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- Sol Hamed ophiolitic has a harzburgite composition as protolith

- The protolith originated as forearc mantle

- Protolith alteration resulted from two stages of fluid-rock interaction

- First stage is due to infiltration of CO_2-rich fluid during isobaric cooling

- Second stage of fluid-rock interactions took place through prograde metamorphism

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6   **Thermodynamic modelling of Sol Hamed serpentinite, South Eastern Desert of**

7   **Egypt: implication for fluid interaction in the Arabian-Nubian Shield ophiolites**

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9

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Key-words: Arabian-Nubian Shield; Forearc peridotite; Ophiolites; Carbonatization; Thermodynamic modelling; T-XCO$_2$.

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1. Introduction

Arabian-Nubian Shield (ANS) in Northeast Africa and West Arabia is the largest tract of juvenile continental crust of Neoproterozoic age on Earth (Patchett and Chase 2002; Stern et al. 2004). This crust was generated when arc terranes were created within and around the margins of the Mozambique Ocean, which formed in association with the breakup of Rodinia ~ 800–900 Ma (Stern 1994; Hassan et al. in-review). These crustal fragments collided as the Mozambique Ocean closed around 600 Ma (Meert 2003), forming arc-arc sutures, composite terranes, the Arabian-Nubian Shield (Fig. 1), and the larger collisional belt known as the East-African Orogen (Stern 1994; Kusky et al. 2003, Stern et al. 2004). Due to this collision processes a supercontinent variously referred to as Greater Gondwanaland (Stern 1994), Pannotia (Dalziel 1997) or just Gondwana (e.g. Abu-Alam et al. 2013) was formed.

Ophiolitic rocks are remarkably abundant in the ANS. They are scattered across most of the shield, over a distance of ~ 3000 km from the farthest north (Gebel Ess) almost to the equator, and from Rahib in the west to Gebel Uwayjah (45º E) in the east (Fig. 1). The abundance of the ophiolites is a further indication that the Arabian-Nubian Shield was produced by processes similar to those of modern plate tectonics (Stern et al. 2004).

The ophiolitic rocks of Eastern Desert (ED) of Egypt (Fig. 1) are interpreted to be formed in a suprasubduction zone (SSZ) (e.g. Ahmed et al. 2006; Azer and Stern 2007) which operated at forearc convergent margin (e.g. Stern et al. 2004; El-Gaby 2005). The East- and West-Gondwana collision led to obduction of the SSZ ophiolitic rocks over a continental margin (Akaad and Abu El Ela 2002; El-Gaby 2005) of the West-Gondwana

3

70   (Abd El-Rahman et al. 2009). Subduction was active while the process of ophiolitic

71   overthrusting was operative along thrust planes (Kröner et al. 1987; Stern 1994).

72   The ophiolite obduction and overthrusting were associated with fluid-rocks interaction

73   and hence alteration, serpentinization and metasomatism took place (e.g. Hamdy et al.

74   2013) but it is often not known whether this alteration occurred before, during or after the

75   obduction (Stern et al. 2004). On other hand, much doubt exists around origin and

76   composition of the fluid during the alteration process. Some authors suggested that the

77   alteration of ultramafic rocks – the dominant component of the ANS's ophiolites –

78   occurred by interaction with hot fluid during seafloor weathering (e.g. Lebda 1995; Li

79   and Lee 2006). Other authors believed that the alteration took place by infiltration of

80   metamorphic and hydrothermal fluid along major tectonic fractures during or after rock

81   exhumation (e.g. Hyndman and Peacock 2003; Hamdy 2004; Hamdy and Lebda 2007).

82   Sol Hamed ophiolite in the southeastern Egypt and northeastern Sudan (Fitches et al.

83   1983) differs from other ophiolites further north in the ED of Egypt in being an elongated

84   belt defining a near-source tectonic facies (Abdelsalam and Stern 1996). To the north,

85   ophiolites occur in tectonic mélanges or as olistostromal debris, indicating a distal

86   tectonic facies. This interpretation implies the ophiolitic rocks north of the Sol Hamed

87   represent a far-travelled ophiolitic nappe, transported to the north away from its

88   corresponding suture. Thus the Sol Hamed rocks represent an uncommon example in the

89   Eastern Desert that might be less deformed by the movement along faults that occurred

90   after the closure of the Mozambique Ocean. In this work, petrological relationships,

91   mineral chemistry, geochemistry and thermodynamic modelling are described and

92   applied to rocks from Sol Hamed ophiolite (Fig. 2). The results help to more clearly

4

93 define the nature of the fluid-rock interactions process occurred in the intraoceanic

94 collision of the ANS.

95

96 **2. Geological setting**

97 Many of the ultramafic outcrops in the Arabian-Nubian Shield are detached, scattered

98 and isolated (Fig. 1) due to intrusion of syn- and post-tectonic plutons. Gass (1977) noted

99 that these ultramafic bodies have tectonic contacts with other Pan-African rocks. Some of

100 these ultramafic are recognized as ophiolites, representing obducted fragments of an

101 oceanic lithosphere that existed between the Proterozoic island arcs (Gareson and

102 Shalaby 1976; El-Ramly et al. 1993). Dixon (1979) estimated that the ultramafic bodies

103 account for 5.3% of all Precambrian outcrops in Egypt. Serpentinized ultramafics

104 constitute the major part of these ophiolite complexes (Ahmed et al. 2001).

105 The Sol Hamed ophiolite is a part of Allaqi-Heiani-Onib-Sol Hamed-Yanbu arc-arc

106 suture (Abdelsalam and Stern 1996; Abdelsalam et al. 2003). This arc-arc suture is

107 considered − along with the Ariab-Nakasib-Thurwah-Bir Umq suture farther south in

108 Arabia and Sudan (Johnson et al. 2004) to be one of the two longest and most complete

109 Neoproterozoic ophiolite-decorated sutures in the ANS (Azer et al. 2013). Stern et al.

110 (1990) proposed that the Allaqi-Heiani-Onib-Sol Hamed-Yanbu suture represents a south

111 verging nappe which was refolded around a subhorizontal east-west trending axes to

112 produce upright antiforms and late-stage southeast verging thrusts. Vergence of the

113 ophiolite nappe was used to infer a north dipping subduction zone along the line of a

114 suture which lies north of the Allaqi-Heiani-Onib-Sol Hamed-Yanbu ophiolite. Ali et al.

5

115   (2010) suggested two stages for the evolution of Allaqi-Heiani-Onib-Sol Hamed-Yanbu

116   suture (~810–780 Ma and ~750–730 Ma).

117   The ultramafic rocks of Sol Hamed (Fig. 2) are composed of serpentinized peridotite

118   forming the base of a dismembered ophiolitic sequence that comprises also metagabbros,

119   pillow lavas and pelagic sediments (Abu El-Laban 2002). They occur as sheets and lenses

120   enclosed within mélange matrix comprises island arc volcanosedimentary assemblages,

121   gneiss and syn- to post-orogenic intrusions (e.g. Kröner et al. 1987; Greiling et al. 1988;

122   Taylor et al. 1993; Abd El-Naby and Frisch 2002; Kusky and Ramadan 2002;

123   Abdelsalam et al. 2003; Zoheir and Klemm 2007; Ali et al. 2010). Serpentinized

124   ophiolitic peridotites form ridges about 20 km long and about 0.4–1.8 km wide, elongated

125   in NE-SW direction. They are mostly massive but are sometimes brecciated and

126   fragmented along their contacts with the mélange matrix. Some portions contain relicts of

127   primary minerals and others are extremely altered, especially along thrusts and shear

128   zones, with the development of talc, talc-carbonate and reddish brown quartz-carbonate

129   rock (listwaenite).

130   Chromitite deposits occur mainly as lenticular bodies of variable dimensions up to 25 m

131   length × 6 m width, trending ENE-WSW. Thick pods are abundant in serpentinites that

132   are mostly derived from dunite. Micro-lenses and thin planar segregations occur in the

133   serpentinized peridotite. Gradual contacts between massive ore and serpentinized dunite

134   over a meter-range are frequently observed. A typical contact shows gradation from fine-

135   grained disseminated chromite in the dunite through nodular, to massive coarse-grained

136   chromite ore. The highly deformed chromite bodies are the most abundant. Magnesite

137   veins occur in shear zones close to the eastern periphery of the serpentinite rocks. Hamdy

138   (2007) based on the C-O isotopes of these veins estimated that carbon was supplied from

139   both geothermal fluids (giving magnesite with $\delta^{13}C$ values from -2.06 to -4.34‰ VPDB)

140   and metamorphic carbonaceous sediments (giving magnesite with $\delta^{13}C$ values from -9.44

141   to -10 ‰ VPDB).

142

143   **3. Petrography**

144   Variable degrees of alterations are observed in the studied ultramafic rocks. Original

145   peridotite minerals have been preserved (Table 1) in partly altered peridotites. The

146   dominant serpentine mineral is lizardite, whereas chrysotile is subordinate. The lizardite

147   forms psuedomorphic mesh and bastite textures after olivine and orthopyroxene and

148   sometimes occurs as interlocking and penetrating grains (non-pseudomorphic). The

149   chrysotile occurs as cross fiber veins traversing the lizardite matrix. Serpentine minerals

150   appear to be accompanied by shedding of fine-grained magnetite, which concentrates in

151   veins cutting zoned chromite (Fig. 3a) or along relict pyroxene cleavages. Pyroxene

152   relicts occur as inclusions in anthophyllite (Fig. 3a). The anthophyllite is a common

153   replacement mineral of orthopyroxene, where it initially grows along cleavage planes and

154   eventually replaces the whole grain. Talc is not abundant in the studied serpentinites. It

155   forms fine shreds, dense fibers and medium grained flaky crystals (0.01–0.04 mm).

156   Perfect cleavage, straight extinction and high interference colours are characteristic

157   features of the talc. The talc is pseudomorphic after anthophyllite. It is homogenous and

158   commonly associated with the alteration of orthopyroxene. All serpentinite samples

159   contain zoned-chromite (Fig. 3b) and sulphide grains. Chromite occurs as disseminated

160   subhedral and anhedral crystals of reddish brown colour. Some chromite grains look

161   homogeneous in reflected light.

162

163   **4. Mineral chemistry**

164   Different mineral phases were examined in the Institute of Geological Sciences of Polish

165   Academy of Sciences (IGS-PAS). The electron microprobe analyses were carried out by

166   JEOL-JXA-840A scanning electron microscope equipped with Link Analytical AN-

167   1000/855 energy dispersive X-ray spectrometer. The analytical conditions were 15 kV

168   accelerating voltage and 35 nA beam current. Mineral formula and activity of the end-

169   members were calculated by AX program (http://www.esc.cam.ac.uk/research/research-

170   groups/holland/ax). The chemical formula of the serpentine minerals was calculated

171   based on 28 oxygen atoms and ignoring the H_2O, pyroxene formula based on 6 oxygen

172   atoms. 11 oxygen atoms and ignoring the H_2O were used to calculate talc formula, 24

173   oxygen atoms were used for the spinel while the carbonate formula was calculated based

174   on 2 cations. The mineral abbreviations which will be used in the following sections are

175   from Holland and Powell (2011).

