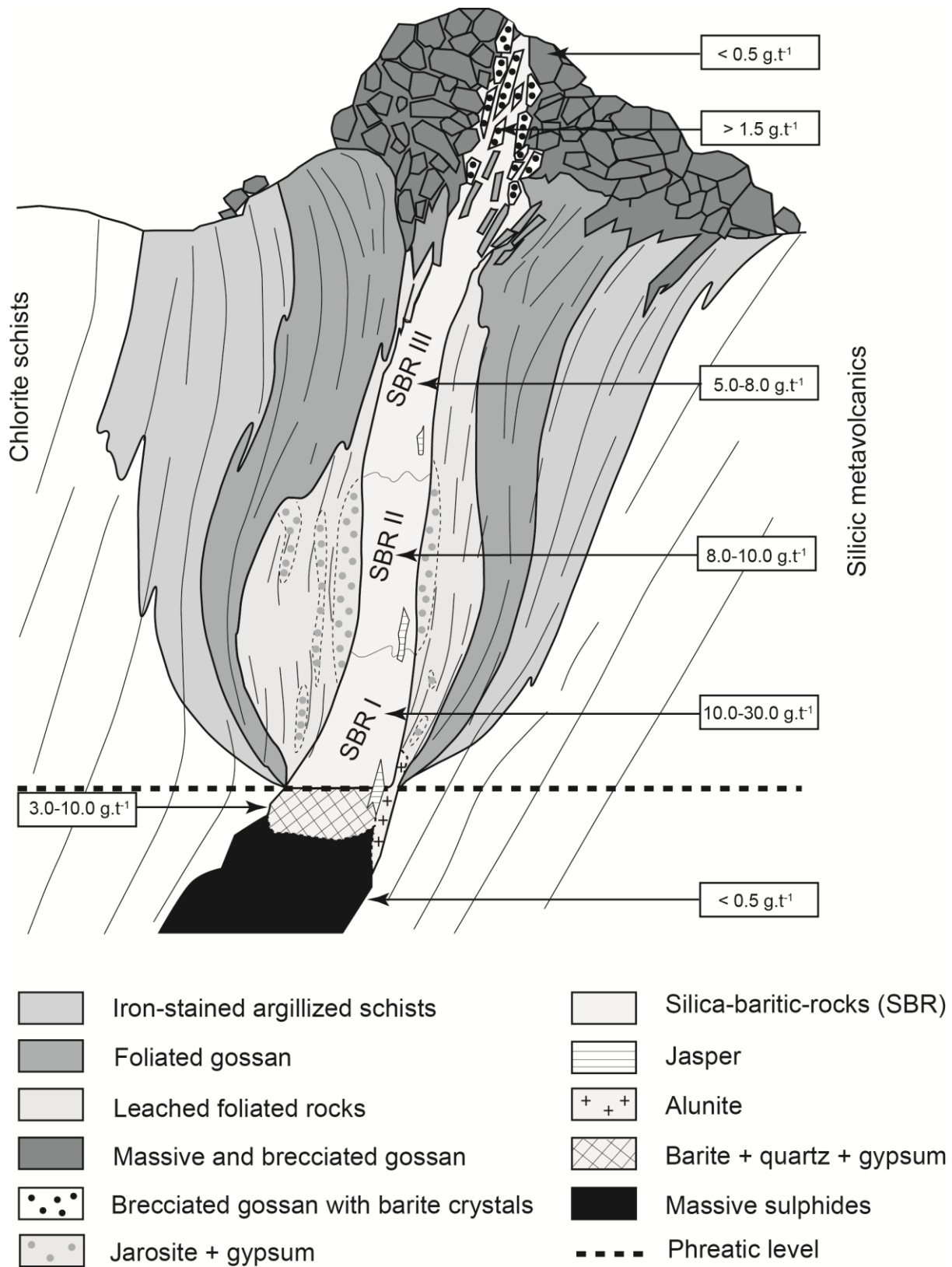
1063 Figure 22. Reconstruction of the late (D3?) P-T-t evolution as recorded by fluid inclusion
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21064 densities (isochores) from three representative areas (Yonim and Haddad Aouatib VHMS;
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51065 Kamoeb quartz veins). The biotite-in isograd is from Yardley (1989, Fig. 3-11, p. 86). The Si-
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71066 in-phengite geo-barometer isopleths (Massonne and Schreyer, 1987) and the Al-in-chlorite
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9
101067 geo-thermometer isopleths (Cathelineau, 1988) have been drawn for relevant compositions.
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121068 Isopleths (dotted dark blue lines) in the system H₂O-CO₂-NaCl are from Weisbrod (1984) for
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141069 a salinity of 6 wt. % NaCl equiv., and a CO₂ content of 10 (a) and 20 (b) mole%. See text for
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16
171070 explanation.
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191071
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221072 The scenario for gold-related fluid circulation in the Ariab area implies the participation of two
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241073 types of fluids. First, circulation of hot (350-400°C) volatile-rich fluids at around 260-200 MPa
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271074 (10 to 8 km depth if lithostatic). Second, the water-dominated fluids (Lw-c FI), enriched in N₂
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291075 occur after a pressure drop from c. 200 MPa to c. 100 MPa at sub-constant temperatures (380-
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31
321076 350°C), This late-stage may correspond to the stage of Bi-Te melts, as discussed in the
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341077 preceding section.
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361078 This evolution finds similarities with the Late Carboniferous gold-forming fluid evolution
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391079 reconstructed in the West Variscan belt (Boiron et al. .1996, 2001, 2003, Vallance et al., 2003)
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411080 and the Limousin area (in the French Massif Central) in particular (Boiron et al. 1990; Essarraj
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43
441081 et al. 2001; Vallance et al., 2004). A similar evolution was also reconstructed for the late-
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461082 metamorphic (Visean) gold introduction in the VHMS deposits of the Iberian Pyrite Belt
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491083 (Marignac et al., 2003, Marignac et Cathelineau, 2006). In the West Variscan Belt, the early
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511084 gold introduction is interpreted to be deposited by similar late aqueous-carbonic fluids. The
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541085 mixing with a “cold” low-salinity fluid is by contrast not recorded at Ariab. Both in Variscan
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561086 terranes, Lc-w, and Lw-c fluids are likely to be of “metamorphic” origin, i.e., equilibrated with

1087 metamorphic rocks (Boiron et al., 2003). In contrast, the latest water dominated fluids would
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21088 record the progressive incoming of more surficial fluids evolving during their infiltration.

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71090 **3.6 Supergene gold enrichment in the VHMS and barite deposits: gossans and SBR**
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101091 **deposits**

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121092 More than 20 gold-bearing gossans derived from the oxidation of the VHMS deposits, with the
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141093 formation of silica-barite-rich rocks (SBR) hosting gold (and silver) concentrations, were
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171094 discovered in the Ariab-Arbaat area (Cottard et al. 1986a), with a cumulative tonnage of 68 t
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191095 gold averaging 9.7 g.t^{-1} (from 4.2 to 17.1 g.t^{-1}) (Abu Fatima 2006). The main deposits are Hadab
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21
221096 Ouatib W ($16.2 \text{ t gold at } 15.1 \text{ g.t}^{-1}$), Oderuk N ($7 \text{ t gold at } 5.0 \text{ g.t}^{-1}$), and Hademayet-Wali N
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241097 ($6.7 \text{ t gold at } 6.2 \text{ g.t}^{-1}$).

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261098 The mineralised gossans are part of the regional weathering developed in the Red Sea Hills
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291099 (Cottard et al. 1993). Usually present on a depth no more than 10-20 m, this alteration was
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311100 enhanced by the intense deformation affecting the VHMS host-rocks and the sulphides' strong
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341101 reactivity (mainly, the pyrite) themselves. As a consequence, extensive weathering at the origin
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361102 of mineralized gossans could penetrate down to 90-100 m depth, being best developed onto
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391103 tectonically verticalised VHMS bodies.



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1105 Figure 23 Cross-section model of a typical Au-bearing gossan developed at the expense of a
1106 VHMS deposit from the Red Sea Hills (redrawn from Abu Fatima 2006, after Recoché 1989).

1107 The gold-bearing gossans structure (in Fig. 23) is mainly based on the results from Hottin and
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21108 Laforet (1986), and Recoché (1989). The different weathering facies form broadly concentric
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4
51109 envelopes, with from outside towards the inside (i) iron-stained clays (mostly kaolinite) and
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71110 iron-oxide impregnation of the schistose host rocks (“foliated gossan”), (ii) leached (bleached)
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101111 schistose host rocks, with patches of jarosite-rich facies, and (iii) the core of silica-barite rocks
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121112 (SBR), that concentrate gold. The jarosite-rich rocks are mainly present as a 5-20 m thick halo
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151113 surrounding the SBR in the lower part of the profile. Jarosite is the main component associated
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171114 with gypsum, alunite, and Fe- and Cu-sulphates. The SBR is a heterogeneous silica-rich micro-
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191115 brecciated rock, with chalcedonic or microcrystalline clastic quartz and a variable matrix
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21
221116 always including barite, the nature of which defines three successive facies from bottom to top:
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241117 (i) white sugary barite-rich SBR I, highly siliceous (amorphous silica and chalcedony) and
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26
271118 containing the highest gold grades, (ii) yellowish-green SBR II, jarosite-rich, and (iii) reddish
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291119 ferruginous SBR III, formed of micro-brecciated microcrystalline quartz and a matrix of
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311120 goethite-hematite and barite (up to 15-20% in volume).
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341121 The base of the SBR defines a redox boundary, with complete leaching of the underlying
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361122 sulphides, pyrite being currently transformed into jarosite, and transformation of the phyllites
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391123 into silica. The underlying VHMS exhibits a cementation zone's typical mineralogy, with
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411124 covellite, digenite, and bornite replacing primary copper minerals, anglesite formed after from
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441125 galena, and barite from the dissolved Ba-phengites. Pyrite is partly dissolved, and sphalerite,
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461126 tennantite, and tellurides are absent. The surrounding rocks, where phyllites are replaced by
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491127 microcrystalline silica, are characterized by the association alunite-oxidised copper minerals
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511128 (native copper, tenorite, and the rare delafossite CuFeO_2). Gold is often concentrated in the
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53
541129 SBR as “invisible” gold, i.e., tiny gold particles (4-20 μm). It is associated with Ag, Pb, Ba, As
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561130 (up to 1240 ppm), Te (up to 43 ppm), Se and Sb, whereas the mobile elements (Zn, Cu, Mg,
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591131 Al, Ca) are leached, being trapped in the jarosite- or alunite-rich haloes.
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1132 At the surface, the three zones defining the oxidised body are prolonged into massive (the zones
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21133 i and ii) and brecciated (the SBR III) gossans. Goethite and hematite are the main minerals of
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4
51134 the massive gossan. Brecciated gossan is a collapsed facies of the massive gossan, with a barite
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71135 matrix.
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101136 According to Recoché (1989), the mineralised gossan-SBR bodies of the Ariab-Arbaat belt
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121137 result from a multi-stage weathering history, recording the climate fluctuations of the Late
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141138 Cenozoic in the Arabian-Nubian Shield area. The upper part of the oxidised profile is
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171139 interpreted as the bottom remnant of an older profile, developed under the Late Pliocene-Early
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191140 Pleistocene's hot and wet conditions (1.1 Ma). The still-active lower part (with the VHMS-BSR
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221141 boundary coinciding with the present-day phreatic level in Fig. 23) records the more arid
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241142 conditions from the Quaternary until now. The collapse in the upper levels (brecciated gossan)
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271143 results from the substantial volume diminution corresponding to sulphide dissolution.
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311145 **4. DISCUSSION AND CONCLUSIONS**

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361147 **4.1 Base metals in an intra-oceanic arc, gold in an accreted terrane**

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391148 in the Ariab-Arbaat belt of the Haya terrane, the main BM and gold mineralising events'
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411149 geodynamic settings are quite different.
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43
441150 The BM-hosting VHMS have formed at ca. 900 Ma in an intra-oceanic arc setting, and may,
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461151 therefore, be classified as Cu-Zn deposits of the Uralian type (Prokin and Buslaev 1999) – the
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48
491152 “bi-modal mafic setting” class of Franklin et al. (2005). Although massive sulphides are present
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511153 at different levels in the lithologic pile, most VHMS of the Ariab-Arbaat belt are located in the
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531154 same felsic unit (C2), at the top of a thick andesite-dominated pile. The VHMS are associated
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561155 with barite-hematite-chert lenses that are their lateral equivalent, and a wrapping exhalative
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581156 horizon appears to close the hydrothermal process. Intense subsequent deformation precludes
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1157 a precise identification of the local tectonic environment of the VHMS. Still, the abundance of
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21158 felsic domes in the C2 unit and the existence of a plutonic complex of similar geochemistry
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41159 (G1), at depth, is suggestive of a caldera environment, as is indeed the case for many
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71160 polymetallic VHMS of the Uralian or the similar Abitibi-Noranda class (Herrington et al. 2005,
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101161 Franklin et al. 2005).

11162 By contrast, all the primary gold occurrences of the Ariab-Arbaat belt, whatever their style
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141163 (VHMS-, quartz lode- or shear-zone-hosted), appear to have been syn-to late-D3, i.e., formed
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171164 in a mature continental arc, already accreted to the northern Gebeit terrane, in the context of the
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191165 final Pan-African collision leading to the Arabian-Nubian Shield (poorly constrained at ca. 620-
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21
221166 ca. 560 Ma). Gold deposition occurred at the final stage of the collision, in a thickened crust
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241167 suffering rapid uplift (from ca. 17 km to ca. 7.5 km) and intense fluid circulation, mixing deep
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271168 hot (~ 350°C) “metamorphic” and shallow cold (< 300°C) “meteoric” fluids.

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311170 **4.2 Gold (re)-mobilisation or gold-(re)introduction?**

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341171 By contrast with recent sea-floor deposits, gold concentration in fossil metamorphosed
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361172 polymetallic VHMS ores is often described as native gold or electrum deposited in late syn- to
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391173 post-metamorphic tectonic traps. The latter are usually sets of micro-cracks crosscutting earlier
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411174 pyrite or arsenopyrite. There, gold is commonly associated with either galena, bismuth
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441175 tellurides, or both, whatever the deposit's age. Recently described examples are in the Archean
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461176 (Hongtoushan, China: Gu et al. 2005), the Paleoproterozoic (Boliden, Sweden: Walter et al.
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48
491177 2007) or the Phanerozoic (Urals: Vikentyev 2005; New Brunswick: Thome et al. 2007; Iberian
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511178 Pyrite Belt: Marignac et al. 2003).

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531179 The most probable host of the early gold pre-concentration is usually considered to be the early
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55
561180 pyrite. In such a model, the tectonic-thermal events only act to free and upgrade earlier
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581181 “invisible” gold incorporated in the hydrothermal sulphides from the VHMS (e.g., Huston et
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1182 al. 1992). Modern polymetallic VHMS from the SW Pacific, with a mean gold content of 3.7
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21183 g.t⁻¹ (Herzig et al. 1993), are gold-enriched by comparison with most fossil equivalents. Gold
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41184 concentrations range between 0.6 and 1.4 g.t⁻¹ for all types (Franklin et al. 2005) as, the 0.6 to
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6
71185 1.7 g.t⁻¹ range in the New Brunswick district, with a mean of 0.66 g.t⁻¹ (Thome et al. 2007,
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101186 McClenaghan et al. 2009) or the 0.6 to 1.2 g.t⁻¹ range of the present study. Yet, according to
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121187 Herzig et al. (1993), the most promising geodynamic environments for gold enrichment in
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141188 VHMS are immature back-arcs or arcs, a setting fitting well with the Ariab-Arbaat occurrences.
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16
171189 At our knowledge, there are only very few studies addressing the actual gold content of early
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191190 hydrothermal phases in the VHMS and monitoring gold behaviour during the subsequent
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221191 events. In the Archean Moberly deposit, Larocque et al. (1993) report ion microprobe data on
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241192 primary (syngenetic) and secondary (metamorphic recrystallization). They show that the
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271193 secondary pyrites are systematically depleted to a 60% reduction of the initial content than the
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291194 first pyrite, which contains 0.4 to 10 ppm “invisible” gold as micro-inclusions < 0.2 μm. In the
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321195 Paleoproterozoic Boliden deposits (Skellefte belt), Walter et al. (2007), using the LA-ICP-MS
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341196 technique, clearly demonstrated the early presence of sulphide-hosted gold and its strong
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361197 remobilisation/exportation in syn-metamorphic veins. Yet, these deposits are exceptionally
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391198 arsenopyrite-rich, and it is indeed early arsenopyrite that appears to be the main gold repository,
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411199 the dominant early pyrite being poor in gold (< 0.2 ppm). In the Tharsis deposit (Iberian Pyrite
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441200 Belt of Tournaisian age), Conde et al. (2009), equally using the LA-ICP-MS technique, found
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461201 gold and correlated arsenic contents up to 8 ppm in the early framboidal or colloform pyrites.
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48
491202 In contrast, syn-kinematically recrystallised euhedral pyrites are strongly depleted. However,
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511203 in the same deposit, a detailed study of gold relationships with (micro)tectonic events could not
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541204 find any evidence for a syn-kinematic gold deposition, as all visible native gold was deposited
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561205 in late-kinematic features and associated with late- to post-metamorphic fluids (Marignac et al.
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581206 2003). In the Bathurst N°12 deposit (New Brunswick), McClenaghan et al. (2009) document
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1207 “invisible gold” mainly located in pyrite and arsenopyrite. Pyrite exhibits relict cores that
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21208 concentrate gold (up to 21 ppm, with a mean of 2.7 ppm), whereas syn-metamorphic
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41209 overgrowths are systematically depleted (down to 0.02 ppm, with a mean of 1.9 ppm).
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6
71210 Arsenopyrite, a syn-metamorphic phase overprinting the pyrite, is equally poorer in gold, with
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101211 a 1 ppm mean, and often relatively lower values, and is always “ (...) depleted in gold relative
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121212 to relict inclusions of pyrite” (p. 545). Whereas McCleneghan et al. (2009) think that at Bathurst
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141213 N°12 “(...) redistribution of gold was limited”(p. 552) during the metamorphism, there is room
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171214 to estimate, from their data, that indeed gold was extracted from the VHMS. At least, late
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191215 metamorphic up-grading is excluded in that case.

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221216 Thus, two arguments favour gold export during the recrystallisation of massive sulphides
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241217 during metamorphic events; i) the limited evidence for strictly syn-metamorphic gold
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261218 remobilisation, and ii) the mere fact of the fossil VHMS deposits are relatively gold-poor when
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291219 compared with their modern equivalents. The observed gold concentrations in late brittle
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311220 structures within the same VHMS bodies (and notably in their quartz-rich feeder stockworks,
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341221 that are prone to late fracturing and fluid circulation) may then be taken as a particular
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361222 occurrence of the orogenic class of gold deposits. Thus, they are the result of gold re-
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391223 introduction by late orogenic fluids and trapping on the pre-existing sulphides (pyrite,
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411224 arsenopyrite) that are universally known to be efficient topochemical traps for gold. Detailed
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441225 observations at Tharsis in the Iberian Pyrite Belt (Marignac et al. 2003) and the VHMS of the
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461226 Ariab-Arbaat belt (this study) are good examples of this process.

