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Overview of the tectonic evolution of the Iraqi Zagros thrust zone: Sixty million years of Neotethyan ocean subduction

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Abstract

The world's largest ongoing collisional orogeny is the Europe Alps-Himalayan-SE Asian belt and is a natural laboratory to understand many processes that have shaped the continents. Due to political instability and conflict throughout this millennium, the Iraq (Kurdish) sector of the Zagros mountain chain is the least studied part of this orogenic system. In Iraq, the Zagros contains the suture between the Arabian subcontinent to the south and west and the Iranian edge of the Eurasian continent to the north and east. The suture zone is marked by several allochthons of Neotethyan ophiolitic and volcanic arc assemblages that were obducted onto the Arabian margin. New geochronological data, including SHRIMP U-Pb zircon, integrated with whole rock geochemistry, indicates that both Cretaceous (~96 Ma) and Cenozoic (~40 Ma) assemblages are present. The relationships between these units are complicated, thus some Cretaceous arc rocks were intruded by Cenozoic arc rocks, and out-of-sequence thrusting has interleaved and juxtaposed assemblages of different ages. Ongoing wrench faulting since continental collision at ~14 Ma has further complicated the pattern of lithotectonic units, particularly those that were obducted out of the Neotethyan realm. The new data indicate that the Iraqi sector of Neotethys was not 'quiet' in the Cretaceous, but contains fragments of arcs of that age, contiguous with those along strike in Turkey, Iran and the Himalayas.

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1 **Overview of the tectonic evolution of the Iraqi Zagros thrust zone: sixty million years of**
2 **Neotethyan ocean subduction**

3
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27 was not 'quiet' in the Cretaceous, but contains fragments of arcs of that age, contiguous with
28 those along strike in Turkey, Iran and the Himalayas.

29 Keywords: Zagros thrust zone; Neotethys; Kurdistan; Cretaceous arc; Iraq

1. Introduction

The European-Himalayan-Southeast Asian orogen is the Earth's greatest natural laboratory for an ongoing collisional orogeny (Şengör, 1979; Rosenbaum and Lister, 2002; Agard et al., 2005; Robertson et al., 2009; Homke et al., 2010; Ali et al., 2016). In the broadest terms, it represents the transfer of continental fragments (such as the Sanandaj-Sirjan Zone of Iran; Fergusson et al., 2016 and references therein) rifted off the edge of Gondwana to the south, that travelled across ephemeral oceans and were then accreted onto the edge of the northern continent Eurasia (Agard et al., 2005; Shafaii Moghadam et al., 2014). This process has so far involved the opening of two intervening oceans – Paleotethys and Neotethys (e.g., Stampfli, 2000; Ali et al., 2016; Aswad et al., 2016; Robertson et al., 2016), that were consumed by several subduction zones with associated arcs, developed from the Jurassic and into the late Cenozoic (Berberian and King, 1981; Alavi, 1994; Vincent et al., 2005; Allen and Armstrong, 2008; Horton et al., 2008; Dargahi et al., 2010; Agard et al., 2011). Over the last few decades, there has been continual refinement and understanding of this model, via more detailed and extensive fieldwork, plus advances in radiometric dating and isotope geochemistry (Şengör et al., 1993; Agard et al., 2005; Azizi et al., 2011; Shafaii Moghadam et al., 2014). However, major questions still remain concerning the mechanisms and synchronicity of these processes. Thus, was arc development and shutting-off of their activity by collisional orogeny piecemeal along the >10,000 km extent of this orogen, or have these events been coeval over a great lateral extent (Şengör, 1990; Şengör and Natal'in, 1996; Agard et al., 2005)?

Along the entire extent of the orogen, the data needed to answer this fundamental geodynamics question on the world's greatest collisional orogeny are of variable density and quality. One domain where there has been a particular paucity of information is the Iraq (Kurdistan) sector of the orogen (Fig. 1). This has been because of the decades-long political instability and conflict in this region. This sector contains part of the Zagros mountain belt and its western foreland (Fig. 1), and within it are allochthons of volcanic arc and supra-subduction ophiolites, resting over predominantly sedimentary strata. Relative to the Arabian subcontinent to the south and west, some of these sedimentary strata are allochthonous, whereas others are parautochthonous or autochthonous packages (Fig. 2; Ali et al., 2012, 2013, 2014, 2016, 2017; Aswad et al., 2013, 2016). Although there is agreement that the arc and supra-subduction ophiolites relate to the

60 consumption of Neotethys (Şengör, 1979; Agard et al., 2005; Azizi et al., 2011; Shafaii
61 Moghadam et al., 2014), there is uncertainty and debate over (i) the age of these assemblages, (ii)
62 their longevity and (iii) how they correlate with the better-documented parts of the orogen along
63 strike to the east in Iran and farther into the Himalayas, and to the west into Turkey and the
64 eastern Mediterranean.

65 Until a few years ago it was considered that in terms of arc activity, the Iraq (Kurdistan) sector
66 was dormant in the Cretaceous but active in the Cenozoic (Jassim et al., 1982, 2006). This is in
67 contrast to the east and west, where there is widespread documentation of Cretaceous arc activity
68 (Agard et al., 2005; Shafaii Moghadam and Stern, 2011, 2015; Shafaii Moghadam et al., 2014).
69 In this paper we assemble data that have been acquired over the last decade to demonstrate that
70 there was actually Cretaceous arc activity in the Iraq (Kurdish) sector, giving rise to a continuity
71 of this activity and subduction along the entire collisional orogeny - from Europe to the
72 Himalayas (Agard et al., 2005; Shafaii Moghadam and Stern, 2011, 2015; Ali et al., 2012, 2013,
73 2014, 2016; Shafaii Moghadam et al., 2014; Ajirilu et al., 2016). Having presented this evidence
74 we can then synthesise the complex mosaic of several subduction zones and arcs in the Iraqi
75 sector of Neotethys over a sixty million year period, and discuss how well these events correlate
76 laterally to the west (Turkey and the eastern Mediterranean) and the east (Iran and the
77 Himalayas). We then use this information to contribute to the big debate concerning long lateral
78 distance synchronicity of events in collisional orogens.

79 **2. General Geology of Iraqi Zagros thrust zone**

80 *2.1. Overview*

81 The NW-SE Zagros Orogen is a result of the closure of the southern Neotethys, located between
82 the Arabian Plate and the Iranian plate, which with the closure of Paleotethys had already
83 accreted onto the Eurasian margin (Alavi, 1994; Talbot and Alavi, 1996; Stampfli and Borel,
84 2002; Ali et al., 2012, 2013, 2014, 2016). Ophiolites, which are Neotethys oceanic lithosphere
85 remnants, occur along the Zagros Orogen. These ophiolites were emplaced along two main belts
86 (Fig 2b; Stöcklin, 1977; Homke et al., 2010; Ghazi et al., 2012; Ismail et al., 2017). The Inner
87 Zagros Ophiolitic Belt through Khoy–Nain–Dehshir –Shahr-e-Babak–Baft (Fig. 2a; Shafaii
88 Moghadam and Stern, 2011) located between the Sanandaj-Sirjan Zone and Central Iran. The

89 Sanandaj-Sirjan Zone (Fig. 2a) contains intensely deformed and metamorphosed rocks in the
90 hinterland of the Zagros Orogen. It is separated from the Zagros thrust zone by the Main Zagros
91 Thrust in the southwest and is bordered by the Urumieh-Dokhtar magmatic arc in the northeast
92 (see Mohajjel et al., 2003; Ferguson et al., 2016). The Sanandaj-Sirjan Zone extends as a NW-
93 SE-trending belt (~1500 km in length and ~150–200 km in width) across southern Iran and
94 consists of Phanerozoic magmatic, metamorphic, and sedimentary rocks (Shafaii Moghadam et
95 al., 2017). The Central Iranian Block (Fig. 2a) is a region of moderate relief surrounded by fold-
96 and-thrust belts, within the Alpine-Himalayan orogenic system of western Asia. The Central
97 Iranian Block records continuous continental deformation in response to the ongoing
98 convergence between the Arabian (Gondwanan) and Turan (Eurasian) plates (Berberian and
99 King, 1981). The Central Iranian Block consists, from east to west, of three major crustal
100 domains: the Lut Block, Tabas Block and the Yazd Block (see for example, Alavi, 1991).

101 The Outer Zagros Ophiolitic Belt through Hasanbag–Piranshahr–Pushtashan–Bulfat–Mawat–
102 Penjween–Kermanshah–Neyriz–Haji-Abad (Fig. 2b; Shafaii Moghadam and Stern, 2011, Ali et
103 al., 2016; Ismail et al., 2017) is located between the Sanandaj-Sirjan and Zagros thrust zones.
104 The Inner and Outer Zagros Ophiolitic Belts resulted from the closure of both Neotethys oceanic
105 crust *sensu stricto* and backarc basin crust (Alavi, 1994; Ghasemi and Talbot, 2006; Casini et al.,
106 2011; Ghazi et al., 2012; Saccani et al., 2013).