176   CaO content is below 0.03 wt% in the orthopyroxene and FeO content is in the range of

177   4.85 – 5.09 wt% while MgO content is around 35 wt% (Table 2). This reveals that the

178   main pyroxene end-member is enstatite. SiO_2 content of the serpentine ranges between

179   40.62 and 44.54 (Table 2). Al_2O_3 is in the range of below the detection limit up to 1.79

180   wt%. FeO shows a wide range (0.83 – 6.21 wt%). MgO ranges between 34.37 to 39.35

181   wt%. The MgO and the FeO ranges indicate ionic substitution between Fe^{2+} and Mg^{2+}.

182   Figure 4a shows limitation of the substitution between Al^{3+}, Mg^{2+} and Si^{4+} in the

183 serpentine crystals. FeO and Cr_2O_3 contents in lizardite increase (0.94 to 6.21 wt% and

184 from below detection limit to 0.36 wt% for FeO and Cr_2O_3, respectively) distinctly with

185 increasing degree of alteration from partly to completely serpentinized rocks (Table 2).

186 Chrysotile shows that Al and Cr are relatively immobile during recrystallization of

187 lizardite and therefore remain in their original crystal lattice.

188 Low Al_2O_3 and TiO_2 contents in talc chemistry reveal limitation in substitution between

189 Si, Ti and Al. The main ionic substitution is between Mg^{2+} and Fe^{2+} (Table 3). The

190 activities of talc and Fe-talc end-members are in the range of 0.68-0.85 and 0.00013-

191 0.00061, respectively. Table (4) shows chemical analyses of the carbonate minerals. The

192 high concentrations of MgO and FeO (35.53-40.14 and 8.46-14.1 wt%, respectively)

193 indicate high activity of the magnesite and the siderite end-members. The CaO content is

194 in the range of 0.04 to 0.27 wt% revealing low activity of the calcite.

195 Three compositional zones are distinguished for the spinel minerals. The composition of

196 core, intermediate and rim zones is given in Table (5) and plotted in Al-Cr-Fe^{3+} triangle

197 of Stevens (1944) (Fig. 4b). Cores and intermediates zones have aluminian chromite to

198 ferritchromite composition. Composition of the outer rim is Cr-magnetite which is nearly

199 devoid of Al and lie along the Cr–Fe^{3+} sideline (Fig. 4b). The studied spinels show

200 metamorphic conditions correspond to that of the upper greenshist to the transitional

201 greenschist-amphibolite facies (Fig. 4b).

202 The variation in the spinel composition can be interpreted as a result of chemical

203 alteration under hydrothermal conditions (Abzalov 1998; Barnes 2000; Proenza et al.

204 2004). The alteration is accompanied by decrease in Al, Mg and Cr contents and

205 consequence increase in Fe^{3+} and Fe^{2+}. Apparently with the increasing of the alteration,

9

206 Fe releases from olivine and orthopyroxene and Cr releases from chromite and are

207 accommodated in the serpentines. In addition, the wide compositional variation

208 (particularly in Al_2O_3) reflects temporal and/or spatial variations in melt types (boninitic

209 and tholeiitic) that were generated from, and emplaced in subarc mantle domains in a

210 suprasubduction zone environment (Hamdy and Lebda 2011).

211

212 **5. Whole-rock chemistry**

213 Representative bulk rock chemistry of Sol Hamed serpentinite is given in Table (6).

214 Chemical analyses of major and some trace elements were carried out at the

215 geochemistry labortory of the IGS-PAS. Concentrations of major and trace elements were

216 determined after microwave-assisted acid digestion with atomic absorption

217 spectrophotometer (AAS-PU 9100xUNICAM). Before digestion samples were heated to

218 1100 to determine loss on ignition (LOI). Analytical precession was better than 0.5% for

219 major elements and 4 ppm for trace elements.

220 Due to the almost complete serpentinization of some of the Sol Hamed peridotites, modal

221 compositions could not be determined. Therefore, normative compositions were

222 calculated from anhydrous analyses using the CIPW norm, assuming a Fe_2O_3/FeO ratio

223 of 0.2 (Melcher et al. 2002), and plotted in Streckeisen (1976) classification diagram (not

224 shown). The normative contents of olivine, orthopyroxene, and clinopyroxene of the

225 studied Sol Hamed serpentinites classify them as harzburgites. Trace element values are

226 typical of residual mantle (e.g. high Cr (2.696–2.742 ppm), Ni (1.650–2.381 ppm) and Co

227 (116.20-166.79 ppm)). In contrast, the contents of Ba, Pb, Sr and, Li are highly

228 concentrated compare to depleted and pristine mantle peridotites (McDonough and Sun

229 1995). This enrichment in the fluid-mobile elements may be directly related to the

230 serpentinization process or due to metasomatism by subduction-related fluids (Hamdy et

231 al. 2013).

232

233

234 **6. Discussion**

235 *6.1. Origin and tectonic setting of the serpentinite protolith*

236 Earth contains two main shallow mantle domains: sub-oceanic lithosphere and sub-

237 continental lithosphere. The Sol Hamed harzburgite falls within the oceanic array (Niu

238 2004) in MgO/SiO_2–Al_2O_3/SiO_2 space (Fig. 5). The oceanic array is parallel to the

239 terrestrial array but offset to lower MgO/SiO_2 values, presumably due to loss of MgO

240 during low-temperature seafloor weathering and not due to the serpentinization process

241 itself (Snow and Dick 1995; Niu 2004). Oceanic peridotites may originate in a variety of

242 tectonic environments including mid-ocean ridge (MOR), suprasubduction zone (SSZ)

243 and rifted margins settings. We term these suprasubduction zone (SSZ) peridotites

244 (Pearce et al. 1984); a group that incorporates peridotites from both island arcs and

245 spreading centers above subduction zones. These discrete genetic types are distinct in

246 mineralogical and geochemical characteristics of mantle residues. Composition of the

247 unaltered accessory spinel is extensively used as a petrogenetic and geotectonic indicator

248 (e.g. Barnes and Roeder 2001). Chromium numbers $[Cr/(Cr+Fe^{3+}+Al)]$ higher than 0.6

249 are usually restricted to subduction-related rocks (Dick and Bullen 1984). Ishii et al.

250 (1992) used the Mg# $[Mg/(Mg+Fe^{2+})]$ and Cr# of the spinel to discriminate between

251 peridotites from MOR, forearc and back-arc settings. Spinels from the Sol Hamed

252   serpentinites lie in the chemical space of the forearc peridotite (Fig.6) and distinctly

253   higher than spinels from MOR and back-arc basin in the Cr#. This indicates that the Sol

254   Hamed serpentinites represent a fragment of oceanic lithosphere that has been

255   incorporated above subduction zone in a forearc.

256   Hellebrand et al. (2001) tested which trace elements correlate with major element

257   indicators of partial melting in central Indian ridge peridotites. The most common of

258   these is the Cr# in spinel. They found a well-defined correlation between moderately

259   incompatible elements, such as HREEs in clinopyroxene with spinel Cr#. Hellebrand et

260   al. (2001) developed an empirical equation ($F = 10 \ln (Cr\#) + 24$) to estimate the degree

261   of melting F (in percent) as a function of spinel Cr#. Using the equation of Hellebrand et

262   al. (2001), the estimated melting in the studied peridotites ranges from 20 to 22%.

263

264   *6.2. Thermodynamic modelling*

265   All the thermodynamic calculations in the following sections were calculated by

266   THERMOCALC (Powell and Holland 1988), PerPle_X (Connolly 1990) and using the

267   internally consistent dataset of Holland and Powell (2011). Lizardite bearing reactions

268   which were proved experimentally (i.e. liz = br atg (Evan 2004), liz = chr (Chernosky

269   1975), liz = ta fo clin H_2O (Caruso and Chernosky 1979)) will be only used (Fig. 7).

270   Figure 7 shows a *P-T* grid in the system CFMASH for the following end-members: atg,

271   chr, en, fs, di, hed, fo, fa, anth, tr, clin, ta, sp, herc, mgts, fta, br, H_2O. Activity of the H_2O

272   is imposed to be the unity therefore all the CO_2 bearing phases are not seen in this grid.

273   The *P-T* grid shows forty six univariant equilibria, five invariant points and three

274   experimental lizardite bearing reactions. All the H_2O bearing univariant reactions show

275   steep slope in the *P-T* space. Consequently these reactions can be used as temperature

276   indicators. Two water absent invariant points (508 ^{o}C-1.08 kbar and 542 ^{o}C-2.2 kbar)

277   involve reactions with notable change in the volume and can be used as pressure

278   indicators. For better reading to the *P-T* grid, only the interesting reactions are shown in

279   Figure 8 using two different scales for temperature axe.

280

281   ***Anthophyllite and talc formation***

282   One of the key petrographic features is the relation between pyroxene, anthophyllite and

283   talc. The anthophyllite is a common replacement mineral of orthopyroxene. The

284   anthophyllite can be formed due to eight metamorphic reactions (Fig. 8), however the

285   absence of clinochlore and the formation of the talc psuedomorphic after anthophyllite

286   make the only possibility to crystallize anthophyllite is due to breakdown of high grade

287   minerals (i.e. pyroxene). Two reactions can produce anthophyllite during a retrograde

288   path at relatively high pressure (> 1.7 kbar) and above the atg-chr-fs-di-hed-fa-tr-ta-herc-

289   fta-br-mgts invariant point, however, these reactions produce clinochlore in considerable

290   values. This makes reaction fa anth=fs en H_2O and the lower pressure part (< 1.7 kbar) of

291   reaction fo anth=en H_2O are preferred way to produce anthophyllite in the assumed fluid

292   composition.

293   Eight reactions can produce talc as a retrograde phase due to breakdown of high grade

294   assemblage that includes anthophyllite. Four reactions can be excluded since they contain

295   clinochlore as a reactant or a product. The petrographic observation "orthopyroxene

296   consumed due to talc growing" makes fa ta=fs anth H_2O, ta sp=H_2O mgts anth reactions

297   (Fig. 8) are the favorable equilibria to produce talc. The two talc producing reactions

298   have a temperature range 630 - 790 ^{o}C in a wide pressure condition. The pressure

299   conditions of anthophyllite formation (< 1.7 kbar) make the upper temperature limit of

300   talc producing reactions is below 730 ^{o}C. Other reactions can produce anthophyllite and

301   talc in the same pressure-temperature range but with different fluid compositions, these

302   reactions will be discussed in the fluid composition section.

303   Talc and anthophyllite formations indicate isobaric cooling path at pressure below 1.7

304   kbar and in a temperature range of 800-550 ^{o}C. The cooling path can be extended to a

305   lower temperature condition based on the presence of lizardite in the studied assemblage.

306   This assumption is in agreement with the greenschist facies conditions of the intermediate

307   zone of the spinel grains (Fig. 4b).

308   Stern et al. (2004) reconstructed the ophiolitic sequence of the Arabian-Nubian Shield

309   and concluded that the ophiolitic successions have crustal thicknesses of 2.5 to 5 km.