48 491227 **4.3 An integrated model for gold in the Red Sea Hills?**

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511228 The late-D3 gold event in the Ariab-Arbaat belt has all the makings of an orogenic gold system,
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531229 being in particular very similar to the Late Carboniferous gold event in the West European
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561230 Variscan belt. The existence of a crustal-scale hydrothermal system may, therefore, be inferred.
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581231 Gold deposits and showings at the scale of the whole Red Sea Hills area (Fig. 1), either cluster

1232 around the main D3 shear zones (notably, Oko and Hamisana SZ) or are dispersed in specific
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21233 terranes (Gabgaba, Haya).
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51234 It is thus possible to propose that, at the very end of the Pan-African collision event (poorly
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71235 constrained at ca. 560 Ma), fluids released from the root of the newly formed terrane (the Red
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101236 Sea Hills area) were channelled towards the upper crust through the major D3 shear zones.
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121237 Then, outpouring from these major drains, they conveyed through fault plays and damaged
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141238 zones up to specific chemical traps. The latter are either the micro-fractured massive sulphide
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171239 ore bodies or mineral and structural trap such as the Kamoeb quartz veins systems. In this
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191240 scheme, occurrences such as the Debikwan prospect testify the fluid transfer reality through
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221241 secondary tectonic features.
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241242 Therefore, the Ariab-Arbaat gold district would be classified as a representative of the shear-
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271243 zone related orogenic gold deposits class (Groves et al. 2005 and references therein).
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291244 The source of the gold remains speculative. Yet, it may be observed that the high gold
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321245 endowment of the Arabian-Nubian Shield is correlated with the abundance of ophiolitic sutures,
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341246 making the shield to be “a massive graveyard of Neoproterozoic oceanic lithosphere” (Stern et
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361247 al. 2004). Thus the mafic-ultramafic rocks could have been a source for the gold, and Stern et
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391248 al. (2004) have drawn attention to the pervasive alteration of the harzburgites (serpentinisation,
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411249 carbonatation, often considered as “post-obduction”), that could have mobilised gold. In this
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441250 connection, the role plaid by the reactivated D1 sutures (such as the Nakasib in the Haya
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461251 terrane) in locating some gold clusters (Fig. 1) could be significant.

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50 511253 **4.4 The Ariab gold deposits in the ANS context**

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531254 In recent years, the orogenic gold deposits from the Eastern Desert terranes were thoroughly
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561255 investigated (Harraz 2000; El Tokhi and El Muslem 2002; Zoheir 2008a, b; Zoheir et al. 2008a,
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581256 b; Zoheir and Akawy 2010; Zoheir and Moritz 2014; Boskabadi et al. 2017; Cheng et al. 2017;

1257 Abd El Monsef et al. 2018; Talmer et al. 2018; Zoheir et al. 2018, 2019b; Perret et al. 2020).
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21258 Most of the deposits are associated with the Najd fault system's second and third-order plays
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41259 (Zoheir et al. 2019a). Based on a few direct c. 600 Ma Ar/Ar or Re/Os ages (Zoheir et al. 2019a
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71260 and references therein), or the observation that the gold lodes are often observed to overprint
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91261 the late alkalic granite plutons (e.g., Harraz 2000, Zoheir 2008b, Zoheir et al. 2008a), the gold
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111262 deposition could be related to the late activity of the Najd system during the ductile to brittle
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131263 transition. The gold lodes are typically syn- to late-kinematic quartz-carbonate-sulfide veins
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151264 hosted in the shear-zones, with bleached alteration zones (quartz-sericite-carbonates-pyrite).
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171265 The dominant fluids in the quartz lodes are described as volatile-rich (CO₂-CH₄-N₂) with
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191266 variable salinity, from low (0.3-3 wt % eq NaCl) to moderate (4-8 w% eq NaCl) salinity
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221267 (exceptionally reaching 17-20 wt % eq NaCl in a few deposits). Similar conclusions are
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241268 proposed for the Hamadi deposit for initial “homogeneous aqueous-carbonic mixture” are
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271269 inferred to form during retrogress metamorphism (Cheng et al., 2016). Gold is thought to have
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301270 been deposited from these fluids, at temperatures and pressures variably estimated between
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331271 225° to 380°C, and 35 to 260 MPa, a pressure drop (possibly associated with fluid unmixing)
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361272 being often considered to have been the key factor in gold deposition (Abu-Alam et al., 2019,
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391273 Cheng et al., 2016). Concerning the origin of ore-forming fluids, Talmer et al. (2018) assert
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411274 that “gold-bearing fluids were generated due to the metamorphic devolatilization processes
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441275 along with the greenschist-amphibolite-facies transition of ophiolite and metasedimentary
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461276 source rocks”. In contrast, Boskabadi et al. (2017) proposed that “carbonation of ANS ophiolitic
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491277 rocks due to the influx of mantle-derived CO₂-bearing fluids caused the breakdown of Au-
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511278 bearing minerals such as pentlandite, releasing Au and S to the hydrothermal fluids that later
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531279 formed the Au-deposits”.

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561280 All these results converge so that the Neoproterozoic gold deposits in the ANS are considered
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581281 as representatives of the mesothermal class (Zoheir et al. 2019a). Our findings in the Ariab belt
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1282 fit rather well with these results, with however significant differences. In the Ariab belt, gold is
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21283 demonstrated to have been deposited late in the tectono-thermal history, clearly at the end of
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41284 the exhumation. The late-D3 gold event in the Ariab belt has all the makings of an orogenic
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71285 gold system, notably similar to the Late Carboniferous gold event in the West European
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101286 Variscan belt (Boiron et al., 2003).
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121287 Does this mean that the Ariab deposits were formed under exceptional conditions not
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141288 encountered elsewhere in the ANS? A careful reading of the published works shows that it is
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171289 not the case and that, to the contrary, the Ariab style of evolution must be considered the rule
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191290 for all ANS. The crucial point is that the workers take the tight association of gold with either
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221291 arsenopyrite or As-rich pyrite as evidence that gold was coevally deposited with these
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241292 sulphides. Consequently, the studied fluid inclusions (FI), mainly representative of the quartz-
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271293 pyrite or quartz-arsenopyrite association in the mineralised veins, are currently considered
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291294 representative of gold deposition stage. The main cited evidence for coeval gold and sulphide
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321295 deposition is that gold is often found as inclusions in As-pyrite or arsenopyrite. Yet, when
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341296 looked carefully at, gold inclusions in the sulphides are later. What is more, in all the published
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361297 occurrences, gold, alone, or in rare cases, associated to galena and or tellurides, is also found
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391298 in microcracks in either quartz or pyrite or arsenopyrite. Therefore, it may be concluded that in
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411299 the other ANS gold deposits, gold was, as at Ariab, introduced later than pyrite or arsenopyrite
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441300 and that, consequently, the studied FI in these deposits cannot be representative of gold
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461301 deposition.
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491302 It is thus possible to propose that, at the very end of the collision events, fluids released from
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511303 the root of the newly formed terrane (Boskabadi et al., 2017; Talmer et al., 2018), were
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531304 channelled towards the upper crust through the major D3 shear zones such as the Najd system
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561305 in the northern terranes, and the Oko shear zone for the Ariab belt. The fluids outpouring from
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1306 these major drains were then conveyed through fault plays and damaged zones, up to specific
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21307 chemical traps like massive sulphide ore bodies or quartz veins systems.

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6 71309 **5. CONCLUSIONS**

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101310 - The existence of a crustal-scale hydrothermal system is linked to a late-D3 event in the Ariab-
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121311 Arbaat belt. Thus, at the end of the Pan-African collision event, fluids released from the root of
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141312 the newly formed terrane, the Red Sea Hills area, were channelled towards the upper crust
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171313 through the major D3 shear zones. Then, the outpouring from these major drains, they conveyed
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191314 through fault plays and damaged zones up. Most fluids are deep-seated CO2 rich fluids mixed
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221315 with a water component, thanks to the system's decompression and subsequent downward
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241316 penetration of waters.

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271317 -Specific chemical traps for gold are either the micro-fractured massive sulphide ore bodies or
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291318 mineral and structural trap such as the Kamoeb quartz veins systems. The high gold endowment
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311319 of the Arabian-Nubian Shield is correlated with the abundance of ophiolitic sutures. It could
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341320 have been a source for the gold after syn to post-obduction serpentinisation and carbonatation.
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361321 Thus, massive sulphides are not necessarily the main source for gold, and deformation of
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391322 VHMS not only the main process of metal extraction.

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411323 Therefore, the Ariab-Arbaat gold district would be classified as a representative of the shear-
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441324 zone-related orogenic gold deposits class. The late-D3 gold event in the Ariab-Arbaat belt has
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461325 all the makings of an orogenic gold system, remarkably similar to the Late Carboniferous gold
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491326 event in the West European Variscan belt.

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1570 **Figure captions**

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51572 Figure 1. Schematic tectonic map of the Arabian-Nubian Shield, with the location of the Ariab
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71573 belt (redrawn and adapted from Johnson et al., 2011, with complementary data from Barrie et
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101574 al. 2016, Bierlein et al., 2016), and Zoheir et al., 2019)
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151576 Figure 2. Diagrammatic cross-section through the Gebeit-Oshib-Haya terranes, showing the
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171577 Amur-Nakasib ophiolitic suture and the Ariab Series (redrawn from Abu Fatima, 2006, after
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191578 Abdel Rahman, 1993 and Wipfler, 1996).
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241580 Fig. 3. Main events affecting the Haya terrane (see text for explanation and references).
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291582 Figure 4. Reconstruction of the Ariab series lithologic pile before the D1 deformation event
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311583 (redrawn from Abu Fatima, 2006).
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361585 Figure 5. Classification of the Ariab belt magmatic rocks in the SiO₂ vs Zr/TiO₂ (A) and Nb/Y
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391586 vs Zr/TiO₂ (B) diagrams (Winchester and Floyd, 1977). The use of the “immobile” elements
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411587 Zr and Ti partly alleviates the bias introduced by the strong hydrothermal alteration affecting
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441588 most volcanic and many plutonic rocks (see text).
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491590 Figure 6. Projection of the Ariab belt magmatic rocks in adapted mineral-chemical diagrams.
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511591 A. The A-B diagram of Debon and Le Fort (1988) for volcanic rocks of the Ariab Series.
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531592 Numbers associated with rock symbols are the LOI. B. The A-B diagram of Debon and Le Fort
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561593 (1988) for the Ariab-Arbaat belt plutonic G1 and G2 suites. C. The Q-P diagram of La Roche
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581594 (1980) for all magmatic rocks of the Ariab-Arbaat belt. Although most of the volcanic and
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1595 many plutonic rocks are affected by an intense alteration (involving an increase of the
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21596 peraluminous index A), a differentiation trend, common to all the volcanic series and, at least,
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41597 the G1 plutonic rocks, remains discernable in all diagrams (see text).
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101599 Figure 7. Isocon plots (Grant 1986) for selected couple of “unaltered”-altered rocks from (A):
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121600 Unit A basalts; (B): Unit B1 andesites; and (C): Unit C2 rhyolites.
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171602 Figure 8. Chondrite-normalized REE plots for the Ariab belt magmatic rocks (C1 chondrite
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191603 normalizing values from Anders and Grevesse 1989). Data from the Augaro back-arc basalts
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211604 in Eritrea (Teklay et al. 2003) have been added to the Unit A plot for comparison (see text).
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271606 Figure 9. MORB-normalized plots for the Ariab-Arbaat belt magmatic rocks, using the
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291607 normalization and ordering scheme of Pearce (1983) (in Rollinson 1993), with LILE on the left
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311608 and HFSE on the right, and incompatibility growing outward from Ba-Th (save the Cr addition
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341609 on the right). Data for comparison with Ariab basalts: Flin Flon arc tholeiites, redrawn from
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361610 Syme et al. (1998); Eritrea arc and back-arc basalts, redrawn from Teklay et al. (2003) (see
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391611 text).
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411612
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441613 Figure 10 Projection of the Ariab-Arbaat belt volcanic rocks in the Th/Yb vs Nb/Yb diagram
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461614 of Pearce and Peate (1995), showing clearly their oceanic island-arc affinity. G1 and G2 suites
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491615 have also been projected for comparison.
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541617 Figure 11. Comparison of C2 felsic lavas with reference series in primitive mantle-normalized
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561618 LILE-HFSE plots (normalizing values from Ringwood 1991). A: Comparison with the FI and
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581619 FII Archean felsic metavolcanic rocks (redrawn from Leshner et al. 1986). B: Comparison with
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1620 the FIIIa and FIIIb Archean felsic metavolcanic rocks (redrawn from Lesher et al. 1986). C:
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21621 Comparison with the Neoproterozoic Cape Breton Island felsic metavolcanic rocks from an
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51622 ensimatic arc (redrawn from Dostal et al. 1972). D: Comparison with the Paleoproterozoic Flin
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71623 Flon Belt felsic metavolcanic rocks from a primitive ensimatic arc (redrawn from Syme et al.
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101624 1998).

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151626 Figure 12. Geologic map of the central part of the Ariab area, with the location of the main gold
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171627 occurrences (redrawn from Abu Fatima 2006).

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191628 A Adasselakh B Baderuk D Dimdim G Ganaet H Hassai HA Hadal Aouatib Hm Hamim
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221629 Ht Hadayamet K Kamoeb M1-M5 Medadip 1 to 5 O Oderuk R Rukab S Shidimam T
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241630 Talaidirut Y Yonim

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291632 Figure 13 Post-depositional transformations in the VHMS deposits. **a.** Porphyroclastic pyrite
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311633 (py) in a quartz-sericite matrix (diffuse stockwork, Talaidirut) Qz1 and 2: two generations of
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341634 syn-kinematic fibrous quartz (“pressure-shadows”). **b.** Porphyroclastic muscovite (mus) and
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361635 pyrite (py) in a schistosed stockwork (Medadip 1) quartz: quartz amygdale. **c.** Differential
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391636 behaviour of massive sulphides during deformation (Yonim): fractured pyrite (py) and ductile
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411637 chalcopyrite (cp) (left); sketch of relationships in the same thin section (right) chl: chlorite; qtz:
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441638 porphyroclastic quartz; S1, T1, and S2, T2: two generations of schistosity and related tension
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461639 gashes. **d.** Differential behaviour of massive sulphides (Hassai): porphyroclastic pyrite (py) and
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491640 sphalerite (sp) fractured and enclosed in ductile chalcopyrite (cp) elongated in S1; Qz: quartz;
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511641 cv: supergene covellite. **e.** Two pyrite generations in the massive sulphides (Medadip 5):
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531642 cataclastic pyrite (py1), quartz (qz), and chalcopyrite (cp) banding (S0+S1) and porphyroblastic
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561643 euhedral pyrite (py2). **f.** Annealed pyrite (py) (Baderuk) Qz: quartz.

1645 Figure 14. Paragenetic evolution in the VHMS and Kamoeb deposits.
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51647 Figure 15. Late events in the massive sulphides. **a.** Quartz (Qtz) and chalcopyrite (Ccp) possibly
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71648 recrystallised with pyrite (Py) annealing (Medadip 5). **b.** Replacement of syn-kinematic fibrous
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91649 quartz (Qtz1) by post-kinematic chlorite (Chl) (Barut) Qtz: quartz; Tit: titanite. **c.** Fracture-
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111650 dissolution in pyrite (Py), filled with chalcopyrite (Ccp), sphalerite (Sp), and tennantite (Hadal
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141651 Ouatib). **d.** Dissolution cavity in pyrite (Py), filled by pyrrhotite (Po), arsenopyrite (Apy), and
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171652 tennantite (Hassai). **e.** Microvugs inside annealed pyrite (Py), filled with sphalerite (Sp),
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191653 chalcopyrite (Ccp), galena (Gn), and altaite (SEM-BSE, Medadip 1); note that the microvugs
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221654 developed independently from the pyrite joints. **f.** Microcrack in pyrite (Py) with altaite sealings
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241655 (SEM-BSE, Hadal Aouatib).
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291657 Figure 16. Gold setting in late microstructures. **a.** Brecciated microcrystalline pyrite (pyrite 1)
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311658 with quartz filling (SEM-BSE, Adassedakh); Gn: galena; Sp: sphalerite; Ccp: chalcopyrite;
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341659 AuAg: electrum. **b.** Detail of a: gold is associated with recrystallised pyrite in relation with the
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361660 brecciation. **c.** Native gold along actinolite (Act) cleavages (SEM-BSE, Yonim). **d.** Gold (Au)
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391661 in brittle fractures affecting magnetite (Mgt) and in re-activated S1 schistosity marked by
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411662 chlorite (Chl) (Ganaet); Brt: barite; Q2: fibrous (syn-D2) quartz (“pressure-shadow” around the
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441663 magnetite). **e.** Gold (Au) in kinked chlorite (Chl) (D2 structure, Ganaet); note that gold is either
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461664 along cleavage or cross-cutting them, demonstrating a post-D2 introduction.
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511666 Figure 17. Massive barite and associated hematite-chlorite schists. **a.** Porphyroclastic barite
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531667 (Brt) (Hamim) Qtz: quartz; Chl: chlorite. **b.** F2 microfold affecting the S1 hematite schistosity
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561668 (Ganaet) Mgt: magnetite. **c.** Early magnetite (Mgt) wrapped in the S1 hematite (Hem)
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581669 schistosity (Medadip 3); Chl: iron-stained chlorite. **d.** Early (pre-S1) microfolds (slumps?)
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1670 (white arrow) (Hamim) S0: primary bedding; S1: main schistosity; Chl: iron-stained chlorite;
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21671 Mgt: early magnetite; Qtz: quartz in “pressure-shadow”. e. Late D3 microcracks and quartz
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51672 (Qtz) infillings in magnetite (Ganaet). f. Gold association with late D3 microcracks in magnetite
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71673 (Mgt) (Ganaet).
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121675 Figure 18. Geological map of the Kamoeb area, displaying the Kamoeb quartz vein setting
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151676 (redrawn from Abu Fatima 2006, after Cortial et al., 1985). KE: Kamoeb east; KN: Kamoeb
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171677 north; KS1, KS2: Kamoeb south, vein 1 and 2; KW: Kamoeb west.
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221679 Figure 19. Microstructures in the deformed Kamoeb quartz and the gold and sulphides settings.
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241680 a. Existence of two successive flattening surfaces (S1, S2) materialised by microcrystalline
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271681 quartz, and the associated stylolites (Sty 1, Sty 2) expressed as either indented grain contacts or
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291682 (more often) as Ba-phengite accumulations. Stylolites 1 are microfolded with S2 as axial planes.
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321683 Sample KAM 9209-3A. b. Schematic relationships in mylonitic quartz (Sample KAM 9309-
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341684 123), with penetrative S2 and relicts of the pre-existing D1 microstructures, as S1 preserved
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361685 within porphyroclastic domains and intensely microfolded stylolite 1. c. Sulphide setting in
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391686 deformed quartz (Sample KAM9209-3C), with D1 relicts (mainly early phengite stylolite 1)
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411687 and penetrative D2 features (S2, phengite stylolite 2). Pyrite is syn-kinematic for D2 (quartz
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441688 pressure-shadows), whereas galena and gold overprint D2 features. d. Detail of c (redrawn from
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461689 SEM-BSE imaging). e. Aspect of d (redrawn from SEM-BSE imaging).
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511691 Fig. 20 : Main types of fluid inclusions in quartz from Kamoeb and VHMS. a: quartz textures
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541692 and simplified types of occurrences of fluid inclusions. b: microphotograph showing fluid
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561693 inclusion FI assemblages in Q₂₋₃, c: deformed fluid inclusions in Q₁, d: aqueous-carbonic
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581694 three-phase Lc-w inclusion, e-f : Lw-c aqueous-carbonic two-phase fluid inclusions
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21696 Figure 21. Calculated bulk compositions (in molar fractions) of representative FI s from the

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41697 quartz at Kamoeb and several VHMS (Yonim, Haddal Aouatib, Hassai). in a N_2/CO_2-H_2O

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61698 diagram.

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101700 Figure 22. Reconstruction of the late (D3?) P-T-t evolution as recorded by fluid inclusion

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121701 densities (isochores) from three representative areas (Yonim and Haddad Aouatib VHMS;

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141702 Kamoeb quartz veins). The biotite-in isograd is from Yardley (1989, Fig. 3-11, p. 86). The Si-

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161703 in-phengite geo-barometer isopleths (Massonne and Schreyer, 1987) and the Al-in-chlorite

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181704 geo-thermometer isopleths (Cathelineau, 1988) have been drawn for relevant compositions.

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201705 Isopleths (dotted dark blue lines) in the system H_2O-CO_2-NaCl are from Weisbrod (1984) for

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221706 a salinity of 6 wt. % NaCl equiv., and a CO_2 content of 10 (a) and 20 (b) mole%. See text for

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241707 explanation.

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281709 Figure 23 Cross-section model of a typical Au-bearing gossan developed at the expense of a

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301710 VHMS deposit from the Red Sea Hills (redrawn from Abu Fatima 2006, after Recoché 1989).

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1712 **Table captions**

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1714 Table 1. Representative analyses of the main rocks types from the Ariab Series and G1 and G2

1715 suites. Italics designate the altered samples.