107 The Inner Zagros Ophiolitic Belt was emplaced by obduction associated with closure of the
108 southern Neotethys back-arc basin that existed between Central Iran and the Sanandaj-Sirjan
109 Zone (Glennie, 1992; Ghazi and Hassanipak, 1999; Golonka, 2004; Ghasemi and Talbot, 2006;
110 Shafaii Moghadam et al., 2009; Ghazi et al., 2011, 2012). The Outer Zagros Ophiolitic Belt
111 crops out along the southern Neotethys suture, and is the result of the Late Cretaceous
112 convergence and collision between the Sanandaj-Sirjan Zone and the Arabian Shield (Fig. 2).
113 With the exception of the Kata-Rash volcanic group, which is located 10 km west of the suture,
114 the Iraqi segment of the Outer Zagros Ophiolitic Belt encompasses four ophiolite complexes
115 (i.e., Pushtashan, Bulfat, Mawat and Penjween) within the Upper Allochthon (Aziz, 2008;
116 Aswad et al., 2011; Aziz et al., 2011) in the North Zagros thrust zone.

117 The North Zagros thrust zone forms the north-east boundary of a wide region of deformation
118 within the Zagros thrust zone (Fig. 2a). The North Zagros thrust zone, in particular, is considered

to be the final suture between the Arabian and Eurasia plates (Alavi, 1994; Agard et al., 2005). Detailed stratigraphic relations show that the folded rocks and accretionary wedge cropping out within the Iraqi segment of the Zagros thrust zone comprise mainly Mesozoic successions that were partially overthrust by the Lower Allochthon thrust sheet composed of Walash-Naopurdan volcano-sedimentary rocks with back-arc/arc affinities (Fig. 2b; Ali et al., 2013). These were both over-ridden by the Gimo-Qandil Group in the Upper Allochthon (Fig. 2b). The contact between the two allochthons is marked by a 25–50 m thick highly sheared serpentinite horizon (Aziz, 2008; Aziz et al., 2011) with K-Ar hornblende ages of 80–110 Ma (Aswad and Elias, 1988). Serpentinite also occurs at the base of the Lower Allochthon. As the post-collision compressive tectonics progressed, there was disruption of the thrusting sequence. For example, in the Hasanbag and Qalander areas, as well as at Choman (Fig. 2b), the Lower Allochthon is dismembered into two slices; the volcano-sedimentary Naopurdan Group which is overlain by the volcano-sedimentary Walash Group (e.g., Jassim et al., 2006; Ali et al., 2017). With these exceptions, in other areas without tectonic partitioning, the Naopurdan Group passes laterally into the volcano-sedimentary Walash Group. These rocks were collectively named the volcano-sedimentary Walash-Naopurdan Groups (Aswad et al., 2013). Thus the Zagros thrust zone reflects shortening of thick sequences of parautochthonous/autochthonous sedimentary rocks from the northeastern margin of the Arabian platform, forming complex accretionary terrane–flexural foreland basin assemblages (Aswad et al., 2013; Ali et al., 2014; Figs 3, 4).

2.2. *Hasanbag complex*

The Hasanbag igneous complex (Figs 2b, 5) is preserved as a nearly 700 m thick volcanic sequence dominated by calc-alkaline andesite and basaltic andesite, crosscut by microgabbroic and dioritic dykes (Ali et al., 2012). The complex tectonically overlies the late Oligocene volcano-sedimentary Walash-Naopurdan Groups. Both these units are allochthonous, and were emplaced over the late Cenozoic molasse (Red Bed Series) during the late Miocene (Ali et al., 2012).

2.3. *Pushtashan complex*

148 The Pushtashan complex (Figs 2b, 4b, 5) is an assemblage of basalt, andesite, gabbro,
149 norite and serpentinite that during the Late Miocene was thrust over both the late Oligocene
150 volcano-sedimentary Naopurdan Group and the late Cenozoic molasse (Red Bed Series) (see
151 fig. 2 in Ismail et al., 2017) The Pushtashan complex is tectonically overlain by the Cretaceous
152 Gimo-Qandil Group (Ismail et al., 2017). Intrusive rocks outcrop with a maximum width of 6
153 km, and are located in the Pushtashan valley (Buday, 1975). They consist mainly of norite
154 with irregular small lenses of gabbro (and hornblende). These form a sill intruded into the
155 volcanic rocks (Buday and Jassim, 1987). Small granitoid intrusions (mostly trondhjemite) that
156 are 5-40 m thick are present in the upper part of the norite body and the volcanic rocks. SHRIMP
157 U-Pb dating of magmatic zircons from a trondhjemite yielded a mean $^{206}\text{Pb}/^{238}\text{U}$ age of 96 ± 2 Ma
158 (Cenomanian; Ismail et al., 2017).

159

160 2.4. Bulfat complex

161 The Bulfat complex (Figs 2b, 5) has experienced medium-grade regional metamorphism,
162 overprinted by a Paleogene contact metamorphic aureole (Jassim and Goff, 2006). In this
163 complex, voluminous and diverse Paleogene gabbro-diorite intrusions crop out at Jabal Bulfat,
164 referred to as the Bulfat Igneous Complex (Jassim et al., 2006). At least two types of intrusions
165 are distinguished (Buda, 1993). An older intrusion is composed predominantly of olivine-free
166 gabbro and diorite intruded by the younger, smaller bodies of olivine gabbro-diorite. A minor
167 body of ultrabasic rocks and very limited occurrence of acid differentiates is also present (Buda,
168 1993; Jassim and Goff, 2006, Aswad et al., 2016). The Bulfat igneous bodies form sheet-like
169 intrusions along lithological layering of the host sedimentary rocks, which are calcareous and
170 argillaceous sediments of Late Cretaceous (Albian-Cenomanian) age (Jassim et al., 2006).
171 Contact metamorphic rocks occur on the highest summits (~2340 m) forming roof pendants
172 down into the Bulfat Igneous Complex (Buda, 1993). Xenoliths of the country rocks with very
173 high-grade thermal metamorphism up to the pyroxene hornfels facies are characteristic of the
174 Bulfat complex (Buday, 1980).

175

176 2.5. Mawat complex

177 In the Mawat ophiolite complex (Mohammad and Qaradaghi, 2016), the ophiolite cover (referred
178 to as the Gimo-Qandil Group; Figs 2b, 3, 5) is represented by meta-volcano-sedimentary
179 sequences upon which rest directly an ophiolitic, gabbroic, doleritic dyke complex of Albian-
180 Cenomanian age (Aswad and Elias, 1988, Aswad et al., 2011), thereby preserving a complete,
181 albeit tectonised, ophiolite sequence. The volcanic units represent an arc magmatic episode,
182 which was dominated by boninitic and tholeiitic arc lavas (Aswad and Elias, 1988). However,
183 the regionally metamorphosed volcano-sedimentary sequence of Albian-Cenomanian age in the
184 Upper Allochthon was intruded throughout the middle-late Eocene by a wide spectrum of arc-
185 related Paleogene plutonic rocks (referred to as Bulfat Arc Plutonism; Aswad and Elias, 1988,
186 Aswad et al., 2011; Mohammad and Qaradaghi, 2016). Due to superimposed thrusting, the
187 Mawat complex stratigraphy is reversed, with pillow basalt at the bottom (10-30 m thick),
188 amphibolite gabbro in the middle intruded by minor silicic, intermediate and mafic bodies (2500-
189 3000 m thick), and ultramafic rocks (harzburgite and dunite) at the top (200-400 m thick).
190 Numerous pyroxenite dykes cutting the ultramafic unit are also observed (Mohammad and
191 Qaradaghi, 2016).

192

193 2.6. *Kata-Rash arc fragment*

194 The volcanic Kata-Rash arc fragment assemblage (Fig. 2b) is located on both sides of the Shalir
195 valley on the eastern side of the Main Recent Fault (marking the Neotethys suture). This
196 volcanic group is isolated east of the Main Recent Fault as a separate slice, overriding the
197 moving Iranian plate as the result of late Plio-Quaternary dextral strike slip (Ali et al., 2016).
198 According to recent geochronological-geochemical data (Ali et al., 2016), it is considered to be
199 another fragment of the ophiolite-bearing Upper Allochthon terranes. The Kata-Rash arc
200 fragment rocks are bimodal in character, with predominantly basalt-andesite and lesser amounts
201 of rhyodacite, rhyolite and granitic dykes (Ali et al., 2016).