310   These crustal thicknesses are equivalent to pressure 0.7 and 1.4 kbar, respectively (Fig. 8)

311   assuming lithostatic conditions and a rock density of 2.84 x 10^3 kg/m^3 (Carlson and

312   Raskin 1984). This constrains pressure conditions of the formation of the anthophyllite

313   and talc process by 0.7-1.4 kbar (the retrograde path as shown by the black arrow in

314   Figure 8).

315

316   *Chrysotile formation and prograde metamorphism*

317   Presence of chrysotile fibers traversing the lizardite matrix indicates that the rocks passed

318   the reaction liz = chr (Fig. 8). Hamdy and Lebda (2007) showed that the magnetite rims

319   of the chromite grains of Malo Grim serpentinites (part of the Sol Hamed ophiolites)

320   equilibrated at a temperature range of 500-550 ^{o}C. These conditions are in agreement

321   with the composition of the rim zones of the spinel grains which show condition of

322   amphibolite facies (Fig. 4b). Neither petrographic observations nor mineral chemistry

323   data allow predicting the pressure conditions of chrysotile formation.

324   The Arabian-Nubian shield ophiolites were obducted within volcanic arc assemblages

325   due to arc-arc collision process (e.g. Stern 1994; Kusky et al. 2003; Meert 2003; Stern et

326   al. 2004). Obducted ophiolites, associated volcanics and sediments may represent an

327   accretionary prism system. Here we will follow the assumption of Valli et al. (2004) that

328   average thermal gradient of ancient and modern accretionary prisms can be in the range

329   of 30 ^{o}C/km and 10 ^{o}C/km, respectively (Fig. 8). El-Naby and Frisch (1999) studied

330   Allaqi-Heiani ophiolite belt and they concluded that these ophiolites record temperature

331   of 700 ^{o}C and pressures up to 8 kbar. These conditions can be converted to a thermal

332   gradient of 25 ^{o}C/km which locates between the two assumed thermal gradient. This

333   thermal gradient cuts the predicted temperature (500-550 ^{o}C) in a pressure range of 5.5-

334   6.5 kbar (Fig. 8).

335

336   ***Fluid composition and T-XCO_2 section***

337   Due to the ambiguity around the pressure condition during the prograde path of the

338   studied samples, the fluid composition will be studied only along the cooling path. Figure

339   9 shows a *T-X*CO_2 grid in the system CFMASH-CO_2 for the following end-members:

340   anth, atg, chr, en, fs, di, hed, fo, fa, ta, sp, herc, mgts, fta, mag, sid, H_2O, CO_2 at 1 kbar

341   (the cooling path of Figure 8). The *T-X*CO_2 grid was constructed in the full *X*CO_2 range

342   (not shown here), however all anthophyllite and talc producing invariant points occur at

343   high *X*CO_2 (> 0.88). In this type of sectioning (*P-, T-X*CO_2)*, mineral phases are produced

15

344   mainly at the invariant point conditions (Spear 1993). The grid includes twenty five

345   univariant reactions and seven invariant points. All of these invariant points occur at

346   temperature range of 450-520 ^{o}C (Fig. 9). All the invariant points above 500 ^{o}C are

347   magnesite-siderite absent invariant points. At 500 ^{o}C and XCO_2 (0.913), magnesite-

348   bearing invariant point appears. With cooling, the carbonate phase (siderite) becomes

349   more stable (at 460 ^{o}C and XCO_2 (0.978)). Below 450 ^{o}C, the magnesite becomes

350   metastable (Fig. 9). These invariant points show sequence of fluid evolution in the Sol

351   Hamed serpentinites.

352   At XCO_2 range (0.88-0.99), the first talc producing reaction (ta sp=mgts anth H_2O (Fig.

353   9)) is at higher temperature than any anthophyllite producing reactions which were

354   discussed in the *P-T* grid. Consequently reaction (herc anth=mgts en fs H_2O) is the

355   preferred anthophyllite producing reaction. Once the rocks started the cooling path, the

356   anthophyllite producing reaction (op. cit.) buffers the fluid composition of the system and

357   the *T-X*CO_2 path (dashed arrows in Figure 9) followed the reaction till the mineral

358   composition arrives the atg-chr-di-hed-fo-fa-sp-fta-mag-sid-CO_2 invariant point (510 ^{o}C;

359   0.998 (XCO_2)). The assemblage stayed at the invariant point conditions until one of the

360   phases (i.e. fs, herc, mgts) was completely consumed or excluded out the equilibrium. At

361   this stage of the path, the rocks follow the isothermal reaction (ta en=anth H_2O) which

362   produces a considerable amount of talc. This reaction crosses all the invariant points at

363   510 ^{o}C with different XCO_2 composition (Fig. 9). Presence of magnesite in the studied

364   assemblage (Table 4) and presence of magnesite-bearing invariant point at 500 ^{o}C and

365   XCO_2 (0.913) make the only possibility to terminate the talc producing reaction (op. cit.)

366   is at the atg-chr-fs-di-hed-fa-sp-herc-mgts-fta-mag-sid-CO_2 invariant point (510 ^{o}C;

367   0.885 (XCO_2)). The assemblage stayed at this invariant point until the anthophyllite was

368   trapped and excluded out the equilibrium, afterward the mineral equilibrium follows the

369   reaction (fo ta=en H_2O) until the magnesite-bearing invariant point at 500 ^{o}C and XCO_2

370   (0.913) which allows the first appearance of carbonate-bearing phase. Forsterite

371   consuming drives the equilibrium to leave the magnesite-bearing invariant point toward

372   the magnesite-siderite-bearing invariant point (460 ^{o}C and XCO_2 (0.978)). Subsequently

373   the reaction (ta sid=en fa CO_2 H_2O) buffers the equilibrium until the magnesite becomes

374   metastable at 450 ^{o}C and 0.984 (XCO_2). Finally, reaction (ta sid=en fs CO_2 H_2O)

375   produces talc and siderite with constant consuming rate of H_2O and CO_2.

376

377   *6.3. Fluid source and Tectonic implications*

378   Decarbonation of altered metabasalts and carbonates of marine sediments at low pressure

379   condition has been considered as a possible mechanism in order to explain CO_2 fluxes at

380   convergent margins (Staudigel et al. 1996; Kerrick and Connolly 1998; Fischer et al.

381   1998; Molina and Poli 2000). When hot geotherms are assumed, CO_2-rich fluids can be

382   transferred from the altered oceanic crust to shallower reservoirs (Fig. 10) in the forearc

383   region (Molina and Poli 2000). This mechanism can account for the CO_2 enrichment of

384   lithospheric mantle on a long-term scale and it may explain the occurrence of carbonates

385   in peridotite xenoliths (Ionov et al. 1993) as well as in some camptonitic lamprophyres

386   (Bea et al. 1999). Here this mechanism can be used to explain the high CO_2 fluxes in the

387   studied ophiolites (XCO_2 = 0.89-0.99 (Fig. 9)). This high CO_2 fluid content reacted with

388   the ophiolitic rocks in the forearc (Fig. 10) under pressure condition of 1 kbar and

389   temperature of around 800 ^{o}C (Fig. 8). Stern and Gwinn (1990) argued on the basis of C

390   and Sr isotopic that carbonate intrusions in the Eastern Desert of Egypt - which could be

391   related to the carbonatizing fluids affecting Arabian-Nubian Shield ultramafic rocks - are

392   mixtures of mantle derived and remobilized sedimentary carbonate. Hamdy and Lebda

393   (2007) concluded the same conclusion based on carbon isotope composition of south

394   Eastern Desert of Egypt.

395   *T-X*CO$_2$ grid (Fig. 9) shows that the fluid composition was buffered all the time by the

396   metamorphic reactions (e.g. Greenwood 1975; Rice and Ferry 1982; Spear 1993; Abu-

397   Alam et al. 2010). Field, petrographical and mineral chemistry evidences support this

398   thermodynamic observation. Majority of the *T-X*CO$_2$ path took place at a temperature

399   range of 450 – 550 ^{o}C. Most of the reactions in this range of the temperature occurred as

400   isothermal reactions which means that the rocks were held at this temperature for a time

401   period enough to consume one phase or more to drive the equilibria toward a lower

402   temperature conditions. Figure 5a of Hamdy and Lebda (2007) shows that spinel minerals

403   of the studied ophiolites were re-equilibrated at temperature condition of 500-550 ^{o}C

404   which is the same range provided by the *T-X*CO$_2$ grid. Presence of magnesite in

405   considerable amount in thin-section scale as well as presence of small pockets and veins

406   of magnesite in outcrop scale, indicate that the rocks were held for a long time at the two

407   magnesite-bearing invariant points (at temperature 500 and 460 ^{o}C (Fig. 9)).

408   The high pressure condition (8 kbar) which was assumed by El-Naby and Frisch (1999)

409   and which was used here to predict the geothermal gradient and the prograde path (the

410   black arrow of Figure 8) as well the predicted pressure range (5.5-6.5 kbar from this

411   study) can be explained in the context of extensive duplex array and thickness of the

412   original ophiolitic sequence (e.g. Hirono and Ogawa 1998; Ueno et al. 2011). Oceanic

413 crust in a forearc setting can be overloaded by obduction of a crust that formed in a mid-

414 oceanic ridge and the thrusting in the forearc crust itself can add more loads (Fig. 10).

415 Original thickness of the Arabian-Nubian Shield's ophiolitic sequence is 2.5 to 5 km

416 (Stern et al. 2004). Following oceanic crust density of 2.84×10^3 kg/m^3 (Carlson and

417 Raskin 1984), the studied ophiolites were overloaded by 20-28 km thickness of obducted

418 and thrusted oceanic crust from both mid-oceanic and forearc settings. This is in

419 agreement with thickness of the original sequence by a factor in range of 5.6 and 11.2.

420 The same thickening factors were suggested numerically by Ueno et al. (2011).

421 One of the opened questions around the ophiolites of the Arabian-Nubian Shield is "*when*

422 *did the alteration take place? Is it before or after the obduction?* (Stern et al. 2004)".

423 Clearly, petrographic observations and thermodynamic modelling that are presented here

424 give an answer to this question. The studied ophiolites show two segments of the *P-T*

425 path; one is the isobaric cooling path at pressure condition of 1 kbar and the second is

426 prograde path from a pressure 1 kbar up to 5.5-6.5 kbar (black arrow of Figure 8). The

427 isobaric cooling path occurred under oceanic crustal thickness of 3.5 km which means

428 that the first stage of alteration took place before the obduction while the second stage

429 occurred during thrusting and obduction processes (prograde metamorphism). At today

430 situation, the ophiolites are thrusted over volcanic arc-assemblage. The volcanic arc-

431 assemblage of the Arabian-Nubian Shield records a peak pressure around 3-4 kbar (e.g.

432 Noweir et al. 2006; Abu-Alam 2005; Abu-Alam and Farahat unpublished data). This can

433 be ensued only if the ophiolites achieved the peak condition (5.5-6.5 kbar) before the

434 final thrusting above the low-pressure arc-assemblage.