1718 Table 2. Summary of the microthermometric measurements in the Kamoeb and VHMS (Yonim,

1719 Hadal Ouatib, and Hassai) quartz. Tm CO₂: melting temperature of the solid CO₂ Tm ice: final

1720 melting temperature of ice. Tm clat: final melting temperature of clathrates. Th CO₂:

1721 homogenisation temperature of the volatile phase. Th: bulk homogenisation temperature. L:

1722 homogenisation in the liquid phase, respectively.

1724 Table 3. Bulk compositions from Raman microanalysis of selected FI from both the Kamoeb

1725 and the VHMS (Yonim, Haddal Aouatib, and Hassai) quartz. Tm CO₂: melting temperature of

1726 the solid "CO₂". Tm ice: final melting temperature of ice. Tm clat: final melting temperature of

1727 clathrates. Th CO₂(L): homogenisation temperature of the volatile phase in the liquid state. Th:

1728 bulk homogenisation temperature, *: decrepitated under heating.

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21732 **Research highlights**

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41733 VMS belong to an ensimatic Tonnian arc and cannot be considered as a gold-rich VMS

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71734 Gold is introduced in a late metamorphic brittle stage, part of the AN Shield gold province

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91735 Gold is mostly exploitable in gossans

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Declaration of interests

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

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

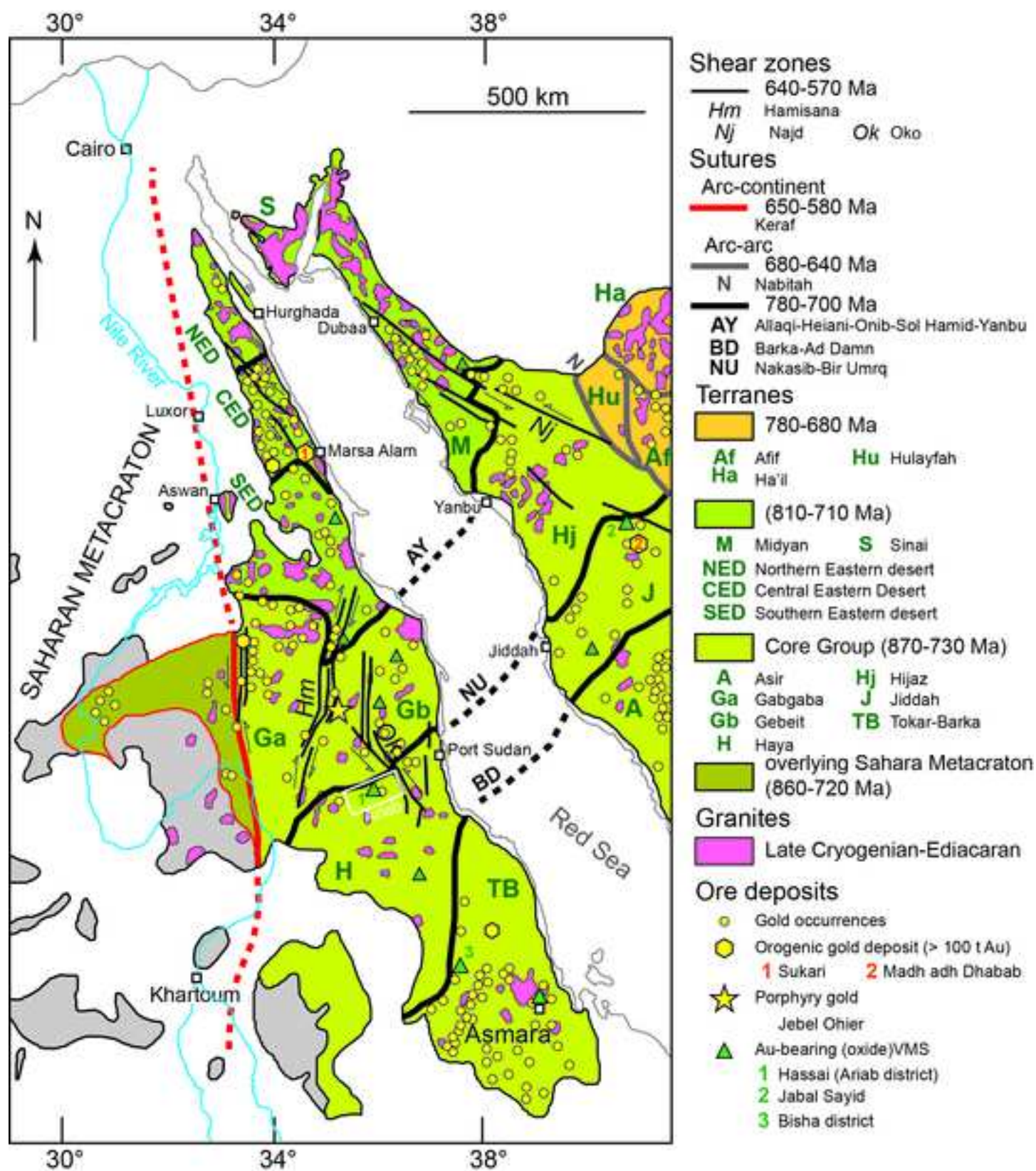
Author contributions

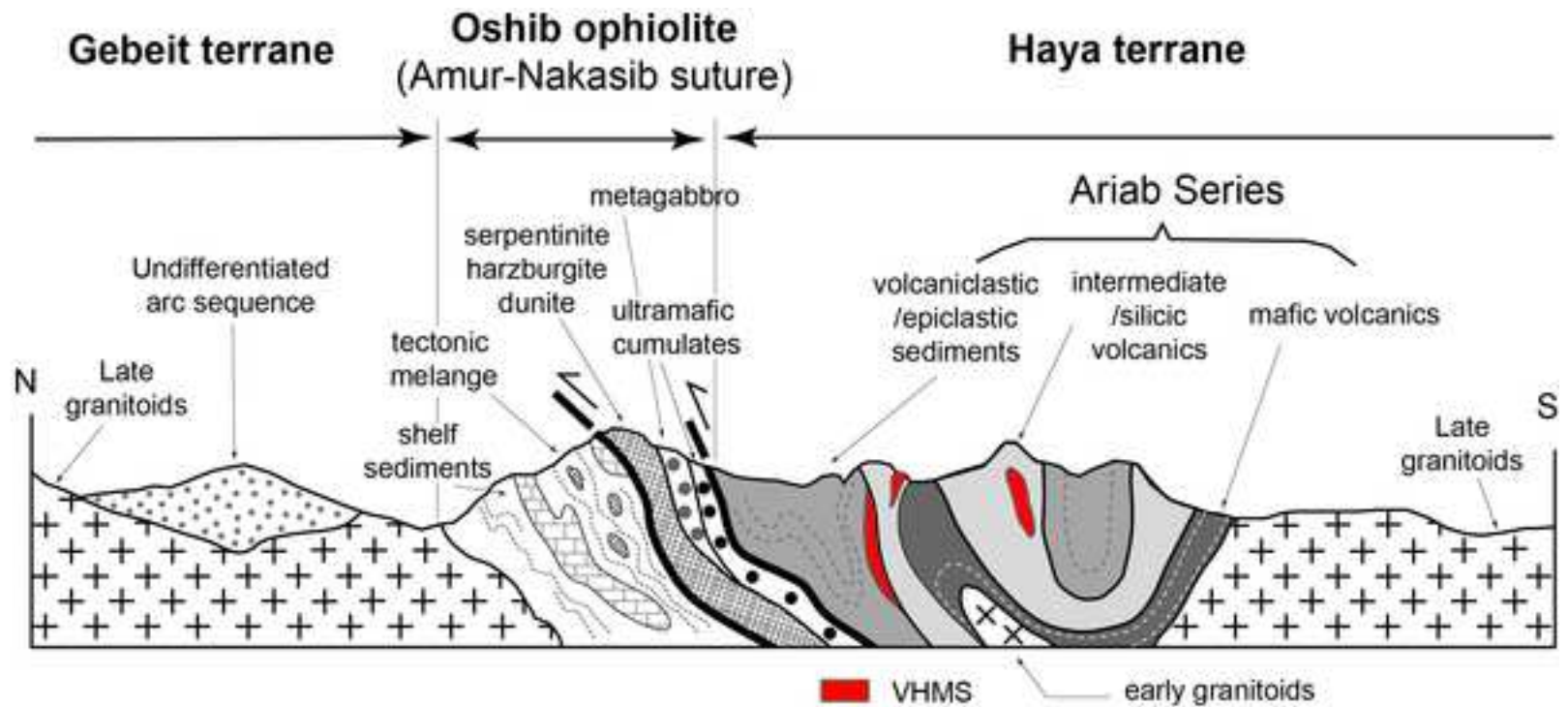
A.F: Investigation

CM: Supervision, Writing (original draft)

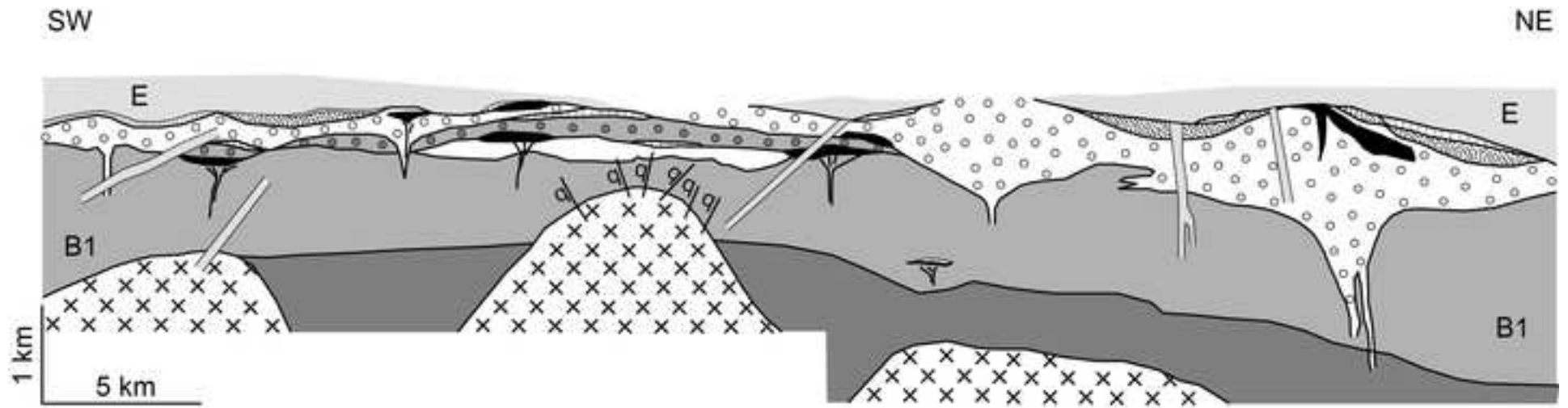
MC: Supervision, Writing (review, editing)













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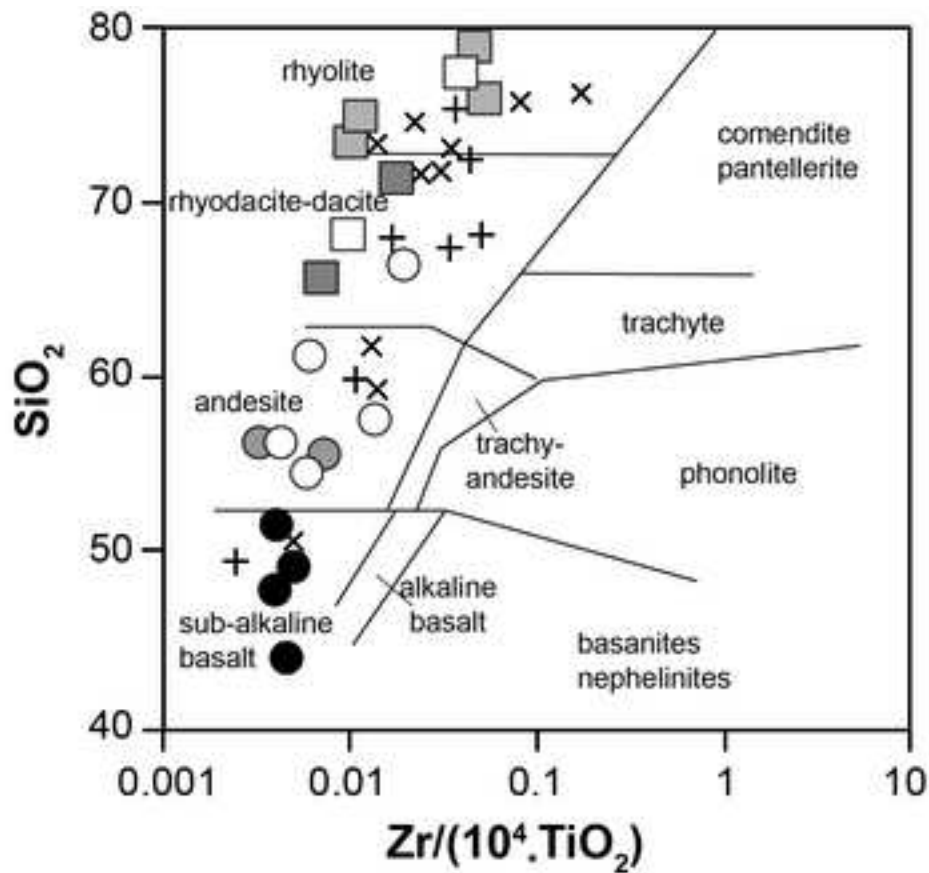




Cycle	Global event	Tectono-metamorphic event	Objects	Age
Pan-African	Juvenile arc in oceanic environment	Hydrothermal sub-sea floor metamorphism	Basalt to rhyolite submarine to subaerial volcanism (Ariab Series) and coeval plutonism (G1 suite) VHMS and barite deposits	ca. 900 Ma G1 granitoids at 888±3.7 Ma
	Arc accretion and thickening <i>unconformity</i>	D1 Obduction and oblique arc collision Low-grade metamorphism	Regional S1 schistosity Recumbent to upright regional N60-70°E F1 folds Transpressive SZ	ca. 770 Ma
		D2 Late shortening Low-grade metamorphism	Upright local N0-30°E F2 folds Regional S2 schistosity Dextral reactivation of D1 SZ	
	Evolved mature arc	Contact metamorphism	Awat-Aseriba calc-alkaline series Coeval to late plutonism (G2 suite)	730-720 Ma 720-680 Ma
Final collision	D3 Transpression	Conjugated sinistral NW-SE to NNW-ESE and dextral NNE-SSW ductile to brittle SZ Faint local S3 schistosity	620-580 Ma	
Gondwana break-up			Anorogenic plutonism	150 ± 4 Ma
Red Sea rift	Red Sea early opening	D4 Fracturation	NE-SW faults	Miocene



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|---|---------------------------------|---|--|
|  | Volcano-sedimentary unit E |  | Felsic pyroclastites and rhyolite domes (C2) |
|  | Early quartz vein (Kamoeb-type) |  | Upper andesites (B2) |
|  | Quartz porphyry (C3) |  | Rhyolites (C1) |
|  | Fe-chert and jasper |  | Lower andesites (B1) |
|  | Massive sulphides (VHMS) |  | Mafic volcanics (A) |
|  | Massive barite |  | Early granitoids (G1) |



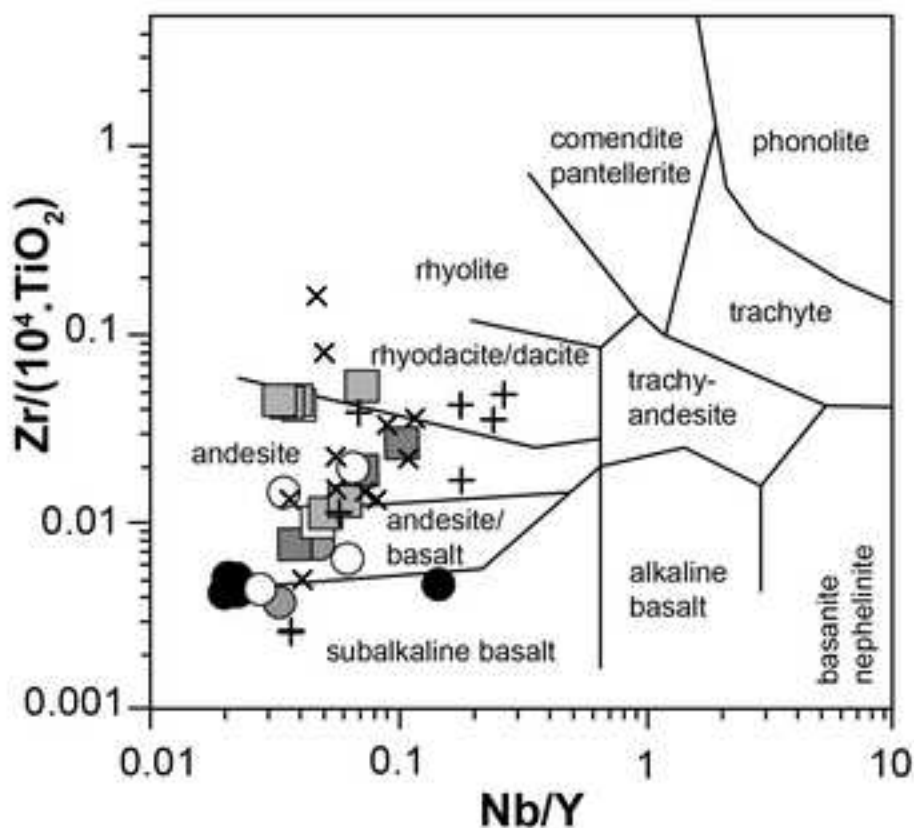
(A)

Volcanic rocks

- C3 quartz-porphyry
- C2 "upper rhyolite"
- C1 "lower rhyolite"
- B2 "upper andesite"
- B1 lower andesite
- A basalts