202

203 2.7. *Penjween igneous complex*

204 In the Penjween region, the Late Cenozoic molasse (Red Bed Series) is overlain by the Penjween
 205 igneous complex (Figs 4a, 5), as a result of syn- to post-collision compressive tectonics. The
 206 Penjween igneous complex is bounded on the east by phyllite and calc-schist of metamorphosed
 207 Gimo-Qandil rocks, and to the west by the Cenozoic clastic Red Bed Series (Fig. 2b). The
 208 Penjween igneous complex consists of ophiolitic Alpine-type peridotite (dunite, harzburgite and
 209 lherzolite) surrounded by highly sheared serpentinite (Aziz et al., 2011) that is underlain by
 210 layered gabbro several kilometres in extent and diorite with small dykes of diorite and pyroxenite
 211 (Figs 2-3; Jassim & Goff, 2006). Amphibolite rocks appear in a discontinuous outcrop or as pods
 212 and lenses in sharp contact with peridotite and serpentinite (Ridha et al., 2013). The complex is
 213 in contact with metamorphosed limestone of the Gimo-Qandil Group at the northeastern end.

214

215 **3. Geochronology and radiogenic isotope tracers**

216 *3.1. At least sixty million years of igneous activity*

217 The recent geochronological results summarised below provide much better constraints on the
 218 age of the allochthonous arc and ophiolitic assemblages. Initial studies employed only K-Ar
 219 and Ar-Ar mineral geochronological methods and documented only Cenozoic magmatic activity
 220 (Aswad and Elias, 1988, Aswad et al., 2011; Ali et al., 2012, 2013, 2016; Aswad et al., 2013,
 221 2016). Over the past decade additional Ar-Ar data and the first U-Pb zircon work has revealed
 222 more detail of the Cenozoic activity and also proved the presence of Cretaceous arc-related
 223 magmatism (Azizi et al., 2011; Shafaii Moghadam and Stern, 2011, 2015; Ali et al., 2012, 2013,
 224 2016; Aswad et al., 2013, 2016; Shafaii Moghadam et al., 2014). These ages can be integrated
 225 with a growing database of whole rock Nd and zircon Hf isotopic data.

226 New geochronological data indicates that the tectonic evolution in the Zagros thrust zone was
 227 much more complex than previously thought (Saura et al., 2011). Ar-Ar geochronological data
 228 from feldspar and hornblende (Aswad and Elias, 1988, Aswad et al., 2011, 2013, 2016; Ali,
 229 2012; Ali et al., 2012, 2013) indicate that the pre-collisional evolution of the Upper
 230 Allochthons can be divided into two stages, apparently with a lengthy hiatus between them.
 231 This geochronological pattern can be related to distinct tectonic and volcanic signatures: 1)
 232 Albian-Cenomanian boninitic and island arc volcanism that continued at least into Campanian

time; and 2) middle Eocene-early Oligocene arc magmatism of the Bulfat Igneous Complex. In the Lower Allochthon, the ~43 to ~24 Ma total duration of arc volcanic activity represented by the volcano-sedimentary Walash-Naopurdan Groups outlasted the 45-35 Ma emplacement duration emplacement of the Bulfat Igneous Complex into the Upper Allochthon (Aswad et al., 2016; Karo et al., in press).

We will first focus on the Albian-Cenomanian ophiolite-bearing Upper Allochthon ($^{40}\text{Ar}/^{40}\text{K}$ age; 118-97 Ma; Aswad and Elias, 1988; Aswad et al., 2011). In the Iraqi segment of the North Zagros thrust zone, the Upper Allochthon incorporates multiple episodes of arc magmatic components including juvenile protoarc and forearc settings with strong boninitic affinities found at Mawat (Farjo, 2006), Bulfat and in the Hasanbag island arc volcanic and subvolcanic rocks ($^{40}\text{Ar}/^{39}\text{Ar}$ age 106-92 Ma, Ali et al., 2012).

3.2. Hasanbag ophiolite

^{40}Ar - ^{39}Ar dates on Hasanbag magmatic hornblende (kaersutite) from arc-related volcanic arc units also indicate an Albian-Cenomanian age (106-92 Ma; Ali et al., 2012). The similarity in ages with other Upper Allochthon ophiolite complexes might indicate that the Hasanbag igneous complex represents a slice of the ophiolite cover.

3.3. Pushtashan ophiolite

Euhedral magmatic zircons in a trondhjemite from the Pushtashan assemblage (Figs 2b, 5) yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ of 96 ± 2 Ma and their initial ϵHf value of $\sim +14$ falls just below the depleted mantle linear evolution line, in the field for arc rocks with minimal contamination by older crust (Ismail et al. 2017). There is no evidence (e.g., corroded ~96 Ma cores with younger mantles) that the zircons are pre-magmatic xenocrysts. The integrated study indicates that 96 Ma is the magmatic age of the trondhjemite, establishing the Pushtashan ophiolite is Cretaceous.

3.4. Bulfat ophiolite

For the Bulfat Igneous Complex (Figs 2b, 5), Ar-Ar geochronological data indicate a lengthy hiatus between Albian-Cenomanian boninitic-island arc volcanism and the subsequent Paleogene gabbro-diorite arc-related Paleogene plutonism that locally intrudes the regionally metamorphosed volcano-sedimentary rocks of the Gimo-Qandil Group (Aswad et al., 2016). There has been an ongoing debate on the tectonic affinity, age and duration of assembly of the Bulfat Igneous Complex. The early geochronological data for the Bulfat gabbros revealed a K-Ar closure (cooling age) for amphibole of ~45 Ma (Jassim et al., 1982). Differentiating the age of the host rocks by K-Ar or Ar-Ar chronology will be complicated by the thermal and metasomatic effects of the large Bulfat intrusions. Aswad et al. (2016) collected samples least contaminated by country rocks from minor intrusions of the granitoid-gabbro suite at Wadi Rashid in the southern part of Bulfat Igneous Complex. This is an offshoot from the main intrusion nearby. Petrographic study of rocks from this minor intrusion shows that the emplacement of gabbroic magma was coeval with cooler granitoid magma. This presumably rapid cooling interrupted the assimilation of country rocks and caused the observed bimodality (Aswad et al., 2013). Ti-rich Fe-biotite and Ti-rich pargasite separates from the granitoid and gabbroic rocks were dated by $^{40}\text{Ar}/^{39}\text{Ar}$ at 38.87 ± 0.24 Ma and 39.23 ± 0.21 Ma (Aswad et al., 2016). The cooling age for contact metamorphism in the Bulfat area was obtained by $^{40}\text{Ar}/^{39}\text{Ar}$ dating of biotite in a metapelite which gave 34.78 ± 0.06 Ma (Karo, 2015; Karo et al., in press).

279

280 3.5. Mawat ophiolite

A similar age for intra-oceanic arc activity is also preserved within the Daraban leucogranite intrusion, a small body in the Mawat ophiolite. This unit displays different ages using different dating methods ($^{40}\text{Ar}/^{40}\text{K}$ age 37.7 ± 0.3 Ma, Mohammad et al., 2014; U-Pb zircon dating 97 ± 6 Ma, Mohammad and Cornell, 2017). According to Miller et al. (2003), the Daraban granitoid is classified as a cold granite type ($< 800^\circ\text{C}$), with an expected inheritance-rich zircon population. Thus, the U-Pb zircon age of 96 ± 6 Ma could be inherited from the country rocks (Mohammad and Cornell, 2017). These data for the Daraban leucogranite intrusion correspond to two thermotectonic events 59 Million years apart (Mohammad and Cornell, 2017).

289

290 3.6. *Penjween ophiolite*

291 The Penjween ophiolite has essentially the same rock sequence as the Mawat ophiolite, apart
292 from the absence of volcanic units. The petrography of the Penjween and Mawat ophiolite
293 bodies was studied by Jassim (1973), Al-Hassan (1975) and Jassim and Al-Hassan (1977).
294 Recent zircon U-Pb dating of Penjween plagiogranite yielded 93.8 ± 0.7 Ma (Abdulla, 2015).
295 This age is statistically indistinguishable from the U-Pb zircon age of a Mawat leucogranite
296 intrusion at 97 ± 6 Ma (Mohammad et al., 2016). One of the key outcomes of these studies is
297 the concordance of age between the comparable sequences of the two distally separated (40-
298 50 km) ophiolitic bodies. By removal of the Miocene superimposed tectonic events (i.e.
299 folding, gravity sliding and wrench displacement), these distal ophiolitic bodies might be
300 derived from a single slab of Mesozoic Allochthons oceanic supra-subduction ophiolite
301 (Aswad, 1999).