435

19

7. Conclusions

The Sol Hamed serpentinised ophiolitic mantle peridotite in the south Eastern Desert of Egypt at the Allaqi-Heiani-Onib-Sol Hamed-Yanbu arc-arc suture formed in forearc setting and later thrusted over low-grade arc-assemblage of the Arabian-Nubian Shield. It shows a *P-T* path of an isobaric cooling at lithostatic pressure of 1 kbar which is equivalent to an oceanic crustal thickness of 3.5 km. The alteration occurred before the thrusting and at high CO_2 fluxes. The decarbonation of altered oceanic metabasalts and carbonates of marine sediments at low pressure condition can be considered as a possible mechanism to explain the high concentrated CO_2 fluid fluxes at the convergent margin. The concentration of the fluid during the cooling path was buffered by the metamorphic reactions. The second segment of the path represents a prograde metamorphism which occurred under extensive duplex array and thrusting of the oceanic crust. The crust in the forearc basin was overloaded by 20-28 km of obducted and thrusted oceanic crust from both mid-oceanic and forearc basin. This is equivalent to thickness of the original ophiolitic sequence by a factor in range of 5.6 and 11.2.

458   includes lizardite. J. Connolly and F. Gallien are thanked for their help with PerPle-X

459   program. B. Evans is thanked for his help with inaccessible papers.

460

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693

694   **List of Figs.**

695   **Fig. 1.** Distribution of the ophiolites in the Arabian-Nubian Shield (modified after Vail

696   1983; Abu-Alam et al. unpublished data).

697   **Fig. 2.** Geological map of Sol Hamed area modified after Abu El-Laban 2002.

698   **Fig. 3.** (a) Semi-schematic drawing showing the ophiolitic ultramafic of Sol Hamed area,

699   lizardite and anthophyllite are metamorphosed after orthopyroxene. Chrysotile is

700   metamorphosed after lizardite. Talc is after anthophyllite. Liz, opx, anth, mag

701   and s are lizardite, orthopyroxene, anthophyllite, magnesite and sulphites,

702   respectively. (b) Zoned spinel in Sol Hamed serpentinite. Darker zones are

703   richer in Cr.

704   **Fig. 4.** Mineral chemisty. a) Substitution in serpentine. Aluminum and chromium are

705   grouped together, as they tend to vary sympathetically. b) Compositional

706   changes in spinels expressed in a triangular $Cr-Fe^{3+}-Al^{3+}$ plot with reference to

707   the fields of spinel types: a-aluminian chromite, b-ferrian chromite, c-chromian

708   magnetit, d-aluminian magnetite, e-ferrian spinel and f-chromian spinel (Stevens

709   1944) and the different metamorphic facies defined by Purvis et al. (1972);

710   Evans and Frost (1975) and Suita and Streider (1996).

711   **Fig. 5.** Whole rock $MgO/SiO_2–Al_2O_3/SiO_2$ plot. The terrestrial array is a compilation of

712   subcontinental peridotites (Hart and Zindler 1986) and represents a melt

713   depletion trend. The Sol Hamed serpentinites plot offset to lower MgO/SiO_2

714   values because of alteration. Compositions of depleted MORB mantle (DMM;

31

715 Workman and Hart 2005), primitive mantle (PM; McDonough and Sun 1995)

716 and seafloor weathering trend (Snow and Dick 1995) are plotted for comparison.

717 **Fig. 6.** Composition of spinels compared with those in modern peridotites. Data are

718 plotted on 100Cr/Cr+Al (Cr#) vs. 100Mg/Mg+Fe (Mg#) diagram, modified after

719 Dick and Bullen (1984). The fields are after Bloomer et al. (1995).

720 **Fig. 7.** *P-T* grid in the system CFMASH for atg, chr, en, fs, di, hed, fo, fa, anth, tr, clin,

721 ta, sp, herc, mgts, fta, br, H_2O. Activity of the H_2O is imposed to be the unity.

722 Note: reactions liz = br atg, liz = chr, liz = ta fo clin H_2O are used here after

723 Evan 2004; Chernosky 1975; Caruso and Chernosky 1979, respectively.

724 **Fig. 8.** A simplified *P-T* grid of Figure 7 shows only the interesting metamorphic

725 reactions. Maximum pressure during the cooling path is the pressure equivalent

726 to the invariant point [atg chr fs di hed fa tr ta herc fta br mgts]. The vertical bar

727 below the op. cit. invariant point shows the pressure equivalents to the Arabian-

728 Nubian Shield's ophiolitic crustal thicknesses as reconstructed by Stern et al.

729 (2004). The two gray arrows show the modern thermal gradient and ancient

730 thermal gradient. The black arrow showing the path of the study samples, the

731 peak pressure is 5-5.7 kbar based on the pressure calculation from Wadi

732 Haimur-Abu Swayel ophiolites (El-Naby and Frisch 1999). Note: the ancient

733 thermal gradient is equivalent to thickening of the sequence by factor of 5.5 as

734 suggest by Ueno et al. (2011). The temperature axe is in two different scales to

735 show the reactions at high temperature condition in more details than Figure 7.

736 **Fig. 9.** A *T-X*CO_2 grid in the system CFMASH-CO_2 for the following end-members:

737 anth, atg, chr, en, fs, di, hed, fo, fa, ta, sp, herc, mgts, fta, mag, sid, H_2O, CO_2.

738   The grid was constructed at 1 kbar. Fluid concentration is buffered by the

739   metamorphic reactions. The grid shows high CO_2 concentration in the fluid

740 **Fig. 10.** A three dimensional model illustrating the tectonic evolution of the studied

741   ophiolites. a) Development of subduction zone. High concentrated CO_2 fluid is

742   released from carbonate rocks in the subduction zone. These fluids re-

743   concentrated in the fore-arc ophiolites. A passive margin is drawn on the flank

744   of the oceanic basin since some authors recorded volcanic and sedimentary

745   rocks in the Arabian-Nubian Shield were formed in a passive margin setting

746   (e.g. Nakasib suture; Abdelsalam and Stern 1993). b) Thrusting and duplex

747   thickening of the ophiolitic sequence. The white star is the position of the

748   studied ophiolites.

749

Primary minerals	Olivine
	Opx
	Spinel
	Chromite (the inner core)
First stage of alteration and metamorphism	Lizardite
	Antigorite
	Anthophyllite
	Magnesite
	Magnetite
	Chromite (inter. zone)
	Sulphides
Second stage of alteration and metamorphism	Lizardite
	Chrysotile
	Talc
	Magnesite
	Magnetite
	Chromite (outer zone)

Table 1
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

	Opx		Serpentine														
Sample	325/1	325/2	323	324	324	324	325	325	333	333	278	278	310	347	347	369	369
SiO_2	58.21	58.37	43.73	41.86	42.62	41.13	42.47	42.6	40.62	43.78	44.54	43.04	43.96	41.56	41.3	43.23	41.94
TiO_2	0.12	0.09	0.12	b.d.l	0.01	0.03	b.d.l	0.01	0.12	0.08	0.07	0.15	0.08	0.01	b.d.l	b.d.l	0.03
Al_2O_3	0.74	0.67	0.48	0.46	1.12	1.79	0.13	b.d.l	0.3	0.06	0.28	0.14	0.14	0.14	0.26	0.21	0.28
Cr_2O_3	0.42	0.36	b.d.l	b.d.l	0.16	0.02	0.02	b.d.l	0.22	b.d.l	b.d.l	0.34	0.12	0.09	0.16	0.15	0.36
FeO	4.85	5.09	2.61	2.88	1.98	3.33	1.81	2.05	4.2	2.66	1.37	1.85	0.94	3.99	3.36	2.11	6.21
MnO	0.03	0.07	0.09	b.d.l	0.04	b.d.l	b.d.l	b.d.l	0.38	b.d.l	0.01	0.09	b.d.l	0.17	0.22	0.01	b.d.l
MgO	34.99	35.1	37.04	37.04	37.67	36.99	38.42	36.68	34.37	37.74	37.78	36.88	38.14	38.06	38.68	37.17	35.98
CaO	b.d.l	0.03	b.d.l	0.1	b.d.l	b.d.l	b.d.l	0.19	0.03	0.07	0.06	0.13	b.d.l	b.d.l	b.d.l	0.18	0.11
Na_2O	b.d.l	b.d.l	0.13	0.01	0.05	0.05	0.02	b.d.l	b.d.l	0.12	0.01	0.01	0.1	b.d.l	b.d.l	0.09	0.08
K_2O	b.d.l	b.d.l	b.d.l	0.02	0.03	b.d.l	0.01	0.03	b.d.l	0.01	0.02	0.03	0.08	0.01	b.d.l	0.08	0.01
Total	99.36	99.78	84.2	82.37	83.68	83.34	82.88	81.56	80.24	84.52	84.14	82.66	83.56	84.03	83.98	83.23	85
Si	2	2	8.381	8.237	8.211	8.027	8.258	8.41	8.283	8.369	8.47	8.385	8.418	8.097	8.036	8.375	8.162
Ti	0.003	0.002	0.017	-	0.001	0.004	-	0.001	0.018	0.012	0.01	0.022	0.012	0.001	-	-	0.004
Al	0.03	0.027	0.108	0.107	0.254	0.411	0.03	-	0.072	0.014	0.063	0.032	0.032	0.032	0.06	0.048	0.064
Cr	0.011	0.01	-	-	0.024	0.003	0.003	-	0.035	-	-	0.052	0.018	0.014	0.025	0.023	0.055
Fe	0.139	0.146	0.418	0.474	0.319	0.543	0.294	0.338	0.716	0.425	0.218	0.301	0.151	0.65	0.547	0.342	1.011
Mn	0.001	0.002	0.015	-	0.007	-	-	-	0.066	-	0.002	0.015	-	0.028	0.036	0.002	-
Mg	1.792	1.792	10.58	10.86	10.82	10.76	11.13	10.79	10.44	10.75	10.71	10.71	10.88	11.05	11.21	10.73	10.43
Ca	-	0.001	-	0.021	-	-	-	0.04	0.007	0.014	0.012	0.027	-	-	-	0.037	0.023
Na	-	-	0.048	0.004	0.019	0.019	0.008	-	-	0.044	0.004	0.004	0.037	-	-	0.034	0.03
K	-	-	-	0.005	0.007	-	0.002	0.008	-	0.002	0.005	0.007	0.02	0.002	-	0.02	0.002
Cations	3.977	3.98	19.57	19.71	19.66	19.77	19.73	19.59	19.65	19.64	19.49	19.56	19.58	19.88	19.92	19.62	19.79
en	0.93	0.92															
fs	0.0055	0.006															
mgts	0.027	0.01															