Granitoids

- × G1 early intrusives
- + G2 late intrusives



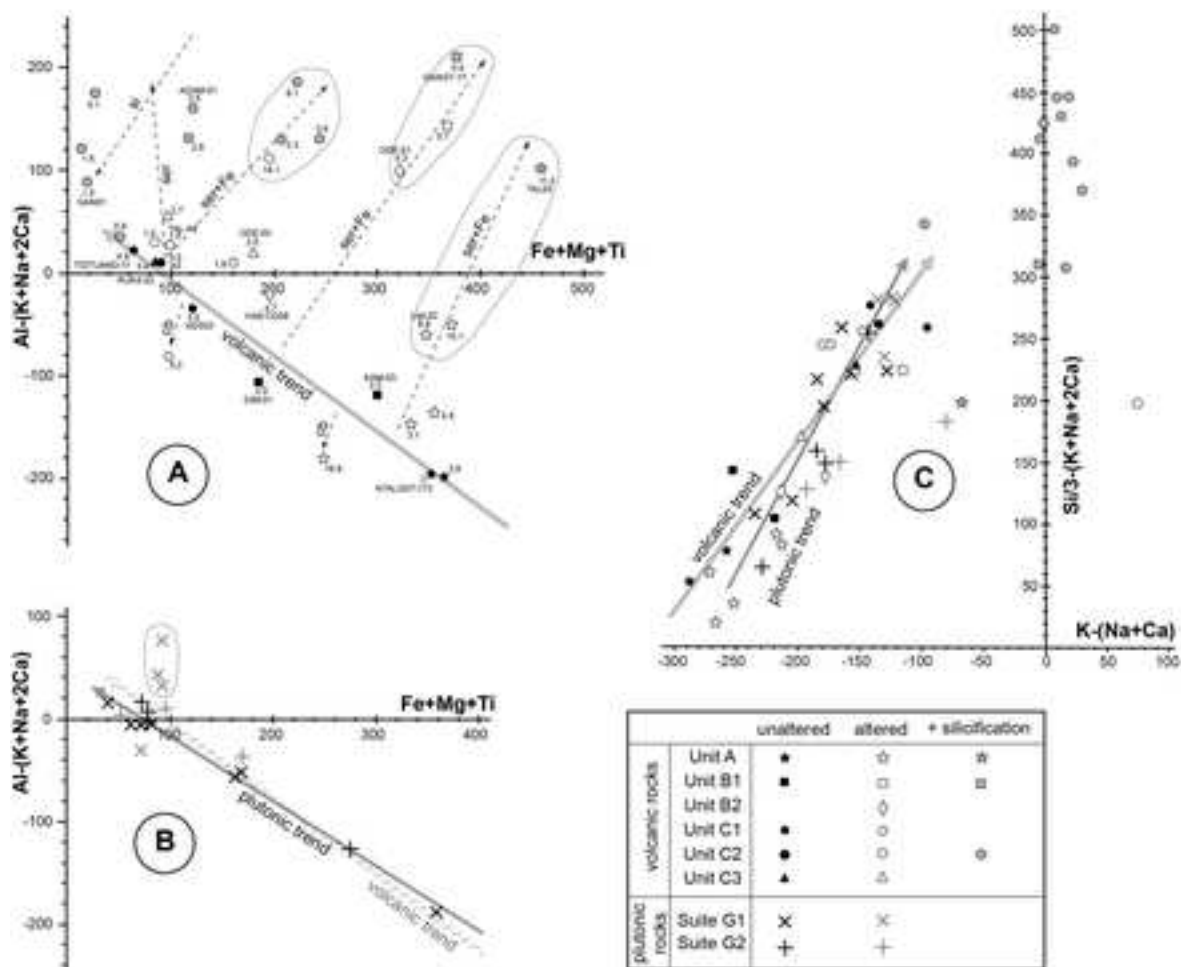
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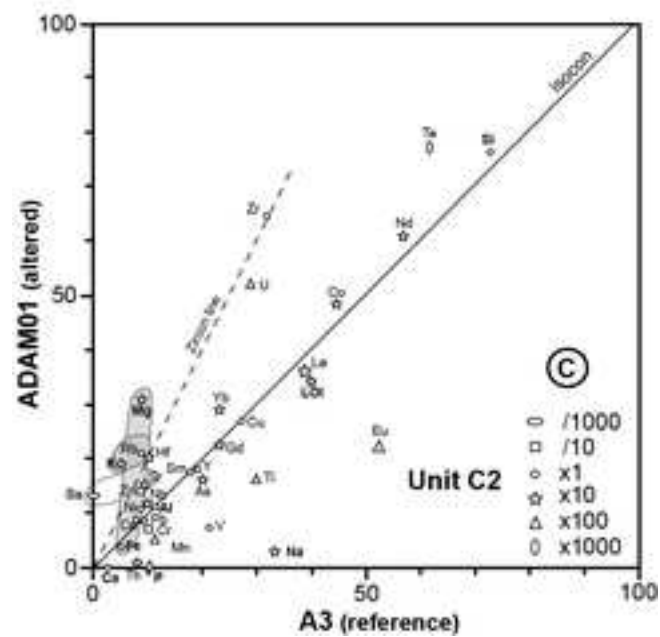
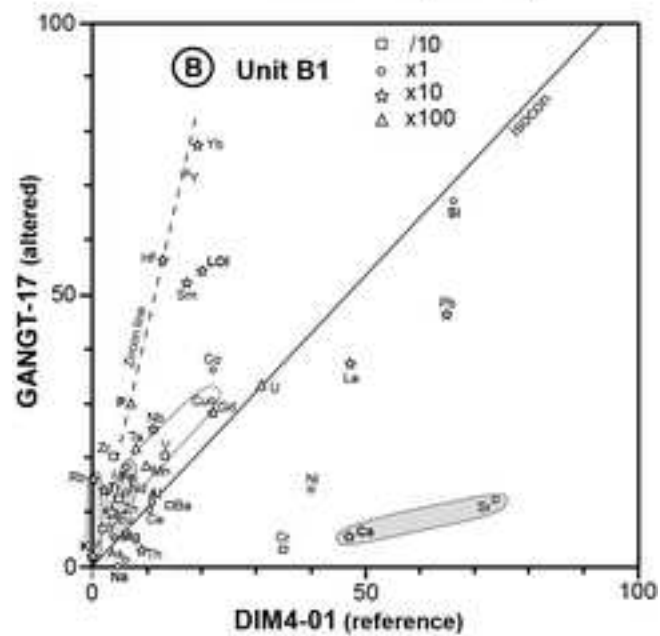
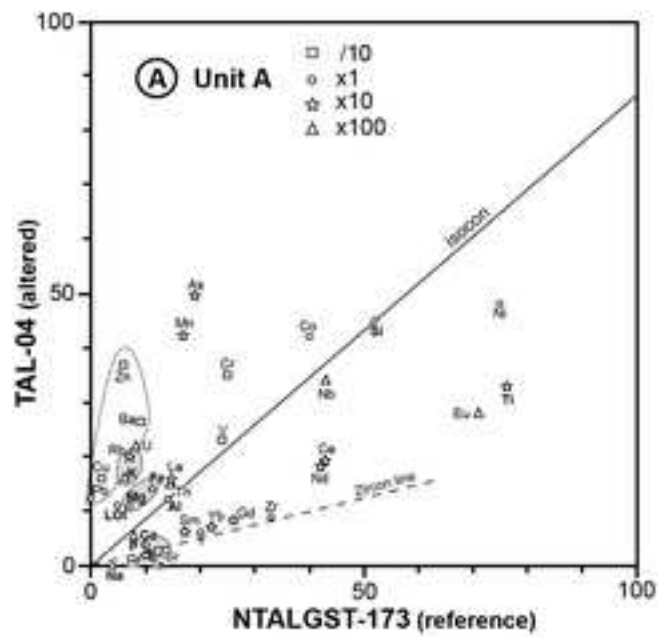
Volcanic rocks

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Granitoids

- × G1 early intrusives
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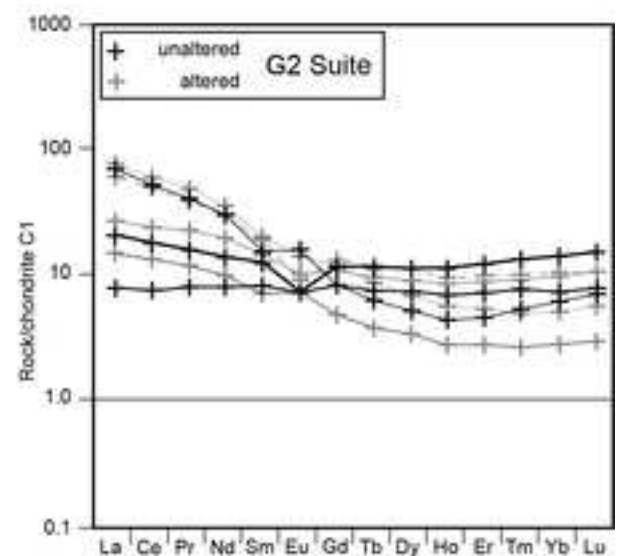
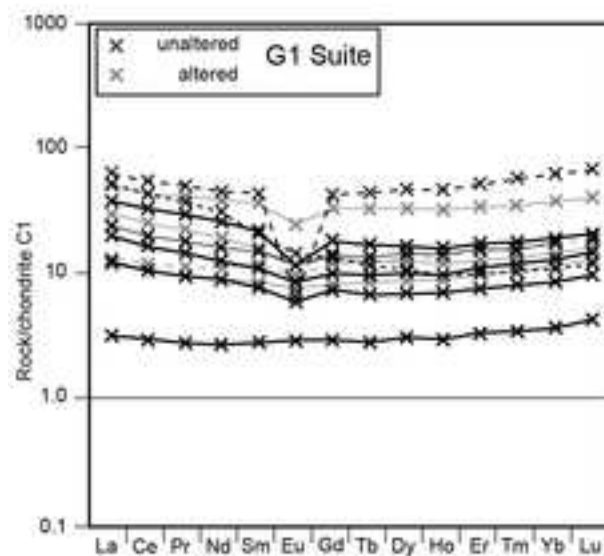
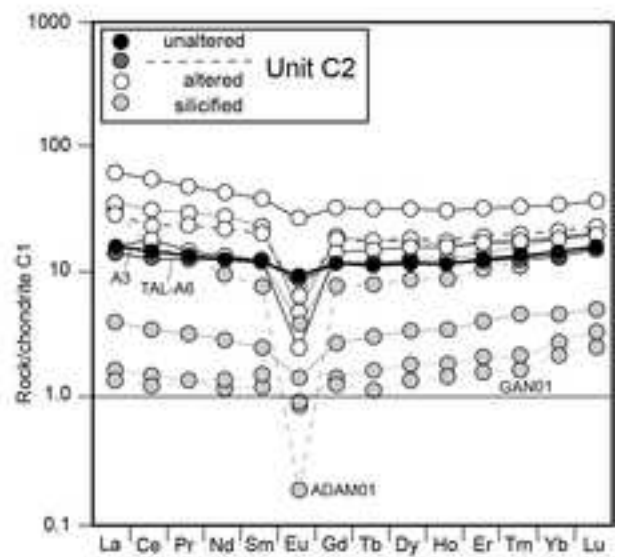
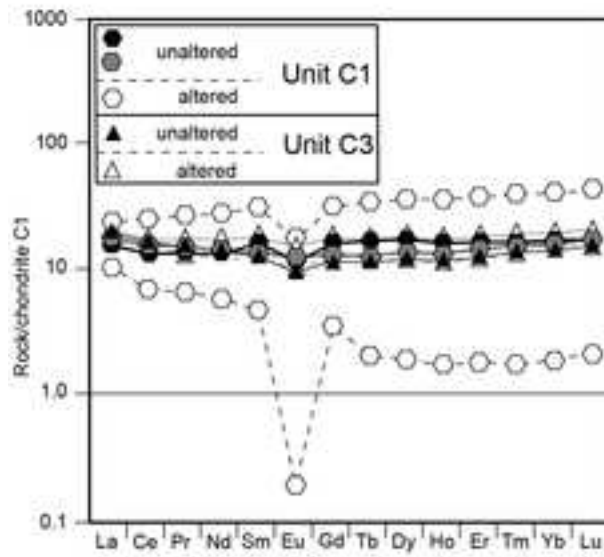
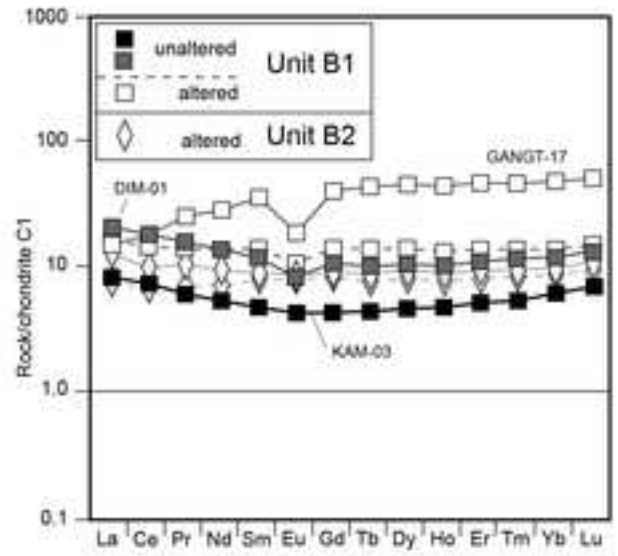
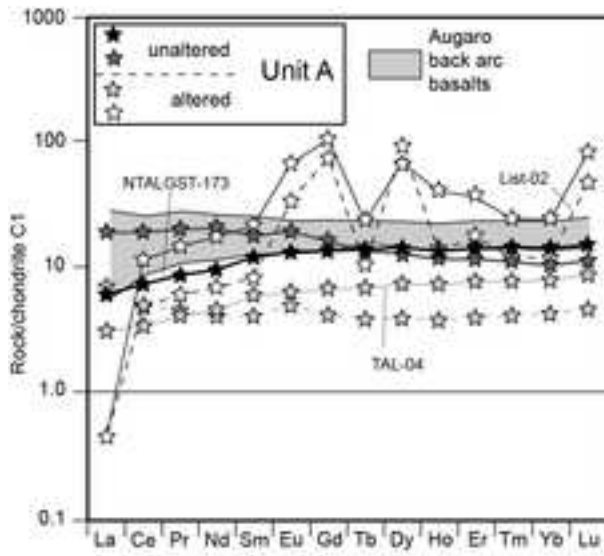
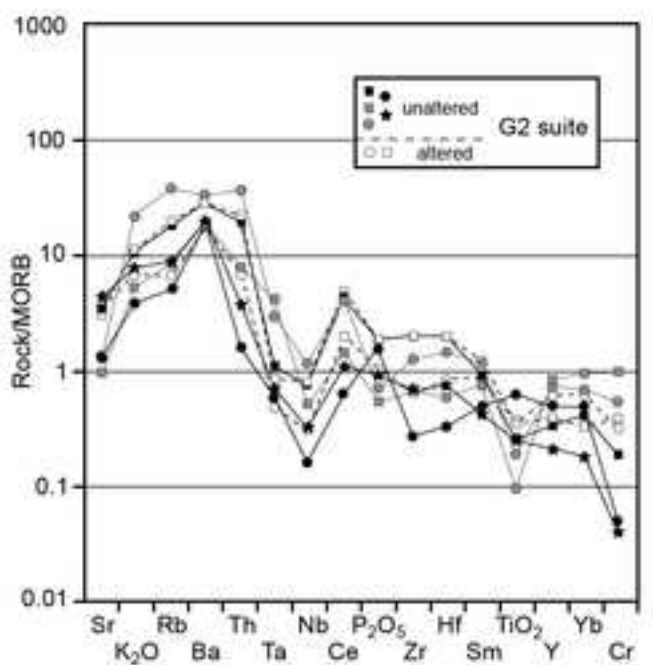
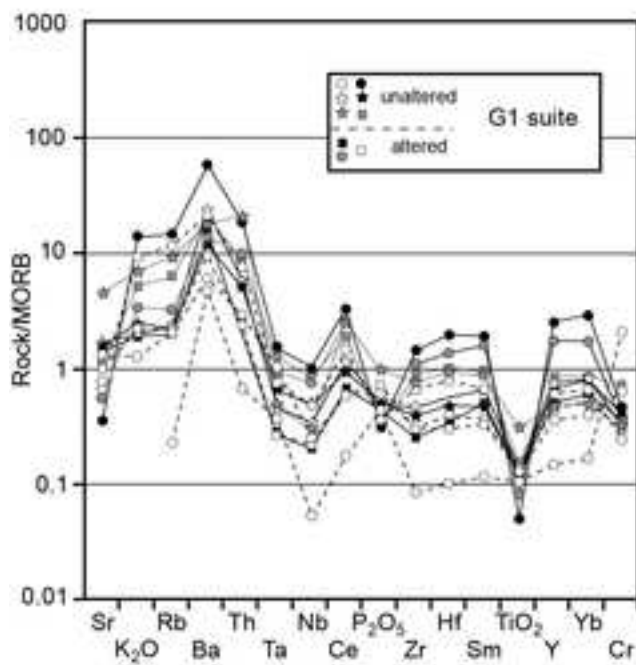
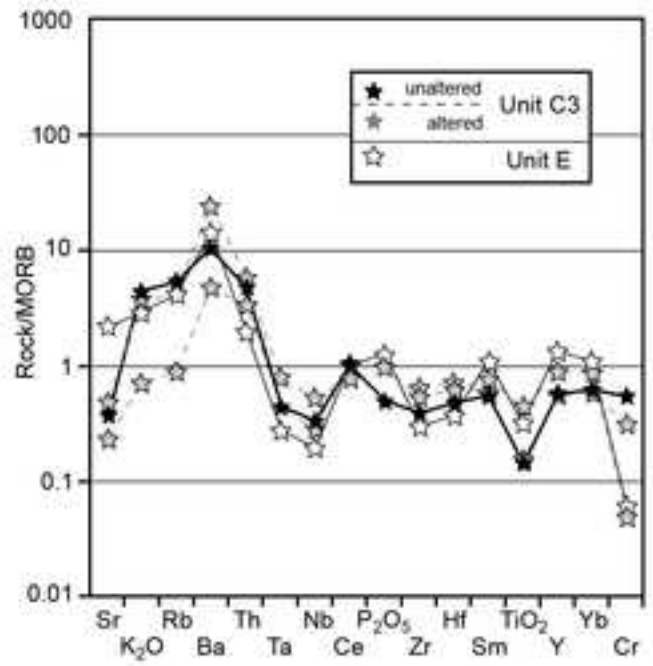
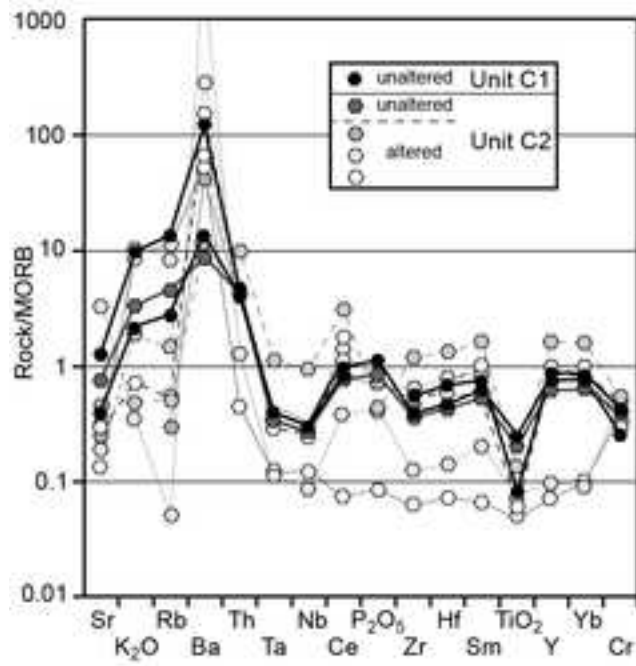
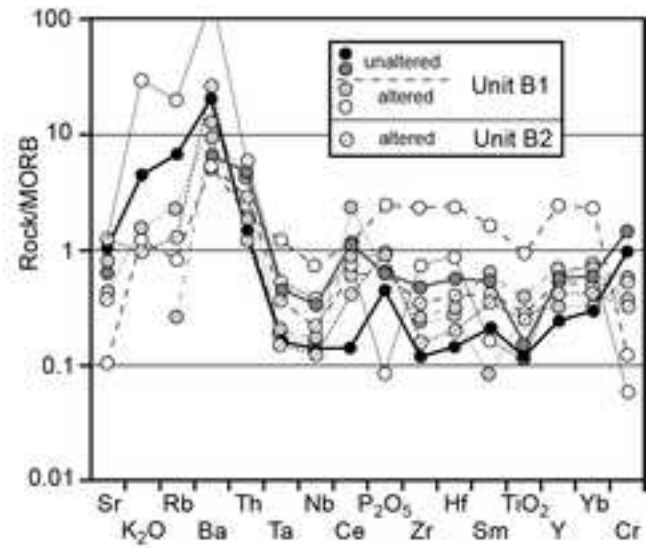
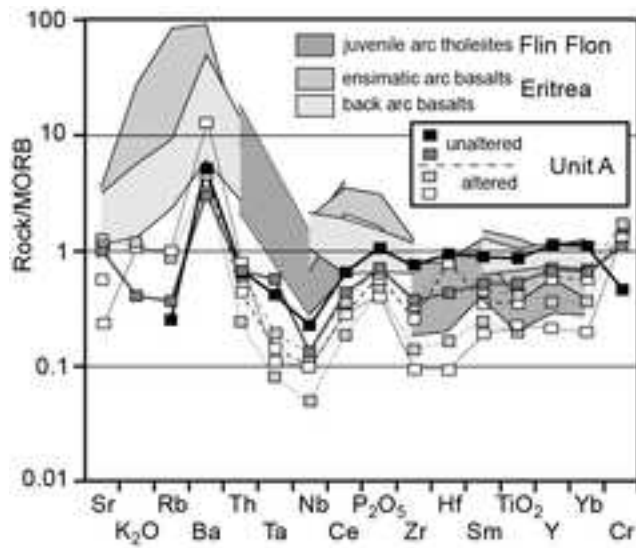
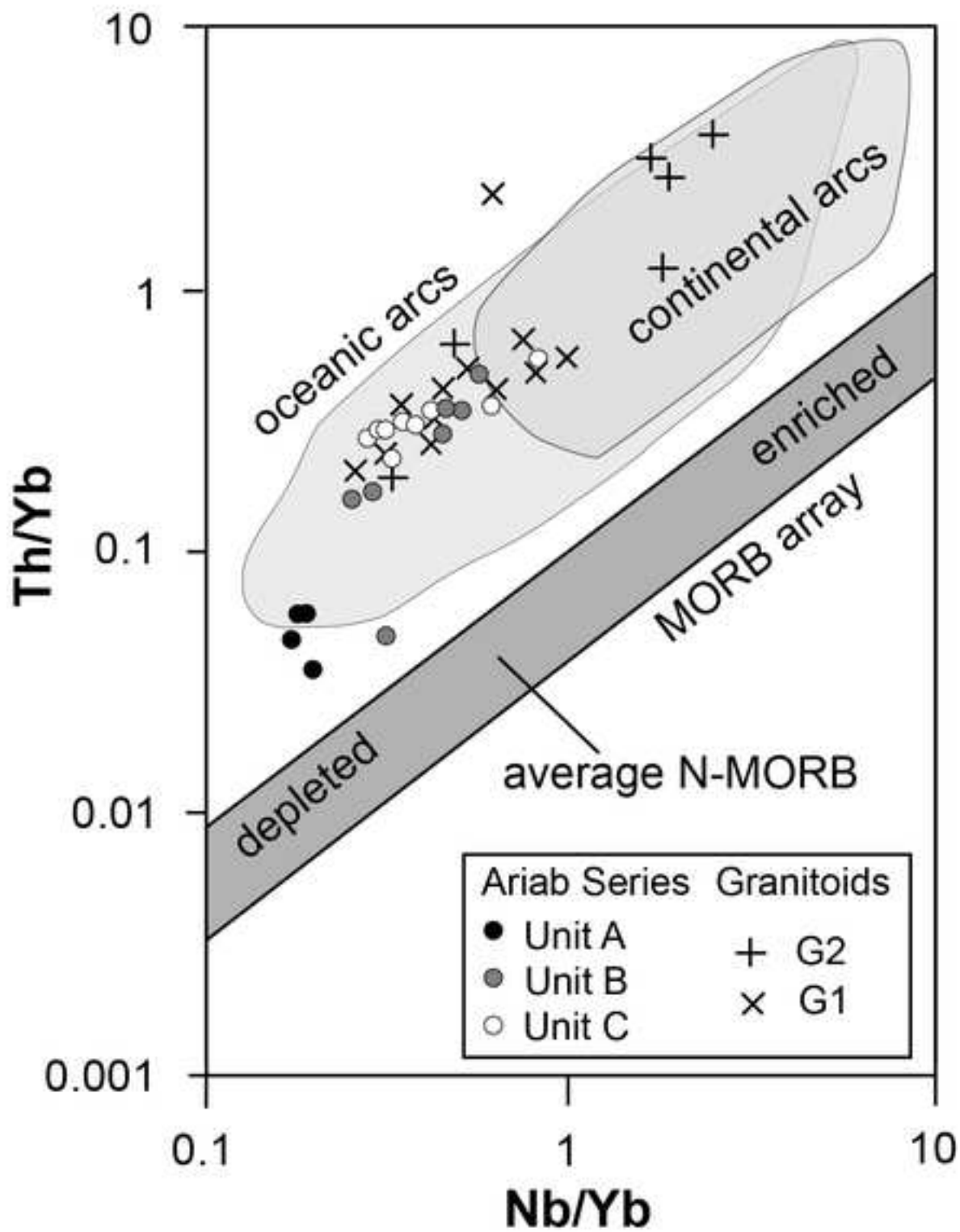
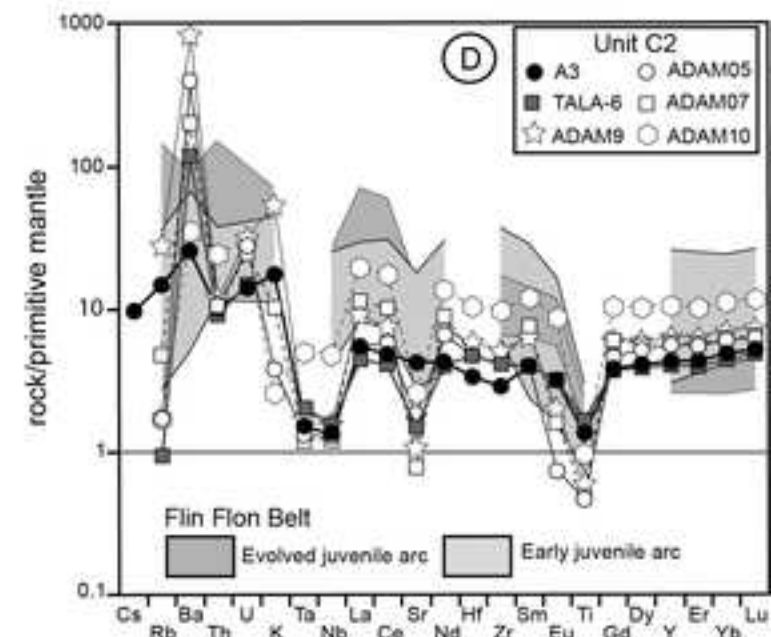
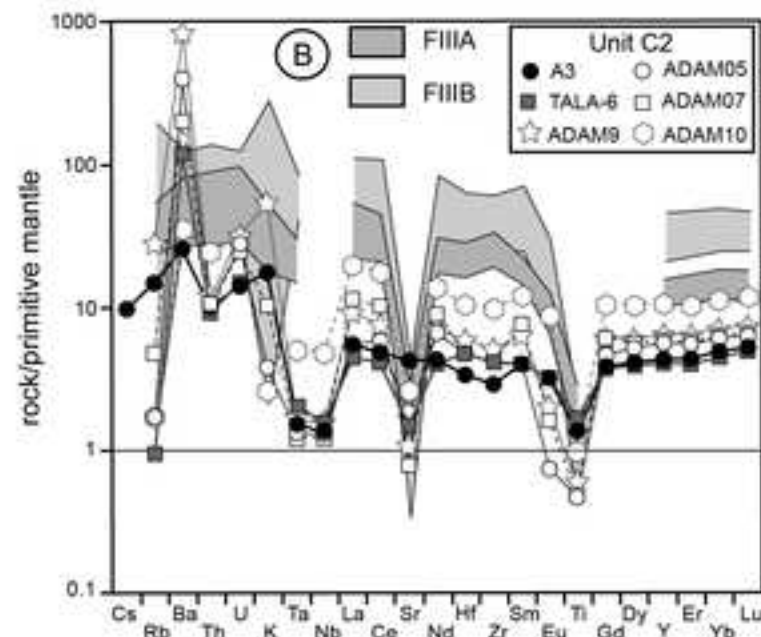
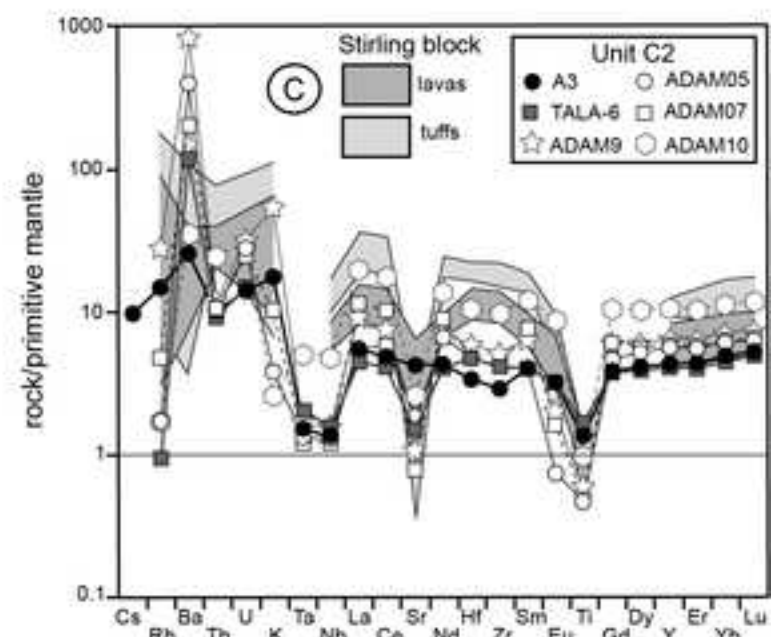
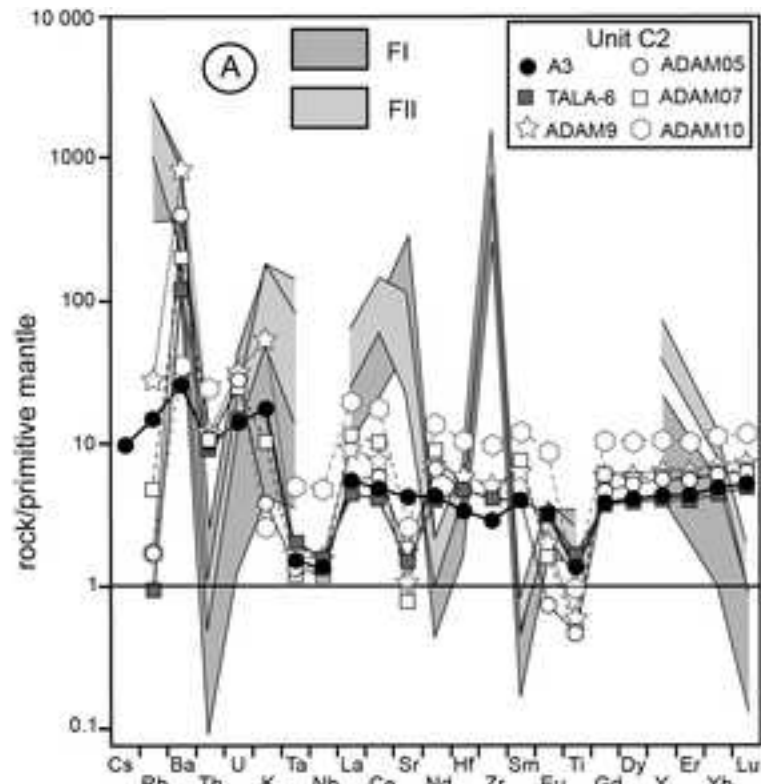