302

303 3.7. *Serpentinite allochthon boundaries*

304 The contact between the Albian-Cenomanian ophiolite-bearing Upper Allochthon terranes with
305 underlying Lower Allochthon rocks is marked by a 25-50 m thick highly sheared serpentinite
306 horizon (Aziz, 2008; Aziz et al., 2011) from which ages of 80-110 Ma have been obtained using
307 Rb-Sr and Sm-Nd methods (Aziz et al., 2011). Ophiolitic mélange serpentinite also occurs on the
308 sole of the Lower Allochthon and gives mixed Rb-Sr ages (150, 200 and 770 Ma) with initial
309 ϵ_{Nd} values down to -30 (Aziz et al., 2011). The negative initial ϵ_{Nd} values are indicative that the
310 analysed samples contain material with prolonged crustal residence probably derived from the
311 Arabian microcontinent margin. The Mesozoic ages of these samples provide only a *maximum*
312 age of ophiolitic mélange serpentinite formation.

313

314 3.8. *Parautochthonous and allochthonous sedimentary sequences*

315 In the accretionary complex terrane, the Qulqula-Kermanshah Radiolarite (Fig. 2b) is highly
316 deformed in most of northeastern Iraq. Therefore, its age is controversial and has not yet been
317 precisely determined. Its chronostratigraphic equivalent in Iran (the Kermanshah Radiolarites)
318 was recently dated using radiolarian biostratigraphy, and provides an age range from early

319 Pliensbachian to Turonian (Gharib and De Wever, 2010) and thus partly overlaps in age with the
320 Albian-Cenomanian ophiolite-bearing Upper Allochthon terrane (Aswad and Elias, 1988).

321 Detrital zircon U-Pb age spectra for the Mesopotamian Foreland Basin, southwest of the Outer
322 Zagros Ophiolitic Belt (Fig. 2), provide new geological information about the provenance of the
323 Neogene molasse deposits (Koshnaw et al., 2017). For example, 31 zircons from sample
324 12KRD-105 in the Injana Formation show age probability peaks at ~18, 28, 42 and 100 Ma
325 (Koshnaw et al., 2017, their fig. 11). These early Miocene-late Oligocene, middle-late Eocene
326 and Albian-Cenomanian ages strongly suggest an allochthon provenance for most of the detrital
327 zircons in this Neogene basin fill. In contrast, Koshnaw et al. (2017) noted that the detrital zircon
328 U-Pb age spectra in the overlying Mukdadiya (Lower Bakhtiari) and Bai-Hasan (Upper
329 Bakhtiari) Formations show nearly exclusive derivation from the Paleogene volcano-sedimentary
330 Walash-Naopurdan Groups near the Iraq-Iran border. This is inferred by the absence of the ~100
331 Ma signature and a decrease in ~300 and ~600 Ma ages which would have been derived from
332 recycled Paleozoic sediments from NNW of the basin (with material ultimately derived from the
333 Gondwana Arabian microcontinent). These rapid shifts were interpreted by Koshnaw et al. (2017)
334 to be indicative of a pronounced Neogene event that altered the topography and drainage to the
335 Zagros foreland in a way that cut off sediment input from ophiolitic rocks.

336 In the Galalah-Choman area (Fig. 2b), SHRIMP U-Pb ages were determined on single zircon
337 grains separated from combined samples belonging to a trachytic andesite suite of Walash arc
338 volcanic rocks (Ali and Aswad, 2013). The two grains recovered reveal two episodes of
339 Paleoproterozoic inherited zircon growth: 1953±39 Ma magmatic cores and low Th/U
340 1777±28 Ma metamorphic rims. The core-rim pair of analyses is interpreted as indicating that
341 the zircons are xenocrysts in the Paleogene magma derived from Arabian crystalline basement.

342 At Mawat (Fig. 2b), Walash-Naopurdan andesites have initial $^{143}\text{Nd}/^{144}\text{Nd} = 0.51073$ and
343 initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.70585$ with a Rb-Sr whole-rock isochron age for Walash volcanic rocks of
344 44.35±0.3 Ma (Koyi, 2009). The distinctive whole-rock isotopic characteristics (i.e. the low Nd
345 and high Sr isotope ratios) in the Walash-Naopurdan andesites indicate that these subduction-
346 related rocks were contaminated by craton-derived sediments from a nearby Arabian platform
347 source (Koyi, 2009).

348

349 *3.9. Kata-Rash arc fragment rocks*

350 For the Kata-Rash arc rocks, SHRIMP zircon U-Pb geochronology on the Harbar volcanic
351 rocks (mostly rhyolite-rhyodacite) and the Aulan intrusion (mostly granodiorite) indicates an
352 Albian- Cenomanian time of igneous crystallisation with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of
353 107.7 ± 1.9 Ma (Ali et al., 2016). These zircons have a $+8.6 \pm 0.2$ (2σ) initial ε_{Hf} signature,
354 indicating the strong juvenile component of these rocks (Ali et al., 2016). The ages of both
355 Pushtashan and Kata-Rash arc rocks are consistent with previous results from the Hasanbag
356 and Mawat areas, perhaps extending the length of a coeval allochthon to >100 km.

357

358 *3.10. Walash-Naopurdan and Bulfat Igneous Complex rocks*

359 In the Lower Allochthon, Ar-Ar hornblende ages of Walash-Naopurdan volcanic rocks decreases
360 towards the northwest, ranging from ~ 44 to ~ 32 Ma at Mawat and ~ 33 to ~ 24 Ma in the
361 Qalander and Sheikhan areas (Figs 2b, 4c), suggesting that the arc volcanicity was propagating
362 northward during the late Oligocene (Ali et al., 2012, 2013). Thus arc-related subduction was
363 diachronous along the Iraqi segment length of the North Zagros thrust zone and took place in a
364 scissor-like fashion with the suture zone younging northwestwards through the Iraqi segment of
365 the North Zagros thrust zone (Ali et al., 2012, 2013; Aswad et al., 2016).

366 The Paleogene magmatic arc episodes are not restricted to the Lower Allochthon; they are also
367 manifested extensively across what is now described as Upper Allochthon. This is referred to as
368 middle-late Eocene (45-33 Ma) Bulfat Arc Plutonism. However, the duration of Walash-
369 Naopurdan arc volcanism considerably outlasted the duration of the latter. Despite the disparate
370 tectonic and petrogenetic characteristics, the ages of the stated arc tectonic domains of the two
371 allochthons point consistently to a dual subduction-zone system with the timing of magmatism
372 peaking between ~ 45 and ~ 33 Ma giving around 12 Ma between ages of these arc tectonic
373 domains.

374

375 **4. Metamorphism**

376 4.1. Seeking convergent plate boundary metamorphic signatures

377 Convergent plate boundaries may exhibit paired metamorphic belts (Miyashiro, 1974) of high
378 pressure- low temperature and low pressure-high temperature. In the Iraqi Zagros, low pressure-
379 high temperature arc metamorphic assemblages are recognised, but the high pressure- low
380 temperature rocks formed in the subduction channel are yet to be found *in situ*.

381 Studies of metamorphism in the Iraqi Zagros thrust zone were carried out by Bolton (1956,
382 1960), continued by Jassim (1973), Al-Hassan (1975) and Buda and Al-Hashimi (1977).
383 Metamorphic events within the Iraqi Zagros ophiolites can be summarised as follows: Firstly,
384 hydrothermal metamorphism of oceanic volcanic rocks occurred in the Early-Late Cretaceous.
385 Secondly, regional metamorphism occurred during Late Cretaceous deformation in the
386 Neotethys segments and in the Sanandaj-Sirjan zone (Fig. 2a). Thirdly, contact metamorphism of
387 country rocks by syn- to post-tectonic gabbros of the Bulfat complex occurred in the Paleogene
388 (Jassim and Goff, 2006). According to Buday and Suk (1978), the Penjween, Mawat and Bulfat
389 igneous complexes were all affected by the same metamorphic history. However, the variation of
390 metamorphic facies from albite-epidote hornfels to pyroxene hornfels is related to the nature of
391 the intrusive rocks emplaced into host volcanic and sedimentary sequences (Jassim et al., 1982),
392 the distance from the igneous intrusion and the amount of denudation (highest at Bulfat and
393 lowest at Penjween).