Table 2
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

	Talc								
Sample	306/2	306/2	306/2	306/2	306/2	347/2	347/2	347/2	347/2
SiO_2	61.07	60.04	60.92	59.98	61.09	56.78	57.36	57.22	57.21
TiO_2	0.05	b.d.l	0.04	0.07	0.22	b.d.l	b.d.l	b.d.l	b.d.l
Al_2O_3	0.03	0.14	0.28	0.05	b.d.l	0.42	0.5	0.37	0.34
Cr_2O_3	0.02	0.14	0.26	0.03	b.d.l	0.28	0.2	0.21	0.17
Fe_2O_3	b.d.l	b.d.l	b.d.l	b.d.l	b.d.l	0.68	0.64	0.64	0.62
FeO	3.31	3.61	3.85	2.71	3.11	5.54	5.22	5.16	5.04
MnO	b.d.l	b.d.l	0.04	b.d.l	b.d.l	b.d.l	0.16	b.d.l	0.2
MgO	27.14	26.91	26.85	27.17	27.6	34.4	34.26	34.41	34.84
CaO	0.09	0.02	b.d.l	0.15	0.16	b.d.l	b.d.l	b.d.l	b.d.l
Na_2O	0.13	b.d.l	0.16	0.17	0.13	b.d.l	b.d.l	b.d.l	b.d.l
Totals	91.84	90.86	92.4	90.33	92.31	98.1	98.34	98	98.42
Si	4.053	4.036	4.034	4.041	4.035	3.646	3.666	3.667	3.654
Ti	0.002	-	0.002	0.004	0.011	-	-	-	-
Al	0.002	0.011	0.022	0.004	-	0.032	0.038	0.028	0.026
Cr	0.001	0.007	0.014	0.002	-	0.014	0.01	0.011	0.009
Fe^{3+}	-	-	-	-	-	0.033	0.031	0.031	0.03
Fe^{2+}	0.184	0.203	0.213	0.153	0.172	0.297	0.279	0.276	0.269
Mn	-	-	0.002	-	-	-	0.009	-	0.011
Mg	2.685	2.696	2.65	2.728	2.717	3.292	3.263	3.286	3.316
Ca	0.006	0.001	-	0.011	0.011	-	-	-	-
Na	0.017	-	0.021	0.022	0.017	-	-	-	-
Sum	6.951	6.955	6.957	6.964	6.962	7.316	7.296	7.3	7.315
ta	0.71	0.72	0.68	0.75	0.74	0.83	0.83	0.85	0.85
fta	0.00023	0.00031	0.00036	0.00013	0.00019	0.00061	0.00052	0.00051	0.00046

Table 3
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

	Carbonate minerals												
Sample	300	300	300	306	306	306	306	306	306	368	368	368	368
SiO_2	0.19	0.15	0.04	0.11	0.23	0.18	0.05	0.04	0.09	0.13	0.16	0.1	0.09
Cr_2O_3	0.11	0.04	0.09	b.d.l	0.07	b.d.l	b.d.l	b.d.l	b.d.l	0.05	0.04	0.02	0.07
FeO	13.47	11.17	13.42	14.1	12.33	8.46	13.03	13.08	13.79	8.97	13.06	13.95	10.63
MnO	0.09	0.25	0.13	0.33	0.33	0.22	0.03	0.08	0.17	0.24	0.25	0.25	0.27
MgO	36.19	38.06	36.82	35.53	36.53	40.14	36.57	36.09	36.63	39.53	36.58	36.08	38.07
CaO	0.11	0.23	0.17	0.15	0.15	0.24	0.04	0.21	0.27	0.17	0.21	0.21	0.15
Totals	50.16	49.9	50.67	50.22	49.64	49.24	49.72	49.5	50.95	49.09	50.3	50.61	49.28
Si	0.006	0.004	0.001	0.003	0.007	0.005	0.002	0.001	0.003	0.004	0.005	0.003	0.003
Cr	0.003	0.001	0.002	-	0.002	-	-	-	-	0.001	0.001	0	0.002
Fe^{2+}	0.343	0.28	0.338	0.361	0.315	0.21	0.333	0.336	0.346	0.224	0.331	0.354	0.269
Mn	0.002	0.006	0.003	0.009	0.009	0.006	0.001	0.002	0.004	0.006	0.006	0.006	0.007
Mg	1.643	1.701	1.65	1.622	1.663	1.772	1.664	1.653	1.638	1.759	1.65	1.63	1.715
Ca	0.004	0.007	0.005	0.005	0.005	0.008	0.001	0.007	0.009	0.005	0.007	0.007	0.005
Sum	2	2	2	2	2	2	2	2	2	2	2	2	2
mag	0.84	0.86	0.84	0.83	0.84	0.89	0.85	0.84	0.83	0.89	0.84	0.83	0.87
sid	0.26	0.21	0.25	0.27	0.24	0.17	0.25	0.25	0.26	0.18	0.25	0.26	0.21

Table 4

Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

	core02	rim03	core05	rim06	rim	core	core-rim	rim	core	core-rim	rim	core	core-rim	rim	core	core-rim	rim
								Spinel									
SiO_2	0,07	0,05	0,07	0,23	0,13	0,24	b.d.l	0,21	0,05	0,02	0,22	0,19	0,12	0,21	0,22	0,04	0,35
TiO_2	0,36	0,23	0,21	0,18	b.d.l	0,22	0,27	b.d.l	0,15	b.d.l	0,02	0,21	0,13	b.d.l	0,08	0,11	b.d.l
Al_2O_3	8,81	9,23	8,28	8,4	b.d.l	6,27	6,32	0,12	5,4	6,53	0,1	6,67	6,5	b.d.l	5,57	6,49	0,15
FeO	34,12	34,47	31,23	32,51	89,78	32,88	31,47	90,89	34,12	32,71	91,34	31,6	32,14	90,03	30,39	30,7	90,16
Cr_2O_3	51,38	51,1	53,76	52,76	1,64	56,51	56,62	1,56	56,25	55,26	1,34	55,82	55	2,1	56,98	56,84	0,75
MnO	0,39	b.d.l	0,36	0,51	0,46	0,18	0,92	0,5	0,02	0,54	0,41	0,66	1,09	0,45	0,85	0,25	0,11
MgO	4,23	4,03	4,96	4,56	1	4,03	4,52	0,88	3,67	4,4	1,05	4,07	4,06	0,75	4,53	5,17	1,03
NiO	b.d.l	0,72	0,48	0,17	0,35	b.d.l	0,22	0,72	0,01	b.d.l	0,23	0,02	0,25	0,38	0,34	b.d.l	0,46
Total	99,36	99,83	99,35	99,32	93,36	100,33	100,34	94,88	99,67	99,46	94,71	99,24	99,29	93,92	98,96	99,6	93,01
Si	0,02	0,014	0,02	0,065	0,052	0,068	-	0,083	0,014	0,006	0,087	0,054	0,035	0,084	0,063	0,011	0,141
Al	2,96	3,084	2,76	2,809	-	2,099	2,113	0,056	1,84	2,207	0,047	2,249	2,203	-	1,889	2,17	0,071
Ti	0,08	0,049	0,045	0,038	-	0,047	0,058	-	0,033	-	0,006	0,045	0,028	-	0,017	0,023	-
Cr	11,56	11,451	12,018	11,83	0,522	12,687	12,692	0,488	12,852	12,526	0,42	12,623	12,499	0,663	12,956	12,744	0,239
Mn	0,094	-	0,086	0,123	0,157	0,043	0,221	0,168	0,005	0,131	0,138	0,16	0,266	0,152	0,207	0,06	0,038
Mg	1,796	1,705	2,093	1,93	0,6	1,708	1,913	0,52	1,583	1,883	0,621	1,737	1,742	0,447	1,944	2,188	0,62
Ni	-	0,16	0,11	0,04	0,11	-	0,05	0,23	-	-	0,07	-	0,06	0,12	0,08	-	0,15
Fe^{2+}	6,11	6,135	5,711	5,907	7,133	6,249	5,816	7,082	6,412	5,986	7,171	6,103	5,932	7,281	5,769	5,752	7,192
Fe^{3+}	1,39	1,402	1,157	1,258	15,426	1,099	1,137	15,373	1,261	1,261	15,44	1,029	1,235	15,253	1,075	1,052	15,549
Mg#	0,227	0,217	0,268	0,246	0,079	0,214	0,247	0,068	0,197	0,239	0,079	0,221	0,227	0,057	0,252	0,275	0,079
Cr#	0,726	0,718	0,754	0,744	0,0321	0,798	0,796	0,030	0,805	0,783	0,026	0,793	0,784	0,041	0,813	0,798	0,015
Fe^{3+}#	0,087	0,087	0,072	0,079	0,967	0,069	0,071	0,965	0,079	0,078	0,970	0,064	0,077	0,958	0,067	0,065	0,980
mt	0.004	0.006	0.005	0.004	0.90	0.0013	0.0015	0.90	-	0.0009	0.89	0.0026	0.0012	0.90	0.0008	0.0018	0.91
cmt	0.39	0.39	0.41	0.40	0.00058	0.48	0.46	0.00052	0.50	0.45	0.0004	0.47	0.46	0.001	0.48	0.45	0.0001

Table 5
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Table

Oxide	306	347	369	310
TiO_2	0.06	0.02	0.01	0.04
SiO_2	41.49	38.41	38.55	38.00
Al_2O_3	0.48	0.31	0.80	0.88
Fe_2O_3	6.79	7.54	8.70	7.49
MnO	0.09	0.09	0.09	0.07
MgO	38.99	38.43	39.83	39.11
CaO	0.18	1.19	0.65	0.12
Na_2O	0.09	b.d.l	0.01	0.17
K_2O	0.01	b.d.l	0.01	0.05
Sum	88.19	86.00	88.65	85.94
LOI	10.60	12.99	10.27	13.14
Cr	2742.20	2717.47	2703.71	2696.98
Co	166.79	154.23	116.20	120.41
Ni	2381.21	2060.51	1650.08	1799.16
Cu	63.83	8.27	46.32	14.97
Zn	57.11	16.55	11.98	13.32
Sr	55.04	89.11	48.08	46.61
V	40.31	14.00	33.93	29.83
Ba	35.24	45.00	20.00	15.00
Pb	4.70	4.63	24.04	13.15
Cd	2.00	3.31	3.43	2.00
Li	10.08	9.93	1.68	4.99
Rb	1.80	0.86	0.17	0.97

Table 6
Abu-Alam and Hamdy (2013)
Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

Fig. 1
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

Fig. 2
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

(a)

~ 5 mm

(b)