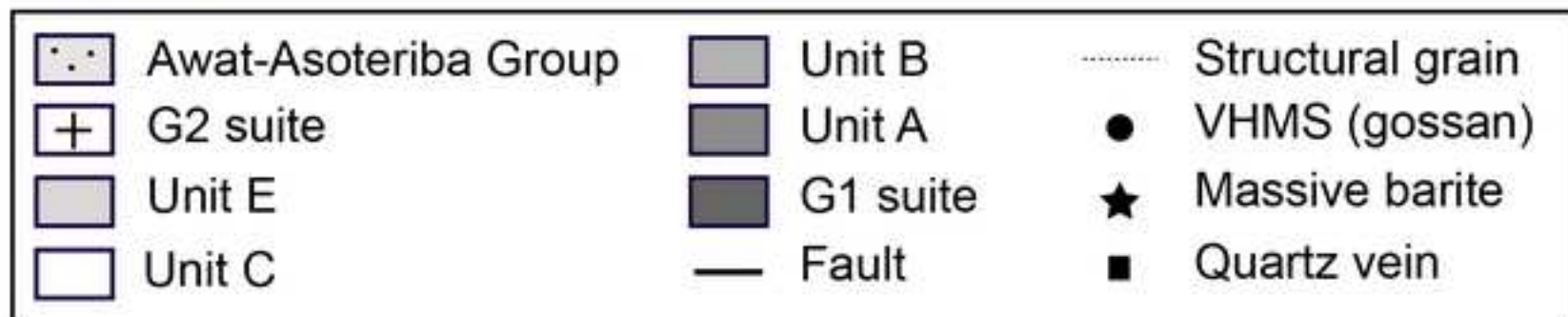
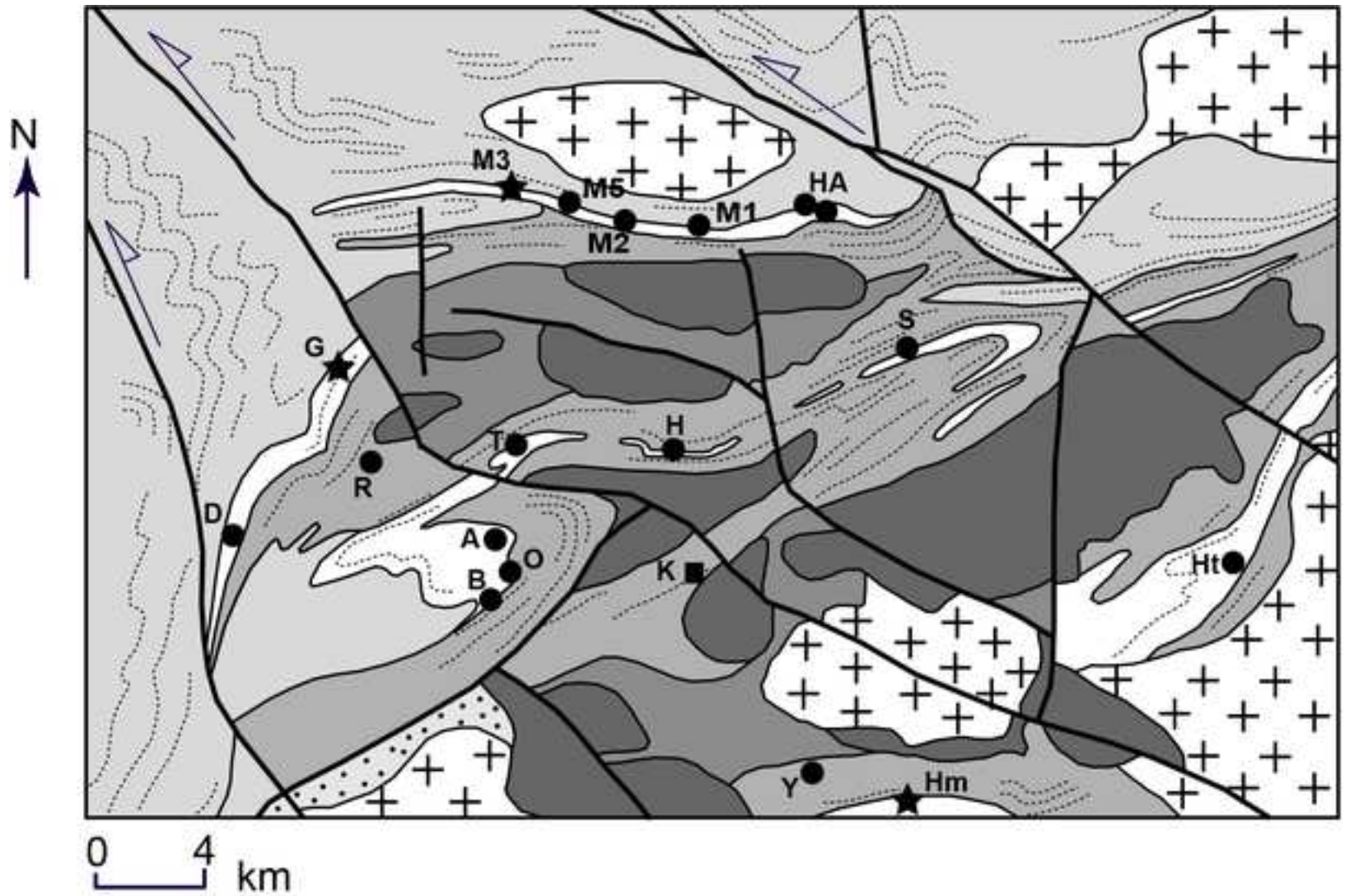
Fig. 9

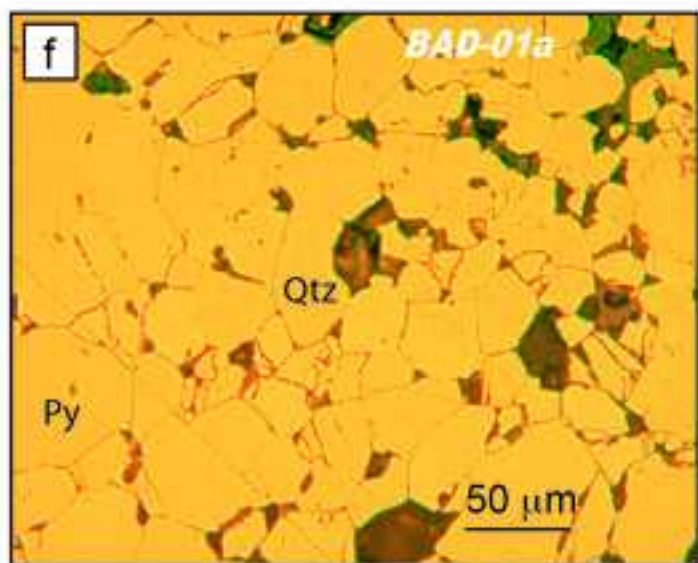
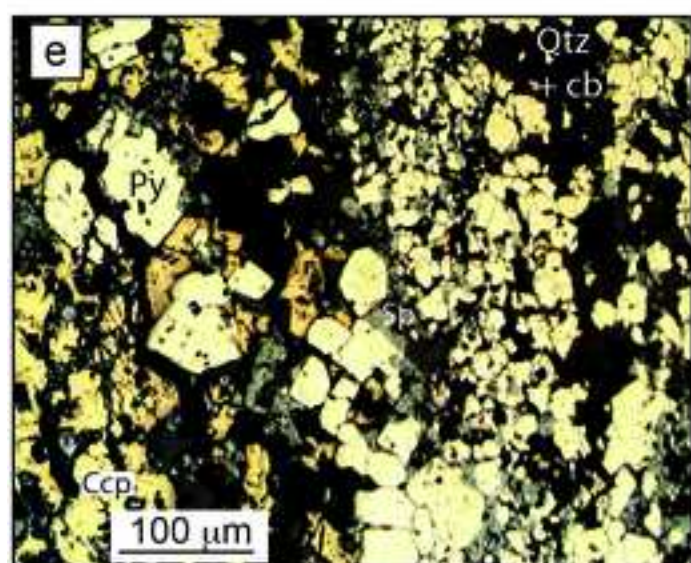
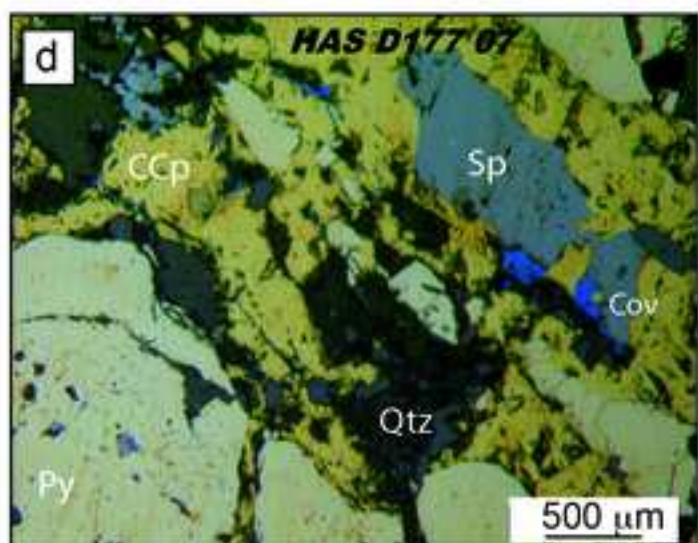
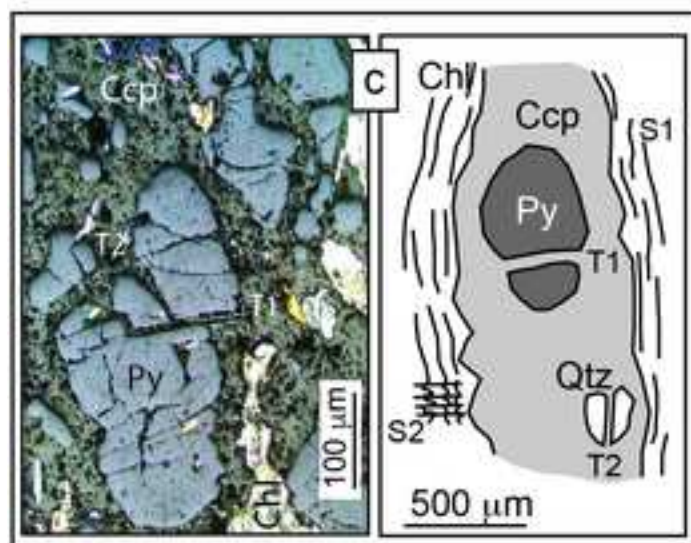
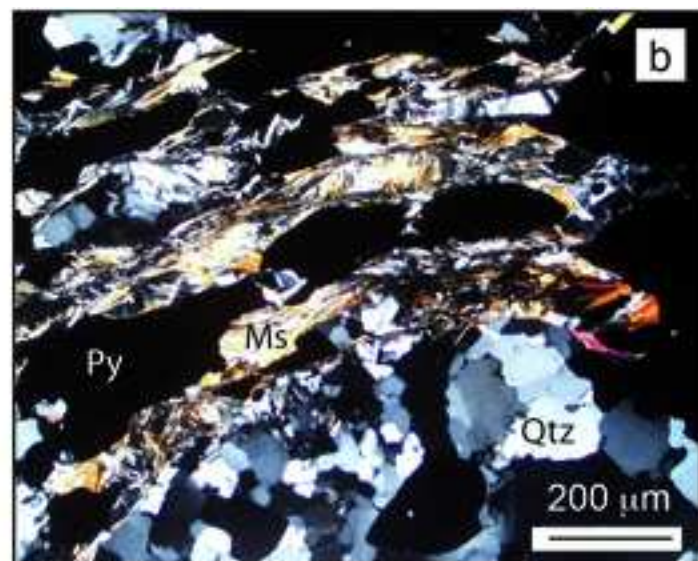
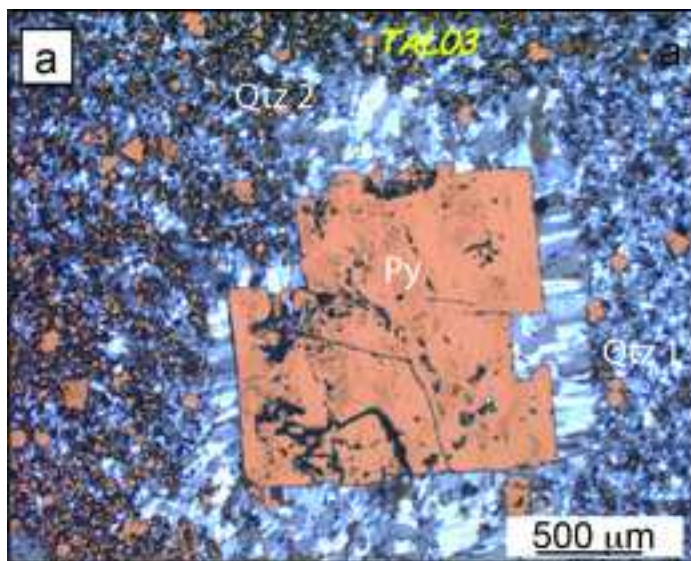
[Click here to access/download;Figure;Fig.9-Soudan-spider-total-vs2.tif](#)