394

395 4.2. Low pressure - high temperature metamorphic rocks

396 Thermal metamorphism around the Mawat and Penjween intrusive rocks has been described by
397 Jassim and Al-Hassan (1977). The high-grade thermal metamorphic influence on the Gimo-
398 Qandil Group caused by the Eocene igneous intrusions around Bulfat was studied by Buda et al.
399 (1978). The width of this high-grade thermal aureole (pyroxene hornfels facies) is about 150 m
400 (Jassim et al., 2006). Aswad and Pashderi (1984) studied impure carbonate xenoliths within the
401 Eocene gabbroic intrusion and they concluded that the temperature of contact metamorphism in
402 the xenoliths reached 850°C. Al-Taweel (2012) studied the regionally metamorphosed rocks
403 (mainly impure and marly limestone) in the Gimo-Qandil Group, near Penjween, and showed
404 that they experienced two metamorphic episodes. The early metamorphic episode was Buchan-

type metamorphism characterized by the static growth of minerals (mainly diopside-hedenbergite, grossular and epidote) without any foliation. The temperature range during this early episode was determined to be between 420-720°C at a pressure ~5 kbar, indicating a steep thermal gradient, whereas thermodynamic data show that the dominant Paleocene peak metamorphism conditions in the Bulfat metapelites are in the range of 605-671°C and 2.8-3.5 kbar (Karo, 2015). The following metamorphic episode was Barrovian-type, foliating phases of the early metamorphic episode with growth of new amphibole and zoisite. Such paleogeothermal gradients are mostly found in the Albian-Cenomanian tectonic regimes. Mineral reaction kinetics in the late episode favour of clockwise P-T paths and were probably due to post-Albian-Cenomanian subduction-related metamorphism. Lacking sufficiently detailed geochronologic data, the interpretations based on stated clockwise P-T paths should be viewed with caution.

Jassim and Al-Hassan (1977) proposed polyphase metamorphism in the Mawat complex. This started with regional metamorphism, followed by contact metamorphism due to emplacement of gabbroic rocks and then dynamic metamorphism related to the subsequent tectonic transport. In the Penjween complex regional metamorphism (mostly giving rise to phyllites and schists) has been documented by Buday and Jassim (1987). Previously Buda and Al-Hashimi (1977) identified hydrothermal, dynamic and contact metamorphism in the Mawat complex.

422

4.3. The search for high pressure - low temperature metamorphic rocks

Due to the low preservation potential of low temperature-high pressure metamorphic rocks that formed in subduction channels (Agard et al., 2001; Korh et al., 2009), the volume of such rocks found in convergent plate boundary orogens is often small. Added to this, the Zagros Orogen is still in active convergence, without the development of orogenic collapse and extensional overprints that are often responsible for exhuming these rocks from depth. Thus as yet, no extant blueschist or eclogite facies rocks have been observed in Iraq. However, there are reports of high pressure indicator minerals such as glaucophane as detritus in drainage systems that traverse serpentinite mélange (Jassim and Goff, 2006; Aziz et al., 2011). Therefore, we are confident that eventually eclogites or blueschists will be located, probably as xenoliths within serpentinite mélange.

434

435 **5. Mesozoic Igneous Rocks**

436 The Iraqi Kurdistan ophiolite associations are generally tectonically partitioned, and consist of
437 some or all of the following components: mantle tectonite (locally lherzolite or harzburgite
438 commonly altered to serpentinite), gabbro, dolerite dyke complexes, locally with trondhjemite
439 dykes and volcano-sedimentary sequences (e.g., pelagic sedimentary rocks interlayered with
440 basaltic volcanic rocks). Post-subduction processes such as allochthon emplacement and
441 subsequent folding and faulting fragmented these ophiolitic components, resulting in a record
442 that is incomplete to varying degrees (Fig. 3). Additionally, fragments of ophiolites of different
443 ages can be interleaved by the out-of-sequence slicing.

444 The Mawat and Penjween complexes represent the largest and most complete ophiolitic
445 sequences, i.e., mantle tectonite, gabbro, spilitised dolerite dyke complexes and volcano-
446 sedimentary rocks are all present (Fig. 3). However, these are severely tectonically disrupted
447 such that their upper mantle and deep crustal components overlie the volcanic parts. The mantle
448 tectonite consists of lherzolite, dunite, troctolite and wehlite, reminiscent of slow-spreading
449 modern oceanic crust (Aqrawi, 1990). Detailed geochemical investigations on spilitised dolerite
450 of the Mawat ophiolite complex led to the recognition of a mid-ocean-ridge affinity (Aswad,
451 1999). The volcanic units at Mawat show strong boninitic affinities (Fig. 6), a signature
452 associated with initiation of subduction (Farjo, 2006). These are associated with rocks of
453 volcanic arc basalt affinity (Farjo, 2006). The entire assemblage may have formed in the proto-
454 forearc environment, which witnessed eruption of MORB-like lavas followed by volcanic arc
455 basalts and boninites (Farjo, 2006).

456 The Pushtashan assemblage has volcanic rocks in its upper part (Ismail et al., 2017). These were
457 intruded by plutonic rocks, which are largely of gabbro and norite, but with rare associated
458 trondhjemitic-granodioritic veins. The gabbro and norite have serpentinised peridotites at their
459 base (Fig. 3). The petrography, mineral chemistry and whole-rock compositions of the
460 Pushtashan gabbros are typical of tholeiite magma crystallised under low-pressure conditions,
461 whereas the norite suite has trace element signatures typical of calc-alkaline magma, suggesting
462 that their source included mantle metasomatised by fluids released from subducted oceanic crust

463 (Ismail et al., 2017). The major, trace and REE geochemistry and field relations indicate that the
464 felsic rocks formed in a supra-subduction zone setting (Ismail et al., 2017).

465 The Kata-Rash volcanic and plutonic rocks (Ali et al., 2016) include basalt, basaltic-andesite,
466 andesite, rhyodacite to rhyolite, while intrusions are mostly granodiorite and granite. SHRIMP
467 zircon U-Pb geochronology on granodiorite indicates a 108 ± 3 Ma crystallisation age (Albian-
468 Cenomanian). On the Nb/Yb-Th/Yb diagram, all Kata-Rash samples fall within the
469 compositional field of arc-related rocks, i.e. above the mid-ocean-ridge basalt ocean island basalt
470 mantle array (Ali et al., 2016).

471 Despite their disparate tectonic and geographic positions, the geochronological-geochemical
472 data from the Kata-Rash volcanic group (in Ali et al., 2016) show a close match with the Mawat
473 Ophiolite Complex (i.e. $^{40}\text{Ar}/^{40}\text{K}$ age 118–97 Ma, Aswad and Elias, 1988), which is located on
474 the western side of the Main Recent Fault. Therefore, the Kata-Rash arc fragment is considered
475 to be part of the ophiolite-bearing Upper Allochthon terrane (designated as the Gimo-Qandil
476 Group). The geochronological-geochemical data from the Kata-Rash arc rocks are quite similar
477 to that of other Late Cretaceous Tethyan ophiolites within the Zagros thrust zone (i.e.,
478 Pushtashan, Bulfat, Mawat and Penjween. Kata-Rash arc rocks, therefore, are considered to be
479 an integral part of Outer Zagros Ophiolite Belt.

480 The Bulfat ophiolite complex is composed mainly of mafic rocks intruded into the Bulfat Group
481 with minor amounts of ultramafic rocks (Fig. 3). The northern part of the complex consists of the
482 regionally metamorphosed Gimo-Qandil Group. Analyses of the Bulfat complex mafic rocks
483 indicate that sub-alkali gabbros are predominant with high-alumina and low contents of TiO_2 ,
484 K_2O , P_2O_5 and MgO . The low Zr in most of the gabbroic rocks may indicate that geochemically
485 they crystallised from primitive magma (Jassim and Goff, 2006). The south eastern part of the
486 Bulfat complex is intruded by Eocene plutonic rocks (Aswad et al., 2016).

487

488 **6. Cenozoic Igneous Rocks**

489 Throughout the middle-late Eocene, the volcano-sedimentary Gimo-Qandil Group was intruded
490 by a wide spectrum of arc-related plutonic rocks (referred to as Bulfat arc plutonism) with ages

491 ranging from ~45 to ~35 Ma (Aswad et al., 2016; Karo et al., in press). The younger date
492 probably marks the end of the tectonic events exclusive to the ophiolite-bearing terranes.
493 Continued convergence in this area resulted in a new subduction zone in proximity to the
494 encroaching Arabian plate within the Iraqi segment of the North Zagros thrust zone, giving rise
495 to the ~43 to ~24 Ma Walash-Naopurdan arc volcanism (Ali et al., 2013; Aswad et al., 2013,
496 2016). It is important to stress that the Paleogene magmatic arc episodes are not restricted to the
497 Upper Allochthon but rocks of these episodes are also present in the Lower Allochthon (Fig. 7).
498 The duration of arc magmatism considerably outlasted the duration of Bulfat arc plutonism.
499 Based on geochemical data, the Iraqi segment of the North Zagros thrust zone is characterized by
500 the presence a dual arc-related Paleogene magmatism, namely the Bulfat arc plutonism and
501 Walash-Naopurdan arc volcanism. The latter is distinguished from the former by its high Nb/Yb,
502 confirming the inclusion of continental crust into this late Paleogene magmatic suite.