Chrysotile   Chlorite

Olivine   Talc

Zoned Chromite   Magnetite and spinel

20 µm

Fig. 3
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

(a) Mg / Fe+Mn / Al+Cr

(b) Fe³⁺ / Cr³⁺ / Al³⁺

Fe-spinels

Lower-amphibolite-facies spinels

Greenschist-facies Cr-spinels

Upper-amphibolite-facies spinels

Granulite-facies Spinels

□ Core   ■ intermediate zone   ○ rim

Figure

Fig. 5
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites
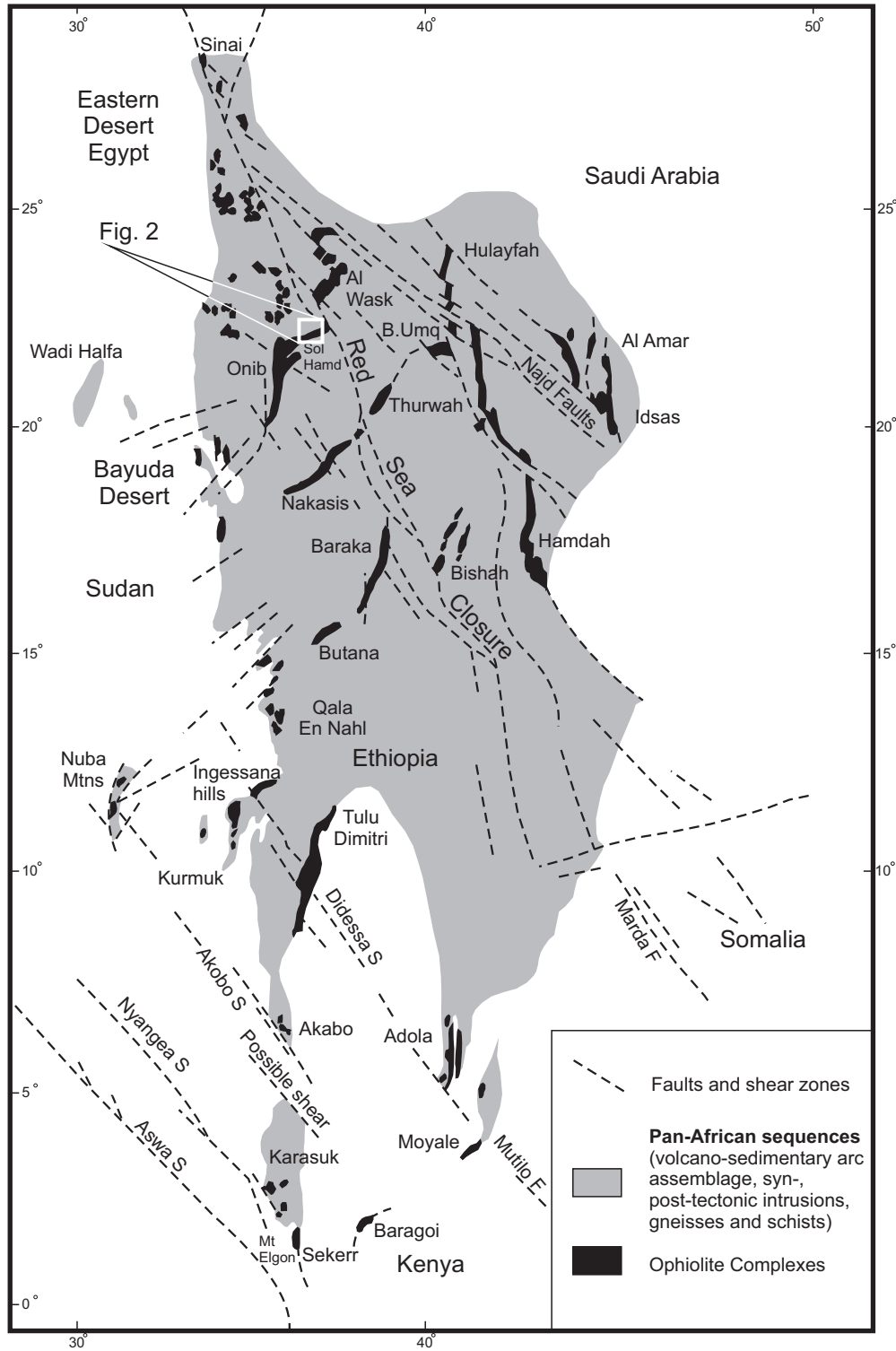
Fig. 6
Abu-Alam and Hamdy (2013)

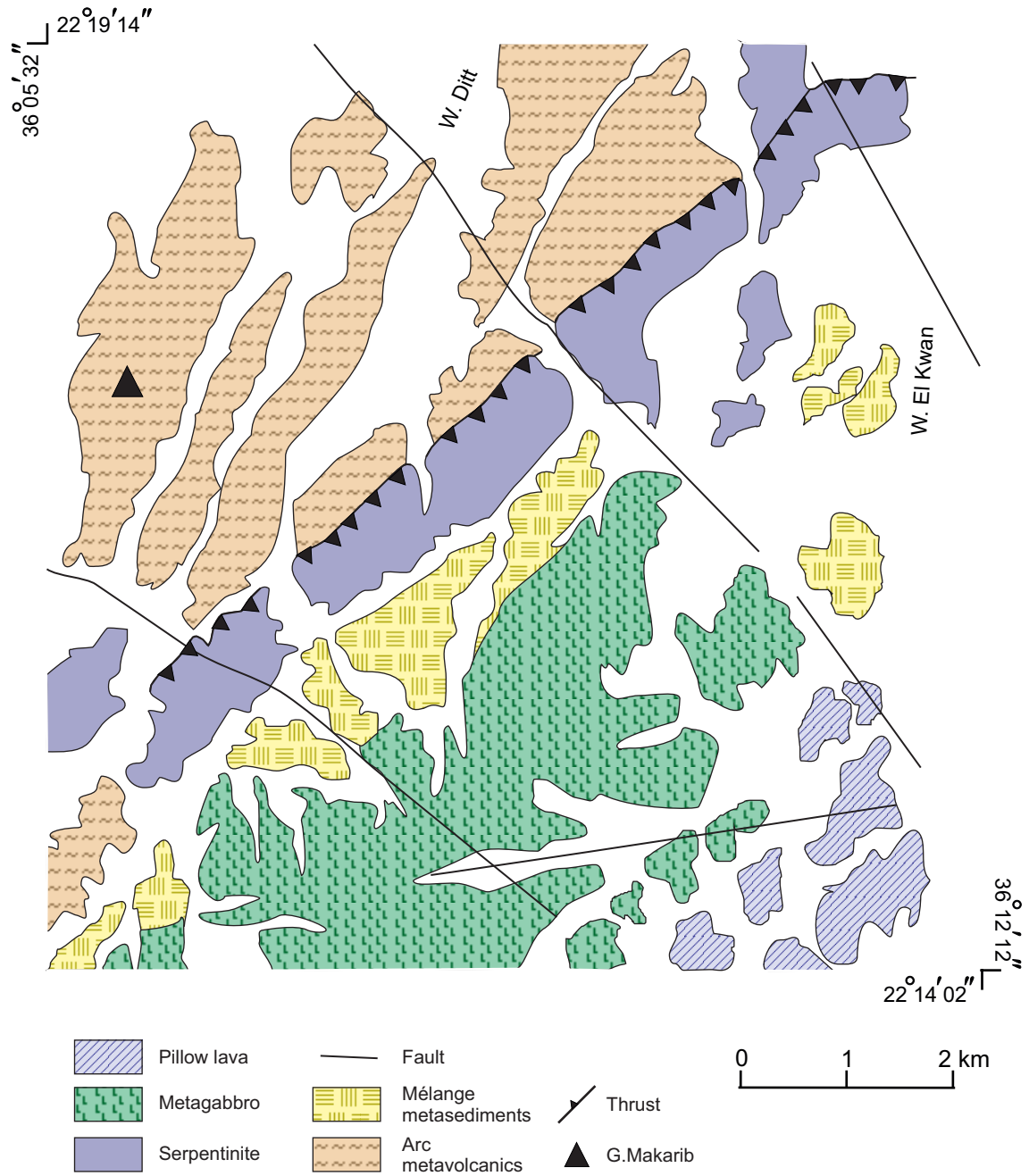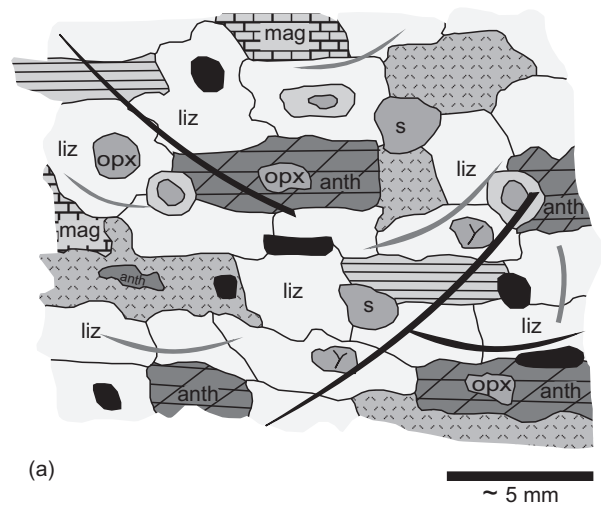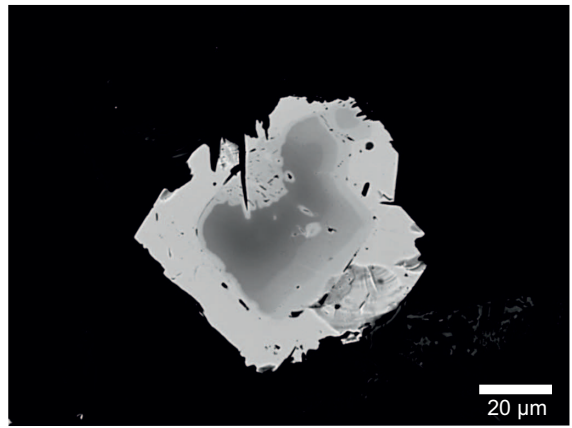Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

CFMASH

P
(bar)

1) tr fo = di en H₂O
2) fa anth = fs en H₂O
3) clin = sp fo en H₂O
4) fo anth = en H₂O
5) clin = sp fo anth H₂O
6) tr fa = di en fs H₂O
7) clin fa = herc fo anth H₂O
8) sp fo ta = clin anth
9) fa ta = fs anth H₂O
10) clin fa ta = anth herc H₂O
11) tr hed fa = di fs H₂O
12) clin anth = sp en H₂O
13) clin fa = herc fo en H₂O
14) di clin fa = herc tr fo H₂O
15) clin ta = H₂O sp anth

16) clin anth fa = herc en H₂O
17) di clin = sp tr fo H₂O
18) fta atg = H₂O ta fa
19) hed br atg = H₂O di fa
20) fa fta = H₂O fs
21) di mgts clin = H₂O tr sp
22) hed br clin = H₂O di herc fa
23) fta clin = H₂O herc ta fs
24) mgts clin = H₂O ta sp
25) br atg = H₂O fo
26) herc ta hed = H₂O mgts tr fs
27) br fa clin = herc fo H₂O
28) clin br = H₂O sp fo
29) herc fta ta = H₂O mgts fs
30) ta sp = H₂O mgts anth

Reaction equations are written with the high T assemblage to the right of the equal sign

6200

5700

5200

4700

4200

3700

3200

2700

2200

1700

1200

700

200

100   150   200   250   300   350   400   450   500   550   600   650   700   750   T °C

clin hed H₂O herc di tr fa

atg fo ta H₂O

fs clin

herc fa ta H₂O

en ta anth

[atg chr fs di hed fa tr clin sp herc fta br mgts]

chr H₂O fo atg

liz = br atg (Evan, 2004)

hed atg ta

H₂O tr fa

hed atg

H₂O di tr fa

hed fa ta tr fs H₂O

clin en sp fo ta

clin hed ta H₂O herc tr fs

liz = chr (Chernosky, 1975)

clin en fa herc fo ta

di atg tr fo H₂O

liz = ta fo clin H₂O (Caruso & Chernosky, 1979)

fo ta anth H₂O

clin en fa herc fo anth

[atg chr fs di hed tr ta sp fta br mgts]