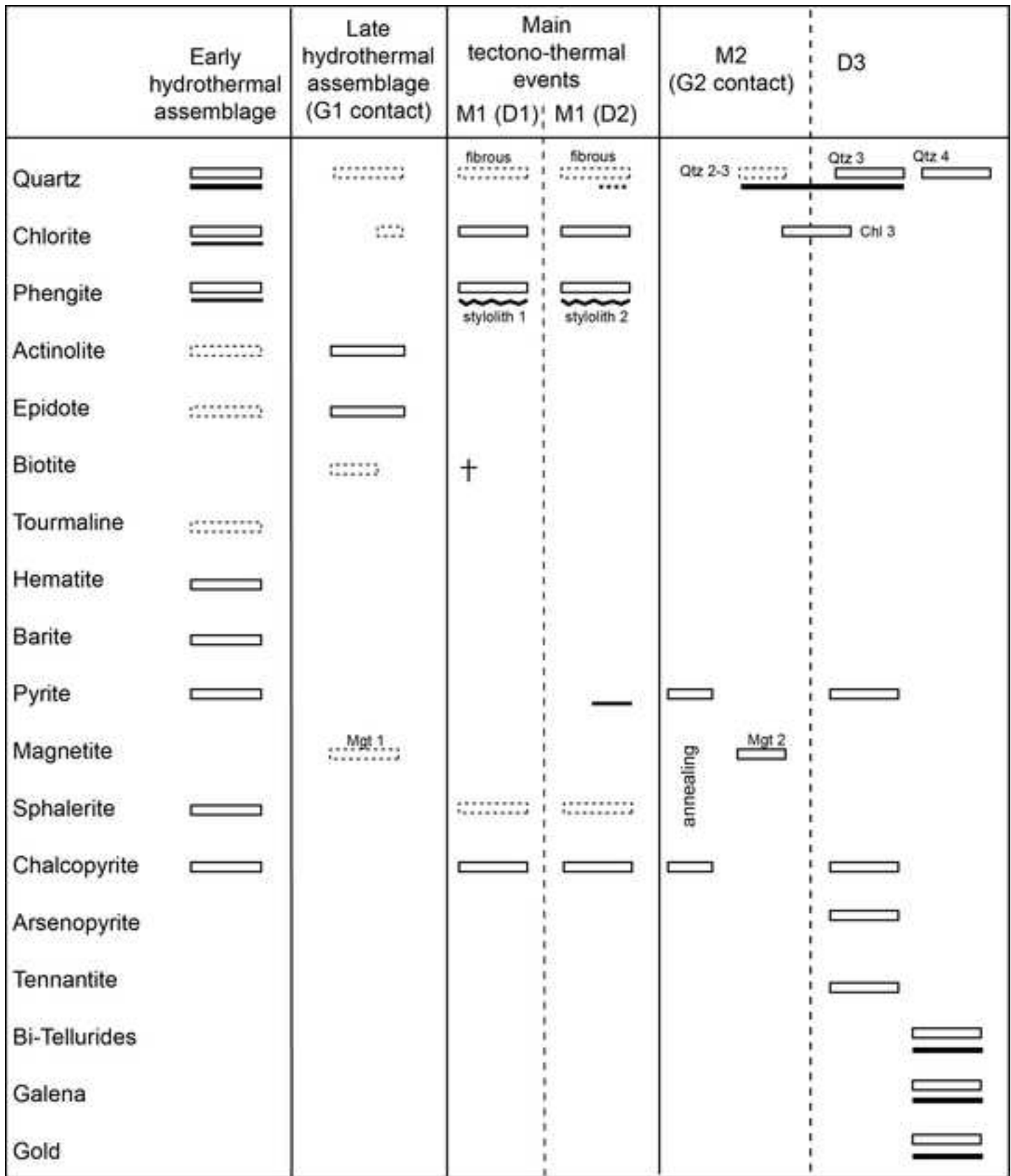






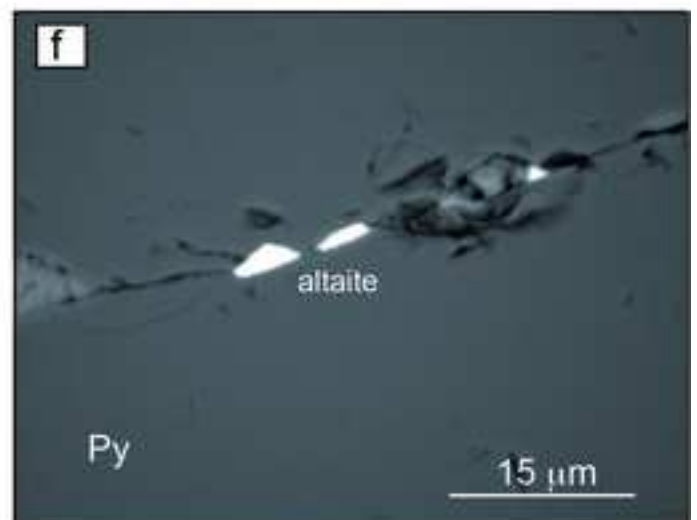
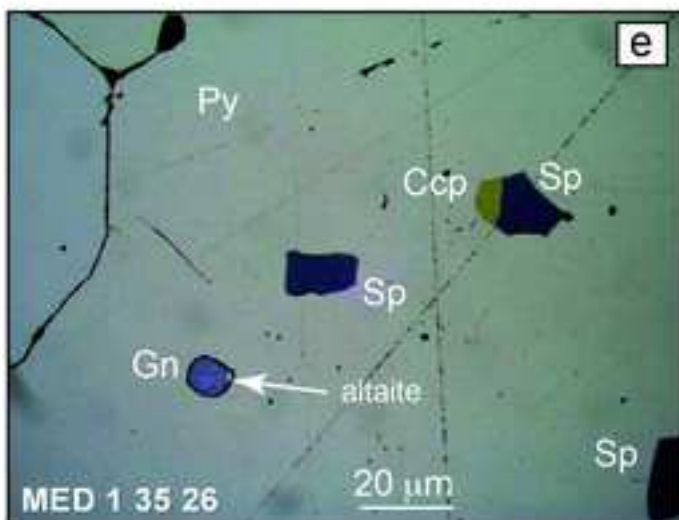
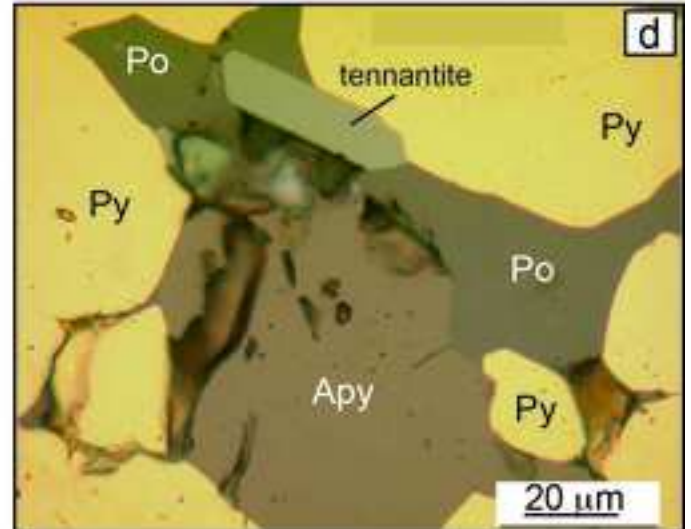
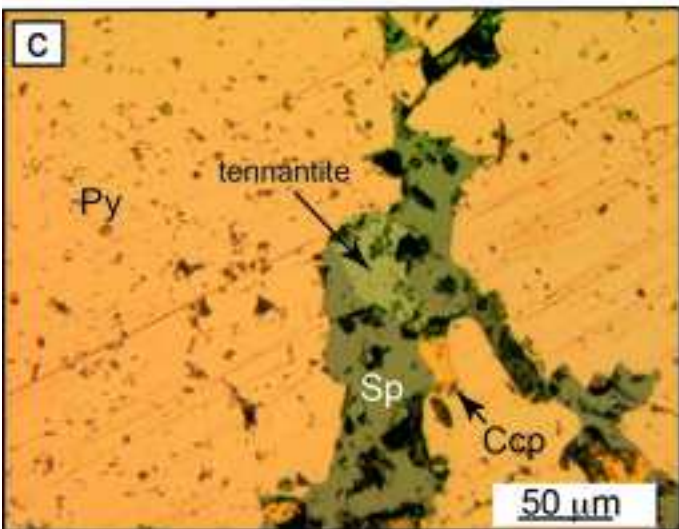
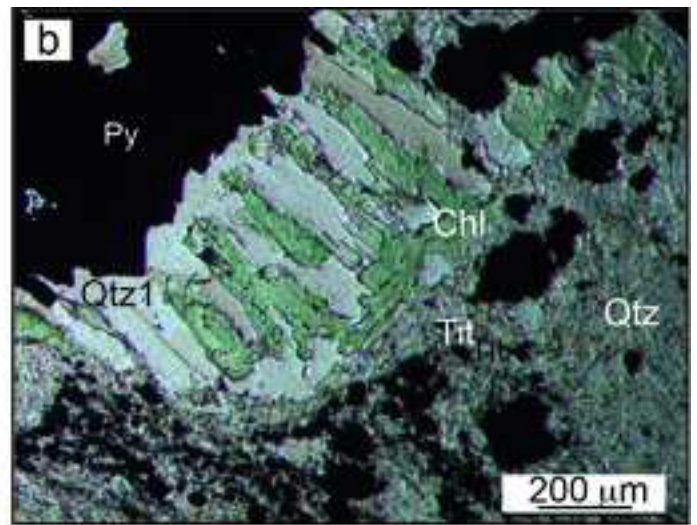
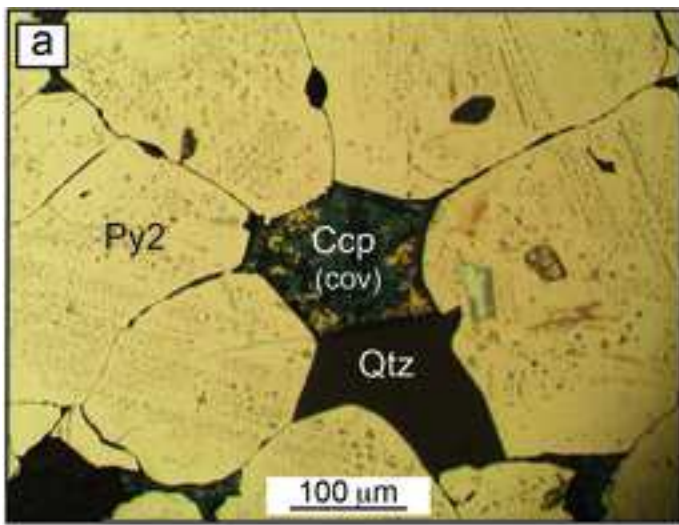


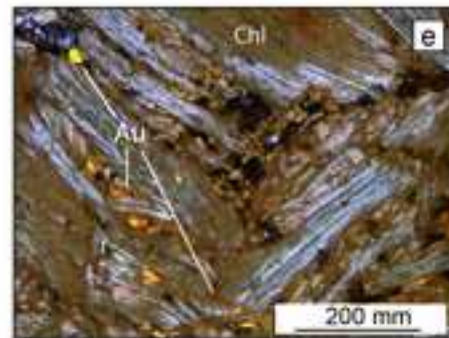
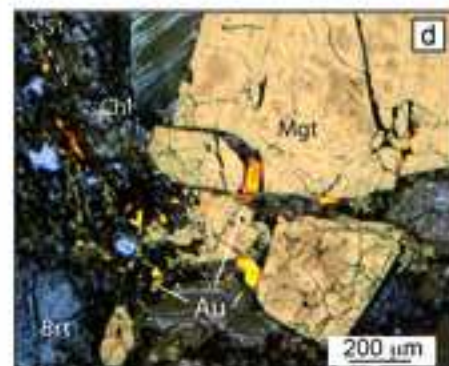
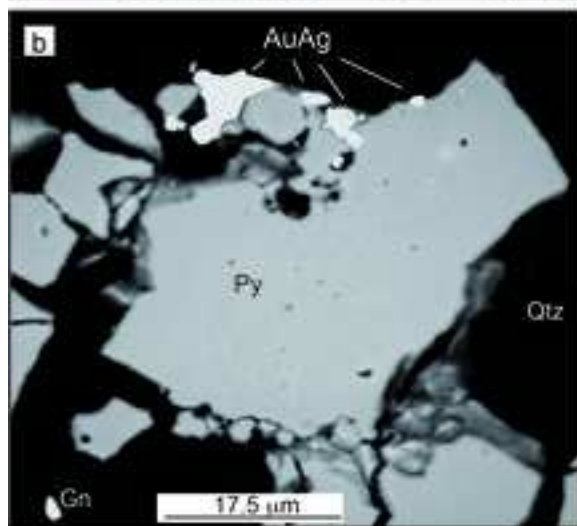
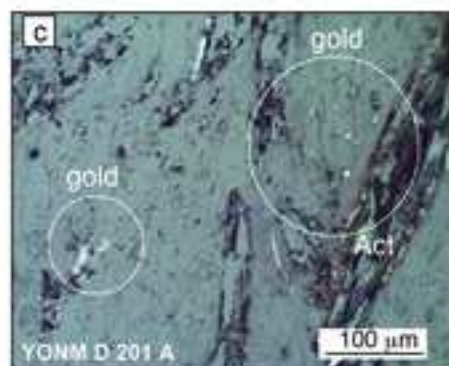
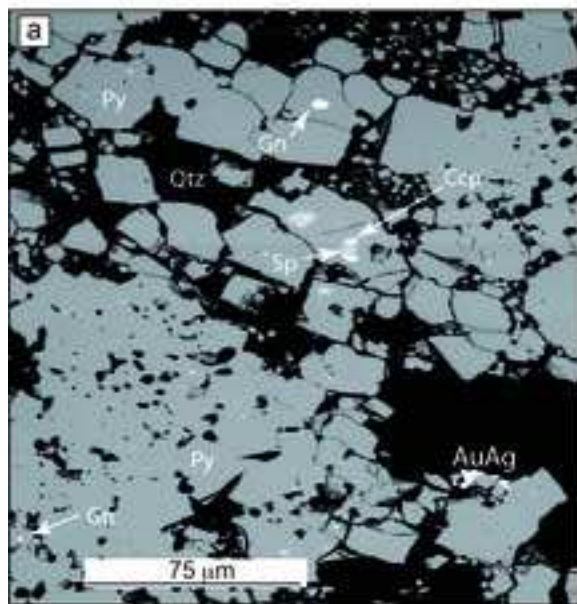