503 The Walash rocks are basalt to basaltic andesites, and alkali basalts to trachytic andesites, with
504 calc-alkaline and alkaline character. The Naopurdan rocks include basalts, andesitic basalts and
505 andesites of sub-alkaline, low-K tholeiitic character. The enrichment in LREE versus HREE and
506 high Th/Nb and Nb/Zr show that the Walash and Naopurdan rocks have distinct subduction-
507 related signatures: specifically calc-alkaline to alkaline and island-arc tholeiite, respectively (Fig.
508 6; Ali et al., 2013).

509

510 **7. Tectonic Evolution**

511 The study region has an added complexity- it is an active war zone. Therefore structural data is
512 not as comprehensive and as dense as in many other orogens. The fundamental findings are that
513 there are both Cretaceous and Cenozoic ophiolitic arc assemblages with a spread of 60 million
514 years in age. Only further work will reveal if magmatism over this period was episodic or
515 continuous, and by fine-scale structural observations, elucidate more detailed information on the
516 tectonic evolution.

517 The tectonostratigraphic components within the Zagros Mountains reflect processes within the
518 Neotethys ocean during convergence of the Arabian and Eurasian continents (Fig. 7) and their
519 subsequent collision prior to the rifting of an Iranian microcontinent from the rest of southwest

Iran. Convergence began in northwestern Iran during the Permian to Triassic (Fig. 7a), but in the Sanandaj-Sirjan Zone it began later during the Late Triassic (Mohajjel et al., 2003; Alavi 1994). During the Early Cretaceous, subduction of part of the Neotethyan oceanic crust toward the east and northeast developed an ophiolite complex (e.g. Hasanbag 106-92 Ma see Fig. 7b; Ali et al., 2012). The Iraqi Zagros ophiolite crops out along the southern Neotethys suture. It includes the Hasanbag-Pushtashan-Bulfat-Mawat-Penjween complexes, and is the result of the Late Cretaceous closure and collision between these ophiolites and the Arabian Plate (Figs. 1-2, 7c; Ali et al., 2012).

During the Eocene, two island arcs were present within the Iraqi Zagros Neotethys (Fig. 7d). An additional deformation event took place in the Miocene (Fig. 7e) as more ocean crust was emplaced onto the Arabian margin and subsequently folded, as documented by Agard et al. (2005) and Ali et al. (2013). The Arabian plate then began to converge closer to the Eurasian continent (Fig. 7e), probably due to rifting of the Arabian continent from Africa along the Red Sea and the Gulf of Aden (Laughton, 1966). Final continent-continent collision (Fig. 7f) occurred in the middle Miocene (Jackson et al. 1995; Ali et al., 2012).

The recent study by Lawa et al. (2013) combined structural, stratigraphic and paleontological data, together with new field observations, to investigate the tectonostratigraphic evolution of the Kurdistan foreland. A major objective of this study was the propagation of the deformation front from the Zagros Imbricate zone in the northeast, towards the Mesopotamian foredeep to the southwest. Lawa et al. (2013) documented six unconformities within the Kurdistan foreland basin succession: Turonian (base-AP9; 92 Ma); Danian (base-AP10; 65 Ma); Paleocene–Eocene (intra- AP10; 55 Ma); late Eocene (top-AP10; 34 Ma); middle-upper Miocene (a local unconformity; intra-AP11; 12 Ma); and Pleistocene. Lawa et al. (2013) suggested that these unconformities can be divided into two groups; obduction-related (Turonian, Danian, and Paleocene-Eocene); and collision-related (late Eocene, middle-upper Miocene, and Pleistocene). Geochronological data presented in Lawa et al.'s study clearly indicates that the arc magmatic episodicity of the Iraqi segment of the North Zagros thrust zone caused an oscillatory compressional deformation of the Tethys continental shelf of the passive Arabian margin, with an apparent interplay between intra-oceanic arc magmatism - compressional dynamics. Consistently, our study confirms the assumption that the Turonian, Danian, and Paleocene-

550 Eocene unconformities were related to ophiolite obduction (See Fig.7c). Based on the above
551 discussion, it is clear that the region experienced long-lived arc-related tectonic activity of at
552 least 60 million years, probably episodic in nature (See Fig.7). This had a profound influence
553 upon the rate of propagation of the deformation front during the evolution of this segment of the
554 Zagros Orogen.

555

556 **8. Discussion**

557 *8.1. Cretaceous convergence*

558 The Arabian passive microcontinent margin (Gondwana provenance) collided with an island-
559 arc in the Late Cretaceous (Golonka 2004; Kazmin and Tikhonova 2006; Allahyari et al.
560 2010; Saccani et al. 2013). This event caused obduction of Neotethys oceanic lithosphere
561 onto the Arabian Plate margin (Fig. 7c). However, there remained a shallow water oceanic
562 basin where Late Cretaceous limestone was deposited (Alavi 1994; Ricou 1974; Haynes and
563 McQuillan 1974; Babaei et al. 2005; Ghasemi and Talbot 2006; Yousefirad 2011). In Iran
564 this is known as the Amiran Formation (Agard et al., 2005; Shafaii Moghadam and Stern,
565 2011) and in Iraq it is represented by the Tanjero Formation (Fig. 7c; Ali et al., 2013, 2014).
566 Erosion of collisional highs carried an enormous volume of detrital materials derived from
567 the accretionary terranes into this northwest-southeast-trending shallow water Mesopotamian
568 foreland basin (Alavi, 1994; Agard et al., 2005).

569

570 *8.2. Cenozoic convergence*

571 After accretion of a Cretaceous arc suite and supra-subduction ophiolites onto the Arabian
572 margin, island arc development continued offshore in the shrinking Neotethys, and lasted
573 until the late Miocene (see Ajirlu et al., 2016, their figs 6-9). During the late Oligocene to
574 middle Miocene a series of island arcs accreted onto the Arabian margin as a consequence of
575 two or more subduction zones that consumed the Neotethys oceanic lithosphere. Emplacement
576 of the upper and lower allochthons, including the Walash-Naopurdan arc volcanic groups,
577 occurred along the North Zagros thrust zone. This marked the final phase of consumption of
578 oceanic lithosphere.

According to Ajirlu et al. (2016), the subduction-related magmatic activities in Iran ceased progressively from northwest to southeast, being early Miocene (~22 Ma) in Meghri (Armenia), middle Miocene (~16 Ma) in Kashan, and late Miocene (~10–6 Ma) in Anar (about 200 km south of Tehran). In contrast, in the northern Zagros thrust zone of Iraq, the Ar-Ar ages of Walash-Naopurdan arc volcanism decreased towards the northwest, ranging from ~43 to ~32 Ma at Mawat-Penjween to ~33 to ~24 Ma in the Qalander and Sheikhan areas (about 80 km northwest of Mawat), suggesting that earliest documented arc volcanicity of the middle Eocene occurred at Mawat and that this activity became younger northwestward in the late Oligocene. The Walash-Naopurdan volcanic units are associated with microfossil-rich oceanic sediments whose biostratigraphy confirm that the ages of volcanic units were not reset due to loss of radiogenic Ar during superimposed arc thermal events. Such Ar-Ar age differences (from southeast to northwest) in the Walash-Naopurdan arc volcanism (equivalent to the Kamyaran magmatic arc rocks in Iran; Ali et al., 2014) are considered to be a consequence of an anticlockwise rotation of the Arabian plate with a highly oblique convergence setting (Aswad et al., 2013). The proposed mechanism causing the plate rotation is that spreading in the Red Sea required the lithosphere of Arabia to move anticlockwise relative to the Eurasian plate (e.g., McClusky et al., 2003), thus accounting for the variation in tectonic timing along the Arabia/Iran (Eurasia) convergent boundary. Therefore, the collision in the Qalander-Sheikhan area in the northeast (Fig. 2b) must have started after the middle Miocene. This timing disagrees with that stated by Agard et al. (2005) that the final consumption of oceanic lithosphere took place slightly after 35 Ma and that collision must have started before ~23–25 Ma in the northern Zagros (Kermanshah–Hamedan area; Fig. 2b). The Ar-Ar ages for the arc volcanism in the Qalander-Sheikhan area at the end of the Oligocene (~24 Ma) suggest that a small oceanic domain had probably persisted up to the early Miocene, separating the advancing nappes from the accretionary complex terrane-foreland basin assemblage (Ali et al., 2013, 2014).

8.3. Correlation of ophiolites between Iran and Iraq

The boundary between Iran and Iraq unfortunately represents a marked political divide; therefore there is little work that transcends this national boundary. This has led to some miscorrelation of ophiolites between Iran and Iraq. The development of the Mesozoic arcs in Iran provides vital

610 clues to the evolution of both the Paleo- and Neotethys oceans. Shafaii Moghadam and Stern
611 (2015) provided a valuable synthesis of these Iranian ophiolites. However, some of the age
612 relationships and origins of the continuation of the Zagros ophiolites northwest from Iran into
613 northeastern Iraq (Kurdistan) necessitates some revision to the correlations and synthesis put
614 forward by Shafaii Moghadam and Stern (2015).