[atg chr fs di hed tr sp H₂O fta br mgts]

clin en sp fo anth

[atg chr fs di hed fa tr ta herc fta br mgts]

clin hed fs

herc tr fa H₂O

[atg chr fs di herc hed tr fa H₂O fta br mgts]

clin anth fa herc fo ta

Fig. 7
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

Fig. 8
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

CFMASH-CO₂

$T\,^{\circ}C$

at 1 kbar
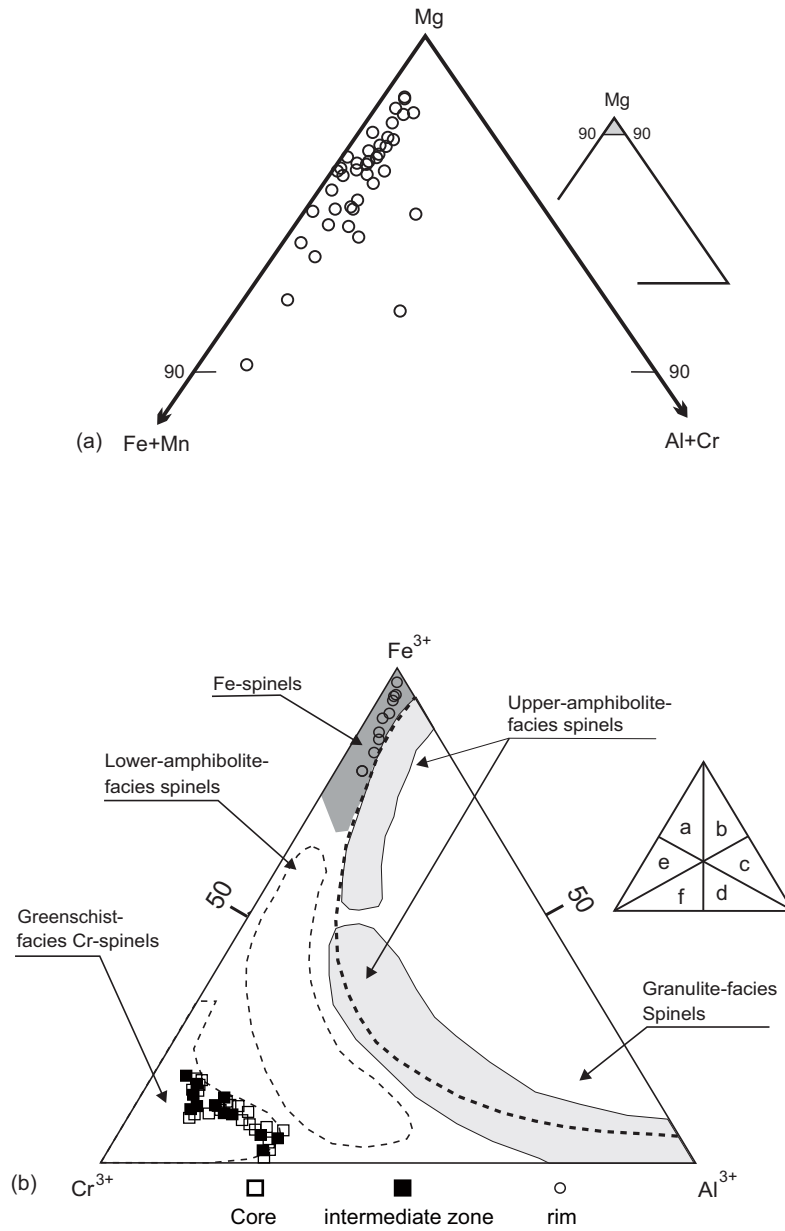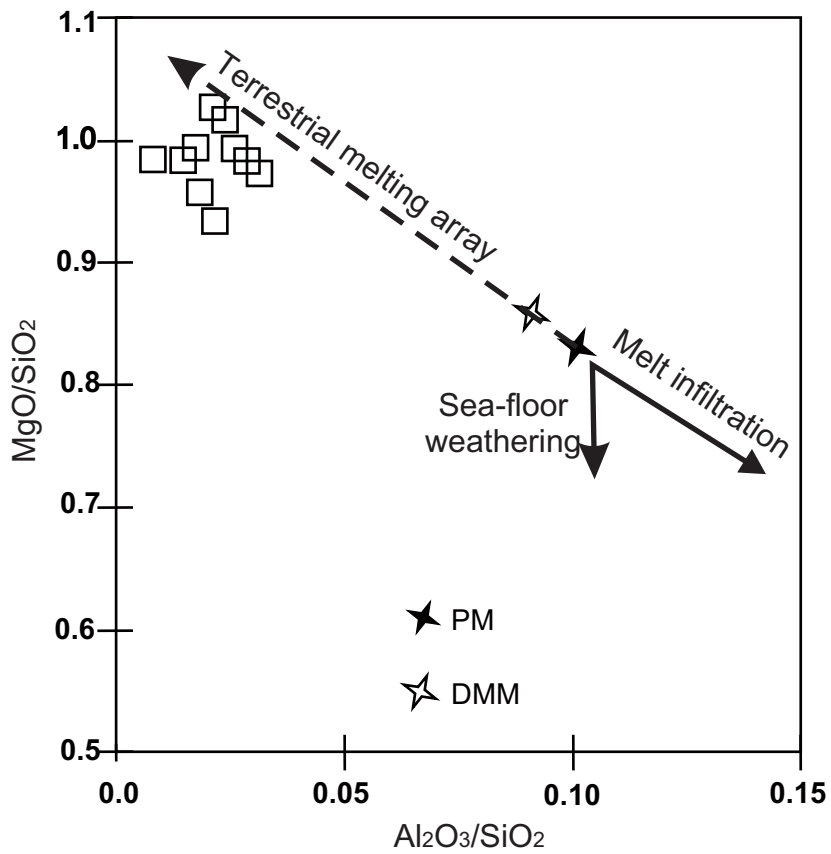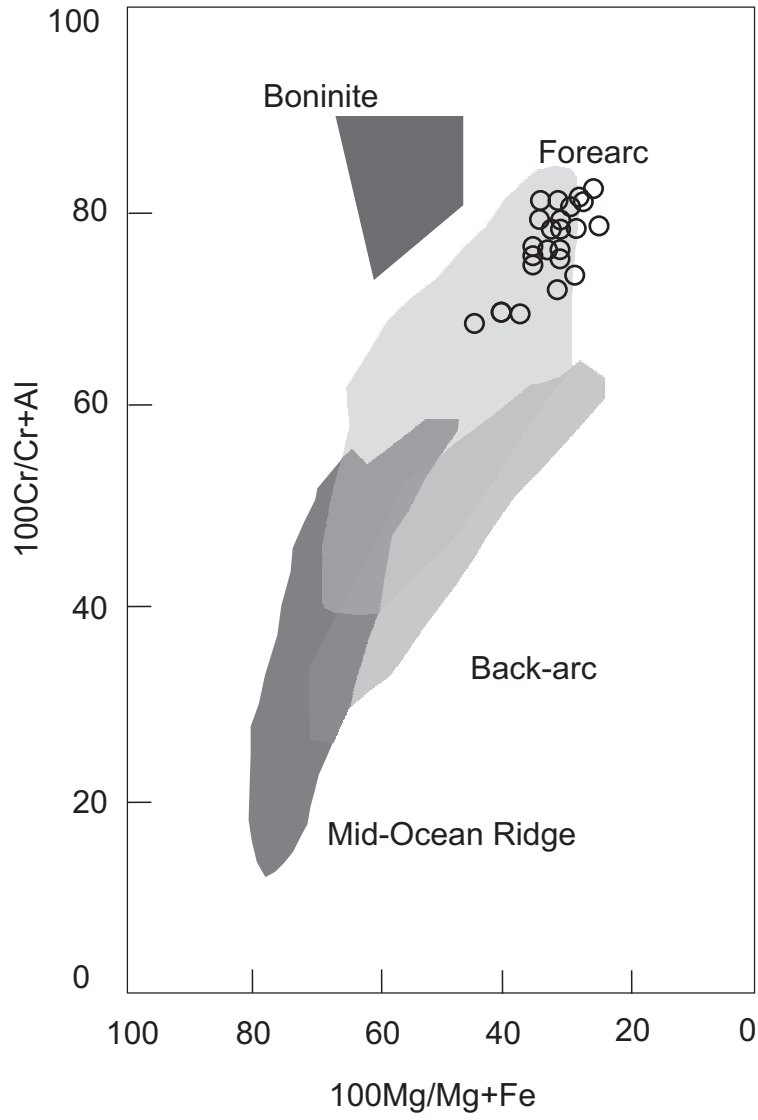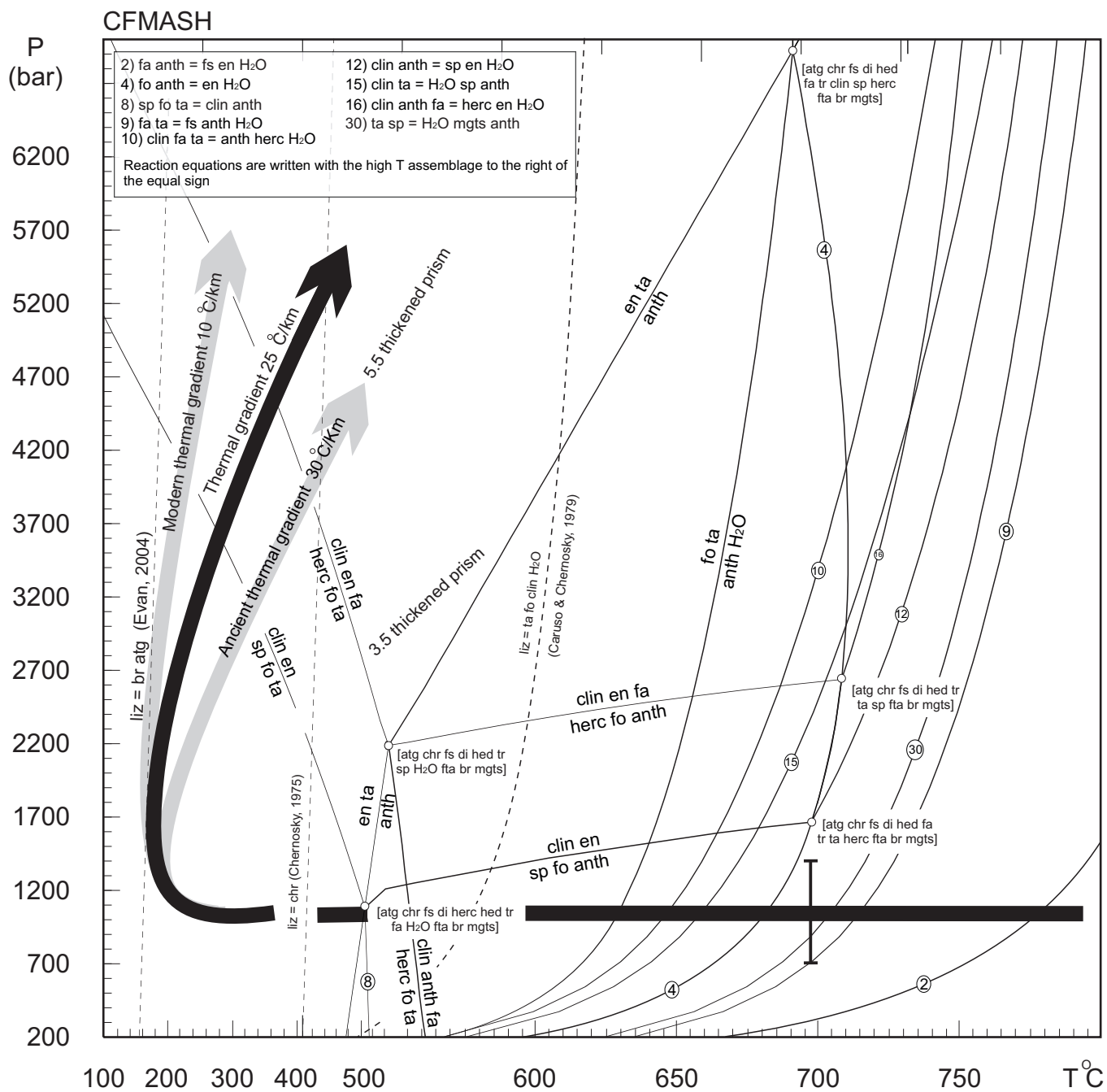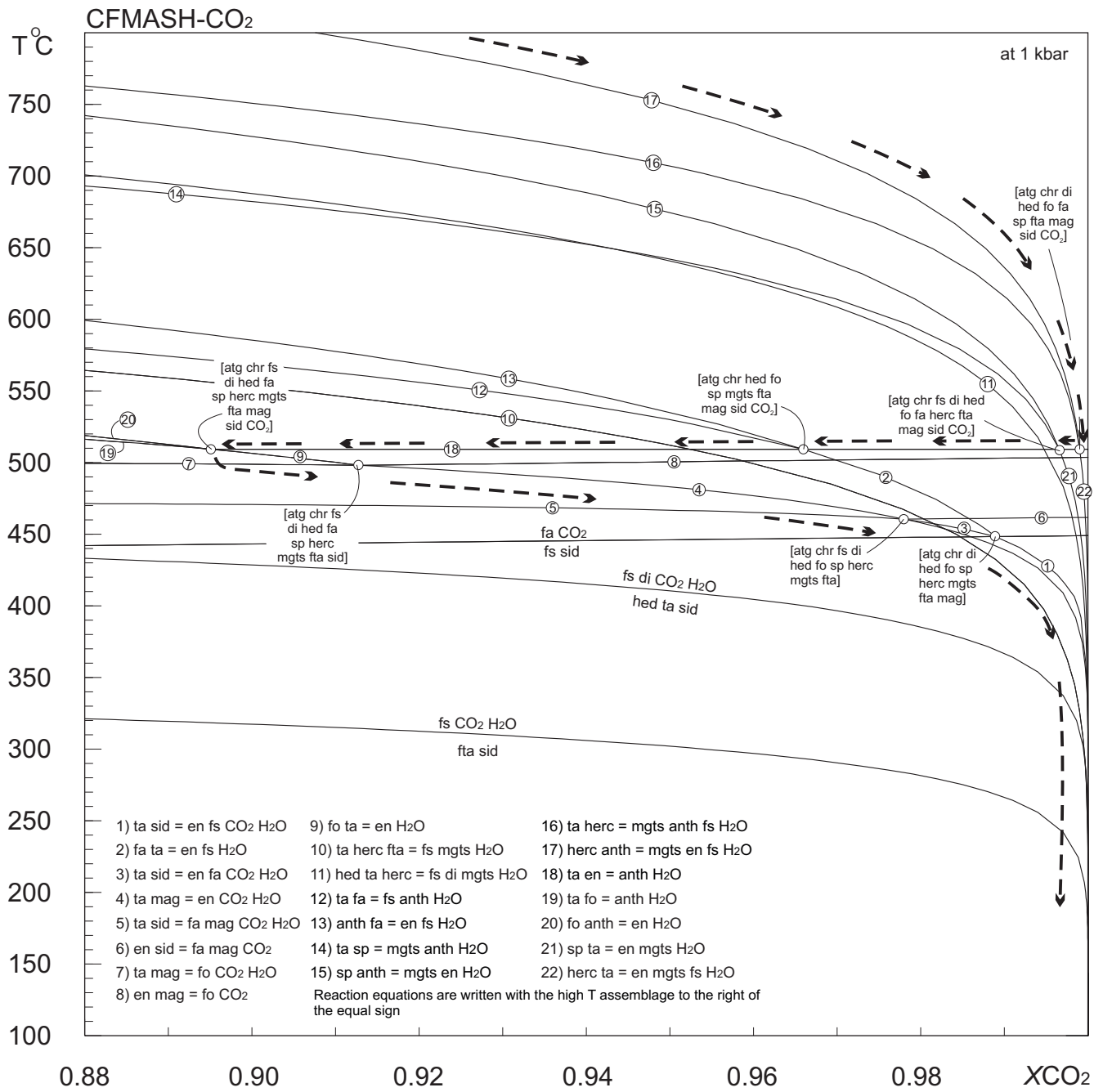
750