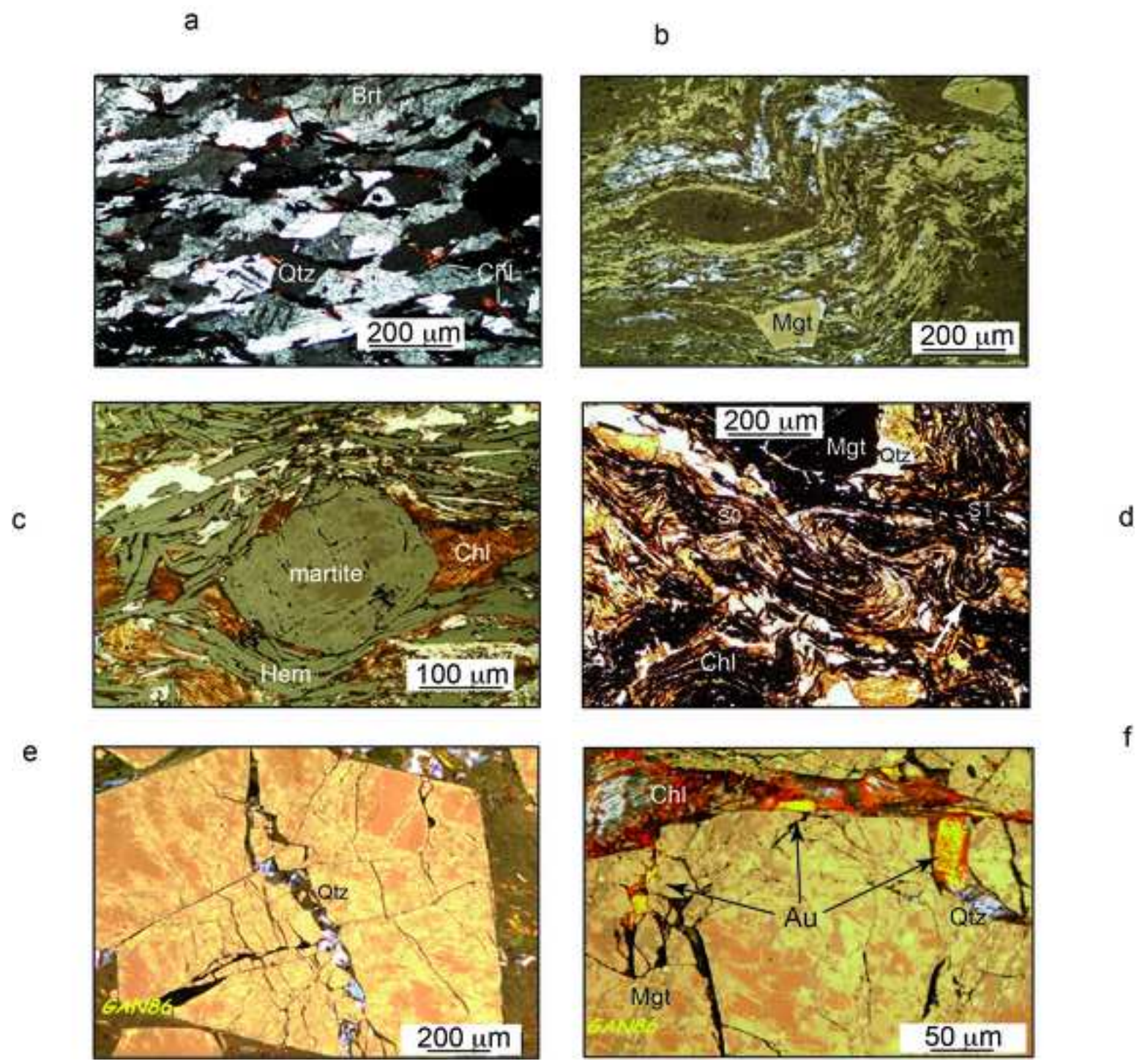


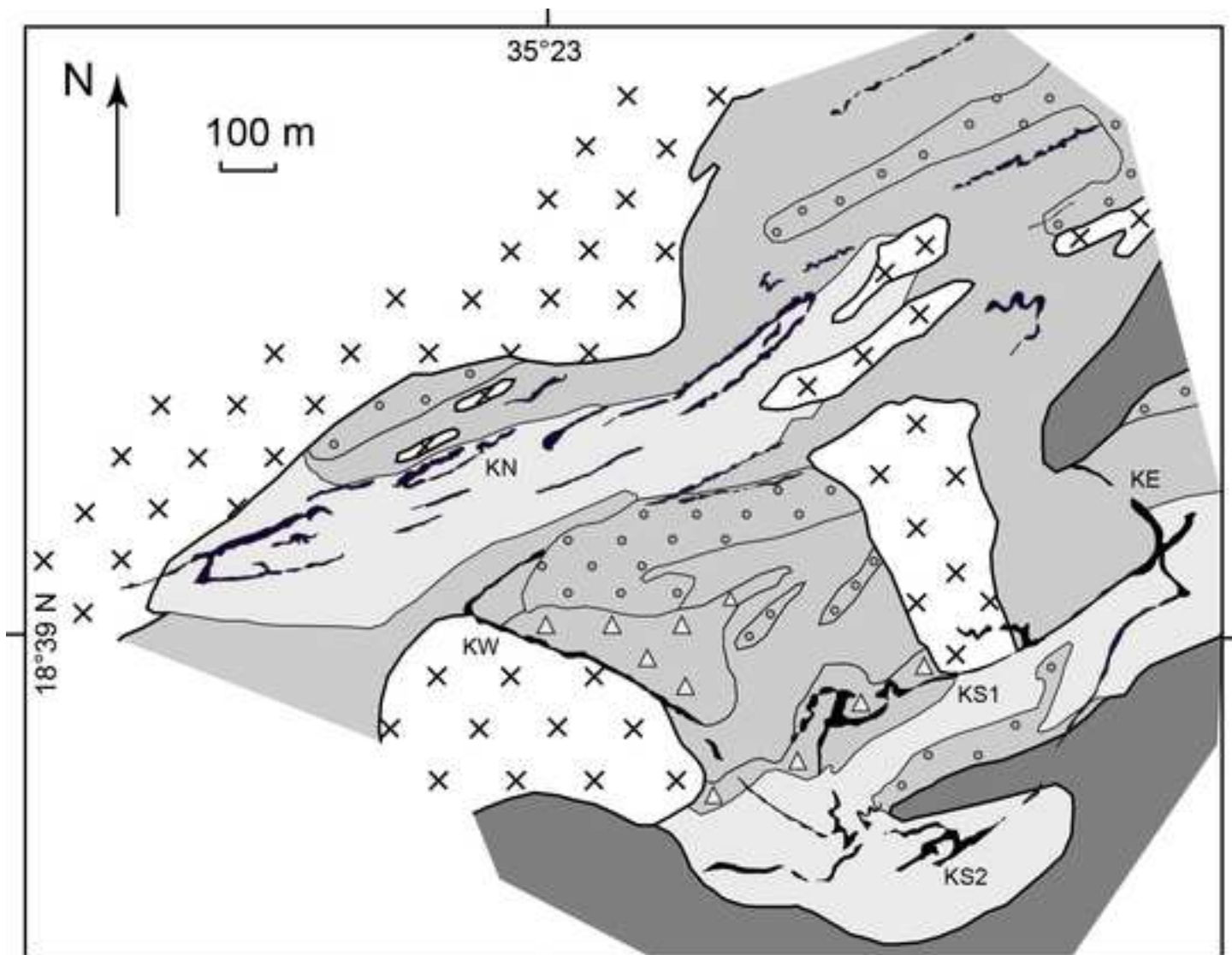
VHMS Major phase
 Minor phase

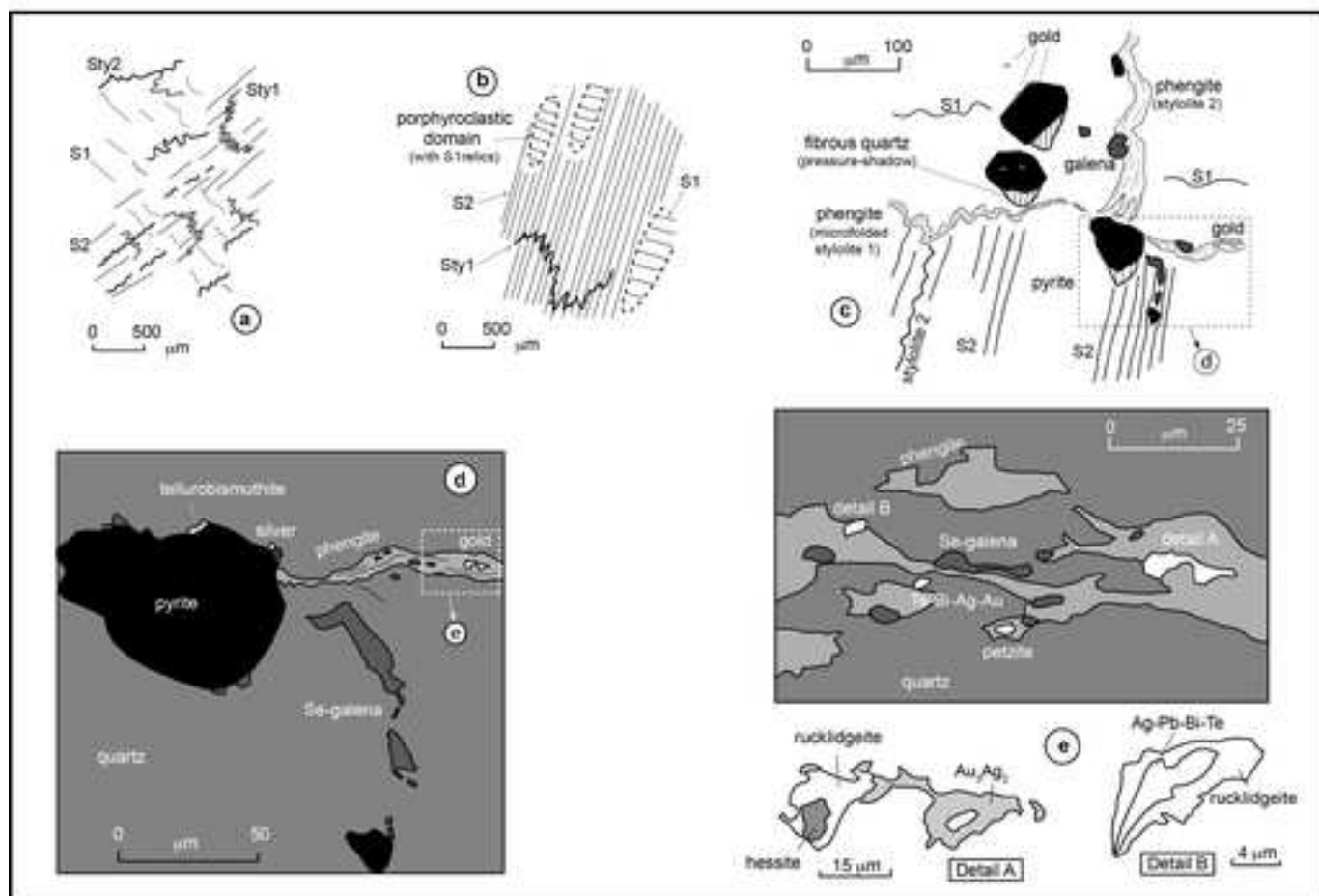
Kamoeb

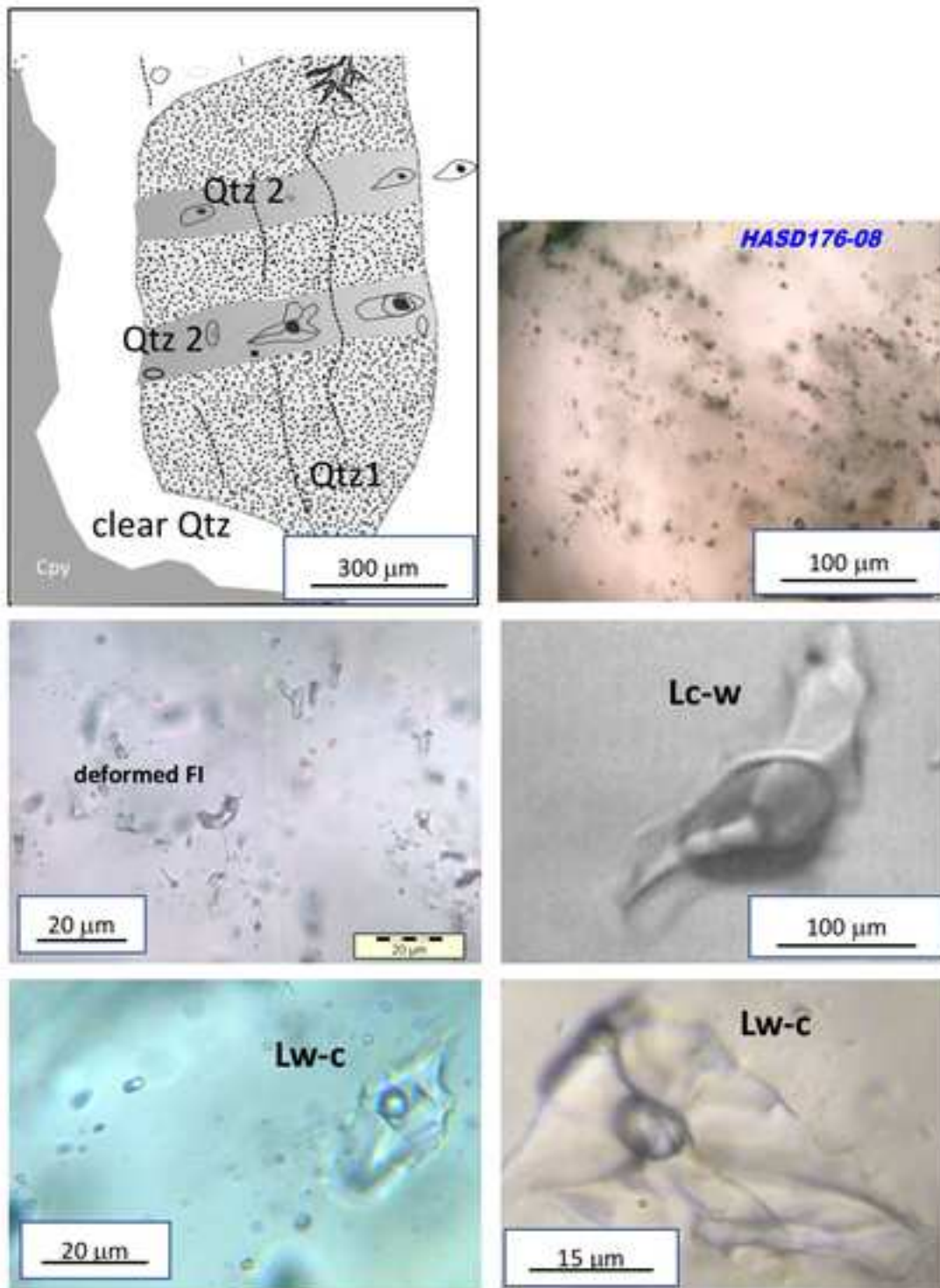


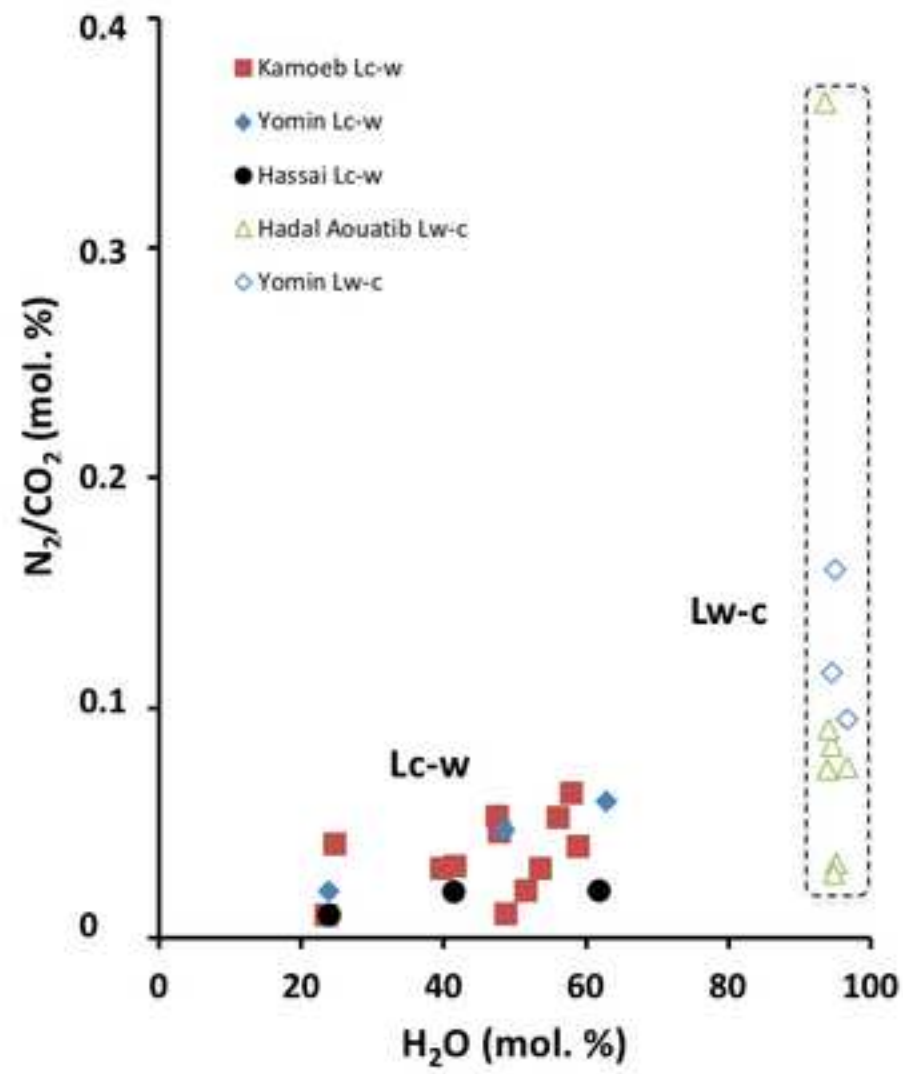


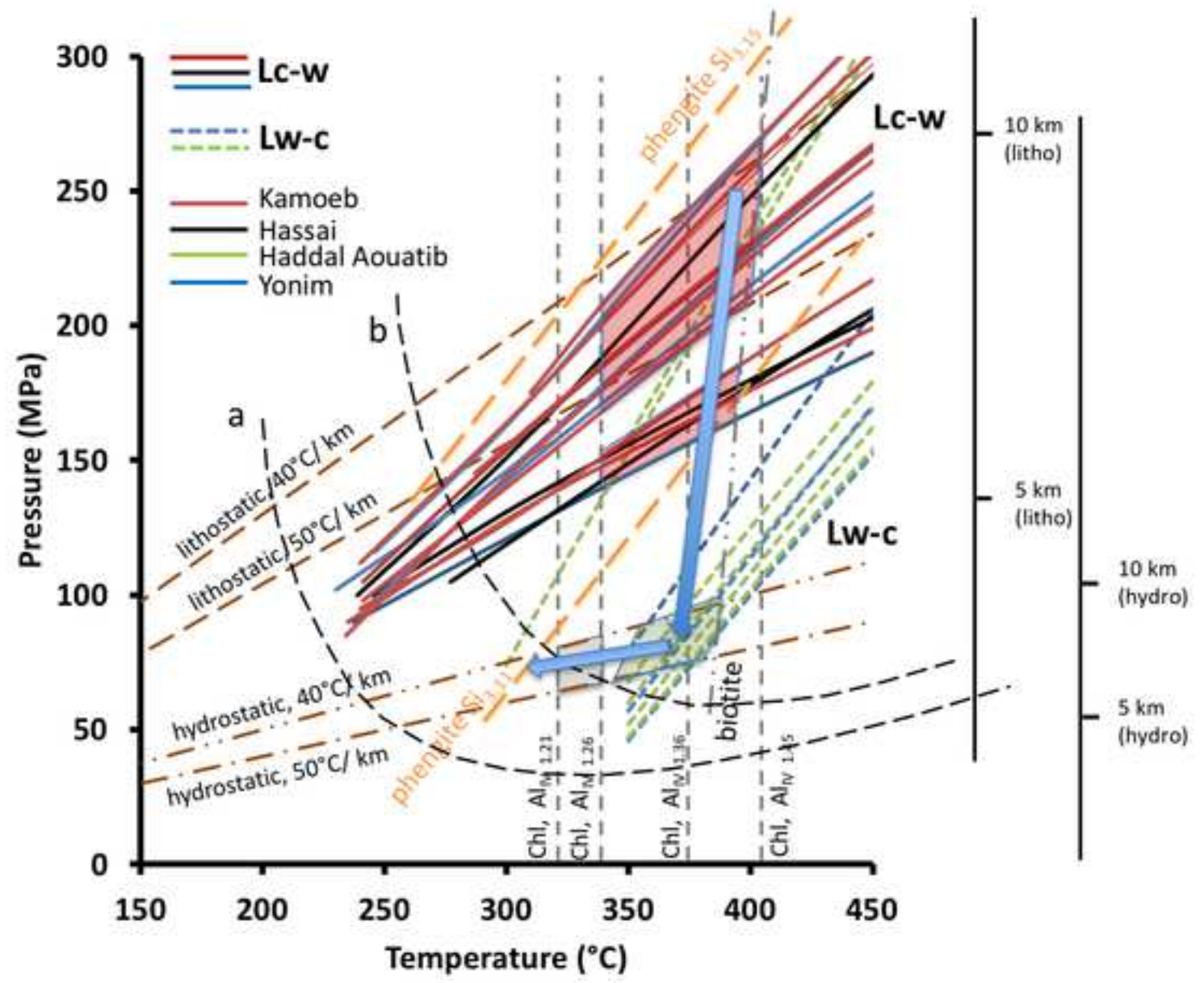


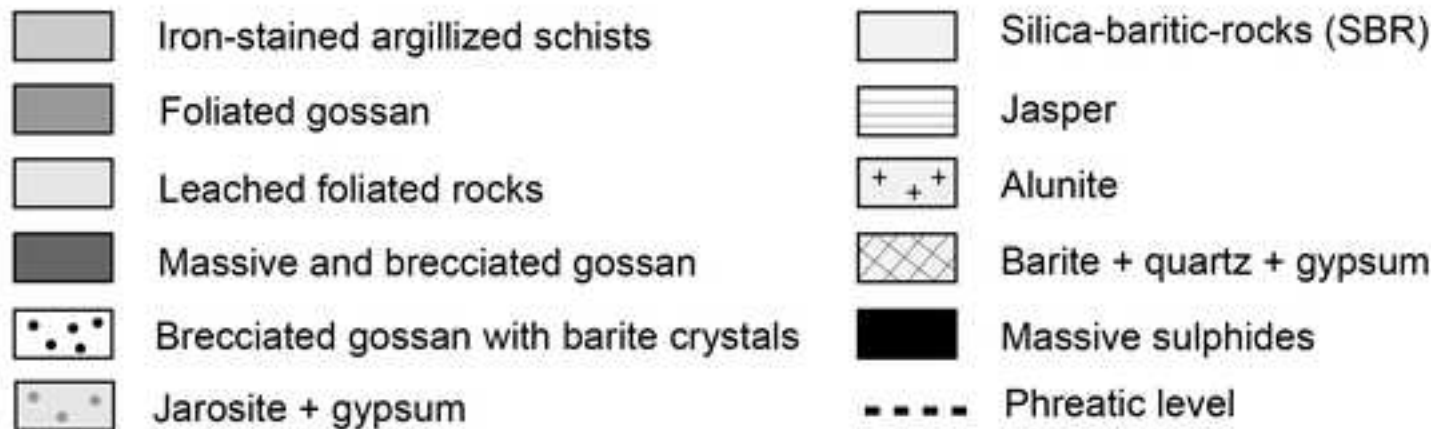
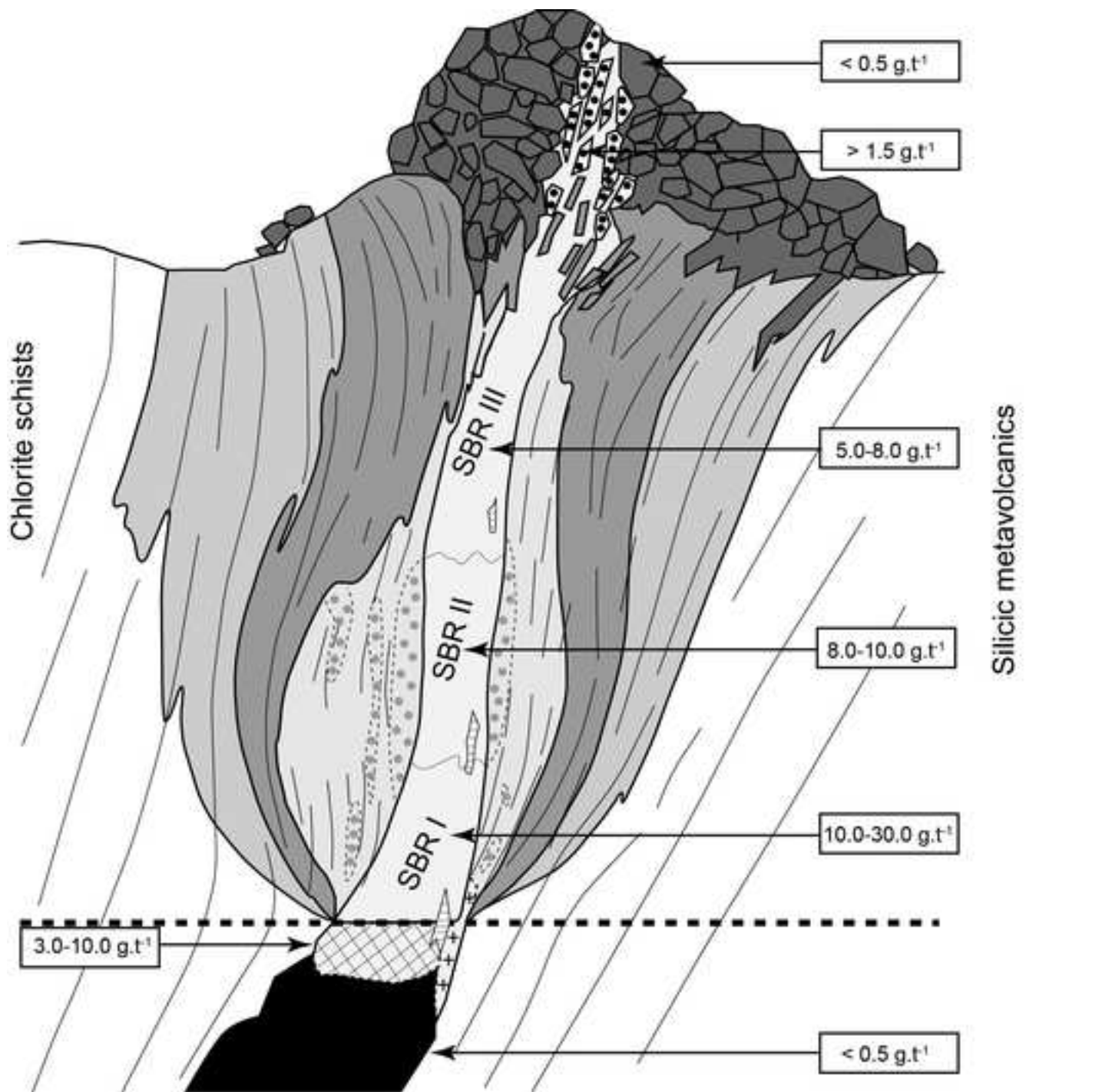












	Unit A			Unit B1			Unit B2		Unit C1		Unit C2				Unit C3		G1 Suite					G2 Suite				
	NTALGST-173	List-02	TAL04	KAM-03	DIM4-01	GANGT-17	HAE-CD08	ODE-01	TIDTUAAND-11	ADS03	A3	TAL-A6	ADAM01	GAN01	RUK4-02	ODE-03	SWKAM137	BAD-25	BIRAJM8	EKITL-2	UMHASH-7	TOMB-01	HADND77-99	KAMLG-131	ADAMG-12	LGRAN
SiO ₂	51.69	47.87	44.71	54.74	66.51	57.63	56.32	55.58	68.18	77.51	73.14	74.99	75.82	88.18	71.69	65.88	50.82	59.21	71.65	76.46	75.79	72.32	49.57	75.34	59.37	67.37
TiO ₂	0.76	0.52	0.33	0.17	0.21	1.39	0.37	0.40	0.34	0.12	0.30	0.36	0.16	0.07	0.20	0.65	0.16	0.48	0.16	0.08	0.12	0.19	0.94	0.16	0.52	0.53
Al ₂ O ₃	13.67	14.84	12.63	14.00	11.48	11.78	15.25	15.47	11.05	9.32	10.80	11.00	10.73	6.98	11.83	12.38	14.11	16.12	13.10	12.02	12.00	11.67	13.47	11.40	16.25	15.65
Fe ₂ O ₃	11.29	9.72	13.98	11.01	6.29	18.20	9.54	10.78	6.42	4.45	5.10	5.03	3.68	1.43	4.97	9.27	9.65	6.87	4.23	1.88	3.51	3.19	12.52	3.33	6.86	4.29
MnO	0.17	0.11	0.42	0.19	0.10	0.17	0.30	0.17	0.17	0.04	0.11	0.06	0.05	<DL	0.08	0.19	0.18	0.12	0.06	0.04	0.06	0.16	0.20	0.08	0.14	0.09
MgO	7.29	8.79	11.24	6.43	4.19	5.29	2.93	7.39	1.47	0.30	0.86	1.29	3.11	<DL	0.90	2.17	9.45	2.86	1.03	0.57	1.88	1.13	4.24	0.61	3.08	1.39
CaO	9.69	7.56	3.95	8.20	4.78	0.48	5.61	1.43	5.58	0.37	2.33	0.69	<DL	<DL	2.14	1.19	12.30	.88	3.35	0.82	0.75	3.97	7.65	2.74	6.52	3.12
Na ₂ O	3.60	2.57	<DL	2.69	5.20	<DL	3.60	4.71	0.77	4.37	3.25	5.06	0.31	0.34	4.02	5.54	0.82	3.29	4.13	4.54	3.69	2.76	3.27	3.49	3.10	4.60
K ₂ O	0.06	<DL	0.16	0.66	<DL	0.18	0.144	<DL	1.34	0.30	0.51	<DL	1.87	1.48	0.70	0.11	<DL	1.03	0.37	2.04	0.51	1.38	0.59	0.81	1.06	1.76
P ₂ O ₅	0.08	0.07	0.05	0.05	0.07	0.31	0.11	0.08	0.13	<DL	0.10	0.09	<DL	<DL	0.06	0.12	0.5	0.12	0.06	0.04	<DL	0.05	0.19	0.07	0.12	0.23
LOI	2.05	8.84	11.21	2.23	1.98	5.36	4.65	4.27	4.92	3.49	3.51	1.53	3.46	1.54	2.36	1.98	3.20	2.45	1.37	1.30	1.69	2.82	7.66	1.50	2.05	0.75
Total	100.35	100.91	98.68	100.37	100.82	100.79	98.83	100.26	100.38	100.29	100.00	100.07	99.19	100.03	99.94	99.47	100.74	99.41	99.52	99.79	99.99	99.63	100.31	99.53	99.43	99.76
Ba	86	101	267	409	130	108	527	105	2539	263	179	850	13,560	1,042	220	98	92	348	234	1,132	269	415	371	396	410	579
Rb	0.7	<DL	2.0	13	<DL	1.6	2.5	<DL	25	5.1	9.3	0.6	21	24	11	1.8	0.5	18	4.3	28	6.4	23	11	18	14	41
Sr	121	68	29	123	130	12	146	43	145	45	87	32	136	44	45	28	84	537	191	43	67	94	151	116	437	348
Cu	25	<DL	155	89	23	30	858	255	37	217	27	29	27	42	272	384	76	70	11	13	5	8	114	125	50	5
Pb	1	2	12	6	7	5	8	3	20	70	9	16	9	339	7	4	5	8	5	8	4	12	7	4	6	7
Zn	62	47	367	98	50	124	1172	408	134	149	87	78	138	21	93	798	99	67	23	29	21	39	129	129	82	66
As	1.9	5.8	4.9	1.5	6.2	1.1	6.4	2.1	2.7	15	2.0	3.7	1.6	37	2.5	<DL	6.8	6.4	1.3	<DL	<DL	1.4	4.2	2.2	1.6	<DL
Cr	273	322	348	254	351	29	79	131	60	100	107	95	75	97	134	12	516	78	91	116	89	62	12	244	78	97
Ni	75	94	48	52	40	14	21	69	6	14	8	9	9	6	7	19	64	18	8	19	7	10	12	14	18	9
Co	40	40	42	46	22	36	41	33	6	6	4	6	5	0.4	8	18	55	18	9	4	5	5	41	137	18	7
V	231	232	229	270	132	197	183	262	14	19	21	21	8	24	43	134	190	169	44	15	17	25	383	16	150	39
Nb	0.43	0.33	0.34	0.46	1.11	2.51	0.42	0.74	0.96	0.95	0.97	1.10	1.22	0.43	1.20	1.01	0.19	1.09	1.71	3.53	2.59	1.73	0.56	1.75	1.09	2.92
Ta	0.10	0.03	0.02	0.03	0.08	0.21	0.03	0.06	0.07	0.07	0.06	0.08	0.08	0.02	0.08	0.08	0.07	0.09	0.12	0.28	0.17	0.14	0.11	0.76	0.09	0.16
Zr	33	22	9	10	42	198	14	30	34	47	32	45	64	5	36	48	8	68	36	129	100	62	24	60	59	188
Hf	1.0	1.9	0.2	0.3	1.3	5.6	0.5	0.9	1.1	1.5	1.0	1.4	2.0	0.2	1.1	1.6	0.2	2.3	1.2	4.7	3.3	1.9	0.8	1.5	2.0	5.0
Th	0.13	0.09	0.16	0.29	0.93	0.37	0.25	0.57	0.76	0.86	0.80	0.73	1.11	0.09	0.96	0.69	0.14	4.07	1.02	3.63	1.94	1.54	0.33	1.65	1.40	4.60
U	0.09	0.04	0.22	0.33	0.31	0.23	0.22	0.16	0.28	0.67	0.29	0.32	0.53	0.27	0.37	0.24	0.07	1.65	0.74	1.20	0.67	0.63	4.01	0.54	0.71	2.65
Y	20	17	6	7	17	73	13	16	21	24	19	18	18	2	18	27	5	15	16	76	52	19	15	25	19	12
La	1.43	0.11	1.62	1.92	4.89	3.57	1.77	3.03	4.07	3.62	3.85	3.21	3.64	0.32	4.34	4.63	0.72	11.8	4.59	14.3	11.2	6.92	2.61	6.81	8.86	25.0