615 The Hasanbag (Kurdistan-Iraq) and Kermanshah (Iran) ophiolites occurring along the Iran-Iraq
616 border (Fig. 2) were thought to be Eocene by Shafaii Moghadam and Stern (2015). These
617 included the Walash- Naopurdan volcano-sedimentary Group in fig. 3 in Shafaii Moghadam and
618 Stern (2015) that were related to intra-oceanic subduction within Neotethys, similar to supra-
619 subduction zone ophiolites (e.g., Dilek and Furnes, 2011, 2014). However, the age of the
620 Hasanbag ophiolite arc complex is 106-92 Ma (Albian-Cenomanian) not Eocene as stated by
621 Shafaii Moghadam and Stern (2015). Therefore, as shown by Ali et al. (2013), it is unrelated to
622 the Paleogene Walash-Naopurdan intra-oceanic subduction arc and back-arc in the northwest
623 Zagros thrust zone. Furthermore, Shafaii Moghadam and Stern (2015, their fig. 3) erroneously
624 classified the volcano-sedimentary Walash-Naopurdan Groups as a Paleogene ophiolite (c.f., Ali
625 et al., 2013). However, field, petrographic and geochemical comparisons between the Kamyaran
626 volcanic/subvolcanic rocks in Iran (Braud, 1980; Moinevaziri et al., 2009; Azizi et al., 2011) and
627 the Walash-Naopurdan volcanic/subvolcanic rocks in Iraq (Ali, 2012; Ali et al., 2013) indicate
628 they are equivalent units unrelated to the Cretaceous ophiolite. Based on our field, geochemical
629 and geochronological studies, the Iraqi rocks that were designated as entirely Cenozoic by
630 Shafaii Moghadam and Stern (2015) in fact consist of several unrelated allochthons containing
631 rocks of different ages. As explained in earlier sections of this paper, we have divided these into
632 Upper and Lower Allochthons (Fig. 2b; Ali et al., 2012, 2013, 2014, Aswad et al., 2011). The
633 Upper Allochthon ophiolite-bearing terrane is part of the Outer Zagros Ophiolite Belt. Unlike the
634 Upper Allochthon, the Lower Allochthon has limited extent. Most of these sequences occur
635 within the Iraqi Zagros thrust zone. Their equivalents in Iran occur in the Kamyaran area along
636 the Sahneh-Marivan line (Moinevaziri et al., 2009; Azizi et al., 2011). The following key pieces
637 of evidence suggest these two allochthons represent two different tectonic domains:

638 (i) The two allochthons have different lithostratigraphies. Thus, in addition to the ophiolitic
639 mantle peridotites and oceanic crust of Albian-Cenomanian age (Aswad and Elias, 1988), the

640 Upper Allochthon encompasses a medium-grade regional metamorphosed volcano-sedimentary
641 sequence of Albian-Cenomanian age, whereas the Lower Allochthon is composed of the
642 volcano-sedimentary Walash and Naopurdan Groups (Ali et al., 2017) which consists
643 predominantly of flysch with volcanic interludes and nummulitic limestone (with abundant early
644 to late Eocene microfossils; Jassim et al., 2006).

645 (ii) The igneous rocks in the allochthons differ markedly in their geochemical tectonic signature,
646 with the Upper Allochthon indicating a proximal to paleo-ridge intra-oceanic arc signature,
647 whereas the Lower Allochthon shows a subduction-related signature characterised by a melt
648 component input from craton-derived sediments (Aswad et al., 2011).

649 Although this division can be recognized for considerable distances along strike, such as the
650 Bulfat and Mawat allochthons (Fig. 2b), they can be disrupted by out of sequence thrusting. This
651 introduces uncertainty in correlations that now are being resolved by the expanding
652 geochronological database (Ali et al., 2012, 2013; Aswad et al., 2011, 2013).

653

654 *8.4. Death of the last arcs and the Arabian-Eurasian collision*

655 On the basis of paleomagnetic data (Ricou, 1974), geology and paleogeography (Berberian and
656 King, 1981), closure of the Paleotethys Ocean started in the Carboniferous with the initiation of
657 northward subduction beneath Eurasia. During the earliest Triassic, this was followed by the
658 development of several micro-continents and the formation of the Neotethys Ocean between the
659 Afro-Arabian and Eurasian plates (Stampfli and Borel, 2002). However, Paleotethys was not
660 completely closed by that time. The final closure of the Paleotethys corresponded to the
661 accretion of Cimmerian terranes (such as the Sanandaj-Sirjan Zone) onto Eurasia at the end of
662 the Middle Triassic (Besse et al., 1998). Throughout the middle Cretaceous, the Afro-Arabian
663 and Eurasian plates converged as the Neotethys Ocean shrank. During the Late Cretaceous, the
664 northeastern passive margin of the Arabian plate started to be subducted under Central Iran. This
665 closure of Neotethys was marked by Late Cretaceous obduction (89.3 ± 1.0 to 83.5 ± 0.7 ;
666 Berberian, 1995; Berberian and King, 1981; Falcon, 1974; Ali et al, 2012, 2013) of ophiolites
667 onto the Arabian continental crust. The evolution of the Zagros Orogen continued after the
668 closure of the Neotethys and continent-continent collision. The separation of the Arabian Plate

669 from Africa along the Red Sea, at a rate of ~1.2 cm/year in a north-northeast direction, together
670 with the pull forces acting on the Neotethys slab below the suture, was probably the engine
671 driving the Arabian plate towards Iran that triggered the final Zagros collision in the late
672 Oligocene and early Miocene (Agard et al., 2005).

673 It is deemed significant that the Neotethyan rocks of Paleogene age in the northern Zagros area
674 remained exposed in front of the advancing ophiolite-bearing terrane of Albian-Cenomanian age,
675 separating the latter from the Zagros accretionary complex. Key evidence for Paleogene
676 Neotethyan oceanic activity includes recognition of the duration and migration of volcanic rocks
677 displaying arc affinities. Sedimentation patterns in the basins on the periphery of the Qulqula
678 Radiolarite (which developed into an accretionary wedge) are characterized by shallowing
679 upward depositional sequences overlying deep-water deposits that strongly suggest a
680 contemporaneous relationship existed between reduction of accretionary prism loading and arc-
681 related Paleogene magmatism (Aswad et al., 2013, 2016; Koshnaw et al., 2017). It can be
682 inferred that continuing stress relaxation, probably triggered by plate convergence through
683 extensive Paleogene arc subduction, caused the shift in deposition style from a deep-water flysch
684 basin into a shallow-water molasse basin. Such a reduction in the compressive force could not be
685 attained without diminishing the Arabia-Eurasia convergence rate, implying that the latter was
686 partly absorbed due to the initiation of Paleogene intra-oceanic arc subduction. In this context,
687 the term “end of collision” is defined as the complete eradication of oceanic crust, together with
688 protracted crustal deformation associated with dextral strike slip offset perpendicular and parallel
689 to the Zagros thrust zone (Aswad, 1999). This deformation period coincided with folding and
690 progressive unconformities in the Arabian foreland (Aswad et al., 2013; Lawa et al., 2013; Ali et
691 al., 2014).

692 Sedimentation within the Mesopotamian Foreland Basin region during the late Miocene-Pliocene
693 shows a disruption of the simple sedimentary pattern with upward coarsening of terrigenous
694 molasse sequences (Adeeb, 2006). Furthermore, detrital zircon U-Pb age spectra for the
695 Mesopotamian Foreland Basin provide additional geological information about the provenance
696 of the Neogene molasse deposits (Koshnaw et al., 2017), suggesting an allochthon provenance
697 for most of the detrital zircons in this Neogene basin fill. The latter tectonic event does not

698 preclude a much earlier collision elsewhere between Arabia and the Sanandaj-Sirjan Zone as the
699 result of Arabian plate rotation and/or indentation tectonics.

700 The available evidence summarised above suggests that spatio-temporal constraints of incipient
701 subduction in the Zagros region were caused by an oscillatory compressional deformation of the
702 Tethys continental shelf along the passive Arabian margin, with an apparent interplay between
703 intra-oceanic arc magmatism and compressional dynamics (Koshnaw et al., 2017).

704

705 *8.5. Correlation of events along the orogen from east and west*

706 The Iraqi Zagros thrust zone represents a small but well exposed section of the closure and
707 subduction of the Neotethys Ocean along the continent-continent collision zone between the
708 African-Arabian-Indian plates and the Eurasian plate during the Cretaceous and Cenozoic. The
709 recent recognition of four Cretaceous ophiolite complexes within the Upper Allochthon and
710 forming the Outer Zagros Ophiolite Belt fills a gap in the Cretaceous ophiolite chain that extends
711 from Cyprus to the Himalayas (Fig. 8).