700

650

600

550

500

450

400

350

300

250

200

150

100

⑰ ⑯ ⑮ ⑭ ⑬ ⑫ ⑪ ⑩ ⑨ ⑧ ⑦ ⑥ ⑤ ④ ③ ② ① ⑲ ⑳ ㉑ ㉒ ⑱

[atg chr di hed fo fa sp fta mag sid CO₂]

[atg chr fs di hed fa sp herc mgts fta mag sid CO₂]

[atg chr hed fo sp mgts fta mag sid CO₂]

[atg chr fs di hed fo fa herc fta mag sid CO₂]

[atg chr fs di hed fa sp herc mgts fta sid]

fa CO₂
fs sid

[atg chr fs di hed fo sp herc mgts fta]

[atg chr di hed fo sp herc mgts fta mag]

fs di CO₂ H₂O
hed ta sid

fs CO₂ H₂O
fta sid

1) ta sid = en fs CO₂ H₂O 9) fo ta = en H₂O 16) ta herc = mgts anth fs H₂O

2) fa ta = en fs H₂O 10) ta herc fta = fs mgts H₂O 17) herc anth = mgts en fs H₂O

3) ta sid = en fa CO₂ H₂O 11) hed ta herc = fs di mgts H₂O 18) ta en = anth H₂O

4) ta mag = en CO₂ H₂O 12) ta fa = fs anth H₂O 19) ta fo = anth H₂O

5) ta sid = fa mag CO₂ H₂O 13) anth fa = en fs H₂O 20) fo anth = en H₂O

6) en sid = fa mag CO₂ 14) ta sp = mgts anth H₂O 21) sp ta = en mgts H₂O

7) ta mag = fo CO₂ H₂O 15) sp anth = mgts en H₂O 22) herc ta = en mgts fs H₂O

8) en mag = fo CO₂ Reaction equations are written with the high T assemblage to the right of the equal sign

0.88   0.90   0.92   0.94   0.96   0.98 $X\mathrm{CO_2}$

Fig. 9
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites

Figure

East-Gondwana

Mozambique Ocean

passive margin

Volcanic arc

Accretionary prism

Forearc basin

Oceanic crust

CO₂ rich fluid

(a)

East-Gondwana

Ophiolites with mid-oceanic ridge setting
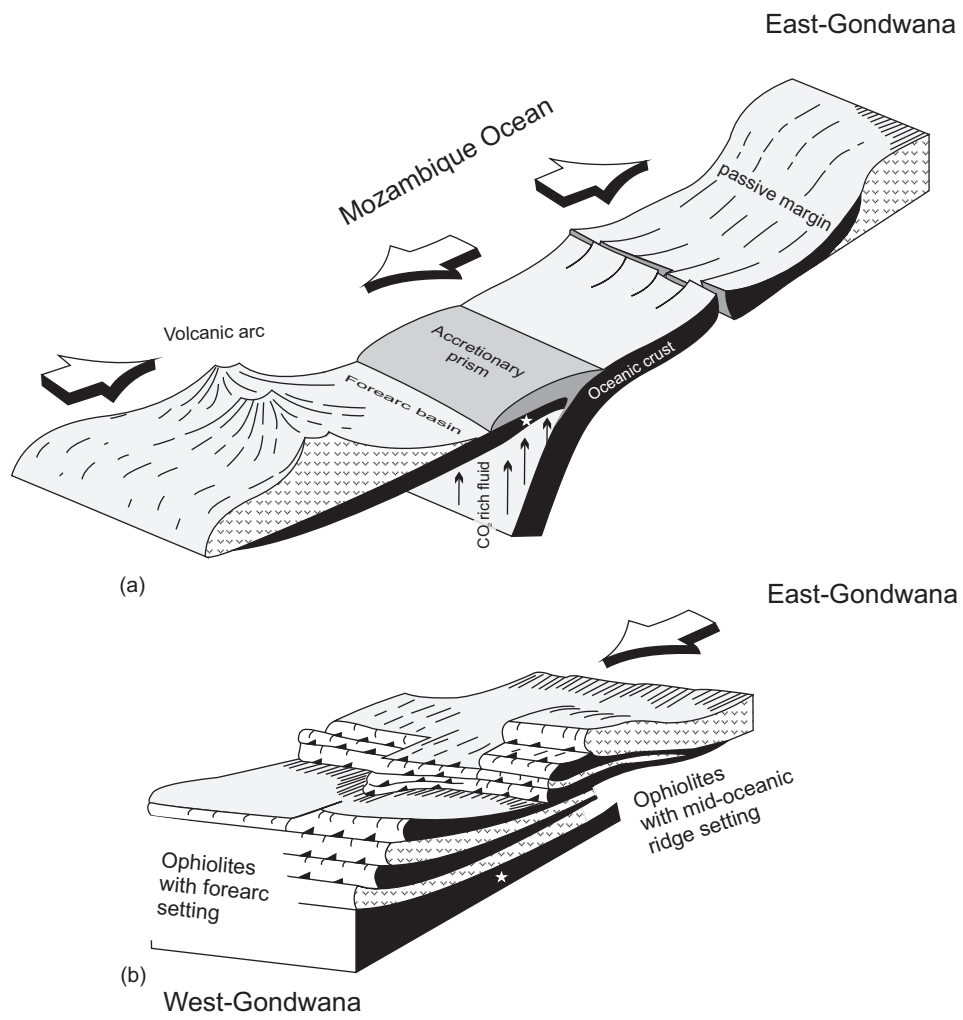
Ophiolites with forearc setting

(b)

West-Gondwana

Fig. 10
Abu-Alam and Hamdy (2013)

Fluid interaction in the Arabian-Nubian Shield ophiolites