Ce	4.30	2.79	2.91	4.38	10.8	11.0	4.04	6.02	9.05	7.97	8.73	7.50	11.1	0.74	9.38	9.79	1.78	25.2	9.68	32.2	25.6	14.9	6.36	15.0	20.3	51.2
Pr	0.74	0.52	0.39	0.54	1.40	2.27	0.62	0.93	1.28	1.19	1.22	1.11	1.26	0.12	1.28	1.49	0.24	3.22	1.26	4.34	3.60	1.98	0.99	1.95	2.85	6.02
Nd	4.22	3.00	1.87	2.36	6.20	12.9	3.23	4.33	6.04	6.11	5.76	5.53	6.01	0.58	5.85	7.50	1.19	13.6	5.49	20.0	17.1	8.47	5.21	8.72	12.6	22.4
Sm	1.72	1.17	0.59	0.69	1.76	5.24	1.16	1.33	1.92	2.28	1.77	1.78	1.87	0.22	1.78	2.50	0.40	2.92	1.58	6.24	5.26	2.30	1.68	2.53	3.01	4.08
Eu	0.72	1.79	0.27	0.24	0.46	1.04	0.45	0.48	0.69	0.63	0.53	0.53	0.22	0.05	0.51	0.83	0.16	0.78	0.47	0.35	1.35	0.62	0.56	0.56	0.81	1.10
Gd	2.62	14.0	0.81	0.84	2.09	7.89	1.60	1.75	2.48	3.12	2.26	2.30	2.22	0.25	2.17	3.37	0.55	2.58	1.92	8.18	6.53	2.56	2.28	3.21	2.96	3.17
Tb	0.49	0.37	0.14	0.16	0.37	1.58	0.28	0.32	0.45	0.59	0.41	0.42	0.41	0.04	0.39	0.60	0.10	0.41	0.34	1.56	1.17	0.44	0.38	0.58	0.49	0.43
Dy	3.36	209	0.95	1.12	2.54	11.0	1.96	2.23	3.22	4.03	2.88	2.85	2.95	0.33	2.78	4.15	0.73	2.51	2.43	11.1	7.94	2.99	2.54	3.91	3.07	2.35
Ho	0.73	0.72	0.21	0.26	0.57	2.45	0.43	0.50	0.72	0.86	0.65	0.63	0.68	0.08	0.62	0.92	0.16	0.53	0.54	2.51	1.76	0.65	0.55	0.88	0.65	0.43
Er	2.20	2.64	0.63	0.82	1.73	7.48	1.31	1.49	2.23	2.56	2.02	1.92	2.19	0.25	1.93	2.78	0.51	1.56	1.71	8.02	5.37	2.01	1.60	2.76	1.94	1.16
Tm	0.34	0.28	0.10	0.13	0.28	1.14	0.20	0.23	0.36	0.39	0.33	0.30	0.37	0.04	0.31	0.44	0.08	0.24	0.28	1.34	0.84	0.32	0.25	0.45	0.31	0.17
Yb	2.27	1.87	0.68	1.00	1.94	7.74	1.43	1.58	2.49	2.66	2.31	2.15	2.87	0.35	2.19	3.04	0.58	1.69	2.04	9.76	5.88	2.26	1.67	3.21	2.19	1.13
Lu	0.36	1.08	0.11	0.17	0.32	1.23	0.23	0.26	0.41	0.40	0.39	0.36	0.49	0.06	0.36	0.48	0.10	0.28	0.35	1.61	0.97	0.38	0.27	0.52	0.36	0.19
SREE	40.21	239.34	11.28	14.63	35.35	76.83	18.71	24.48	35.41	36.41	33.11	30.59	36.28	3.43	33.89	42.52	7.30	67.32	32.68	121.51	94.57	46.80	26.95	51.09	60.40	118.83

FI type	Tm CO ₂ (°C)	Tm ice (°C)
Lc-w		
Kamoeb (n=66)	-58.3 to -56.8	-5.6 to -2.2
VHMS (n=118)	-59.7 to -57.2	-7.6 to -2.5
Lw-c		
VHMS (n=10)		-7.1 to -2.2

Tm clat (°C)	Th CO ₂ (°C) (L)	Th (°C)
2.6 to 7.9 1.8 to 8.3	16.3 to 26.0 22.8 to 29.7	220 to 310 237 to 278
1.9 to 8.0		260 to 355

	Microthermometry (°C)		
	Tm CO ₂	Tm ice	Tm clat
Lc-w			
<i>Kamoeb</i>			
KAME02-11	-57.1	-4.2	7.6
KAME02-12	-57.3	-4.3	3.3
KAME02-13	-57.0	-5.3	2.6
KAME02-14	-58.1	-2.2	4.8
KAME02-16	-57.1	-3.4	7.1
KAME02-18	-57.0	-3.0	7.1
KAME02-110	-57.2	-4.1	6.3
KAME02-111	-57.0	-5.1	7.2
KAME02-114	-57.8	-4.2	4.5
KAME02-23	-57.2	-5.6	7.5
KAME02-25	-57.0	-4.4	5.5
KAME02-26	-56.8	-3.1	7.9
<i>Yonim</i>			
YONM-202-N81	-56.9	-2.5	4.3
YONM-202-N82	-57.2	-4.7	6.4
YONM-202-N83	-57.2	-5.5	5.8
YONM-202-N84	-58.3		4.4
<i>Hassai</i>			
HASD176-081	-57.3	-3.6	4.2
HASD176-084	-56.8		7.7
HASD176-088	-57.3	-2.7	7.6
Lw-c			
<i>Hadal Ouatib</i>			
HDND77-111		-5.1	4.2
HDND77-112		-5.0	8.0
HDND77-113		-5.6	7.6
HDND77-114		-5.1	6.5
HDND77-117		-2.3	4.3
HDND77-118		-7.1	6.3
HDND77-119		-2.2	6.6
<i>Yonim</i>			

YONM 202-N2-1	-56.8	-7.1	
YONM 202-N2-2	-56.6	-3.6	1.9
YONM 202-N2-3	-56.6		1.7

		Raman data (mol.%)			Bulk composition	
Th CO ₂ (L)	Th	fvp (%)	ZCO ₂	ZN ₂	XH ₂ O	XCO ₂
23.2	241.5	75	97.0	3.0	48.8	49.3
20.1	248.3	65	95.0	5.0	56.2	38.5
23.6	234.1	70	98.0	2.0	51.7	44.3
21.8	190*	80	97.0	3.0	39.7	56.8
26.0	240.3	80	98.0	2.0	41.7	54.8
23.6	235.7	90	96.0	4.0	24.7	71.6
22.6	220.5	90	99.0	1.0	23.5	74.9
20.6	241.3	75	95.5	4.5	47.9	48.3
22.2	240.4	65	94.0	6.0	58.0	36.7
19.8	310.1	65	96.0	4.0	58.9	37.8
22.2	287.6	75	95.0	5.0	47.6	47.8
22.1	281.9	70	97.0	3.0	53.6	43.7
25.4	237.1	90	98.0	2.0	23.9	73.5
25.4	230.3	70	96.5	3.5	48.5	47.3
25.4	243.2	75	95.5	4.5	48.5	47.3
23.6	246.2	60	94.0	6.0	62.8	32.0
23.3	239.5	60	98.0	2.0	62.0	34.3
22.8	245.3	90	99.0	1.0	24.0	74.7
29.7	277.7	80	98.0	2.0	41.5	55.4
	285*	30	89.0	11.0	95.2	3.1
	341.9	20	89.5	10.5	94.9	3.6
	260.5	30	86.0	14.0	94.0	4.1
	354.7	30	83.0	17.0	94.6	3.6
	275.7	30	84.5	15.5	96.7	2.7
	278.3	20	80.0	20.0	94.1	3.3
	345.9	20	50.0	50.0	93.7	3.3

250 *	30	65.0	35.0	95.1	2.5
240 *	35	71.5	28.5	94.6	2.6
220*	25	70.5	29.5	96.7	2.1

tion (mol. %)	
XN2	XNaCl
0.5	1.4
2.0	3.3
0.9	3.1
1.7	1.8
1.7	1.8
2.9	0.8
0.7	0.9
2.2	1.6
2.3	3.0
1.5	1.8
2.5	2.1
1.3	1.4
1.5	1.1
2.2	2.0
2.2	2.0
1.9	3.3
0.7	3.0
0.7	0.6
1.1	2.0
0.1	1.6
0.1	1.4
0.3	1.6
0.3	1.5
0.2	0.4
0.3	2.3
1.2	1.8

0.4	2.0
0.3	2.5
0.2	1.0