712 In the middle Cretaceous a major period of north-dipping subduction occurred along the Alpine-
713 Himalayan belt with the Neotethys oceanic lithosphere being subducted along the margin of the
714 Eurasian plate. Ophiolites and their associated igneous intrusions are prominent in the western
715 part of the belt while I-type granitoid rocks represent the same event in the Himalaya region, for
716 example in southern Tibet (e.g., Wen et al., 2008). The age of this subduction event is now
717 known throughout the region from Cyprus to the Himalayas (Fig. 8). The oldest part of this age
718 range is in Iraq, Iran and Oman (95-108 Ma). To the west in Turkey (Kizildag) and Cyprus
719 (Troodos) the ages are all 90-92 Ma while to the east in the Kohistan arc of the Himalayas the
720 ages range from ~80-100 Ma. On the basis of the limited number of available analyses this may
721 suggest that subduction started in the central part of the Alpine-Himalayan belt and then spread
722 both to the west and east, as suggested by Shafaii Moghadam et al. (2013).

723 A second period of subduction is recognised in the central and eastern part of the Alpine-
724 Himalayan belt with major activity in the Paleocene and Eocene culminating in the final closure
725 of Neotethys in the Early Miocene. This phase of I-type igneous activity is mainly recognised in

726 volcanic sequences (e.g. Walash-Naopurdan Groups in Iraq, Sahneh in Iran and the southern
727 Gangdese magmatic activity in Tibet) and their associated intrusions (Wen et al., 2008; Shafaii
728 Moghadam et al., 2013; Ali et al., 2013).

729

730 **9. Conclusions**

731 (1) In the Kurdistan (Iraq) section of the Zagros thrust zone, lithologically diverse allochthons
732 of supra-subduction zone ophiolite and arc rocks sourced from Neotethys have been thrust over
733 the northwestern edge of the Arabian microcontinent.

734 (2) These allochthons are a product of two collisional events, one in the Cretaceous and one in
735 the Cenozoic, which together with out-of-sequence thrusting and post-collision partitioning with
736 wrench faulting have previously hampered their correlation.

737 (3) Recent programmes of U-Pb zircon and Ar-Ar amphibole and mica dating have led to
738 progress in deconvolving the ages and relationships of the allochthons, and enabled more robust
739 linkages to be made with Neotethyan assemblages to the east in Iran and to the west in Turkey
740 and Cyprus.

741 (4) Compilation of U-Pb zircon ages indicates that from the Troodos ophiolite of Cyprus in the
742 west to the Kohistan arc of Pakistan in the east, the oldest Cretaceous ages occur in Iran and Iraq.
743 This confirms that initiation of subduction was diachronous along the Neotethyan realm.

744 (5) Cenozoic intra-Tethyan assemblages appear to be restricted to Iran and Iraq, suggesting
745 closure of a remnant ocean basin of limited lateral extent.

746

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751

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1092 **Figure Captions**

1093 Figure 1. Location of the Zagros Ophiolitic Belt as part of a 3000 km long belt of Upper
1094 Cretaceous ophiolite between Cyprus and Oman (after Shafaii Moghadam and Stern, 2011) that
1095 continues eastwards into Asia.

1096 Figure 2. Geological maps of the Zagros thrust zone along the Iraq-Iran border, showing the
1097 location and tectonic subdivision of the study area (after Ali et al., 2012). (a) Inset showing the
1098 Sanandaj-Sirjan Zone and Inner and Outer Zagros Ophiolitic Belts (adopted from Shafaii
1099 Moghadam et al., 2013). (b) Lithostratigraphic units and section localities in the Iraqi sector of
1100 the Zagros thrust zone (adopted from Ali et al., 2013). Panel (a) is the southeastern continuation
1101 of (b). Note the position of Kermanshah on both (a) and (b).

1102 Figure 3. Schematic and generalised geologic cross-sections of the Hasanbag ophiolite (A-A');
1103 the Mawat ophiolite (B-B'); the Avroman-Bisuton carbonates south of the Penjween ophiolite
1104 (C-C', after Ali et al., 2014); the Pushtashan ophiolite (D-D'); the Mawat ophiolite (E-E',
1105 modified after Al-Qayim et al., 2012); and the Penjween ophiolite (F-F', modified after Al-
1106 Qayim et al., 2012; for locations see Fig. 2b).

1107 Figure 4. Three views across the Zagros thrust zone. (a) The Cretaceous Penjween igneous
1108 complex occurs as an allochthon over younger Cenozoic Red Bed Series rocks. View looking
1109 NW from F in Fig. 2b. (b) Contact between the allochthonous Cretaceous Pushtashan igneous
1110 complex and the Cretaceous Gimo-Qandil Group. View looking east near D' in Fig. 2b. (c)
1111 Contact between the Naopurdan island-arc rocks (Qalander Mountain) and the Red Bed Series.
1112 View looking SE near A in Fig.2b.

1113 Figure 5. Simplified stratigraphic columns displaying idealized internal lithologic successions of
1114 the Late Cretaceous Iraqi ophiolites from Hasanbag in the northwest to Penjween in the
1115 southeast.

1116 Figure 6. Trace element compositional variations of Iraq Mesozoic ophiolites and Tertiary
1117 igneous rocks. (a) Th/Yb vs. Nb/Yb diagram (after Pearce, 2008) and (b) V vs. Ti/1000 diagram
1118 (after Shervais, 1982). Data for Hasanbag after Ali et al. (2012), Pushtashan after Ismail et al.

1119 (2017), Walash-Naopurdan after Ali et al. (2013), Bulfat igneous rocks after Aswad et al. (2013),
 1120 Mawat after Jassim and Goff (2006) and Merza (2008), and Penjween after Abdulla (2008).
 1121 Figure 7. Schematic model for the formation and evolution of the Iraqi Zagros Mesozoic
 1122 ophiolites and Cenozoic igneous rocks (A, B, C, E and F after Ali et al., 2013, D after Ali, 2017).
 1123 Figure 8. Summary of the U-Pb zircon ages of the Neotethyan arc and supra-subduction zone
 1124 ophiolites from Cyprus in the west to the Himalayas in the east. Data are sourced from Mukassa
 1125 and Ludden (1987) and Konstantinou et al. (2007) for the Troodos ophiolite; Karaoglan et al.
 1126 (2013) for the Kizildag ophiolite; Ali et al. (2016), Mohammad and Cornell (2017), Isamil et al.
 1127 (2017) and Ali (2017) for the Kurdistan Iraq ophiolites; Shafaii Moghadam et al. (2013),
 1128 Mazhari et al. (2009) and Ao et al. (2016) for the Iranian ophiolites; Tilton et al. (1981) and
 1129 Roberts et al. (2016) for the Samail ophiolite; and Schaltegger et al. (2002) for the Kohistan
 1130 ophiolite (Pakistan).

(a)

Penjween igneous complex (93 Ma)

Red Bed Series (≤ 60 Ma)

(b)

Gimo-Qandil Group (≤ 100 Ma)

Pushtashan complex (96 Ma)

(c)

Qalander mountain

Naopurdan Group (24 Ma)

Red Bed Series (≤ 60 Ma)

Hasanbag ophiolite
(106-92 Ma)

Pushtashan ophiolite
(96 Ma)

Bulfat ophiolite
(? Ma)

Mawat ophiolite
(96 Ma)

Penjween ophiolite
(93 Ma)

Red Bed Series (Cenozoic)
Naopurdan Group (Oligocene)
Nepheline syenite (Cenozoic)
Olivine-gabbro dyke (39 Ma)
Walash Group (Eocene)
Gabbroic dyke (Cretaceous)
Granodiorite (Cretaceous)

Volcanic rock (Cretaceous)
Pillow lava (Cretaceous)
Volcanic rock (Cretaceous)
Norite (Cretaceous)
Gabbro (Cretaceous)
Peridotite (Cretaceous)
Gimo-Qandil Group (Cretaceous)

Dunite
Podform chromite
Granitic pegmatite (96 Ma)

Quartz vein
Pyroxenite dyke
pegmatite gabbro

* dated unit

A) Permo-Triassic, ~250 Ma : Eurasia-ward subduction in Paleo-Tethys

B) Cretaceous, 106-92 Ma : After closure of Paleo-Tethys, establishment of the Hasanbag arc in Neo-Tethys

C) Late Cretaceous: emplacement of Hasanbag arc onto Arabian margin, initiation of the Walsh-Naopurdan arc system

D) Eocene : Two different intra-oceanic arc domains

E) Miocene, 23-16? Ma: Emplacement of Walsh-Nopurdan arc onto Arabian margin

F) Middle Miocene-Present; Ongoing Arabian-Eurasian Continental collision